Climate Change and Global Warming

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This class will presented as 5 distinct topics in an effort to give a reasonably complete picture of climate change and global warming.

The 5 topics are:

1 - Needed Background Material

Chaos and Fractal Ideas

2 - Climate Change - 1st Pass

Comprehensive Picture

3 - Climate Change - 2nd Pass

Different Approach - Many More Details

4 - Global Warming Science

Quantitative Introduction to Climate Change and Its Consequences

5 - Simple Models (for those interested)

(a) To see how it is done in Detail.

(b) Two Professional Simulations

Part 1

Needed Background Material

We need to study some aspects of Chaos and Fractals.

Our study will be limited to understanding what these two ideas imply

so that we can use the concepts to try to describe weather behavior

as a prelude to tackling the Climate Change - Global Warming problem.

I will make multiple passes through the ideas relevant to chaos and

fractals to reinforce your learning of the important concepts.

Those of you that have taken some previous classes with me will

recognize some of this material.

Chaos is everywhere

Smoke swirling, flag waving, dripping faucets, traffic jams, the weather, etc











Seemingly very different phenomena - but maybe not really different as we will see!

Chaos is the science of the **GLOBAL** nature of systems

Study of chaos looks for patterns **independent of scale**, explores **randomness**, **complexity**, **jagged** edges and **sudden** changes(discontinuities)

Many so-called “well understood” systems are, as we will see, are **not really** understood at all!

Even some **“simple”** systems will have **complex** behavior leading to **unpredictability.**

Recent i**nvestigations** imply that a **new** kind of order arises spontaneously(emerges).

Chaos and order are **intermingled**

Traditionally, in science and life, complex causes imply complex results.

BUT IT TURNS OUT THAT IS JUST NOT 100% TRUE!

Typically, if a scientists **observes** a random relationship between input and output for system, then they will **build** randomness into the theory (artificially)

—> they usually have **added** noise or error to initial conditions

Method used for decades!!! But as we will see, this is the **wrong** approach!!!

What is correct way to approach these systems?

We first consider the work of Ed **Lorenz** <—> he had a computer model…….



Figure 1.1: Lorenz Computer Model - Run 1

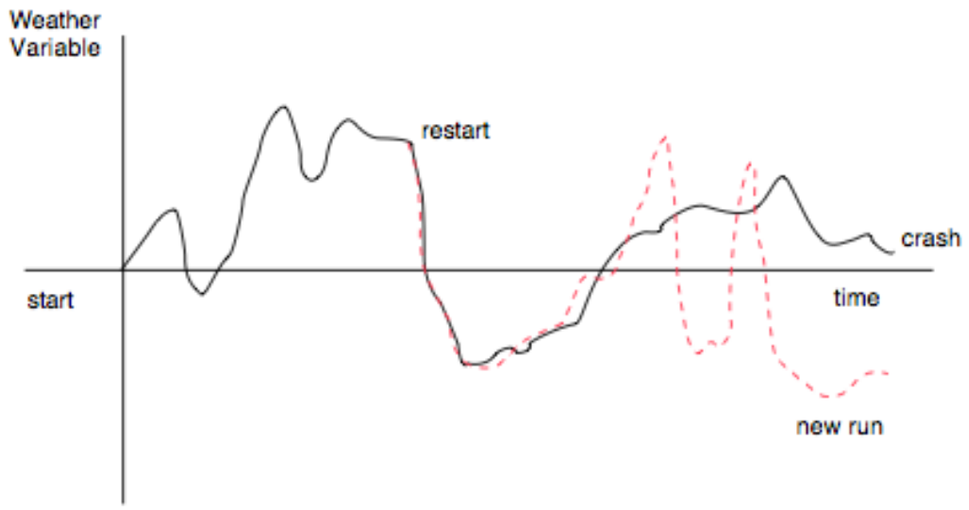


Figure 1.2: Lorenz Computer Model - Run 2

The above figures show what happened.

N**ow let me explain what the graph are telling us.**

In general, was always assumed that for a **deterministic** classical system(equations exist),
approximate initial conditions --> small differences in final results -- **no** large deviations

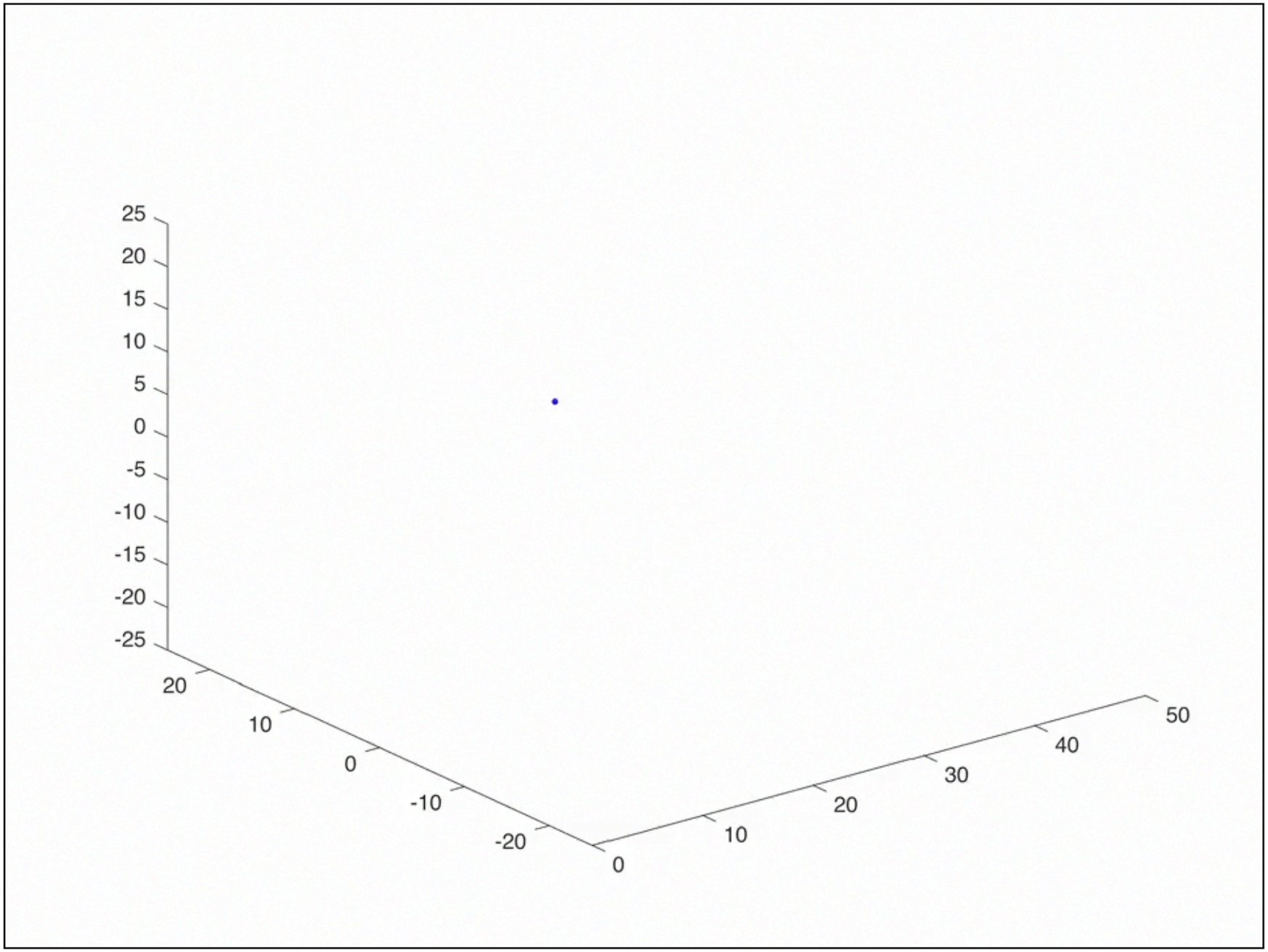
H**owever**, in many systems, actually observe both small and **catastrophic** deviations!
—> **no** predictions are possible (as in example above)

Phenomenon has been named the **butterfly effect.**

IDEA: Any sensor grid location with **missing** values(missing coverage - broken or not there) hides fluctuations —-> leads to approximate initial conditions

—> **No** prediction is possible - i.e., a **butterfly** might have flapped its wings in the gap!!

Seemed that predictability was giving way to randomness (no predictability) in some systems.



class demo

No randomness involved in this motion

Standard Belief: if can determine initial conditions without disturbance.

⟶

Predict trajectory (path of motion)

⟶

know all future behavior

BUT……

General Belief: accurate predictions
limited only by accuracy of initial conditions + knowledge of forces

Generally works for macroworld - after all, we did land on the moon!

Fly in the ointment = **chaos** ⟶ Sensitivity to initial conditions ⟶ not predictable

However, the Lorenz results imply **more** than randomness

They **imply**(as we will see) some kind of **geometrical** structure at a fine scale

Some new kind of **order** masquerading as randomness

Will need to study **aperiodic** systems
i.e., systems that **almost** repeat themselves

—> **link** between aperiodicity and unpredictability exists

Butterfly effect is **no** accident -- studies implied it is **necessary** feature,

i.e., suppose small perturbations remain small instead of generating observed chaotic effects.

This **implies** that when weather system came arbitrarily close to state had passed earlier,
would **stay** arbitrarily close to all patterns that followed earlier state which

means **could predict** future weather patterns and cycles but we **cannot**!

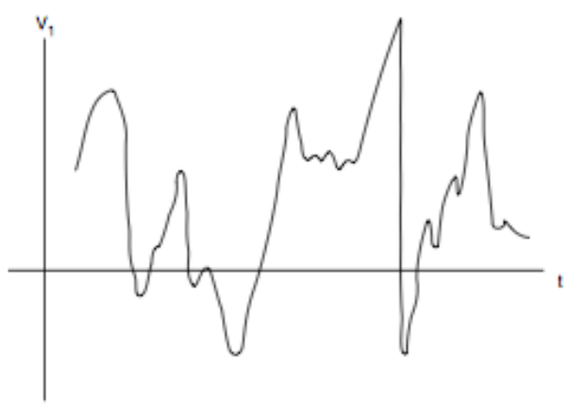
Butterfly effect **must** exist! where butterfly effect

equivalent to **sensitivity to initial conditions**

But **how** does it happen?

How could such unpredictability - such **chaos**
arise from a simple **deterministic** system of (**non-linear**) equations

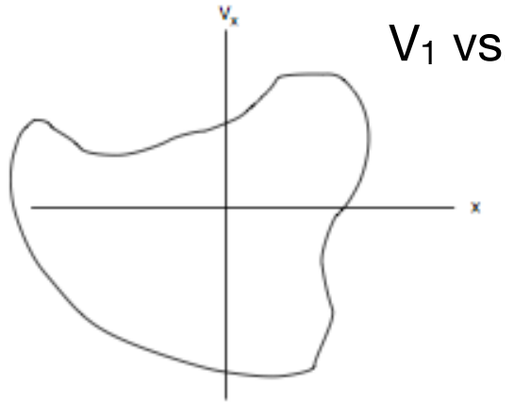
To **visualize** system evolution(in time) can plot variables as functions of time (—> time series)



V_1 vs. t

Figure 1.4: 1 Variable Time Series

Alternatively, can plot evolution in **phase space** where variables label axes, velocity versus position for a particle in this plot.



V_1 vs. V_2

Figure 1.8: Not so simple periodic motion

Meanings of different paths in phase space:

If path **stops** at point,
this corresponds to variables remaining **constant** or a **steady-state situation**.

If path **loops** (closed path), implies **periodic** (repeating) motion.

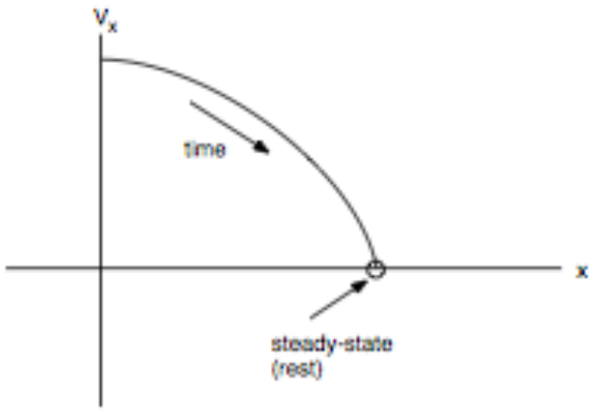


Figure 1.6: Particle with friction

Examples

Figure 1.8: Not so simple periodic motion

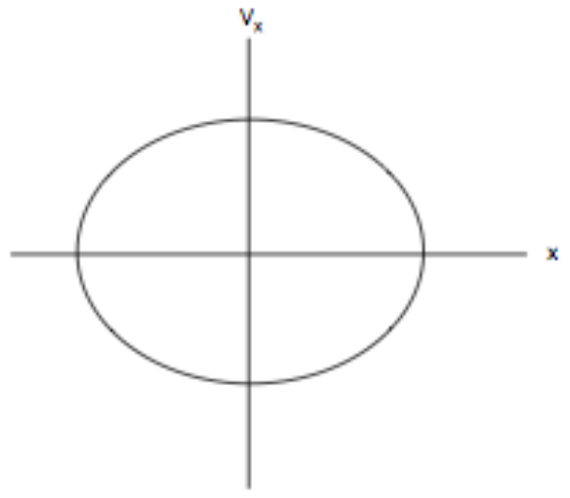
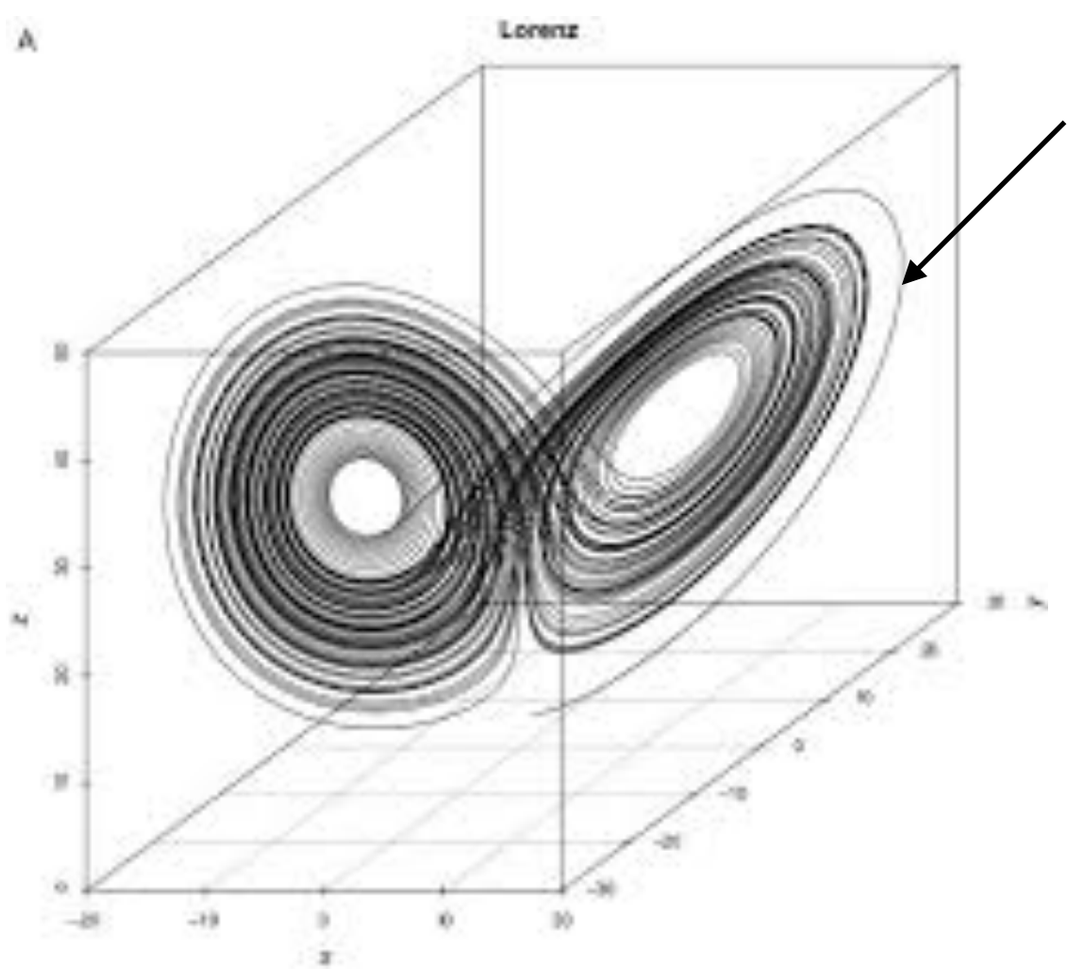


Figure 1.7: Simple periodic motion

What did Lorenz see?

No steady states and **no** loops implying periodicity.

Only what he called “infinite complexity”



Found that(listen carefully):

1. Path **stayed** in finite volume (within well-defined bounds).

2. Path **never** repeated itself no matter how long was watched.

3. Path **traveled** along strange, distinctive shape - kind of
   double spiral in 3D space of variables
   (like pair of butterfly wings (saw earlier)).

Implied pure disorder, **since** no points ever repeated!

However, it actually contained a **new** kind of order.

Physicists **understood** short term motion of such systems,

but could not **extend** understanding to long term motion when **non-linearities** are present.

Microscopic (fundamental equations or rules) pieces were **clear**,

but **macroscopic** behavior (complex patterns) **remained** a mystery!

Tradition of looking at systems **locally**

isolating small scale mechanisms

and then **adding** local effects **together** to figure out whole system

was beginning to **break down**.

Standard methods of **reductionism**

where one understands **smallest** building blocks

and **builds** larger system up from behavior of smaller blocks was **failing**.

Whole seemed to be more than sum of its parts!

Knowledge of fundamental equations no longer **seemed** to be right kind of knowledge at all.

Physicists had to learn **again** - had to learn to **see** things they had **learned not to see**!

Chaos and instability are not the same at all!

Chaotic system could be **stable** (means **equilibrium**)

if type of observed irregularity(pattern) **persisted** in face of small disturbances

(—> **traditional** definition of equilibrium).

Chaos of pendulum is as **stable** as **equilibrium** of **marble** sitting at bottom of bowl!

Can **add** noise, jiggle, stir up, interfere with motion and when everything has settled down -

i.e., **transient behavior**(due to our messing around) dying away

system **returns** to same peculiar pattern of irregularities as before —> **stable**.

Systems, it turns out, are often **locally** unpredictable, but **globally** stable.

Real dynamical systems play by more **complicated** set of rules than anyone imagined.

The **idea** that systems behave in certain quantitative ways that depend

not on detailed physics or model description,

but **rather** only on some general properties of system is called **universality**.

Traditionally, physicists study problem

—> equations with variables

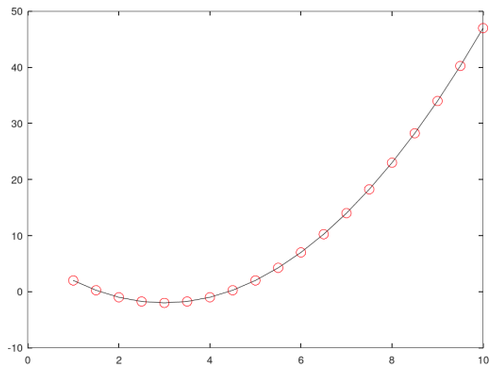
(**continuous** functions of time) (differential equations) —> **make** predictions.

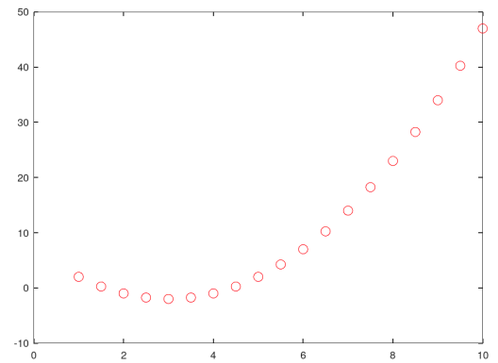
If **cannot** solve equations analytically —> use **computers**.

Numerical solutions return answers at **discrete** values of time variable

(not **continuous** values) - one **assumes** continuity.

might not be true!





Biologists also use **experiment** to study systems.

Experimental data naturally comes at **discrete** time intervals (**not** continuously).

Equations relating variables **known** only at **discrete** time values

are called **difference equations**.

In these equations, the process **jumps** from state to state

with **no** knowledge about **"in between"**.

Changes at discrete intervals become

more important than what happening at **every** time point

(**maybe** more important EVEN than any assumed continuity).

For discrete systems,
variable(s) **now** depends only on the variable(s) **last** year(last month,.....) i.e.,

$$x_{next} = f(x_{now}) \longrightarrow x_{n+1} = f(x_n)$$

n^{th} step $\longrightarrow$ $(n+1)^{st}$ step

In such a simple equation, we **proceed** in this way:

1. pick a starting value $x = x_0$

2.x_1 (value afterΔt (one time interval)) $= f(x_0)$

3.x_2 (value after $2\Delta t$) $= f(x_1) = f(f(x_0))$

4. and so on .....

This process is called **iteration**

Can observe **both** stable and unstable behavior in time in these systems.

Resulting set of numbers generated by iteration process is a **discrete history**

or **time-series depending** on nature of the function $f(x)$

Simplest equation (used for population growth studies) that exhibits chaos is

$$x_{n+1} = \lambda x_n(1 - x_n)$$ **Logistic Map**

Model has 0 ≤ λ ≤ 4, where λ is the growth parameter and 0 ≤ x ≤ 1 always.

$(1 - x_n)$ term **inhibits** growth because as x_n approaches 1, $(1 - x_n)$ **approaches** 0.

Plotting x_{n+1} **versus** x_n . ⟶ nonlinear relationship as shown below.

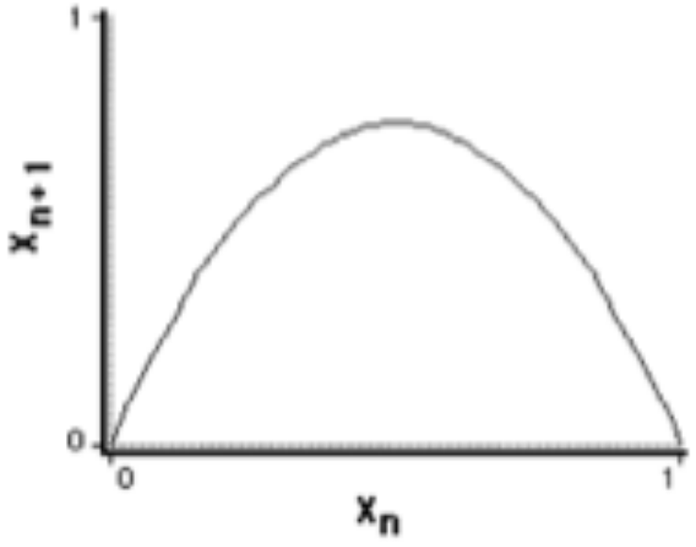


Figure 1.38: Limited Growth Model: x_{n+1} versus x_n , $\lambda = 3$

Have to **iterate** this function to see how it will **behave**.

Suppose λ = 3 and $x_1 = 0.1$

Then **get**

$$x_2 = \lambda x_1(1 - x_1) = 0.270$$
$$x_3 = \lambda x_2(1 - x_2) = 0.591$$
$$x_4 = \lambda x_3(1 - x_3) = 0.725$$

as **shown**

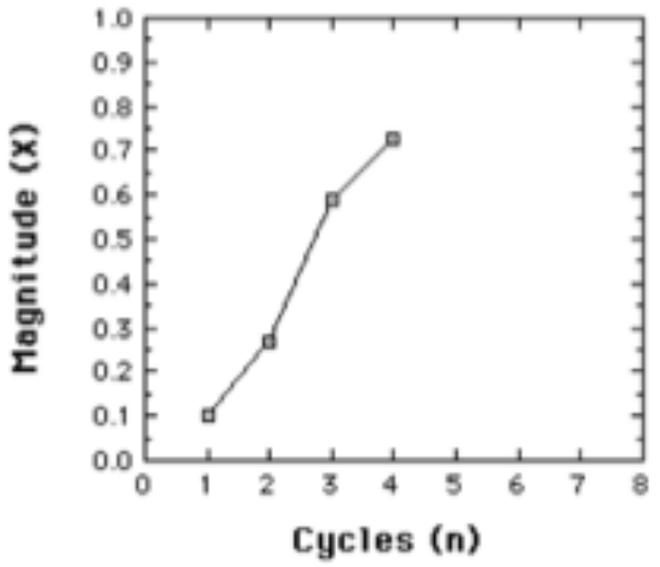


Figure 1.39: Behavior of the Logistic Map for $\lambda = 3$, $x_1 = 0.1$ iterated to give x_2, x_3 and x_4

Logistic map is very **different** animal, **depending** on value of **control** parameter λ

i.e., if **control** parameters are **allowed** to change, **dynamic** system **also** changes.

To **see** this, next **examine** time series produced at **different** values of λ,

starting near 0 and ending at λ = 3.99.

See **very different** results, **revealing** and **introducing** major features of a chaotic system.

We plot **time series:** or **set** of measures of behavior over time

(iteration step number n is a measure of time).

λ < 1:

x ⟶ 0 **independent** of initial value.

Illustrates a **one-point attractor**.

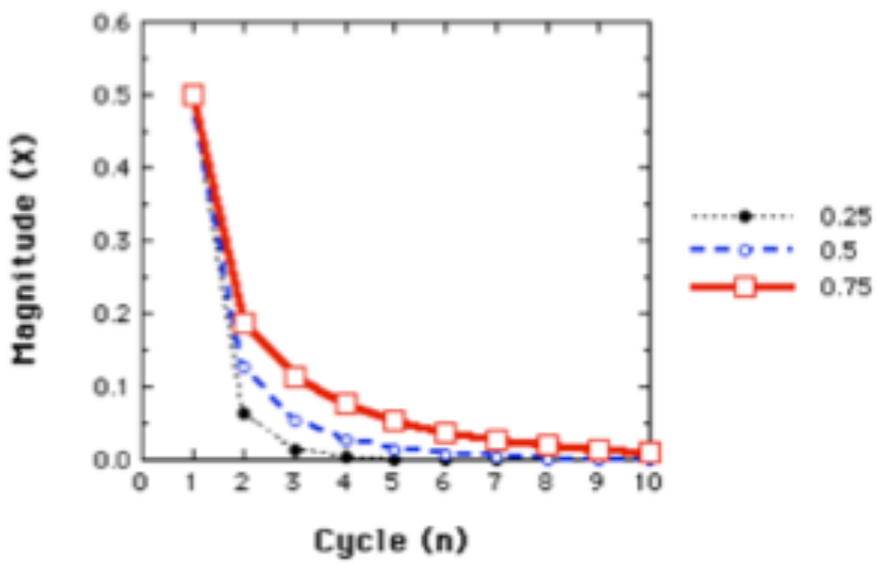


Figure 1.40: Behavior of the Logistic Map for $\lambda = 0.25, 0.50, and 0.75$. In all cases $x_1 = 0.5$

Attractor = state that dynamic system **evolves** into.

or in more complicated system,

Attractor = **set** of values in phase space(states) to which system **migrates** as we iterate.

Attractor can be **single** fixed point, **collection** of points regularly visited, a **loop**,

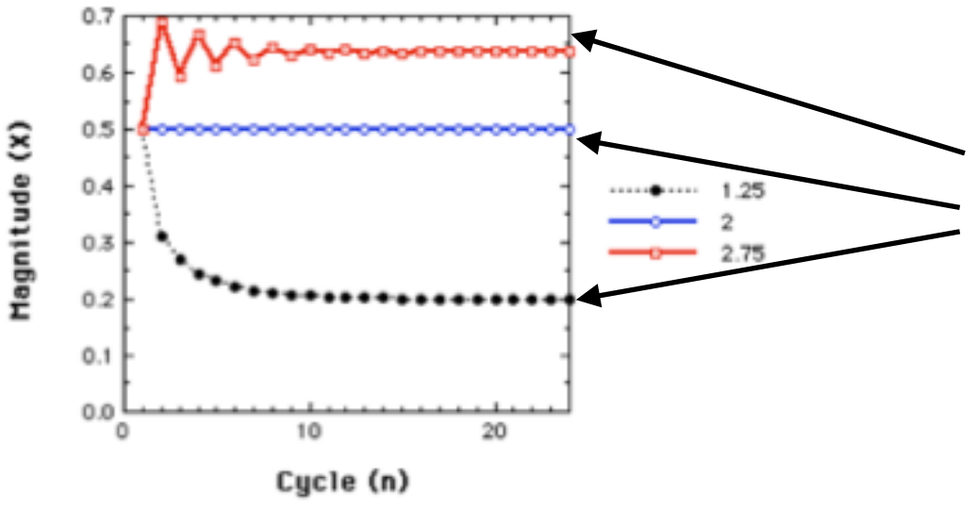
complex orbit, or **infinite** number of points as we will see.

You can probably guess which one is associated with chaos.

Attractors **Need not** be one- or two-dimensional …….

Will find that the **Dimension** of attractor ≤ number of variables that influence system.

When 1 < λ < 3

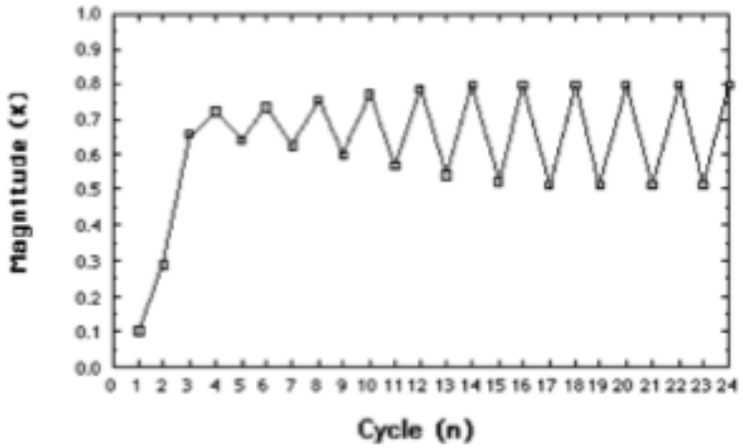


Regardless of the starting value ⟶ get a non-zero **one-point attractor**.

It is not zero now!

Figure 1.41: Behavior of the Logistic Map: for $\lambda = 1.25, 2.00,$ *and* 2.75. In all

When $\lambda > 3$ get:



Behavior of the Logistic Map for $\lambda = 3.2$.

i.e., **beyond** $\lambda = 3$, system settles down to **alternation** between two points.

—> Have a **two-point attractor**.

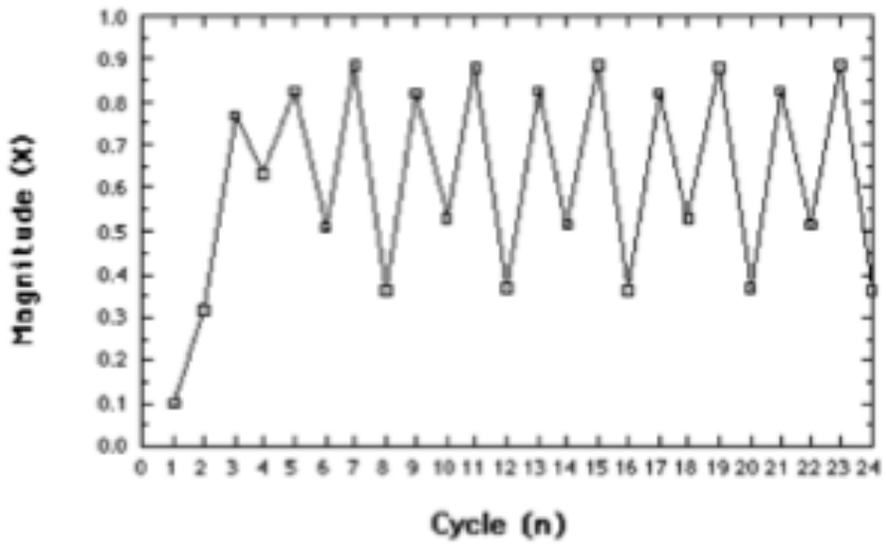
Figure ⟶ a **bifurcation**, or **period doubling (1 to 2) of attractor has occurred**.

bifurcation = qualitative change in behavior (of attractor) of dynamic system associated with **change** in control parameter.

Increasing λ further

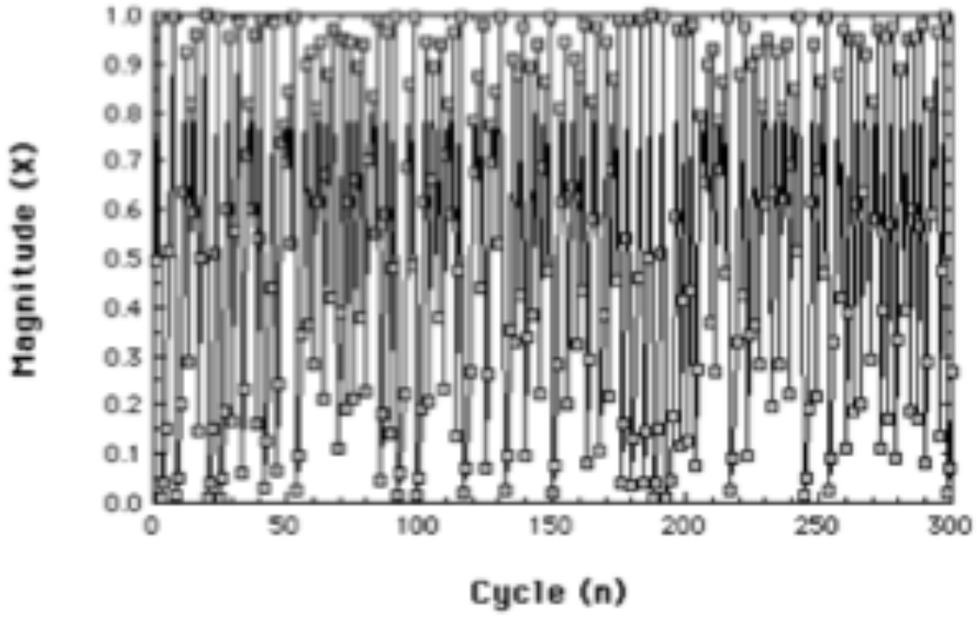
Another **bifurcation** has occurred
⟶ 4-point attractor.

Increasing λ further one **gets** higher order attractors.



Behavior of the Logistic Map for $\lambda = 3.54$. A four-point attractor

Illustrated below is **chaotic** behavior of system (no pattern at all!).



for $\lambda = 3.99$

e 1.44: Chaotic behavior of the Logistic Map at $\lambda = 3.99$

So, **what** is an **attractor**?

⟶ **whatever** geometrical object the system **settles** down to.

A very **important** concept from nonlinear dynamics:

System always eventually **settles** down.

But **what** it settles down to, its attractor, **need** not have stability.

Can be very "**strange**".

Strange attractor = an **N-point attractor** in which **N** equals **infinity —> never repeats**.

Lorenz example!

Usually (perhaps always) is **self-similar** (discussed shortly) in form.

Can see entire structure of logistic map in plot below.

Plot of large number of x values in a sequence for fixed λ value versus λ value.

Guess what it will look like? Using this type of plot can see all of previous plots.

—> **bifurcation diagram**: Visual summary of succession of period-doubling produced as control parameter is changed.

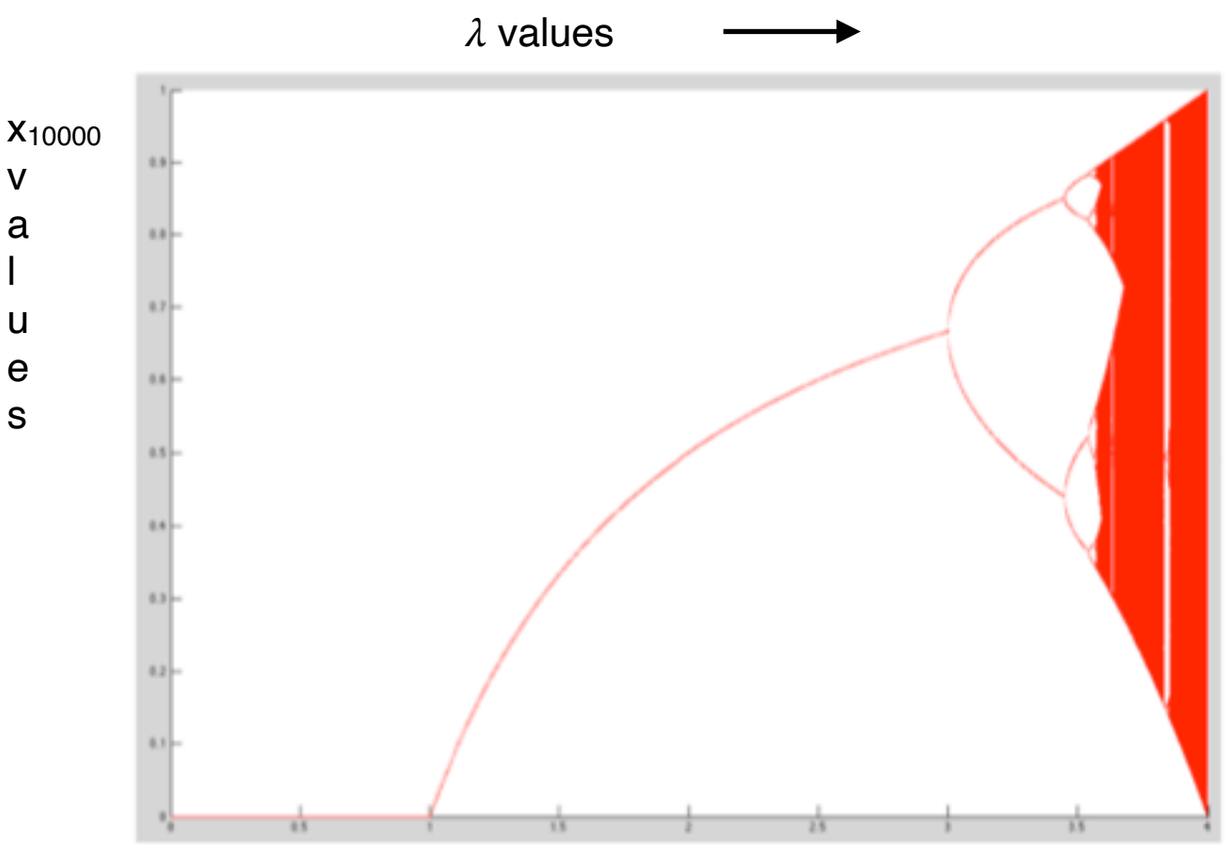


Figure 1.59: A Bifurcation Diagram

Pick a λ value;

Pick an initial value of x

Iterate 10000 times;

Keep last 5000 iterations only (removes transients)

Plot all 5000 values versus that value of λ;

Then repeat for each λ value.

Clearly, can see the period-1, period-2, period-4, etc regions, bifurcations

or period doublings, chaotic regions and many other strange features.

Blowup region from 3.545 to 3.575

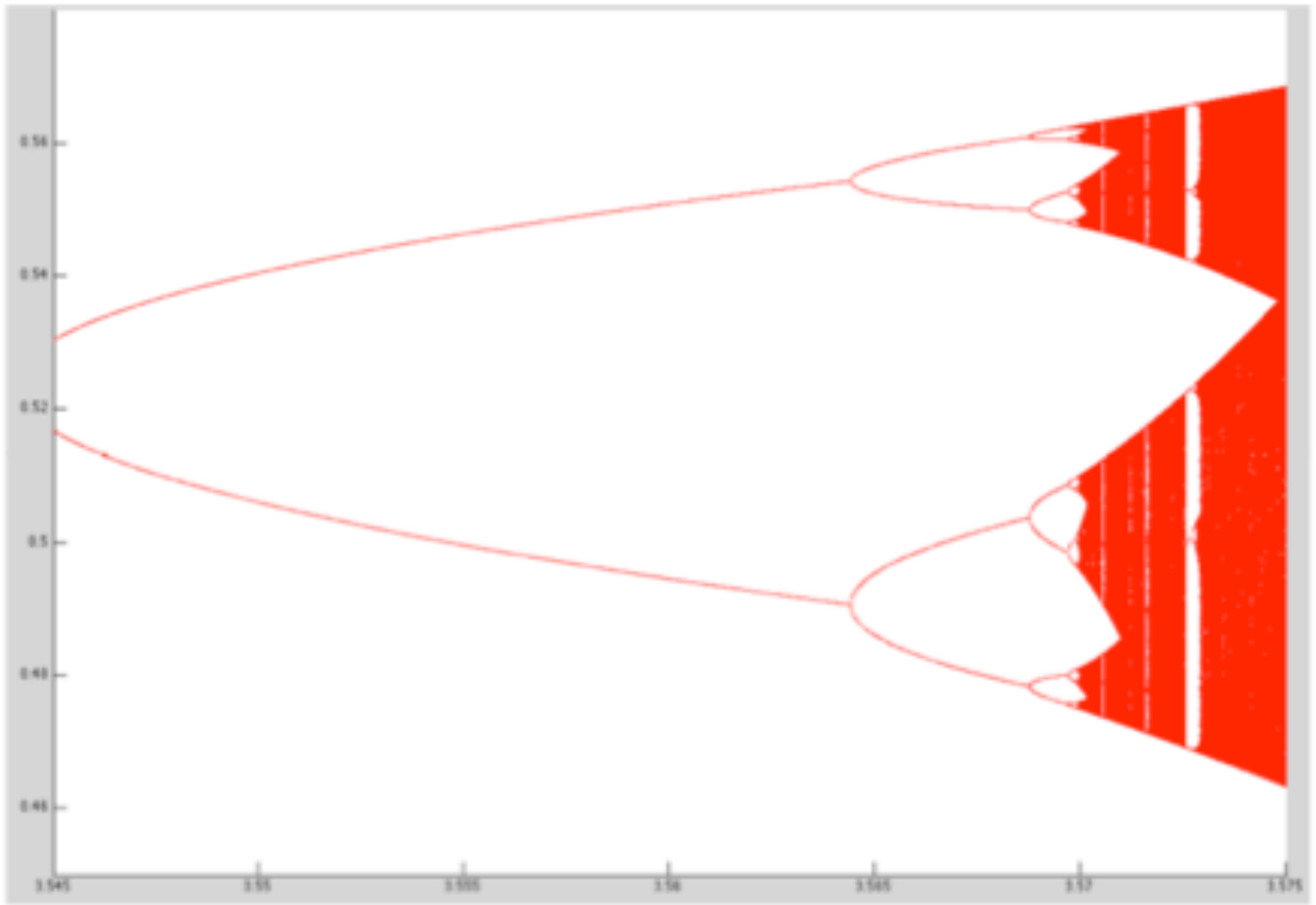
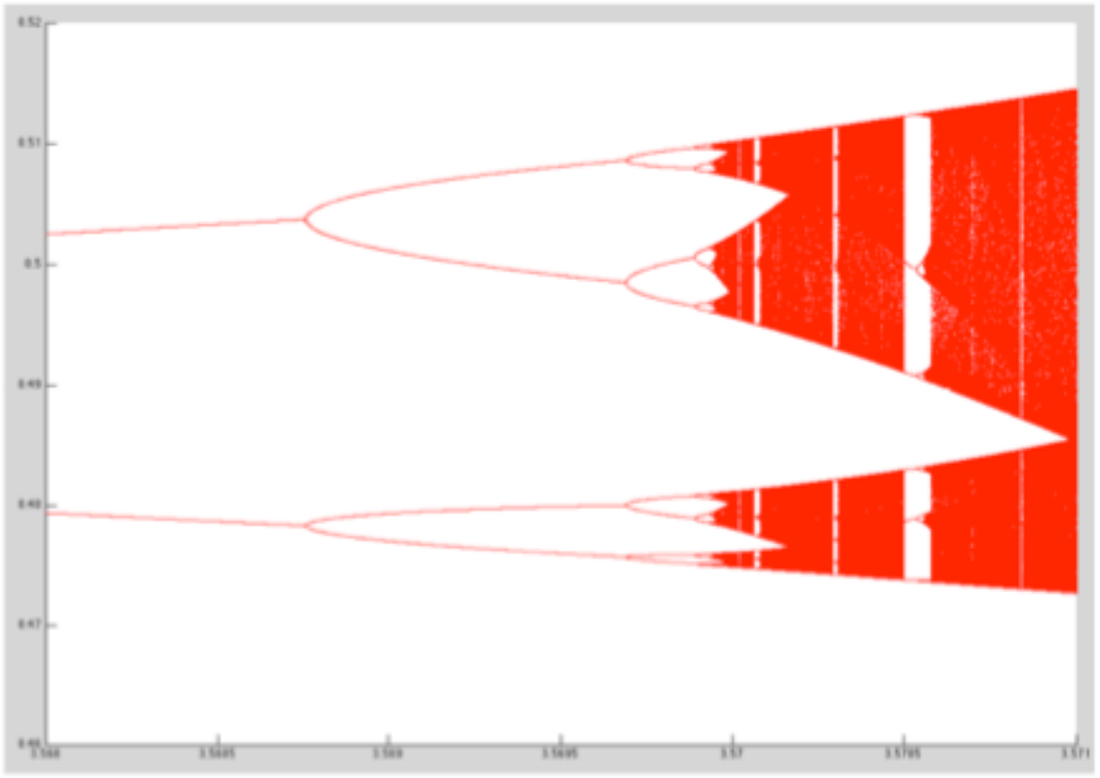


Figure 1.60: Blowup of region 3.545 to 3.575

Discuss details in class

Finally blowup region 3.5680 to 3.5710



Blowup of region 3.5680 to 3.5710

Each bifurcation looks **similar** up to a magnification factor.

This **scale invariance** or **self-similarity** plays an important role in transition to or onset of chaos and in structure of a strange attractor.

chaos: Behavior of dynamic system that has

(a) a very large (possibly infinite) number of attractors and
(b) is sensitive to initial conditions(solid red regions).

Note from pictures above that $\lambda_c = 3.56994...$ attractor set for many (but not all) values of λ shows no periodicity at all(are chaotic); solid —> get all values with no repeats(ever).

For these values of λ quadratic map **exhibits chaos and has a strange attractor**.

In region $\lambda_c < \lambda < 4$ there are windows where attractors of small period reappear.

Important property of chaotic motion is extreme sensitivity to initial conditions (mentioned earlier).

To express sensitivity **quantitatively** introduce **Lyapunov exponent**.

Lyapunov Number:

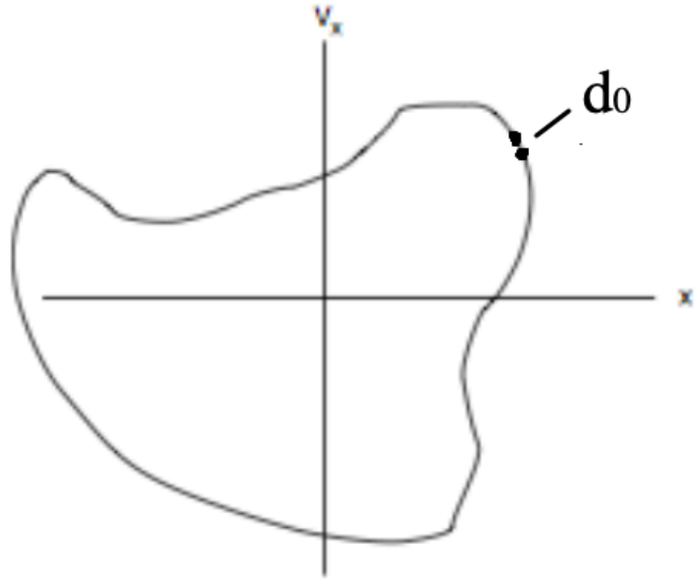
value of an exponent or coefficient of time, that reflects rate of departure of a pair of dynamic orbits —> a measure of sensitivity to initial conditions.

Consider two points on a path in phase space separated by distance d_0 at time t = 0. (see below).

If motion is regular (non-chaotic) these two points will remain relatively close(along the path), separating at most according to a power of time.

In chaotic motion, the two points separate exponentially with time as $d(t) = d_0 e^{\lambda_L t}$

like slightly different initial conditions as shown



Parameter λ_L is Lyapunov exponent.

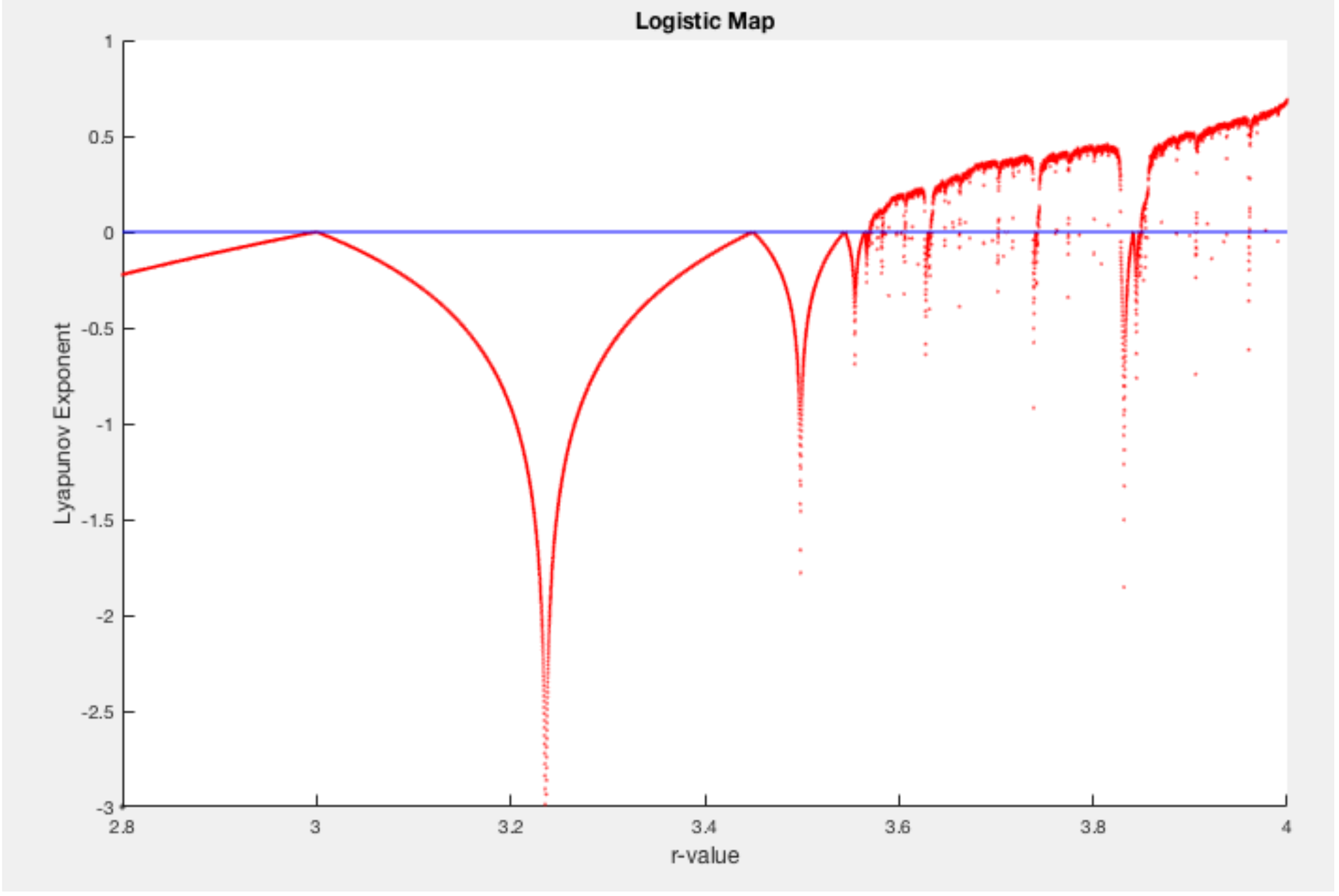
If λ_L is positive motion is chaotic(separation increasing).

A zero or negative coefficient indicates non-chaotic motion.

For logistic map there is one Lyapunov exponent.

We plot Lyapunov exponent as function of λ, logistic map parameter, below.

Lyapunov exponent as a function of λ



Lyapunov exponent is negative

whenever map is stable and positive whenever map is chaotic.

Value of λ is zero when bifurcation occurs and solution becomes unstable.

A superstable point occurs where $\lambda = -\infty$.

Can see this clearly in plot on next slide:

This make the correspondences clear.

Can see clearly from plot that when λ goes above zero,

there are windows of stability where λ goes negative for a while

and period orbits occur amid the chaotic behavior.

Relatively wide window just above λ = 3.8 is apparent.

Figure 1

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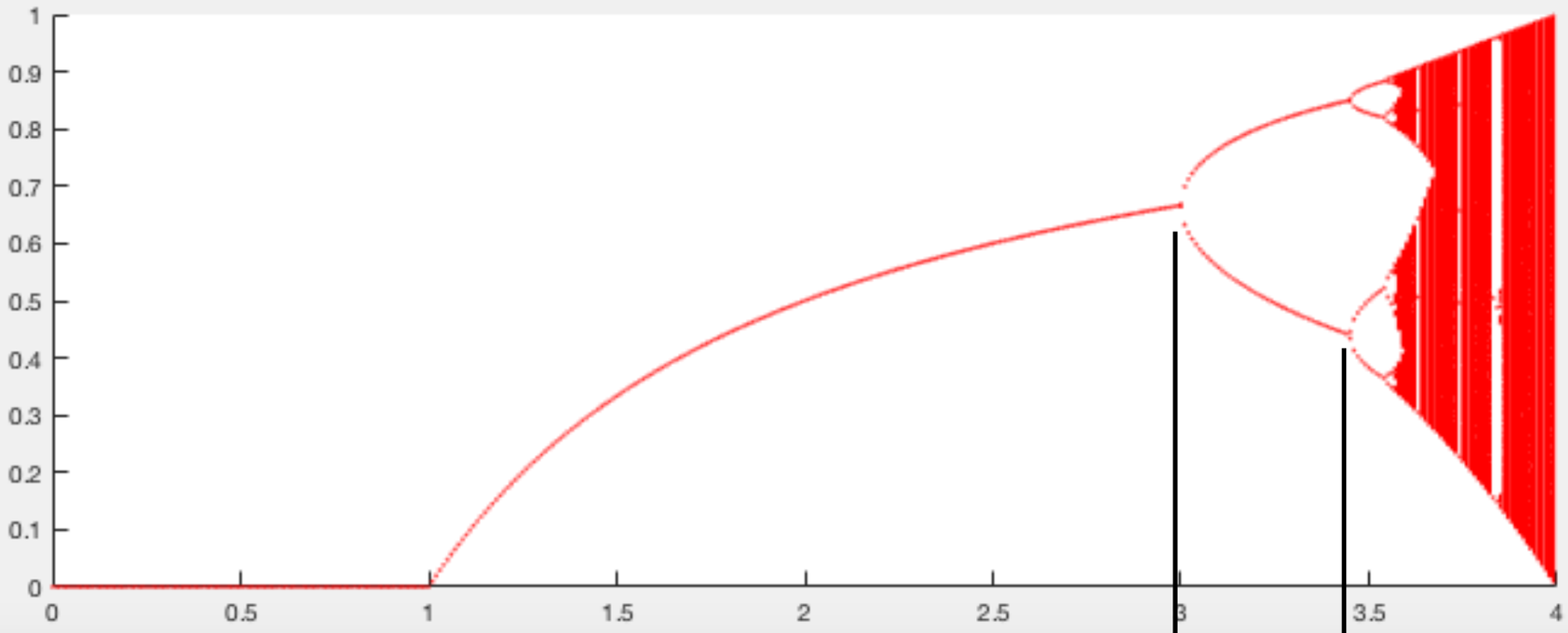
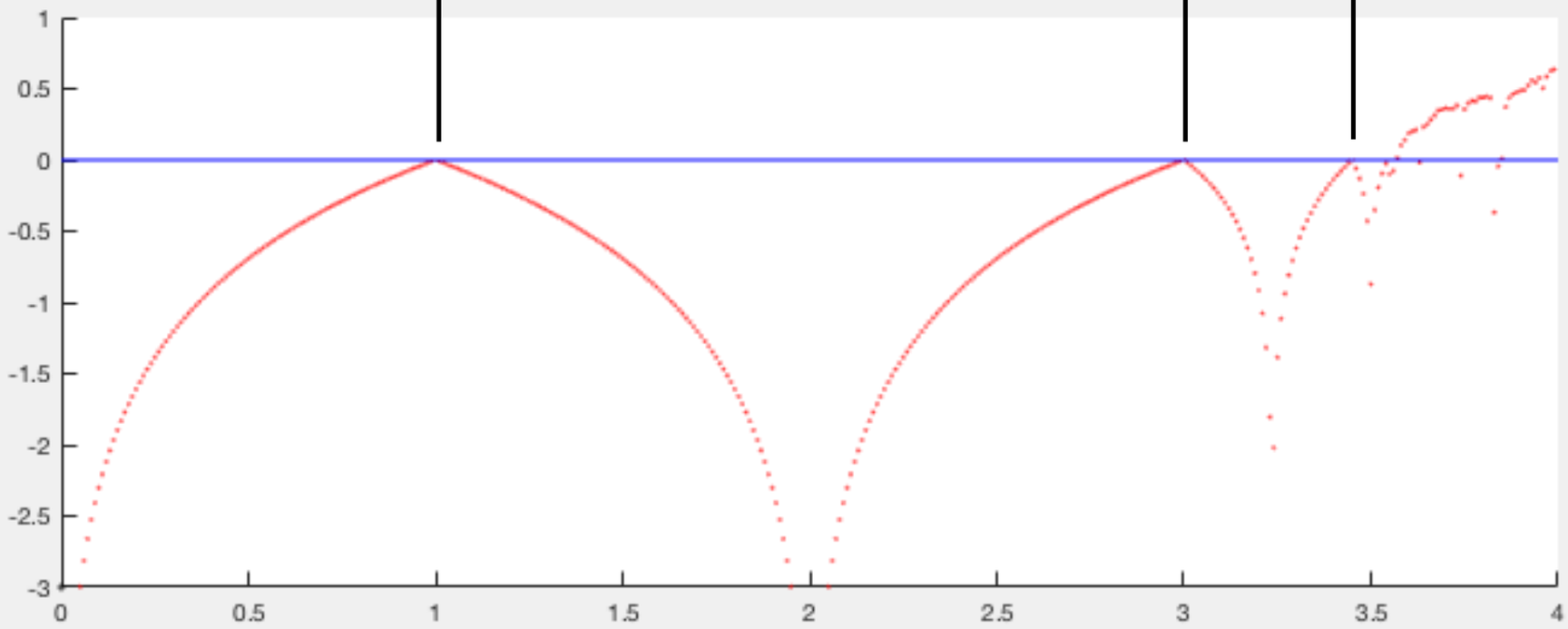


Figure 2

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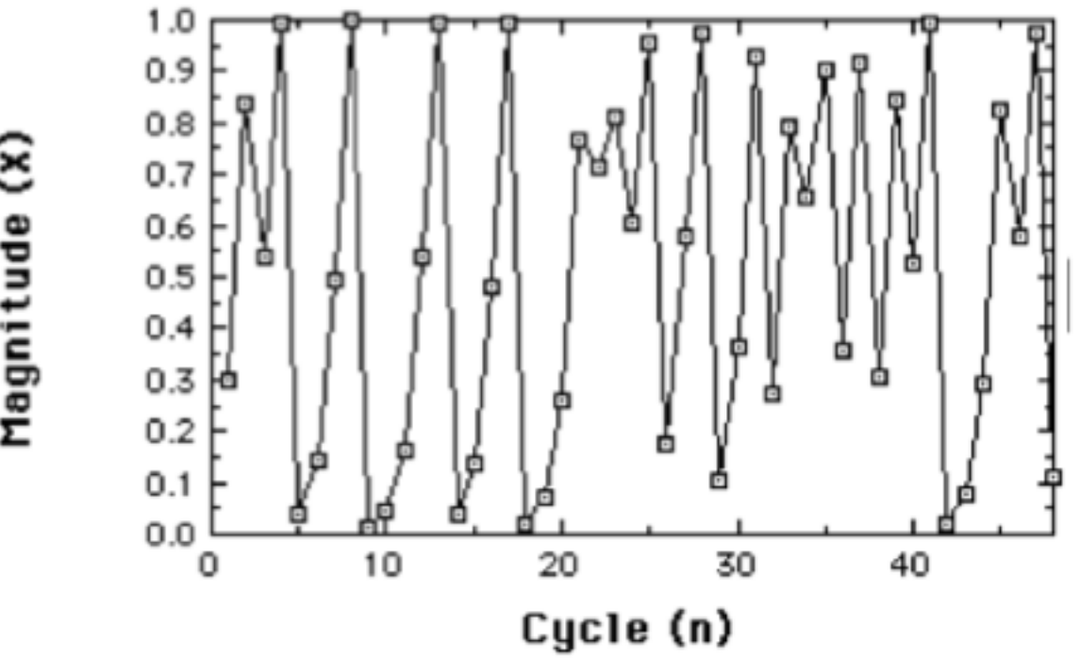


Sensitivity to Initial Conditions is the hallmark of a chaotic system

This important feature emerges in the chaotic regions ...

To see it, set λ = 3.99 and begin at $x_1 = 0.3$.

Next graph shows time series for 48 iterations of logistic map.



Time series - 48 iterations - $\lambda = 3.99$

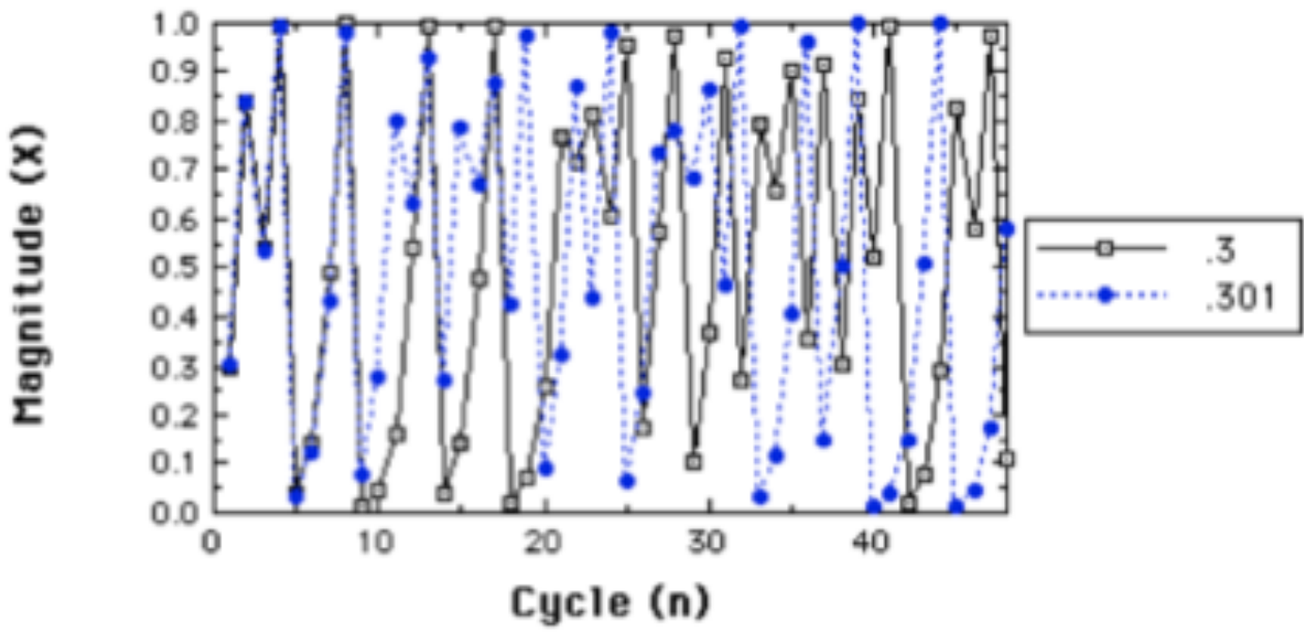
Two time series stay close for about 10 iterations.

But after that, they are pretty much on their own - they diverge from each other(exponentially).

Now suppose we alter starting point a bit.

Next figure compares time series for $x_1 = 0.3$

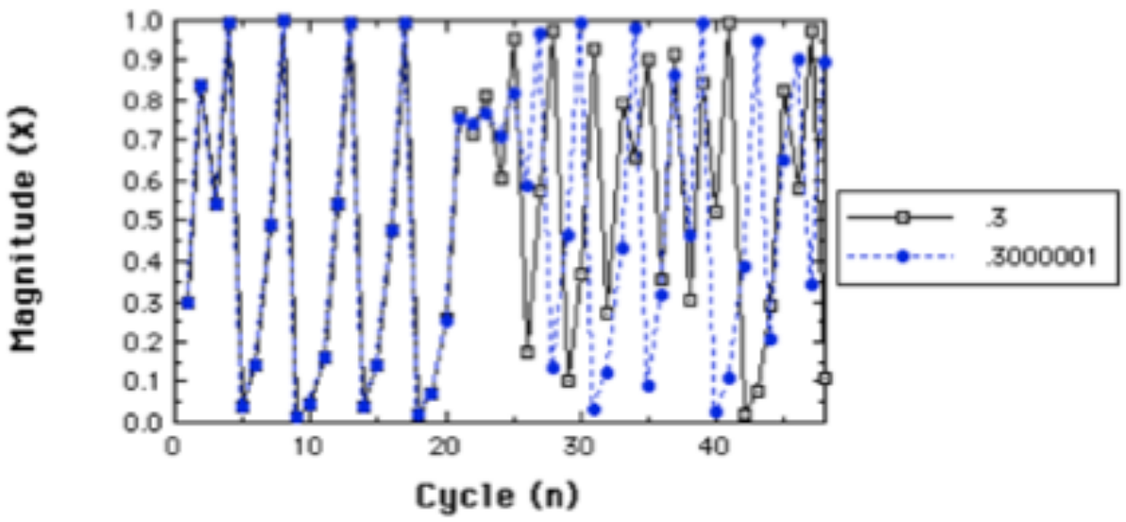
(open squares) with that for $x_1 = 0.301$ (solid dots).



Time series - 48 iterations - $\lambda = 3.99$ - Different starting

Let us try starting closer even together.

We next compare starting at 0.3 with starting at 0.3000001...

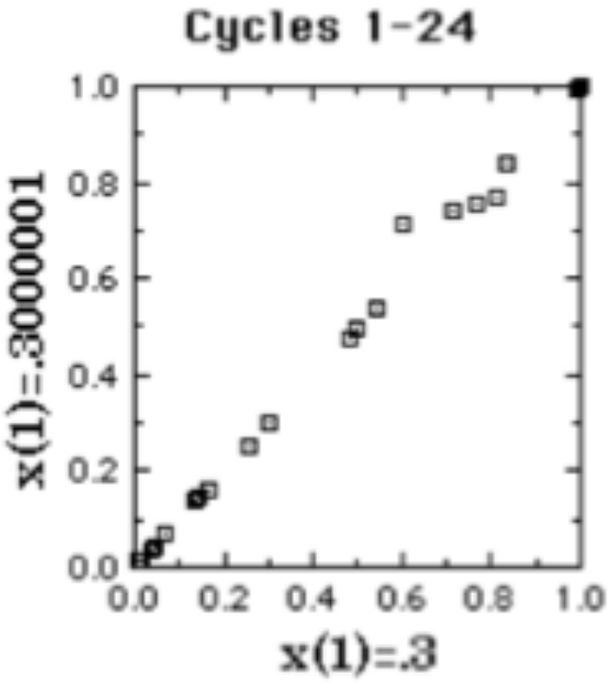


me series - 48 iterations - $\lambda = 3.99$ - Closer starting points

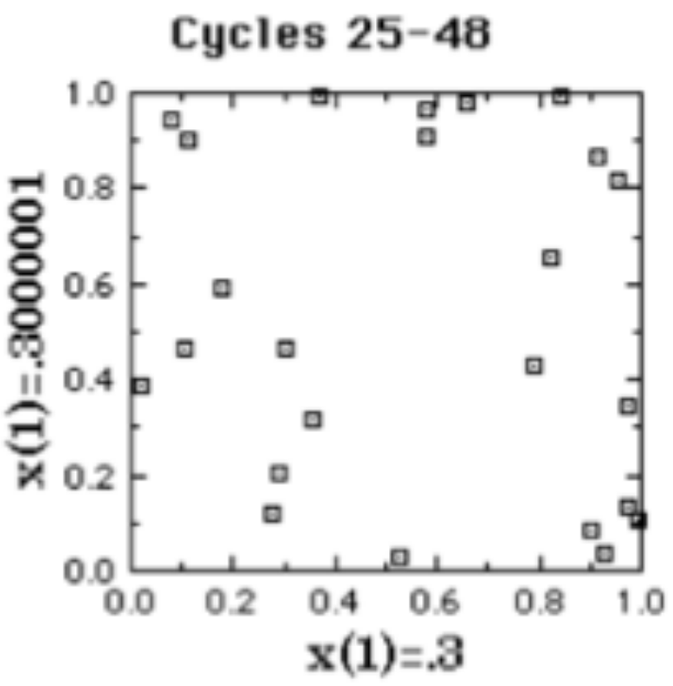
This time stay they close for longer time, but after 24 iterations they diverge.

To see how independent they become, next figure provides scatterplots for two series before and after 24 iterations.

perfect correlation



zero correlation



Scatterplots to compare time series

Correlation after 24 iterations (right side), is essentially zero.

Unreliability has replaced reliability.

Have illustrated one symptom of chaos - **sensitivity to initial conditions**.

A dynamic system has sensitivity to initial conditions
when very small differences in starting values result in very different behavior.

If orbits of nearby starting points diverge, system has sensitivity to initial conditions.

A chaotic system is one for which ”distance” between two trajectories or orbits
starting from nearby points in its state space diverges over time.

Magnitude of divergence increases exponentially in a chaotic system.

What is the meaning of all these statements?

It is unpredictable, in principle because in order to predict its behavior into future
must know its current value precisely.

In example, slight difference in sixth decimal place,
resulted in prediction failure after 24 iterations.

Note that six decimal places far exceeds kind of measuring accuracy
typically achieve with natural biological systems.

Now to look at a Real System

Newtons second law for pendulum (variables = θ = angle of swinging pendulum)

The Nonlinear Damped Driven Oscillator

$$m\frac{d^2\theta}{dt^2} + \alpha\frac{d\theta}{dt} + \beta\sin\theta = \gamma\cos(\omega t)$$

explain terms in class

Corresponding difference equation: $m\theta_{n+2} + (\alpha - 2m)\theta_{n+1} - (m - \alpha)\theta_n + \beta\sin\theta_n = \gamma\cos(\omega n)$

$$b = 1.0 = \frac{\beta}{m} = \frac{1}{m}\sqrt{\frac{g}{\ell}} \quad , \quad \ell = \text{ pendulum length}$$

$$c = 0.5 = \frac{\alpha}{m} \quad , \quad \alpha = \text{ damping parameter}$$

$$\omega = 0.66666666 = \text{ driving frequency}$$

Let us now to illustrate the features and ideas of this system.

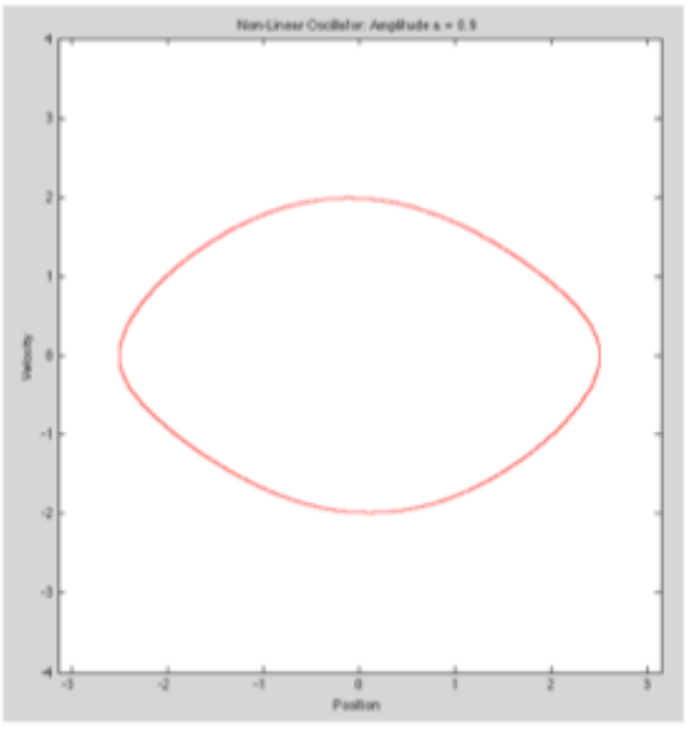
First investigate motion of physical system in phase space.

Fix some parameters as indicated on last slide.

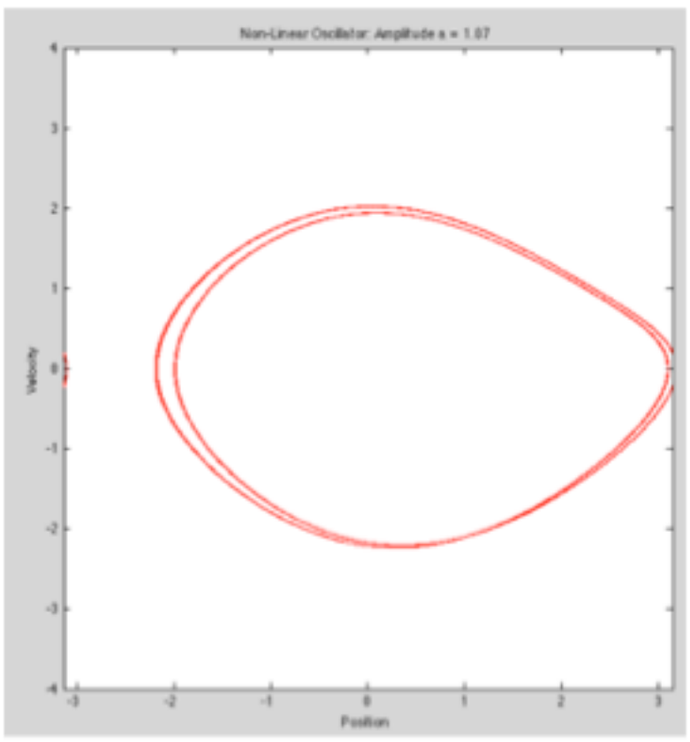
Use as variable (control) parameter (like λ in logistic map) constant a where $a = \gamma/m$. In particular, look at

strength of driving force

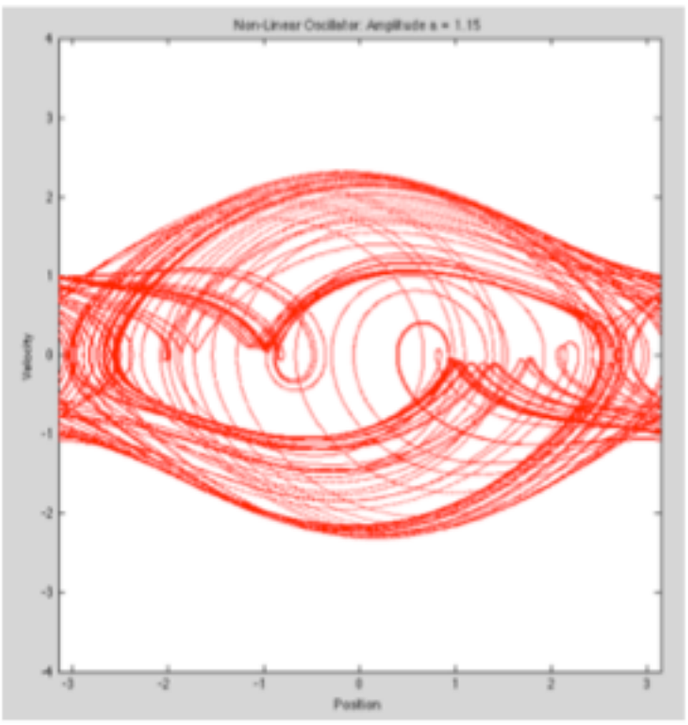
$$
\begin{aligned}
a = 0.90 &\rightarrow \text{ periodic motion} \\
a = 1.07 &\rightarrow \text{ periodic doubling} \\
a = 1.15 &\rightarrow \text{ chaotic motion} \\
a = 1.35 &\rightarrow \text{ periodic motion} \\
a = 1.45 &\rightarrow \text{ periodic doubling} \\
a = 1.47 &\rightarrow \text{ periodic doubling} \\
a = 1.50 &\rightarrow \text{ chaotic motion}
\end{aligned}
$$



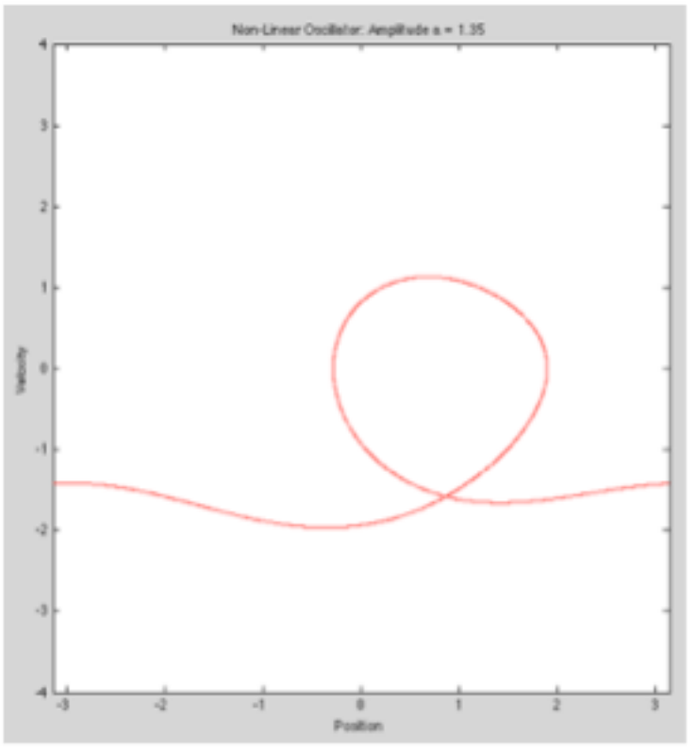
$a = 0.90$ — periodic motion



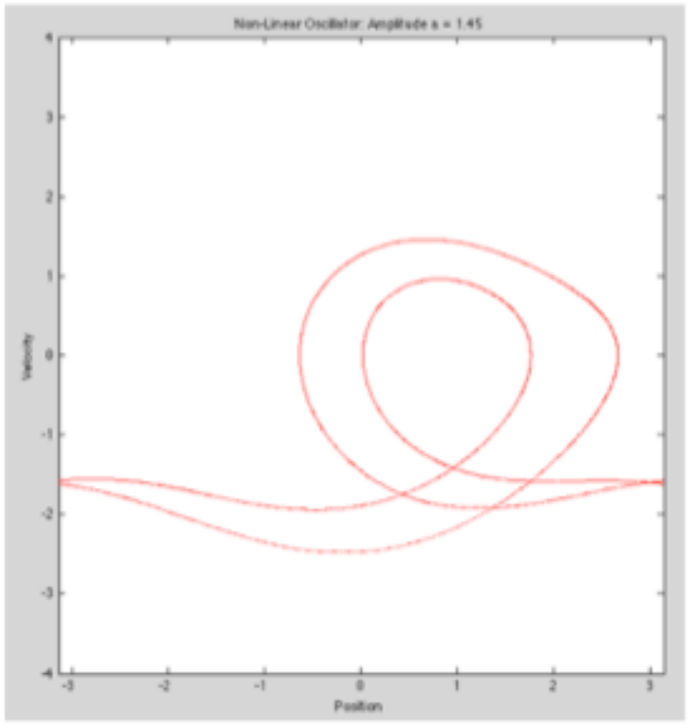
$a = 1.07$ — periodic doubling



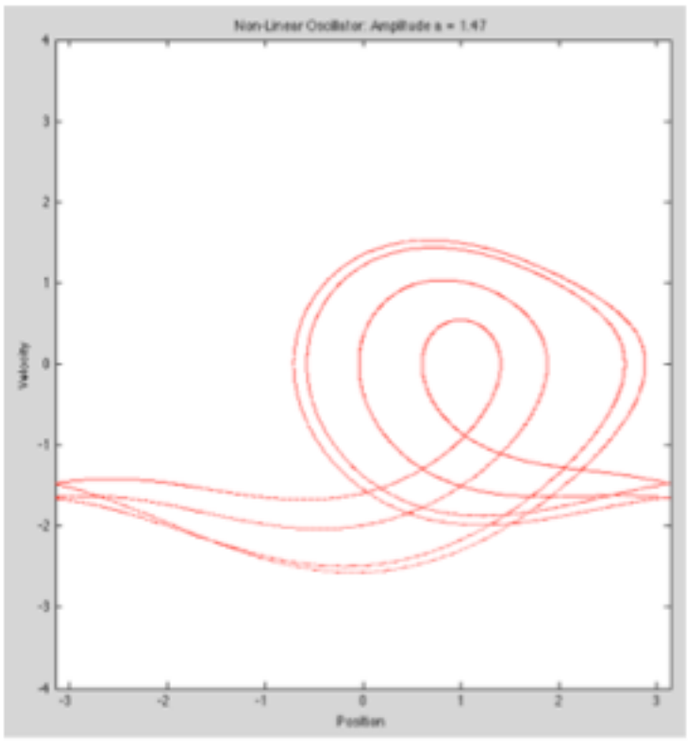
$a = 1.15$ — chaotic motion



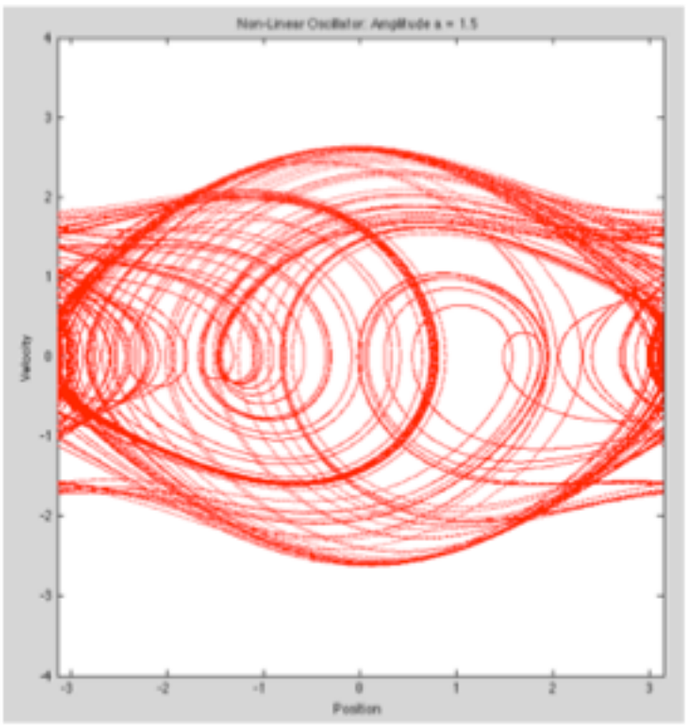
$a = 1.35$ — periodic motion



$a = 1.45$ — periodic doubling



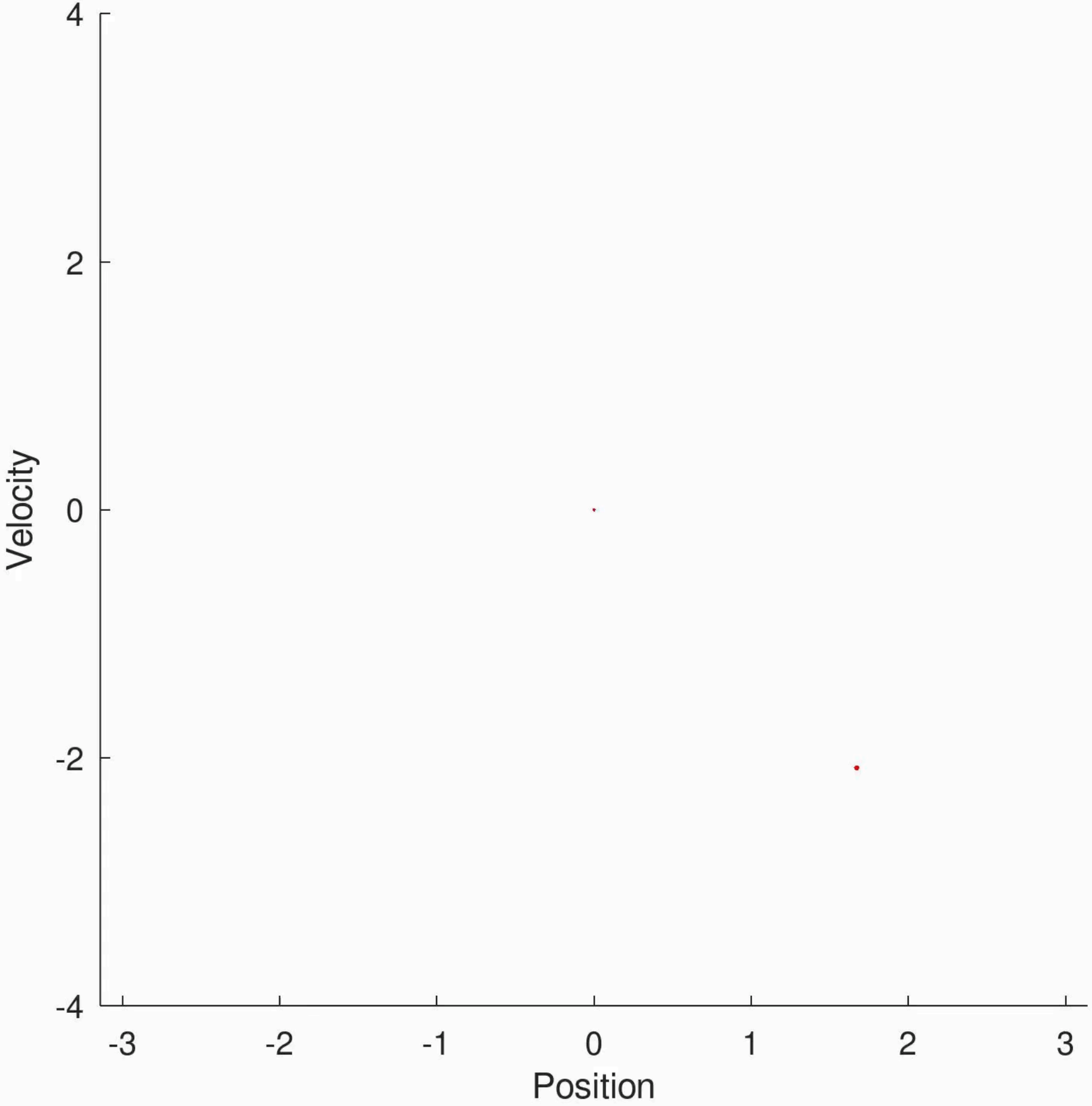
$a = 1.47$ — periodic doubling

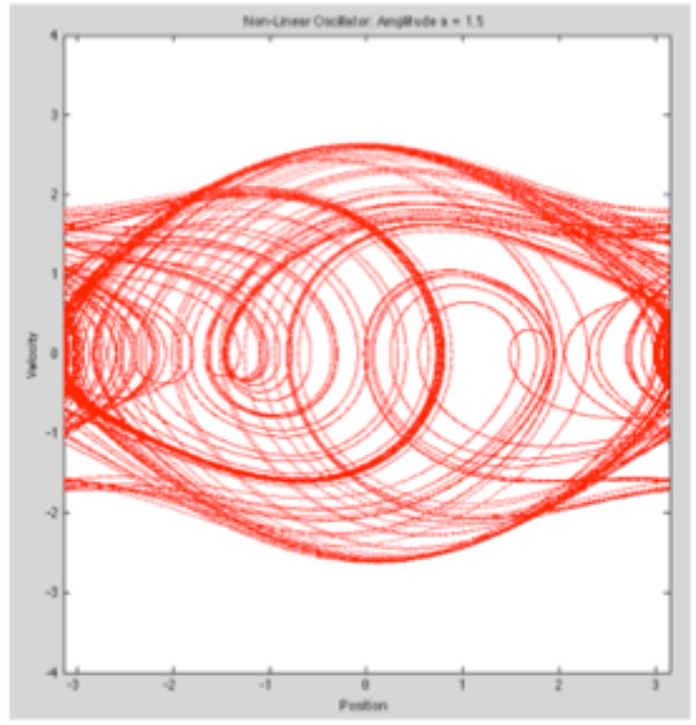


$a = 1.50$ — chaotic motion

Dependence of system motion on amplitude is clearly very complex and sensitive to value.

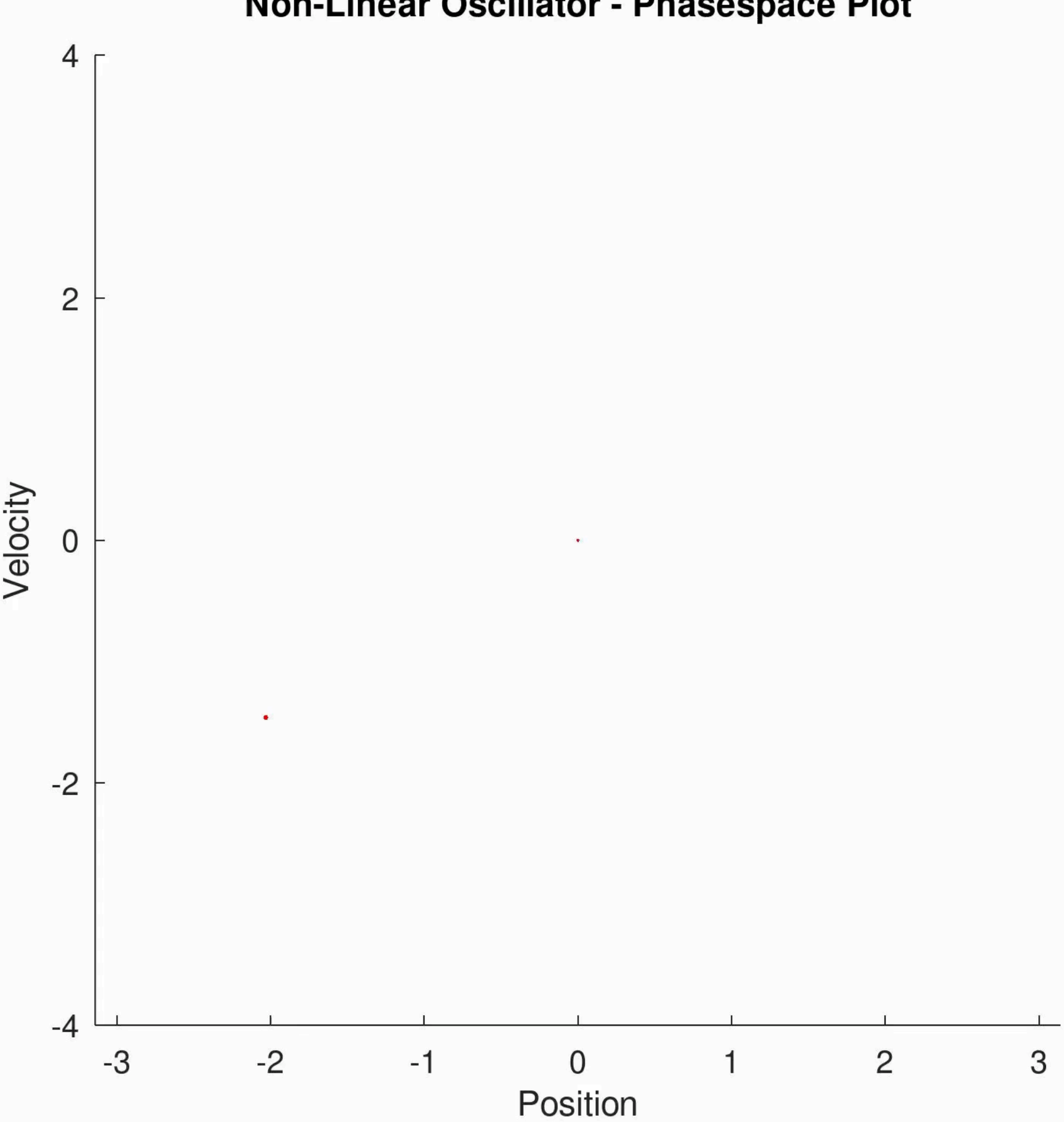
Non-Linear Oscillator - Phasespace Plot

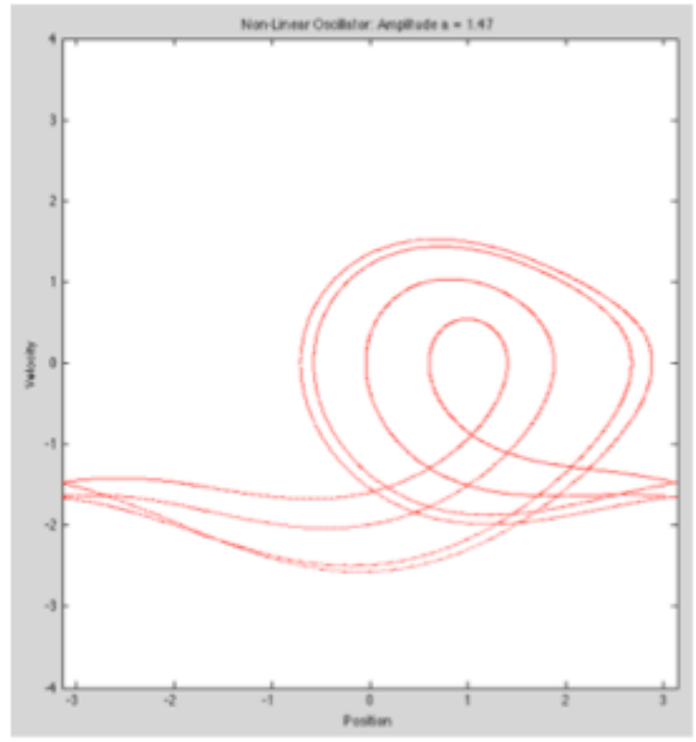




$a = 1.50$ — chaotic motion

a = 1.50





$a = 1.47$ — periodic doubling

a=1.47

Another way to visualize behavior of systems is via **Poincare Plots**.

Poincare plot is same as using a stroboscope on motion.

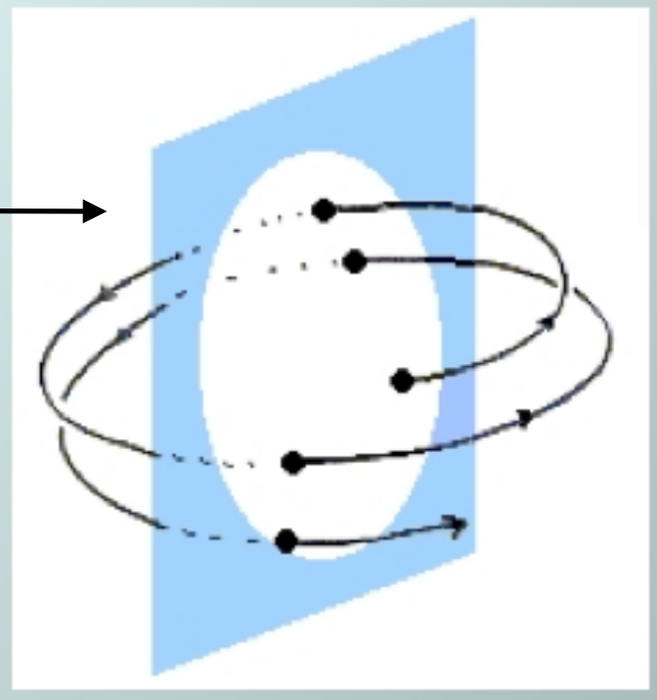
Flash strobe once every cycle of driving force (frequency = ω).

Idea shown here:

see only DOTs →

Poincaré Surface of Section

$$m\frac{d^2x}{dt^2} = -kx + A\cos 2\pi\, ft$$



System:

$$\frac{dx}{dt} = v$$

$$\frac{dv}{dt} = -\frac{k}{m}x + A\cos\theta$$

$$\frac{d\theta}{dt} = 2\pi f$$

Regular orbit movie (Henon-Heiles equations)

Poincare Plots

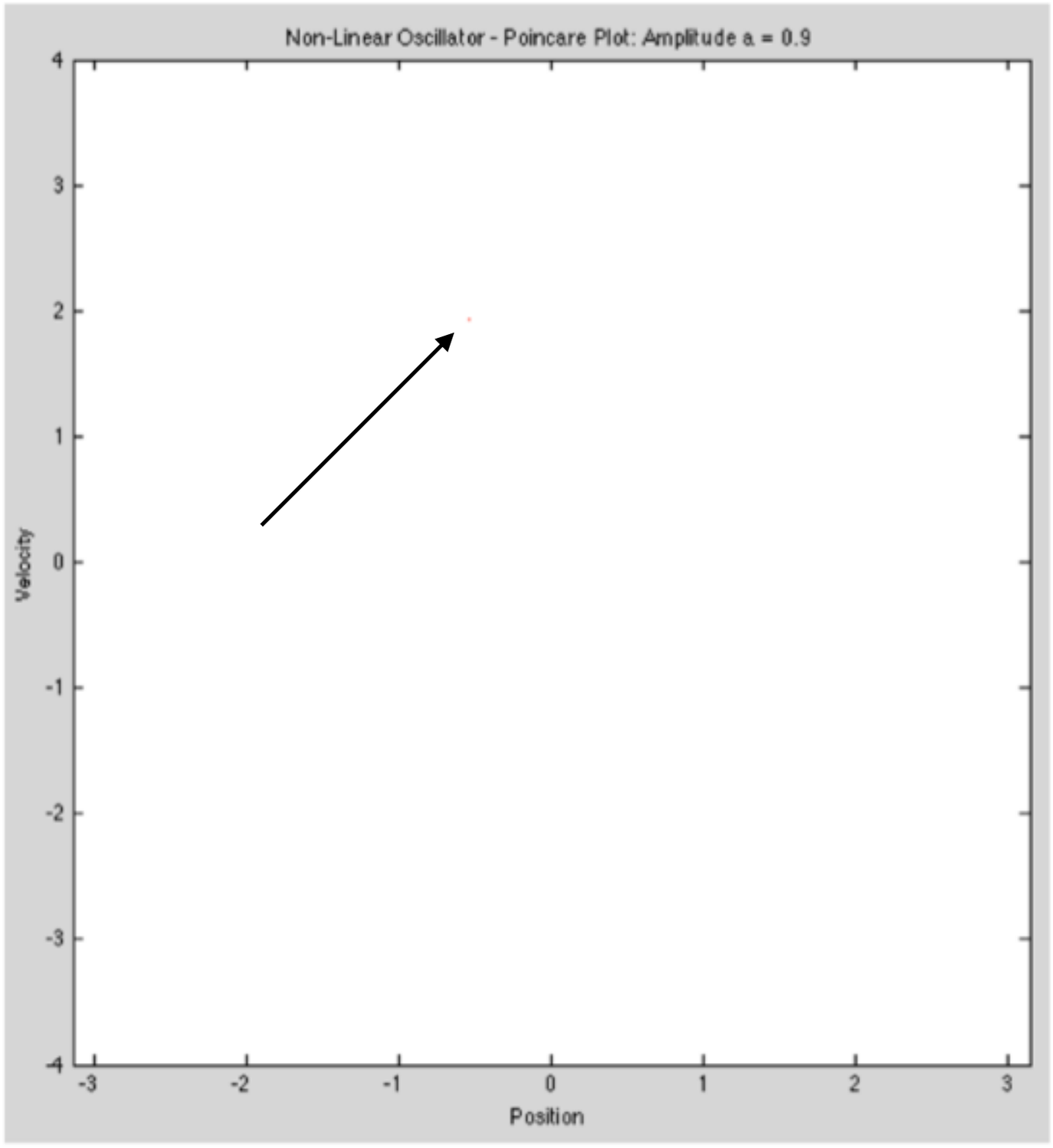


Figure 1.85: $a = 0.90$ — periodic motion

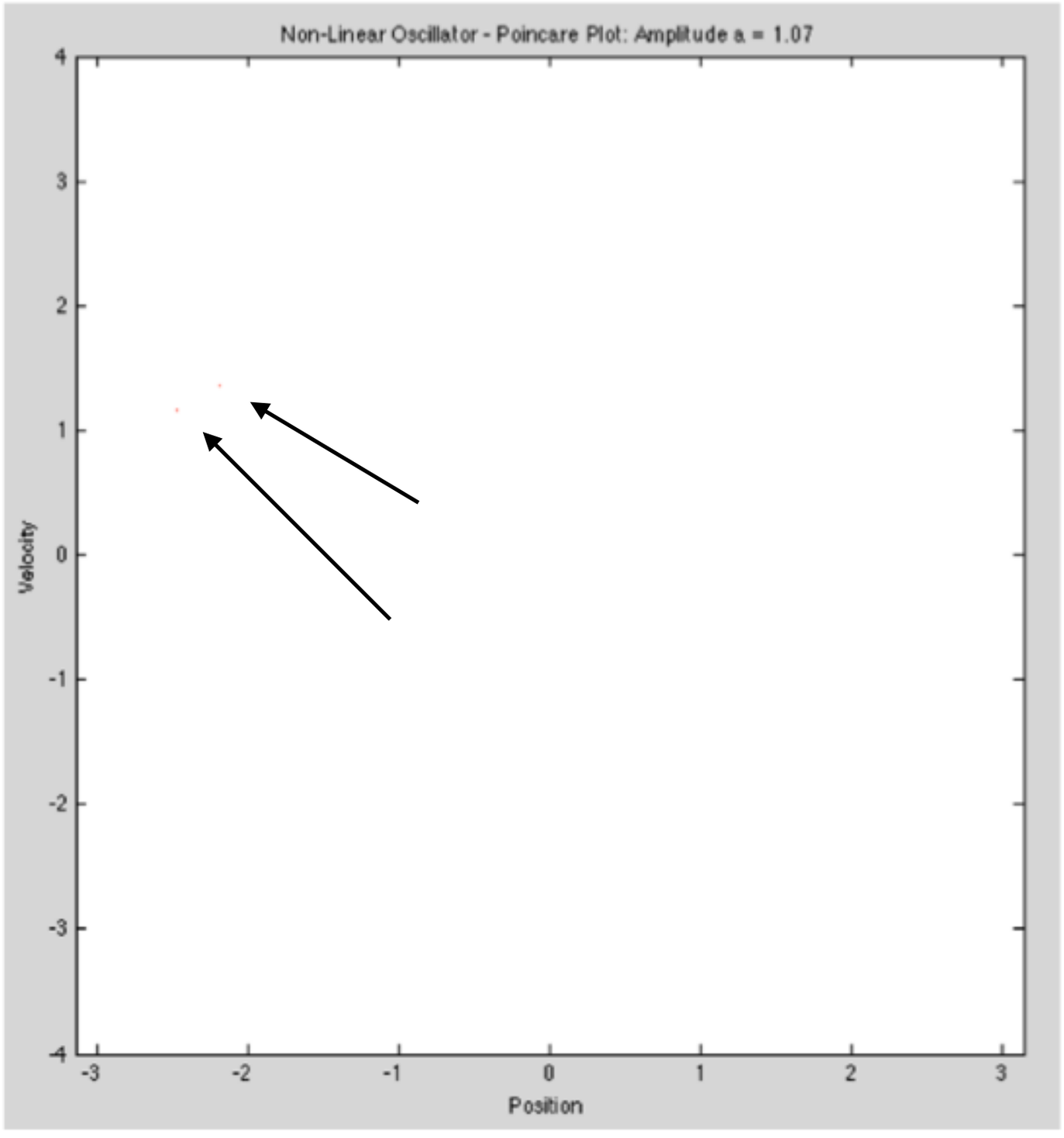


Figure 1.86: $a = 1.07$ — periodic doubling

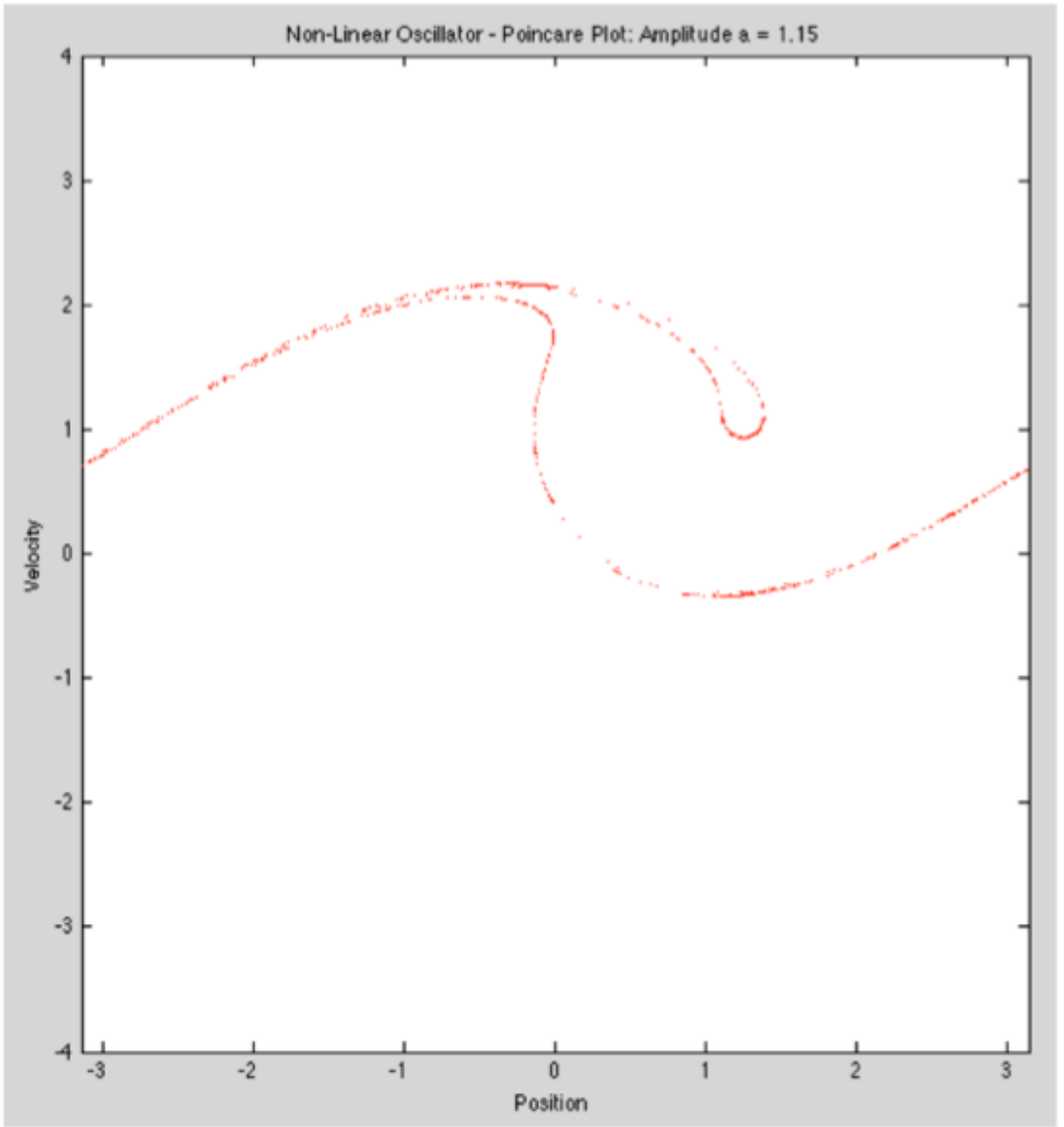


Figure 1.87: $a = 1.15$ — chaotic motion

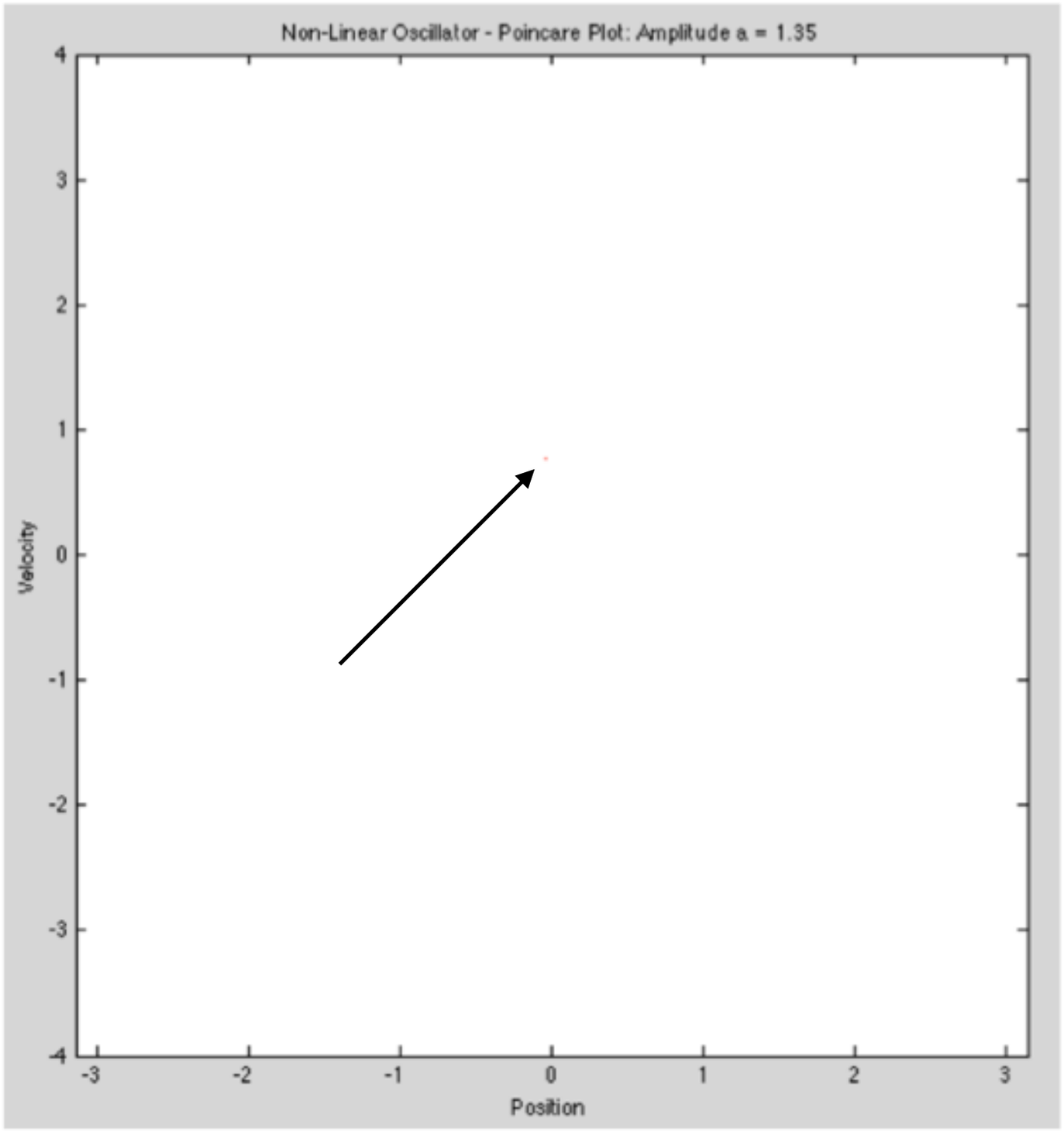


Figure 1.88: $a = 1.35$ — periodic motion

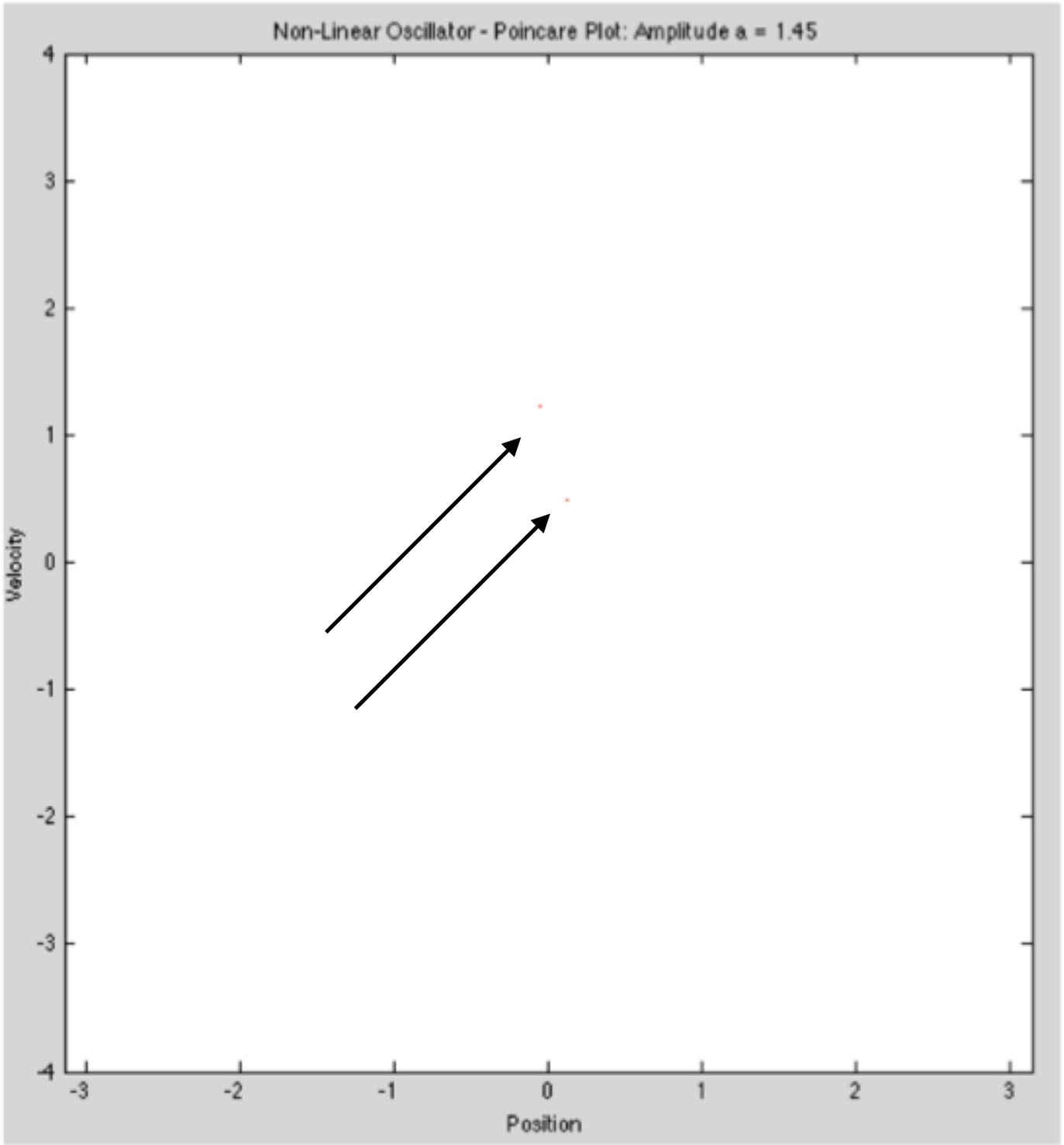


Figure 1.89: $a = 1.45$ — periodic doubling

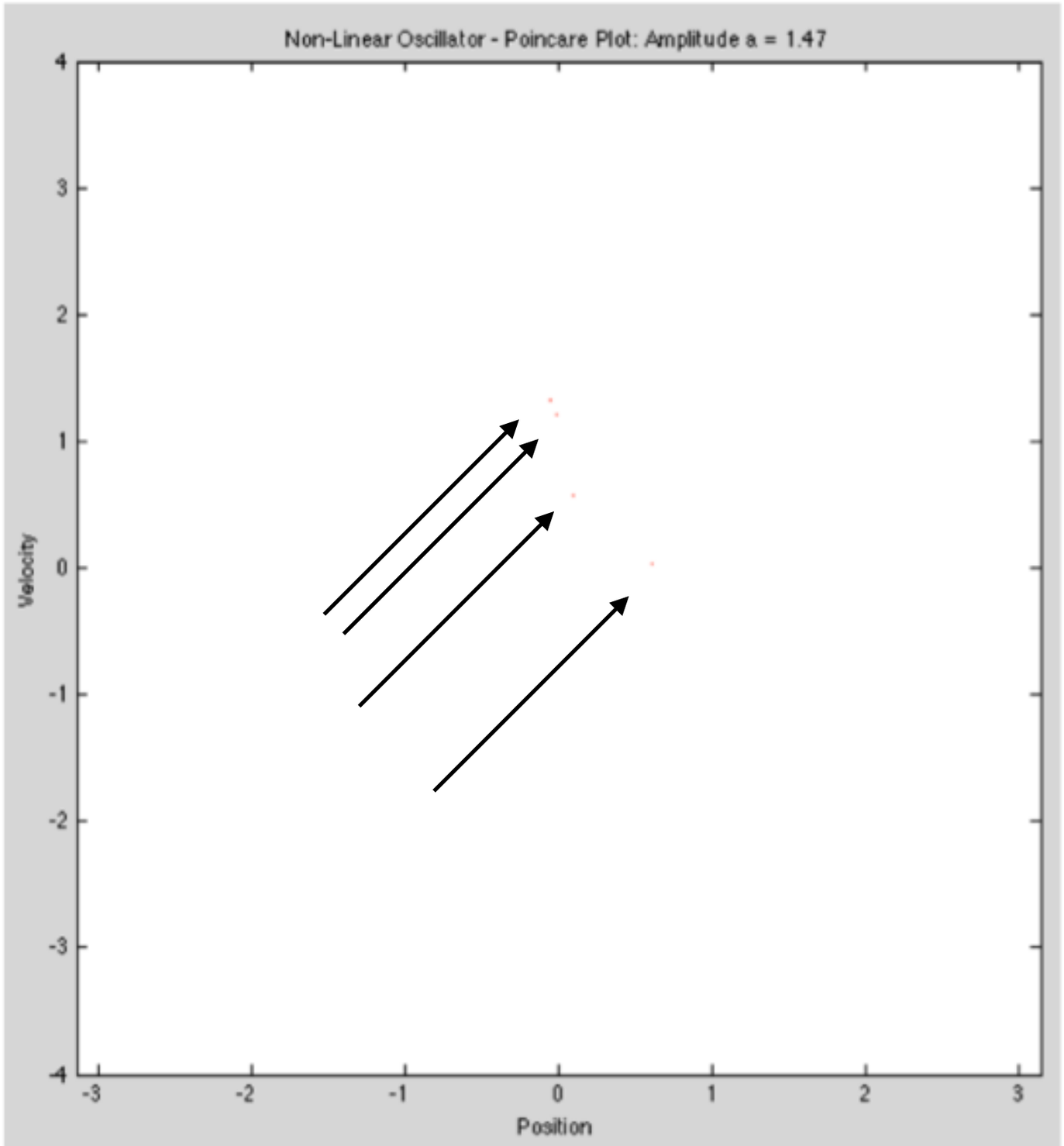


Figure 1.90: $a = 1.47$ — periodic doubling

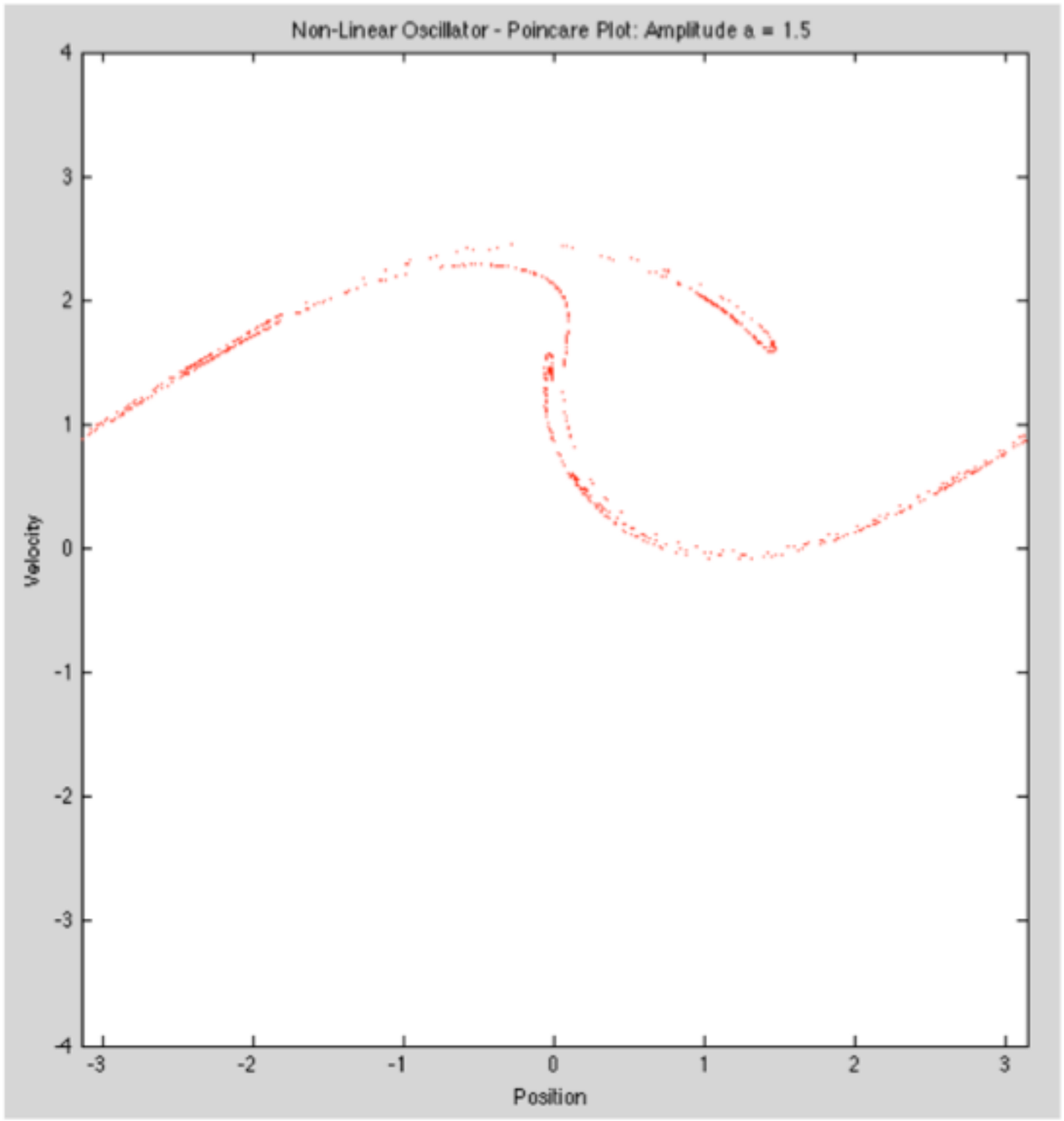


Figure 1.91: $a = 1.50$ — chaotic motion

Poincare plot in this case is attractor with infinite number of points.

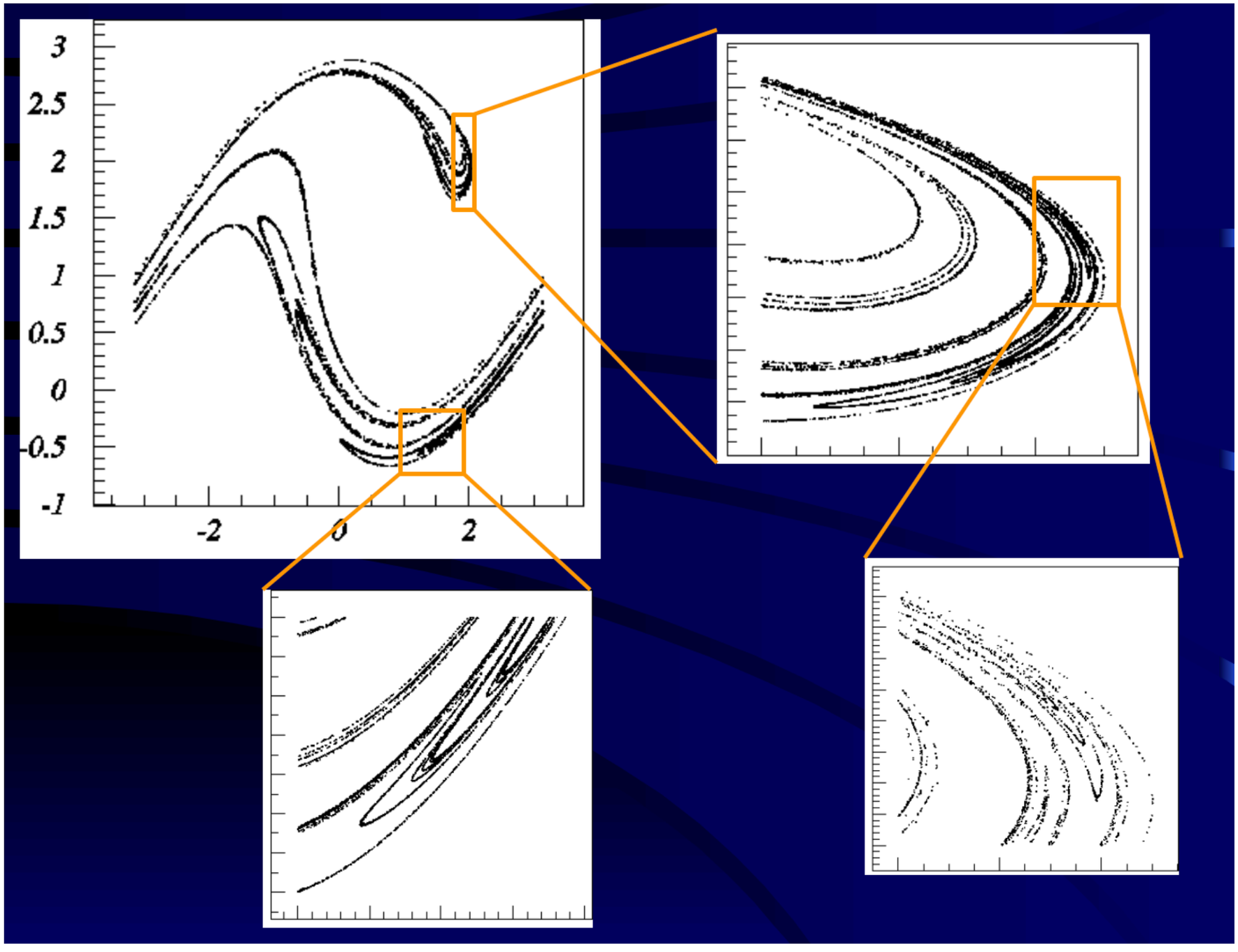
It is **fractal curve**(meaning later) with non-integer dimension.

Steady state motion of oscillator at this a and ω is not periodic at all; motion is chaotic.

Attractor of this sort is known as **strange attractor**.

Its infinity of points are arranged in strange **self-similar (fractal)** manner as we will see.

Zooming in



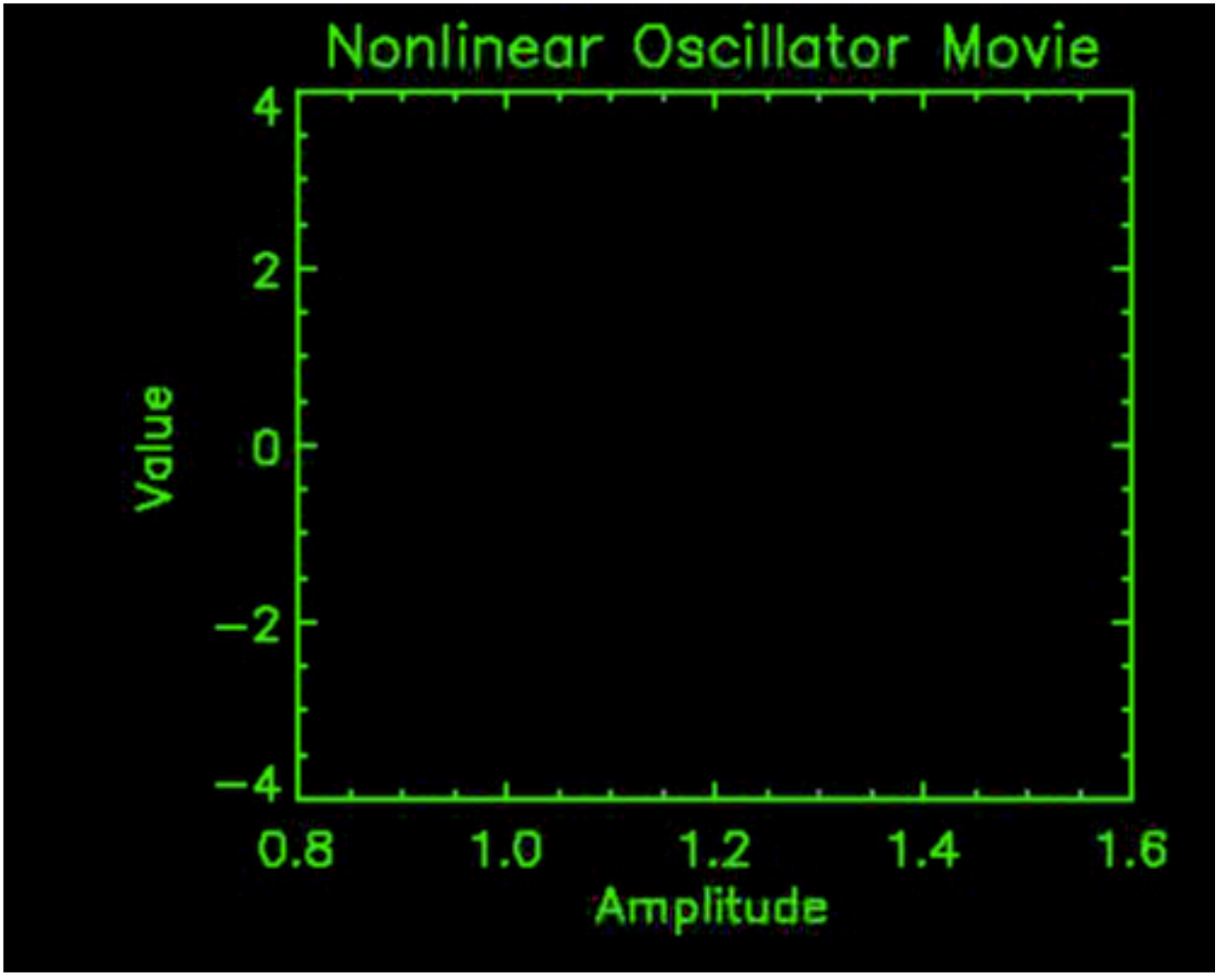
Zoom in on strange attractor.

Consider Poincare plot for driven oscillator when a = 1.50.

Zoomed images are as before.

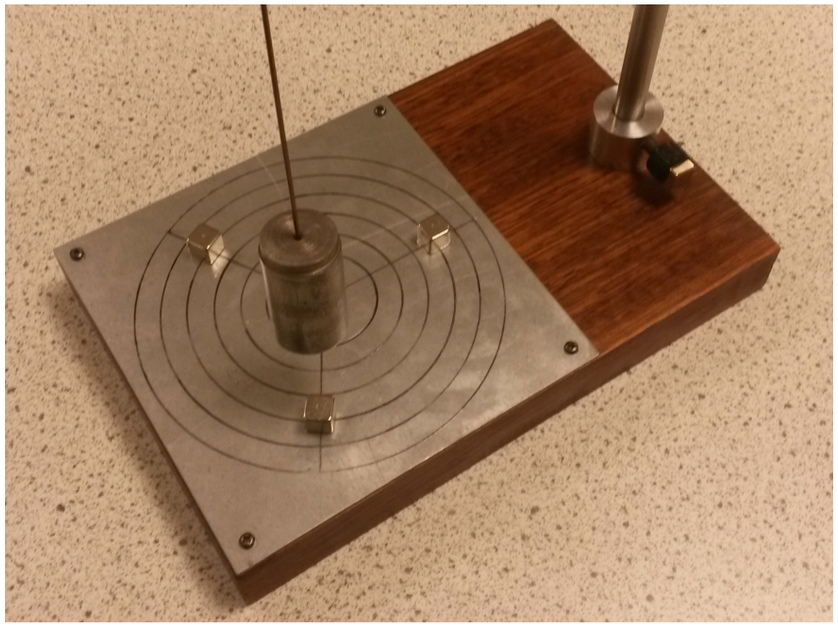
Clear that there is complex structure(self-similarity) in strange attractor or **fractal** (more later) at all levels.

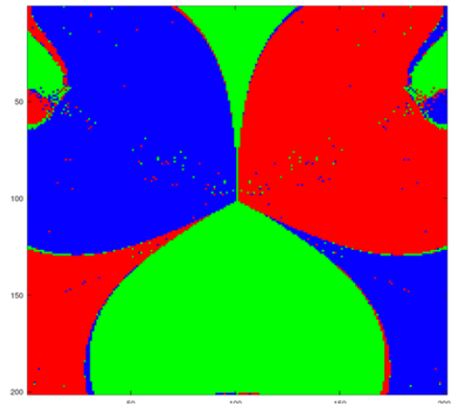
Fractal dimension of this strange attractor is D = 1.4954....(more later).



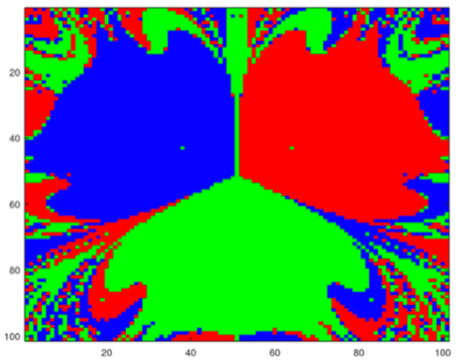
Some features should be familiar from earlier examples

One more example of a chaotic system……

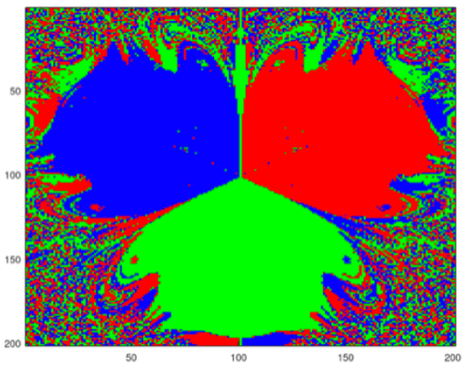




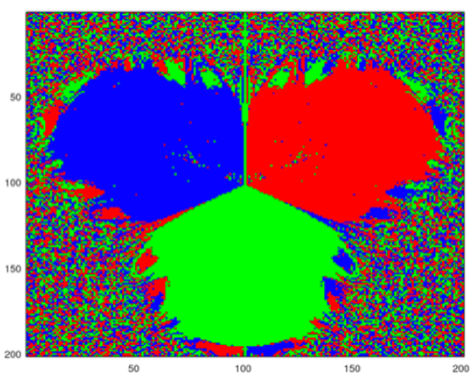
0.2 0.20



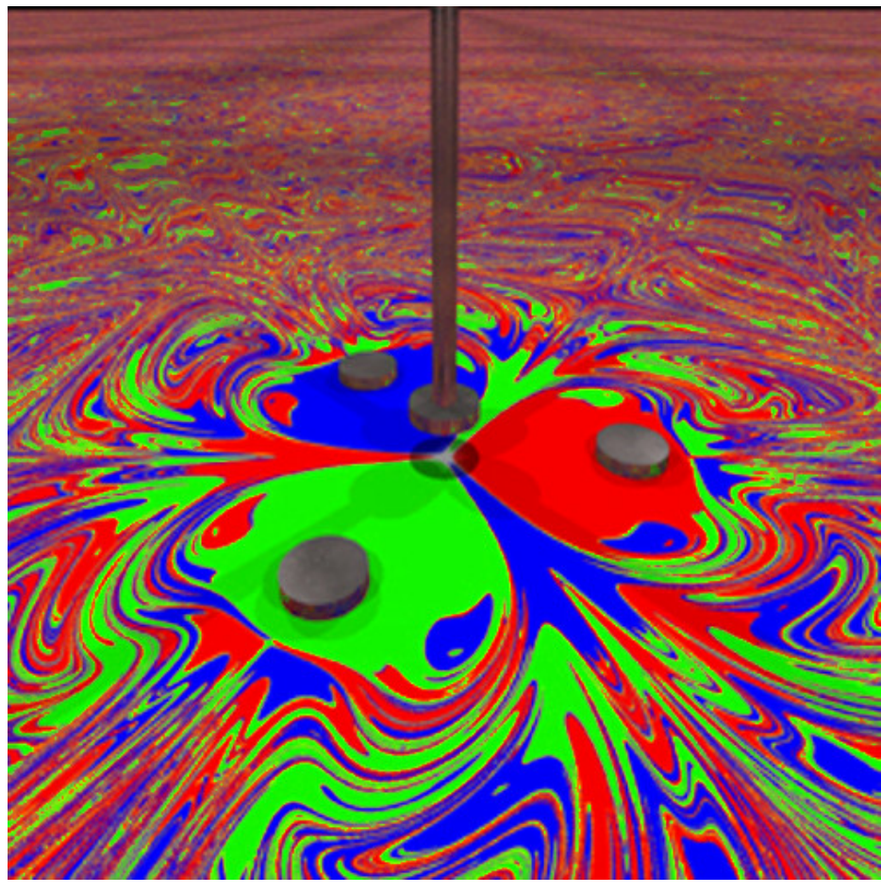
0.2 0.10



0.2 0.05



0.2 0.3



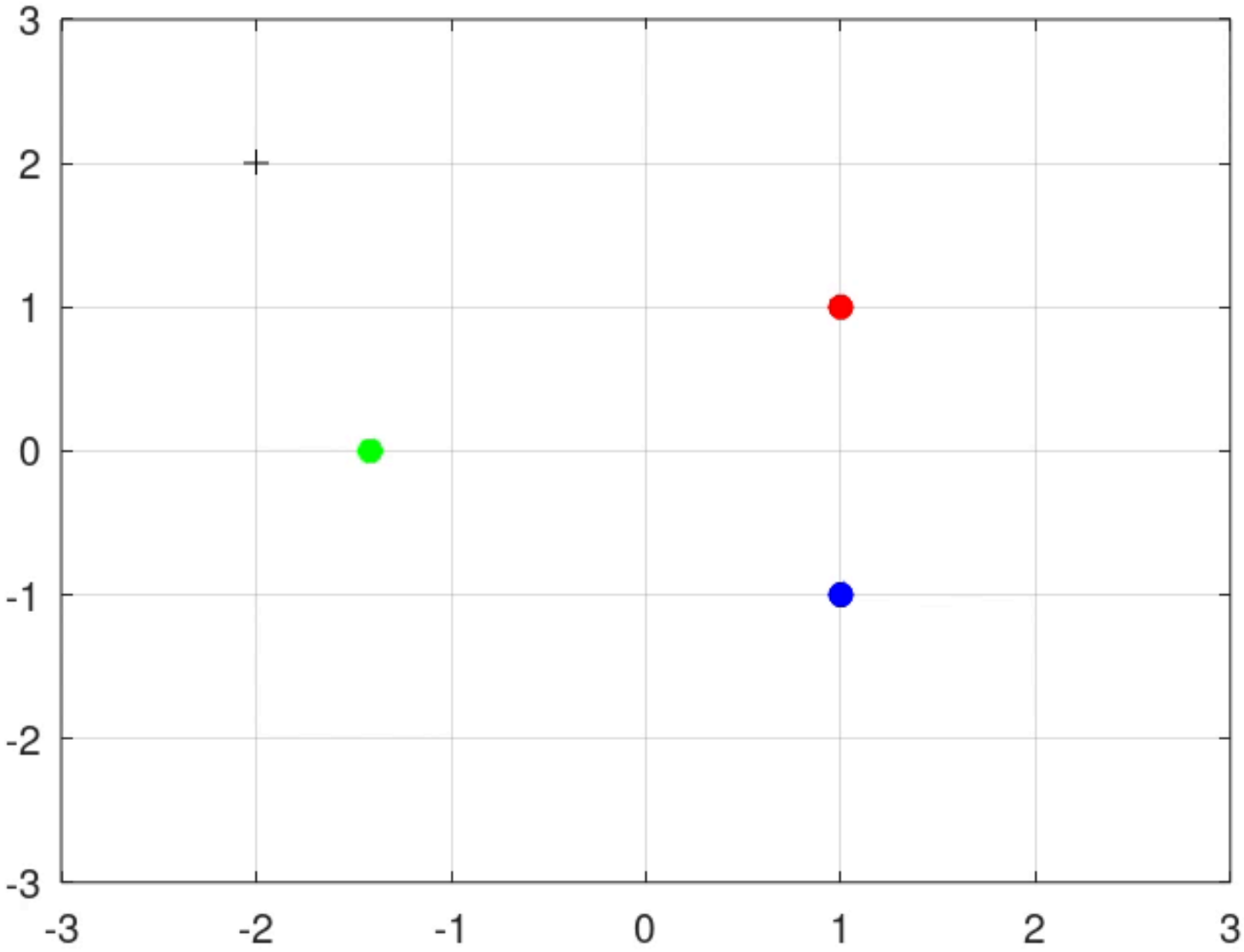
Pattern superimposed on image of device

```
xm1=[sqrt(3),1];
xm2=[-sqrt(3),1];
xm3=[0,-2];
d=0.2;
C=0.2;
R=0.03;
tmax=60;
n=100;
hh=3*ones(2*n+1,2*n+1);
colormap(prism)

couplode=@(t,y)[y(2);-R*y(2)-C*y(1)+4*bigf1(y(1),y(3),xm1,xm2,xm3,d); ....
 y(4);-R*y(4)-C*y(3)+4*bigf2(y(1),y(3),xm1,xm2,xm3,d)];

for k=-n:n
 k
 for j=-n:n
  [t,y]=ode45(couplode,[0:.1:tmax],[0.0+(3/n)*k,0,0.0-(3/n)*j,0]);
  c1=sqrt((y(tmax*10+1,1)-sqrt(3))^2+(y(tmax*10+1,3)-1)^2);
  c2=sqrt((y(tmax*10+1,1)+sqrt(3))^2+(y(tmax*10+1,3)-1)^2);
  c3=sqrt((y(tmax*10+1,1))^2+(y(tmax*10+1,3)+2)^2);
  if ((c1 < c2) && (c1 < c3))
   col=1;
  endif
  if ((c2 < c1) && (c2 < c3))
   col=5;
  endif
  if ((c3 < c2) && (c3 < c1))
   col=4;
  endif
  jj=j+n+1;
  kk=k+n+1;
  hh(kk,jj)=col;
 endfor
 hhpl=hh';
 image(hhpl)
 drawnow
endfor
```

Octave computer code


3
2
1
0
-1
-2
-3
-3
-2
-1
0
1
2
3

Now we aske: what are the shapes we have been observing?

Now must discuss Fractals

Lines, planes and circles are powerful abstractions of reality.

This kind of abstraction eliminates complexity and prevents any understanding of it from being developed.

clouds ≠ spheres

mountains ≠ cones

lightning does not travel in straight lines

Mandelbrot proposed:

There is geometry associated with complexity and it implies strange ”shapes” that have meaning in the real world.

Euclidean objects are idealizations and **cannot** represent the essence of geometry of an object in the real world.

Mandelbrot asked aq question(to illustrate the idea)

How long is coastline of Norway?

Now the coastline is not a true fractal
but is still a good example of some of its properties.

He said, in a sense, that coastline is infinitely long

or, as a better statement

that answer depends on size of your ruler because of the complex structure of the coastline..

Common sense says that a smaller ruler sees finer detail and measures longer length.

But there must be some limiting value!

If coastline was an Euclidean abstraction(a known curve),

the limiting value would be equal to its length(as it is for a real coastline).

Thinking about these ideas leads us to think about **dimension**.

Coastline idea suggests

A possible new kind of curve called **fractal** whose length is “infinite”!

Now proceed to flesh out this idea.

We must learn how to determine dimension of these fractals.

Fractal dimension will be way of measuring quantities

such as roughness or irregularity in object

A twisting coastline, despite not being able to measure its length,

has a certain degree of roughness.

We will see fractal dimension remains constant over different scales

indicating a property called **self-similarity**

Example of fractals that we have seen so far are:

the **strange attractors** in the Poincare plots of the oscillator system
and

the Lorenz “butterfly” path.

The Nature and Properties of Fractals

Fractals are mathematical sets of point with fractional dimension.

Fractals are not simple curves in Euclidean space.

A Euclidean object has integral dimension equal to dimension of space the object is being drawn in.

In addition, if Euclidean line connects 2 points in 3-dimensions

and also stays within finite volume,

then length of the line is **finite**.

A Fractal is a line that stays within finite volume,

but length = infinity.

This implies there must exist complex structure(hidden) at all levels of magnification.

In comparison, a Euclidean curve eventually

looks like straight line at some magnification level.

To be able to describe this property of fractal, we must generalize the concept of dimension.

Standard dimensionality D of space

defined as number of coordinates needed to determine unique point in space.

Defined this way,

it only allows values for D that are the non-negative integers 0, 1, 2, 3, ...

Several ways concept of dimension is redefined

—> still takes on non-negative integer values for systems described above,

but also takes on non-negative real number values for some real systems!

Adopt a simplified version of Hausdorff dimension called box counting or capacity dimension.

In box-counting scheme,

dimension of object determined by asking

how many “boxes” are needed to cover object.

Where , the appropriate “boxes” for coverage are lines, squares, cubes, etc (as we shall see).

Size of “boxes” is repeatedly decreased

and dimension of object is determined by

how number of covering “boxes” **scales** with length of the side of box.

Repeating:

fractal: Irregular shape with self-similarity

it has infinite detail

and has no derivatives.

Wherever chaos, turbulence, and disorder are found, fractal geometry is at play.

fractal dimension:

Measure of geometric object - can take on fractional values.

Usually meant Hausdorff dimension.

Fractal dimension is now more a general term for measure of

how fast length, area, or volume increases with decrease in scale.

The Concept of Dimension

Have used concept of dimension in two ways:

1. 3 dimensions of Euclidean space (D = 1, 2, 3)

2. Number of variables in a dynamic system

Fractals(irregular geometric objects) require a third meaning.

1 dimension: consider line of length ℓ .

To cover it, we need

1 box(a line) of length ℓ
2 boxes of length $\ell/2$
.....
2^m boxes of length $\ell/2^m$
.....



.....

If we define $\delta_m =$ length of the m^{th} box, ≈ $\delta_m = \dfrac{\ell}{2^m}$ m=0, 1, 2, ...

Thus, number of boxes $N(\delta_m)$ scales (grows) as $N(\delta_m) = \left(\dfrac{\ell}{\delta_m}\right)$ ← Note: This is $(\ldots)^1$

In two dimensions, consider square of side ℓ .

To cover it, need

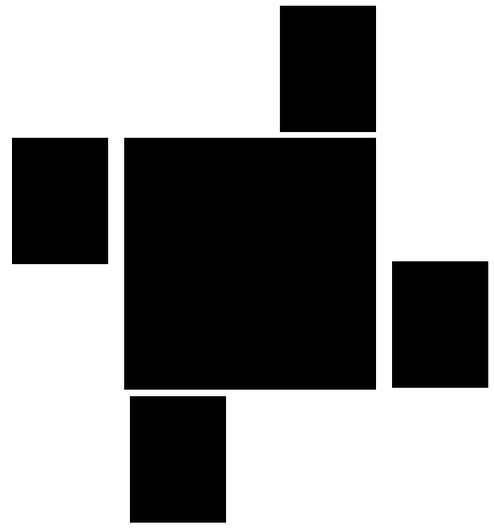
1 box(square now) of area ℓ^2

4 boxes of area $(\ell/2)^2$

.....

2^m boxes of area $(\ell/2^m)^2$

.....



Thus, number of boxes $N(\delta_m)$ scales (grows) as

$$N(\delta_m) = \left(\frac{\ell}{\delta_m}\right)^2$$

Noting power 1 and 2 that appear,

we generalize to D integer dimensions, where we would have

$$N(\delta_m) = \left(\frac{\ell}{\delta_m}\right)^D$$

Now after some algebra

details of algebra are not important —— the final result is important!

$$D = \lim_{m \to \infty} \frac{\log\left(N(\delta_m)\right)}{\log\left(\frac{1}{\delta_m}\right)} = \lim_{\epsilon \to 0} \frac{\log\left(N(\epsilon)\right)}{\log\left(1/\epsilon\right)}$$

where $N(\epsilon) =$ number p–dimensional cubes of side ϵ

needed to completely cover set.

⟶ Hausdorff or fractal dimension D of set of points in p–dimensional space.

Wow!

A very mathematical/tedious derivation.

We will make meaning of result clear by examples.

An **alternative** way to think about it.

Always good to look at a new concept in several ways!

If take object residing in Euclidean dimension D

and reduce linear size by $1/r$ in each spatial direction,

its measure (length, area, or volume) would increase to $N = r^D$ times original.

Pictured in figure below.

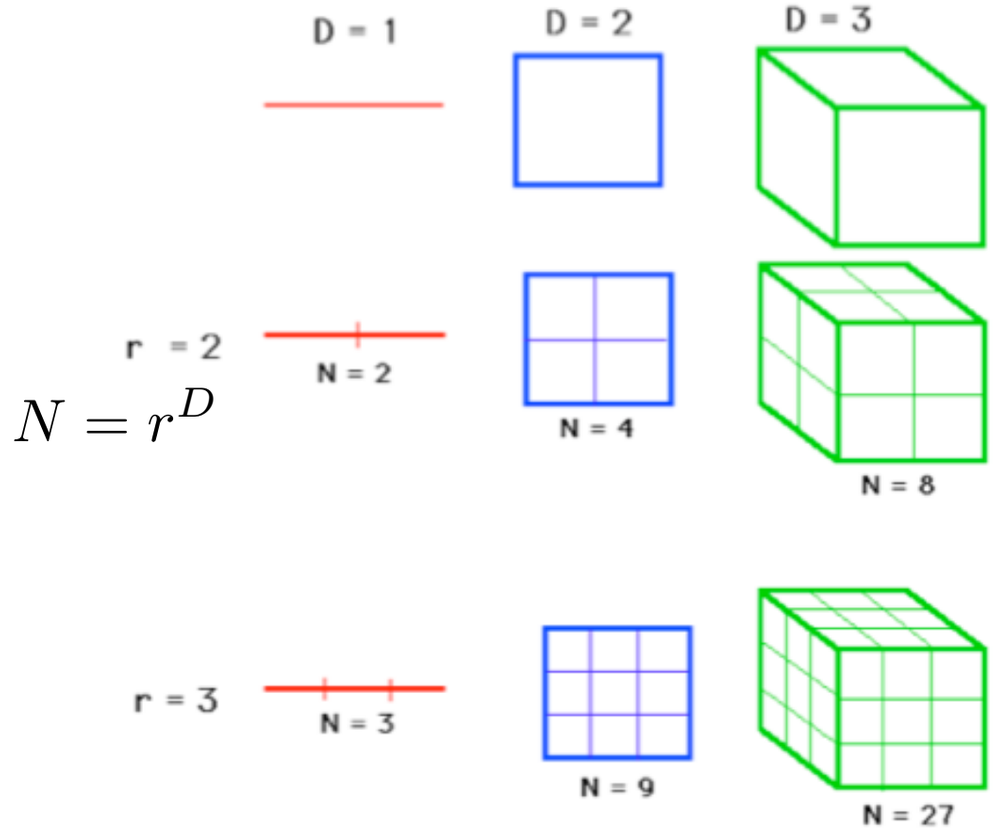


Figure 2.1: size reductions

Consider $N = r^D$, take log of both sides and get

$$\log(N) = D\log(r)$$

Solve for D to get

$$D = \frac{\log(N)}{\log(r)}$$

Plot log(N) versus log(r) —> D is slope

Examples:

(1) Single point: only one cube is required. This means that

$$N(\epsilon) = 1 \rightarrow \log N(\epsilon) = 0 \rightarrow D = 0$$

as expected.

(2) A line of length ℓ : Number of cubes of side ϵ required = number of line segments of length ϵ . Therefore

$$N(\epsilon) = \frac{\ell}{\epsilon}$$

and

$$D = \lim_{\epsilon \to 0} \frac{\log\left(\frac{\ell}{\epsilon}\right)}{\log\left(\frac{1}{\epsilon}\right)} = \lim_{\epsilon \to 0} \frac{\log\left(\frac{1}{\epsilon}\right) + \log\left(\ell\right)}{\log\left(\frac{1}{\epsilon}\right)} = \lim_{\epsilon \to 0} \left(1 + \frac{\log\left(\ell\right)}{\log\left(\frac{1}{\epsilon}\right)}\right) = 1$$

as expected.

(3) A square of area ℓ^2 : Number of cubes of side ϵ required = number of squares of side ϵ . Therefore

$$N(\epsilon) = \frac{\ell^2}{\epsilon^2}$$

$$D = \lim_{\epsilon \to 0} \frac{\log\left(\frac{\ell^2}{\epsilon^2}\right)}{\log\left(\frac{1}{\epsilon}\right)} = \lim_{\epsilon \to 0} \frac{2\log\left(\frac{1}{\epsilon}\right) + \log\left(\ell^2\right)}{\log\left(\frac{1}{\epsilon}\right)}$$

$$= \lim_{\epsilon \to 0} \left(2 + 2\frac{\log\left(\ell\right)}{\log\left(\frac{1}{\epsilon}\right)}\right) = 2$$

as expected.

Our definition of dimension works for Euclidean objects

and thus clearly makes sense.

Now for some mathematical world and real world examples.

The Cantor Set

Consider figure below.

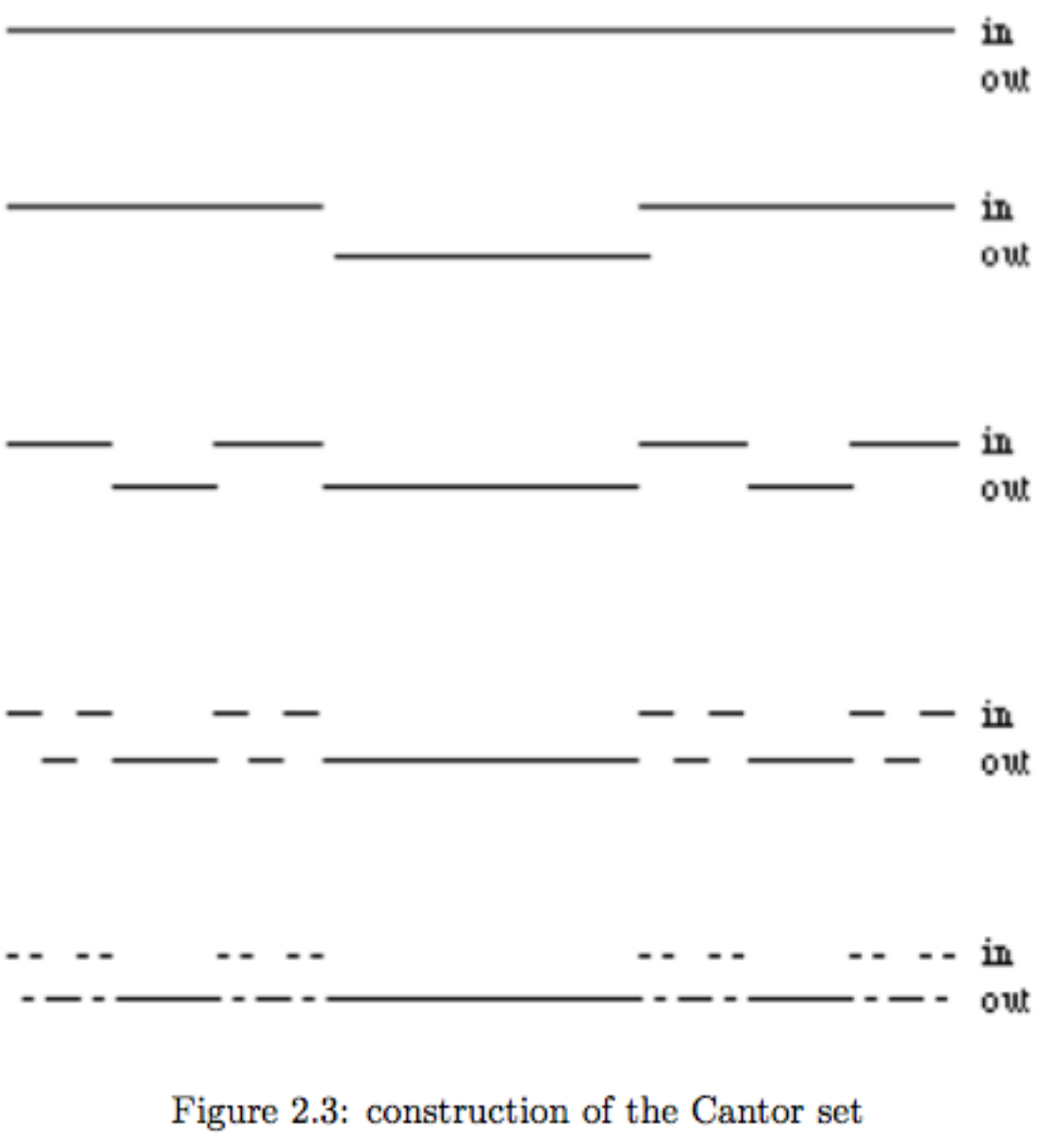


Figure 2.3: construction of the Cantor set

In diagram, points in interval (0,1) tracked for 4 iterations.

1st iteration, middle third of points leave interval (0, 1).

Middle thirds of the two **in** segments leave interval and so on.

Bottom row(after infinite iterations) is Cantor set.

Cantor set has some stunning properties.

For one thing, in infinite limit it has zero measure (it is the empty line),

but set is far from empty,
in fact, has an uncountable number of points.

called “Cantor dust”

To cover the set at step K we need

$$N(K) = 2^K \text{ cubes of side } \epsilon = \left(\frac{1}{3}\right)^K$$

Therefore,

$$D = \lim_{K\to\infty} \frac{\log\left(2^K\right)}{\log\left(\left(\frac{1}{3}\right)^K\right)} = \lim_{K\to\infty} \frac{K\log\left(2\right)}{K\log\left(3\right)} = \frac{\log\left(2\right)}{\log\left(3\right)} = 0.631$$

Dimension of Cantor set is not an integer!

You might think that this mathematical oddity

has no chance of appearing in a real world situation

it does - in the theory of evolution!! - See below

It appears as the Cantor function,

which operates on the entire unit interval

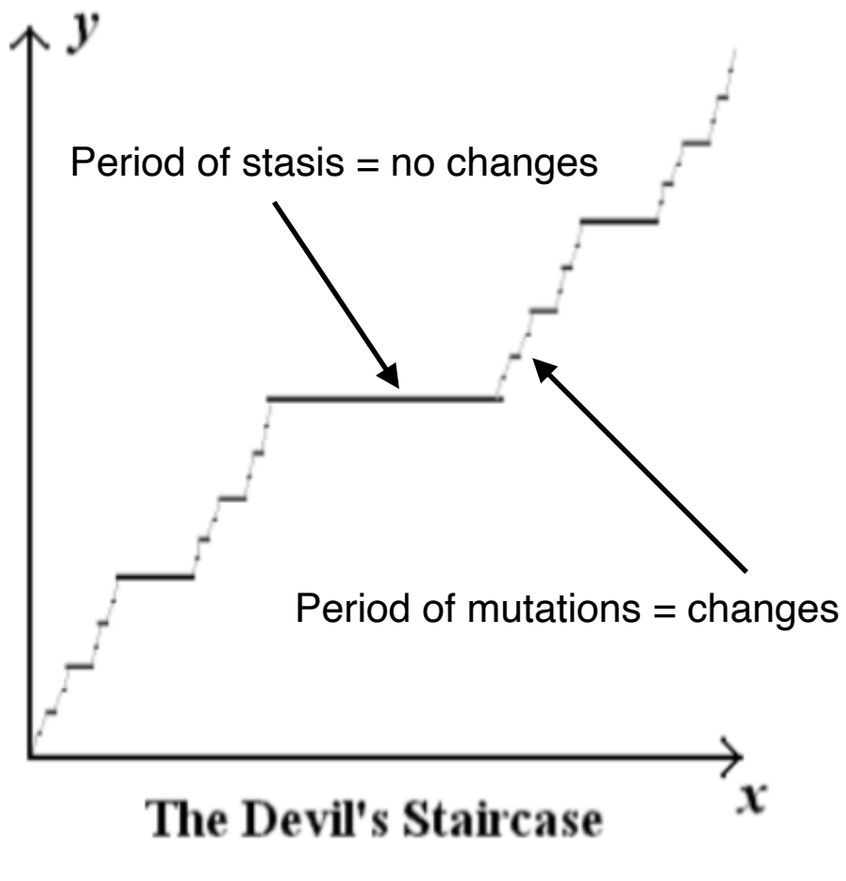
and takes it onto itself

Suppose we complete the function values

by assuming its value on missing intervals be constant values

which equal values of original function on endpoints of those intervals.

If graph this function, get:



The Devil's Staircase

—> a very accurate representation of how evolution works!

The flat parts are the images of all of the “middle thirds,” and these are all connected by the images of the Cantor set itself.

This construction has been called the “Devil’s Staircase” since it has **infinitely** many “steps.”

The Cantor set is a simple example of a fractal, demonstrating

that our geometrical intuitions about space

(even such simple spaces as the unit interval)

can fail to capture much of the deep structure inherent in those very intuitions.

The Koch Snowflake

Koch curve is constructed by recursion as exhibited in figures below. At each step middle-third of each segment is replaced with ”V” shaped bulge.

Curve turns out to have infinite length, while enclosing (together with its natural base: original line segment) a finite area.

Begin with straight line of length 1, called initiator. Then remove middle third of line, and replace it with two lines that each have same length (1/3) as remaining lines on each side.

This new form is called the generator, because it specifies rule that is used to generate a new form.

Rule says to take each line and replace it with four lines, each one-third the length of original.

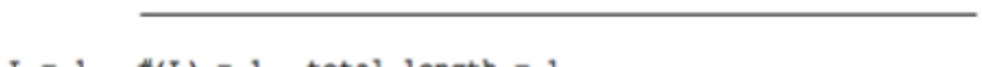


Figure 2.4: Original Line



Figure 2.5: Step #1



Figure 2.6: Step #2

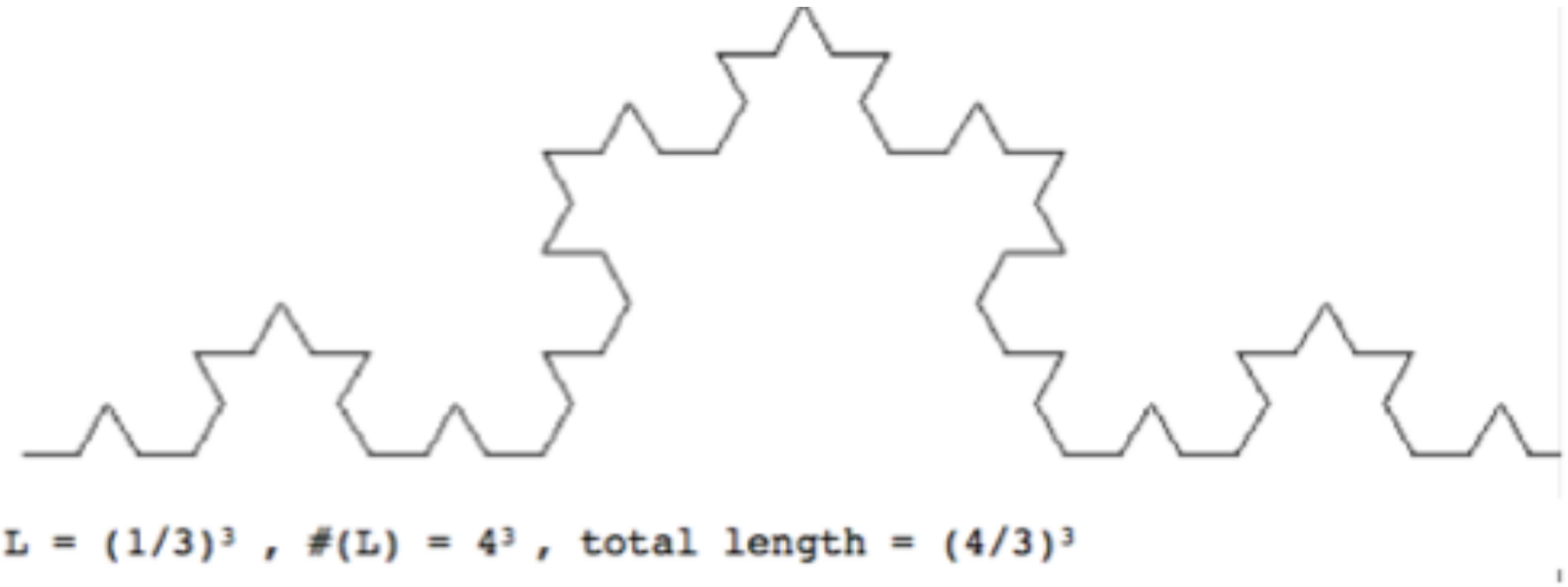
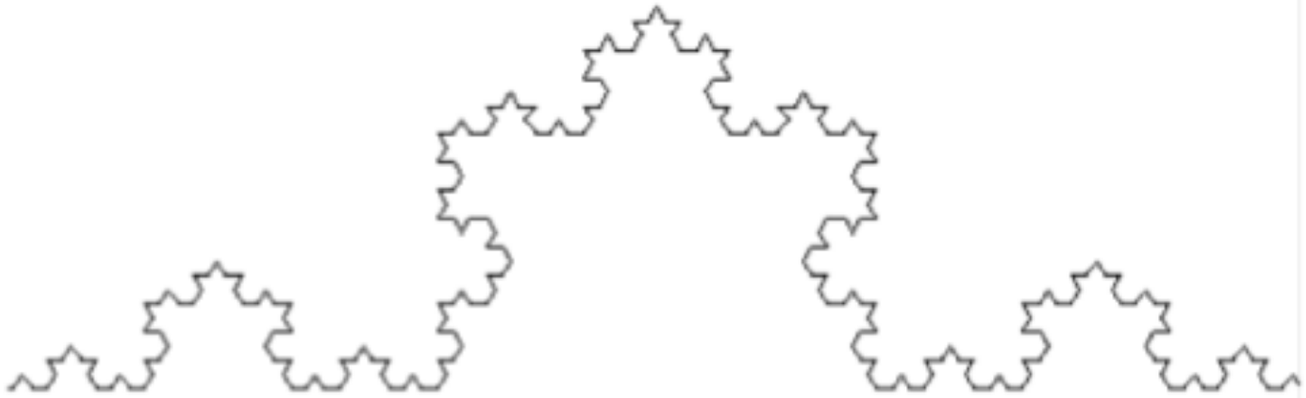


Figure 2.7: Step #3



L = (1/3)⁴ , #(L) = 4⁴ , total length = (4/3)⁴



L = (1/3)⁵ , #(L) = 4⁵ , total length = (4/3)⁴

Figure 2.9: Step #5

Dimension of Koch snowflake given by

$$D = \lim_{n\to\infty} \frac{\log\left(4^n\right)}{\log\left(\left(\frac{1}{3}\right)^n\right)} = \lim_{n\to\infty} \frac{n\log\left(4\right)}{n\log\left(3\right)} = \frac{\log\left(4\right)}{\log\left(3\right)} = 1.26....$$

Get dimension between 1 and 2. Clearly, final length of the fractal line

$$\lim_{n\to\infty} \left(\frac{4}{3}\right)^n \to \infty$$

is infinite even though it stays within a finite area of the 2-dimensional Euclidean plane!

Fractal Dimension Program

Program that carries out box dimension calculations on complex fractal structures.
Idea of covering fractal with boxes shown below.

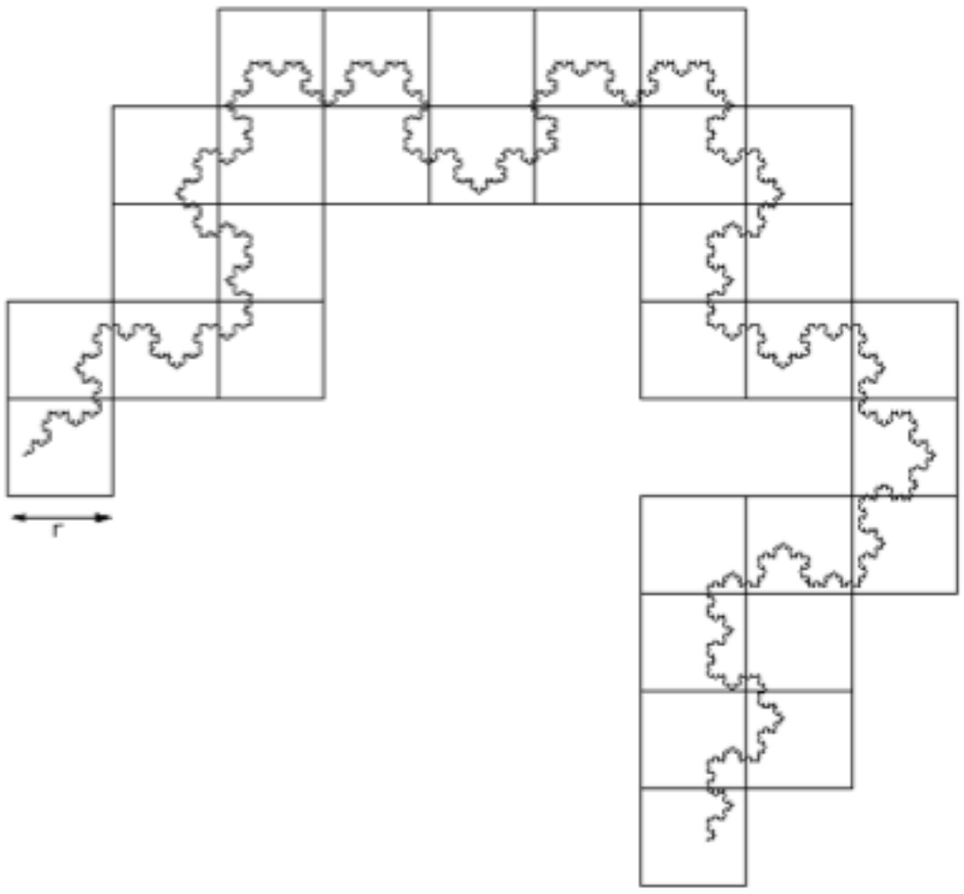


Figure 2.16: Koch curve covered by boxes of size

Program simply chooses a ”box” size, covers the fractal and counts the number of boxes.

It then reduces the box size and repeats until the box is small.

At that point it can recognize the limit of plot log N(r) versus log r and determine the slope which corresponds to the dimension.

If apply the program to Koch snowflake which looks like



Figure 2.17: Koch snowflake curve

we get result

```
reality:fdc_mac boccio$ ./fdc  -i snowflake.tga -b1 200 -b2 8 -n 1 -bf 1.2 -p
FDC - Fractal Dimension Calculator
Initialising GLUT
Creating windows
Creating menus (Right mouse button)
Reading file "snowflake.tga"
   Image dimensions: 1188 x 695 x 24
Creating screen version
   Full screen size: 2560 x 1440
   Display size:     1188 x 695
   Scale factor:     1
Processing starting
   Box size:  200 [  1], Total count = 15
   Box size:  166 [  2], Total count = 19
   Box size:  138 [  3], Total count = 22
   Box size:  114 [  4], Total count = 30
   Box size:   95 [  5], Total count = 31
   Box size:   79 [  6], Total count = 45
   Box size:   65 [  7], Total count = 58
   Box size:   54 [  8], Total count = 70
   Box size:   45 [  9], Total count = 82
   Box size:   37 [ 10], Total count = 109
   Box size:   30 [ 11], Total count = 115
   Box size:   24 [ 12], Total count = 183
   Box size:   20 [ 13], Total count = 225
   Box size:   16 [ 14], Total count = 308
   Box size:   13 [ 15], Total count = 424
   Box size:   10 [ 16], Total count = 547
   Box size:    8 [ 17], Total count = 747
Processing finished
Estimated fractal dimension:  1.211
```

The log(N)/log(r) graph looks like:

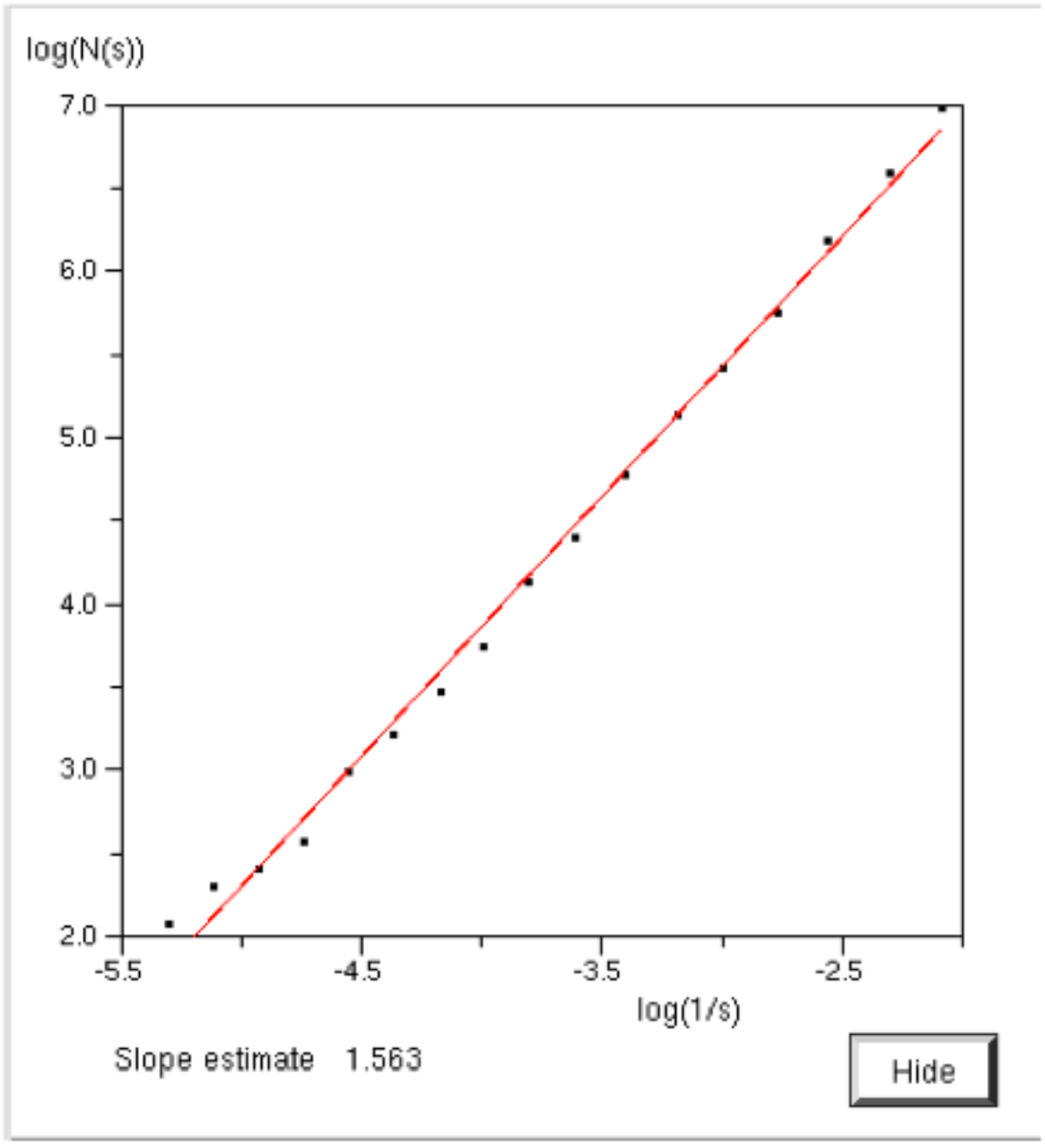


Figure 2.18: log(N)/log(r) plot

Even more dramatic, we apply program to fractal tree(created on computer) as shown below



Figure 2.19: fractal tree

The log(N)/log(r) graph looks like:

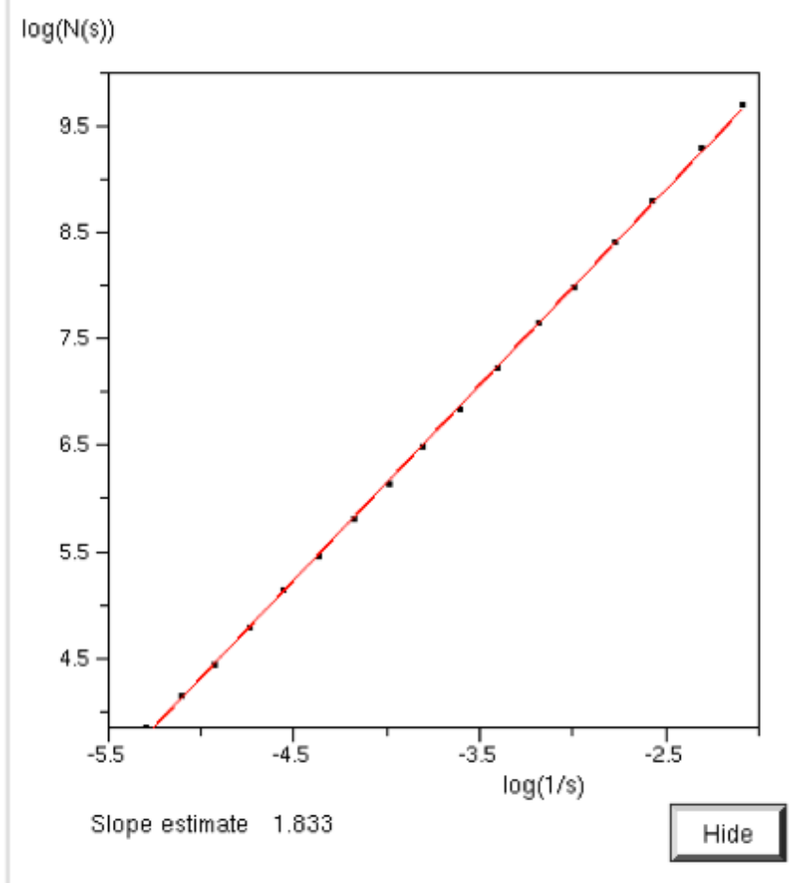


Figure 2.20: log(N)/log(r) plot

and get result

```
reality:fdc_mac boccio$ ./fdc  -i tree.tga -b1 200 -b2 8 -n 1 -bf 1.2 -
FDC - Fractal Dimension Calculator
Initialising GLUT
Creating windows
Creating menus (Right mouse button)
Reading file "tree.tga"
   Image dimensions: 2048 x 1536 x 24
Creating screen version
   Full screen size: 2560 x 1440
   Display size:     1024 x 768
   Scale factor:     2
Processing starting
   Box size:  200 [  1], Total count = 47
   Box size:  166 [  2], Total count = 63
   Box size:  138 [  3], Total count = 85
   Box size:  114 [  4], Total count = 120
   Box size:   95 [  5], Total count = 170
   Box size:   79 [  6], Total count = 232
   Box size:   65 [  7], Total count = 333
   Box size:   54 [  8], Total count = 467
   Box size:   45 [  9], Total count = 653
   Box size:   37 [ 10], Total count = 928
   Box size:   30 [ 11], Total count = 1384
   Box size:   24 [ 12], Total count = 2100
   Box size:   20 [ 13], Total count = 2947
   Box size:   16 [ 14], Total count = 4511
   Box size:   13 [ 15], Total count = 6638
   Box size:   10 [ 16], Total count = 10842
   Box size:    8 [ 17], Total count = 16332
Processing finished
Estimated fractal dimension:  1.833
```

What about a picture of a real tree?

I found a picture of a real tree and put it into program.



Figure 2.21: real tree

and result is

```
reality:fdc_mac boccio$ ./fdc  -i tree-569.tga -b1 200 -b2 8 -n 1 -b
FDC - Fractal Dimension Calculator
Initialising GLUT
Creating windows
Creating menus (Right mouse button)
Reading file "tree-569.tga"
   Image dimensions: 450 x 326 x 24
Creating screen version
   Full screen size: 2560 x 1440
   Display size:     450 x 326
   Scale factor:     1
Processing starting
   Box size:  200 [  1], Total count = 6
   Box size:  166 [  2], Total count = 7
   Box size:  138 [  3], Total count = 11
   Box size:  114 [  4], Total count = 17
   Box size:   95 [  5], Total count = 19
   Box size:   79 [  6], Total count = 23
   Box size:   65 [  7], Total count = 34
   Box size:   54 [  8], Total count = 50
   Box size:   45 [  9], Total count = 57
   Box size:   37 [ 10], Total count = 79
   Box size:   30 [ 11], Total count = 105

   Box size:   24 [ 12], Total count = 172
   Box size:   20 [ 13], Total count = 227
   Box size:   16 [ 14], Total count = 341
   Box size:   13 [ 15], Total count = 486
   Box size:   10 [ 16], Total count = 761
   Box size:    8 [ 17], Total count = 1073
Processing finished
Estimated fractal dimension:  1.563
```

The log(N)/log(r) graph looks like:

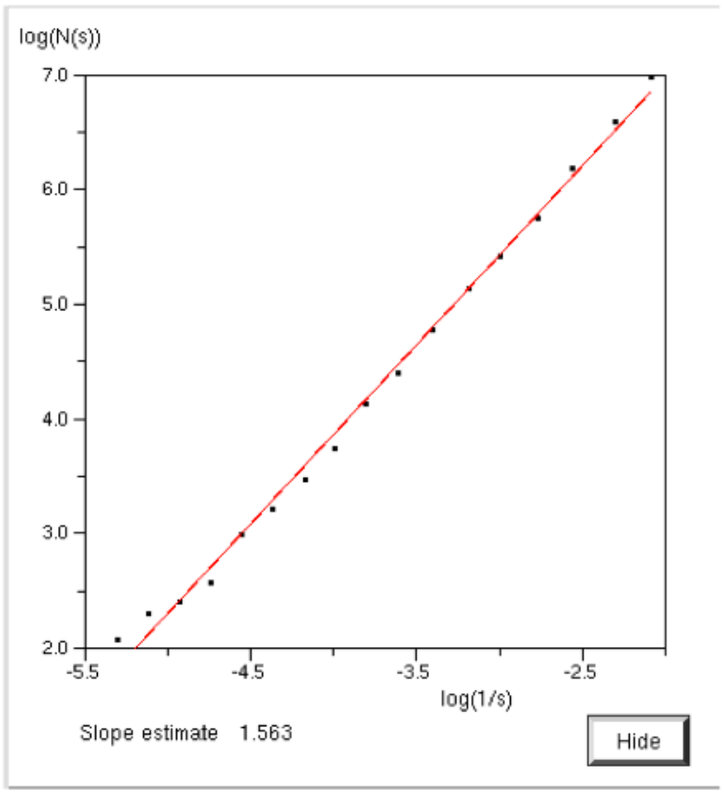
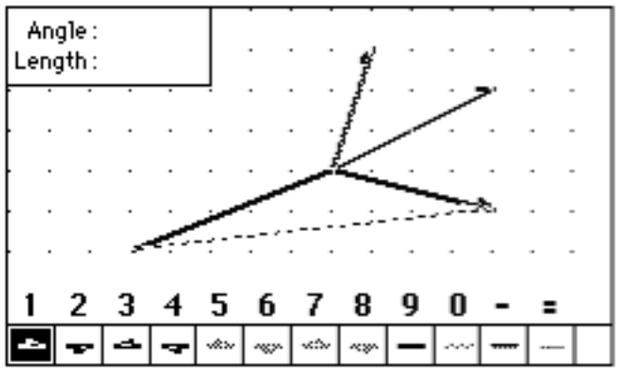


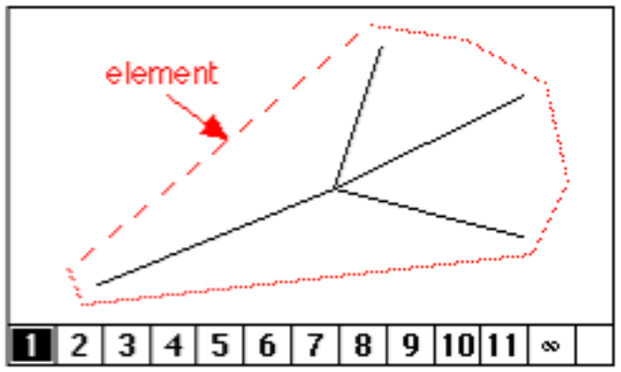
Figure 2.22: log(N)/log(r) plot

Creation of Fractals as Self-Similar Objects

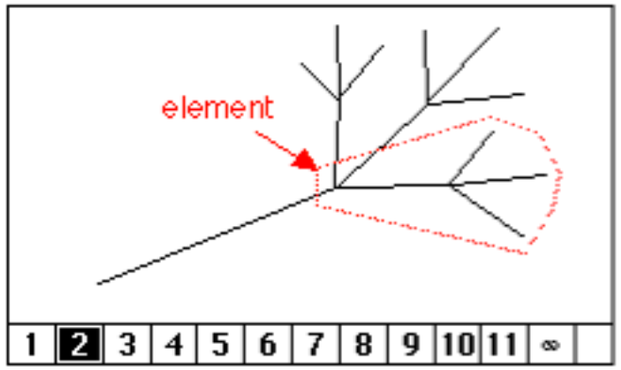
Examples



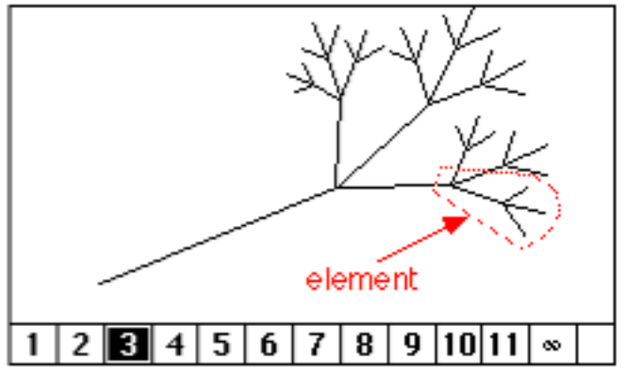
Template (element)



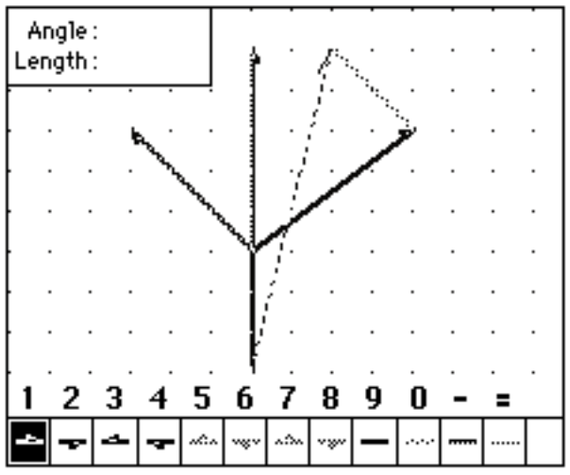
Level 1



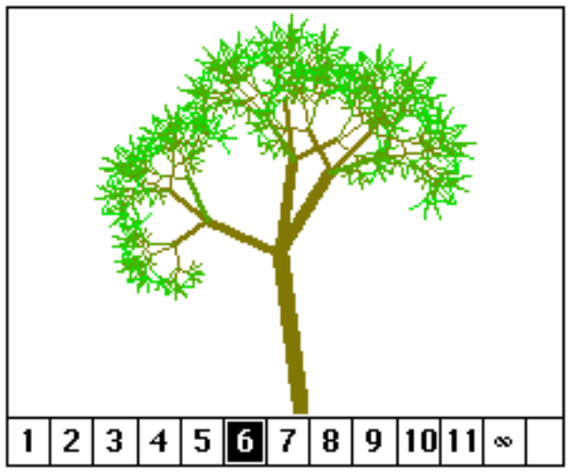
Level 2



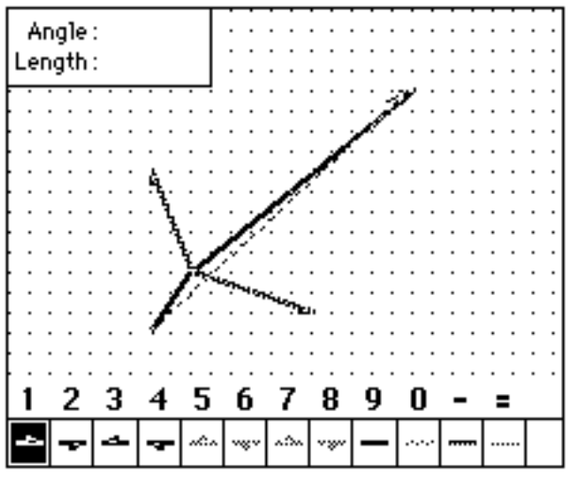
Level 3



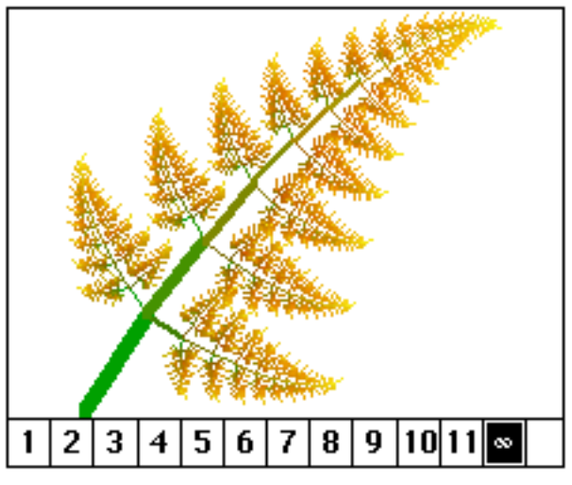
Template



Fractal

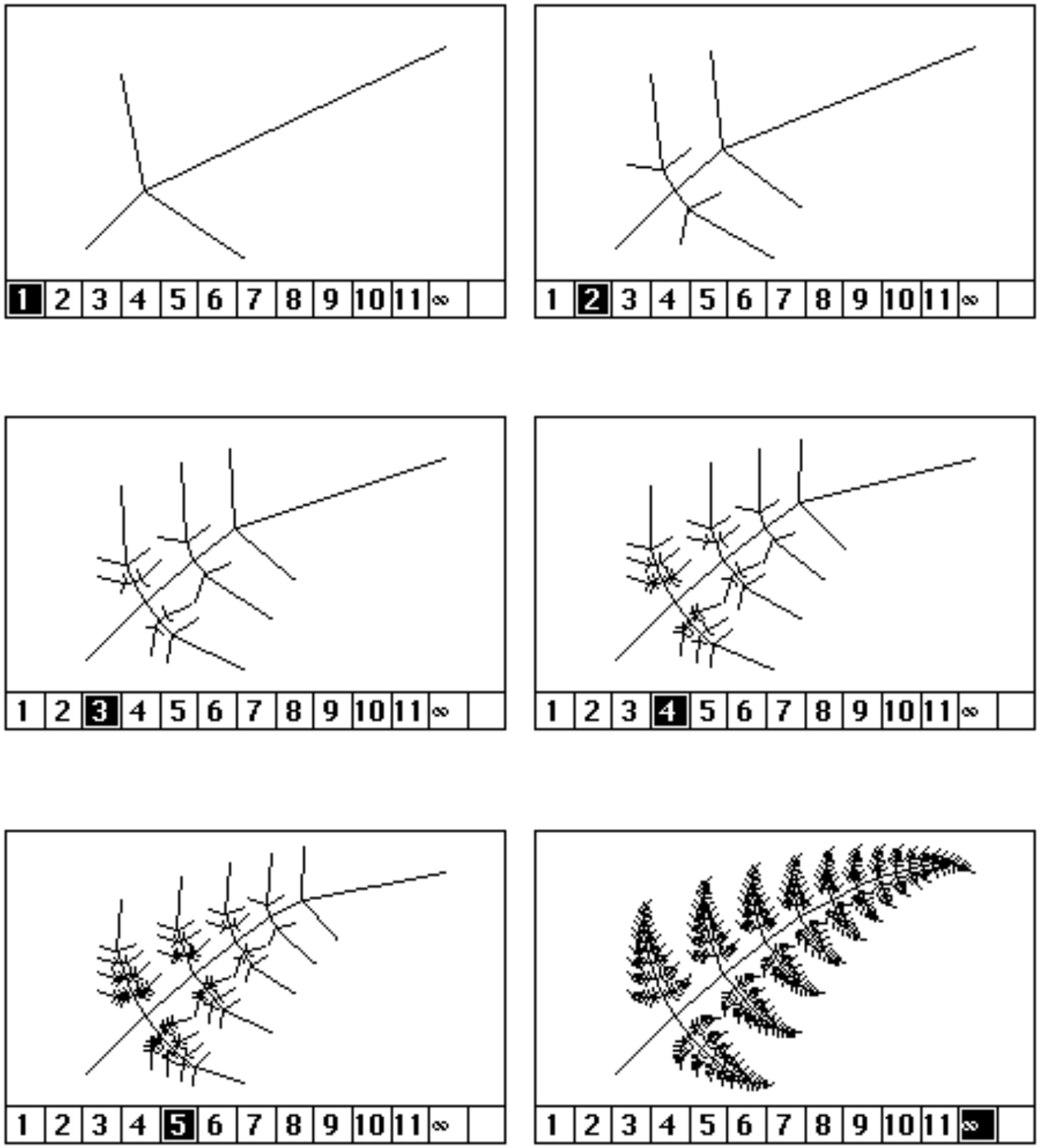


Template



Fractal

slight changes in lengths and angles produces everything!!



Growth of the fern

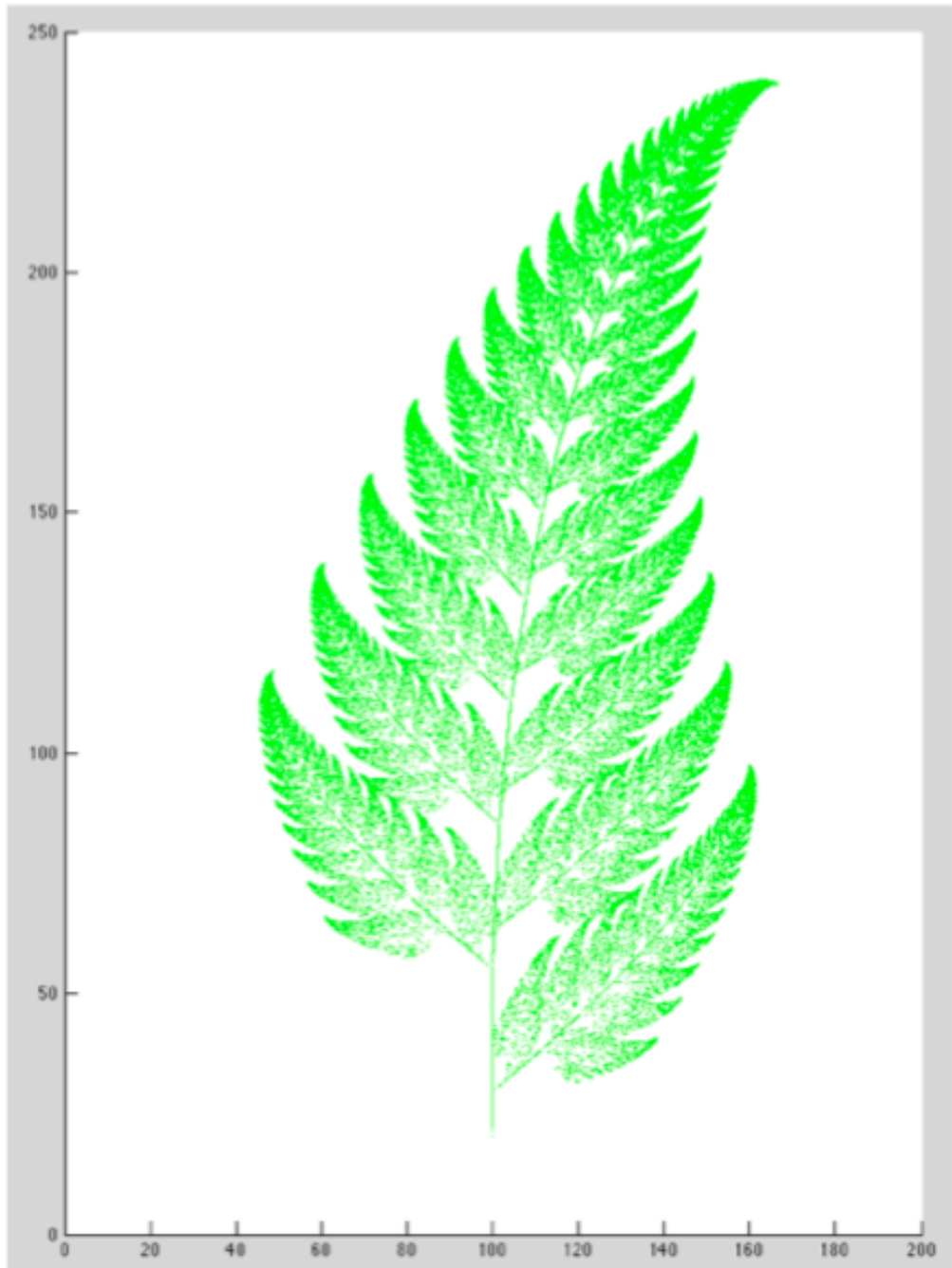


Figure 2.15: A Fractal Fern

Complex Maps

To see self-similar nature of fractals even more dramatically

we study 2-dimensional complex maps.

In particular, look at the Mandelbrot fractal boundary

Mandelbrot Sets

Key mapping is quadratic map given by: $z_{n+1} = z_n^2 + c$

where z_n are complex numbers and c = complex parameter.

For different values of c,

a trajectory **either** stays near origin **or** escapes to infinity.

Turns out that if $|z_n| > 2$ further iteration of this quadratic map is unbounded - can stop!

Thus, when calculations are carried out, we stop iteration when this condition is satisfied.

When investigating these maps, one iterates equation as follows:

(a) start with complex value for c inside rectangular boundary

(b) iterate equation until either

(1) $|z_n| > 2$ or

(2) preset number of iterations exceeded

In case (1) color the starting point (c) in red.

In case (2) leave the point (c) black.

Boundary between two regions is Mandelbrot fractal as shown.

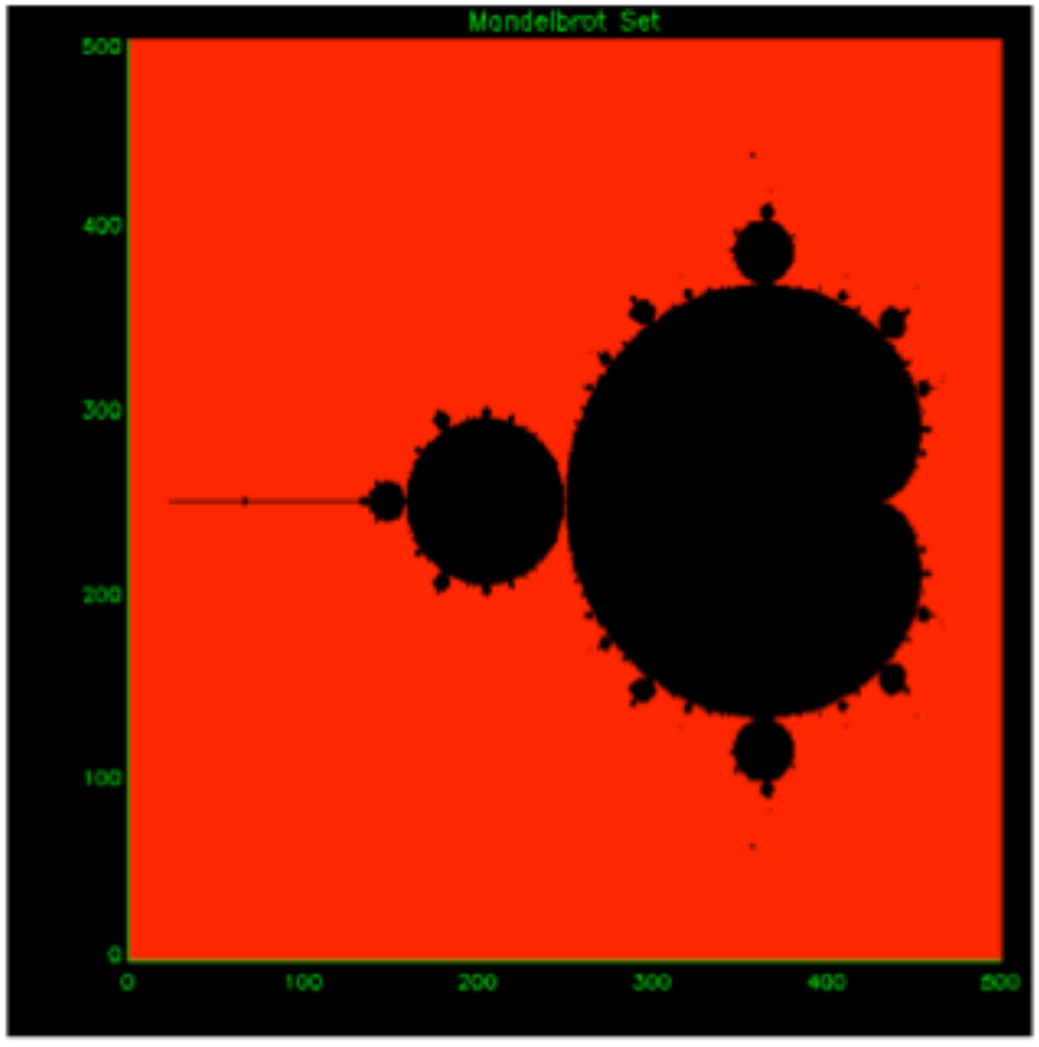


Figure 2.23: The Mandelbrot Fractal Boundary

Alternately, in case (1) color starting point (c) by number of iterations it took to escape.

In case (2) leave point (c) a single color (say dark blue).

Boundary between two regions is still Mandelbrot fractal as shown below left, but wide variety of spectacular images can be generated by clever choice of colormap.

mandelcplxz.m

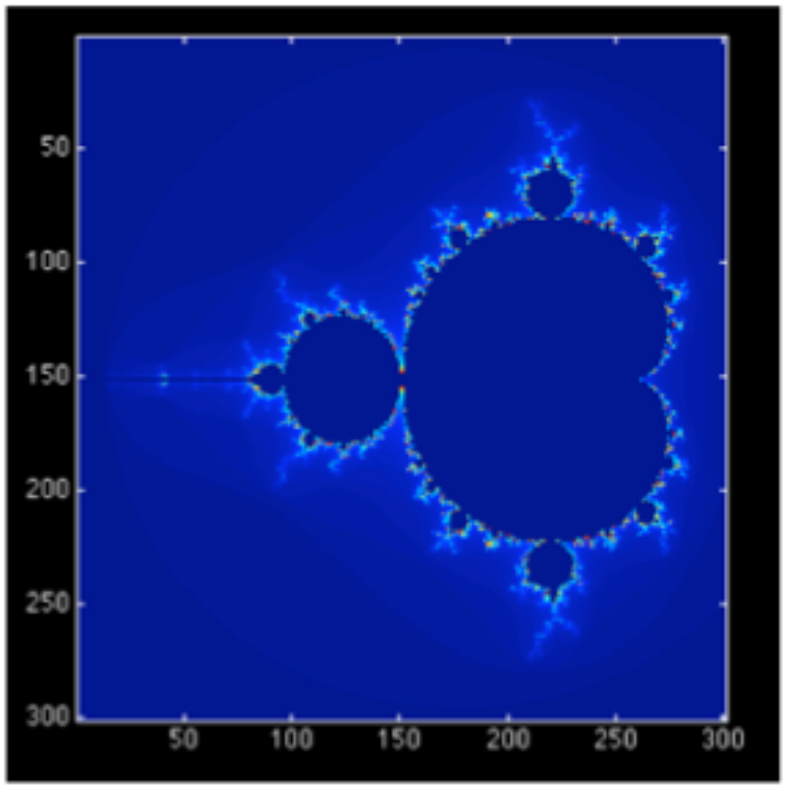


Figure 2.24: The Mandelbrot Fractal Boundary - Different Coloring

or choosing different colormap (no change in information) get

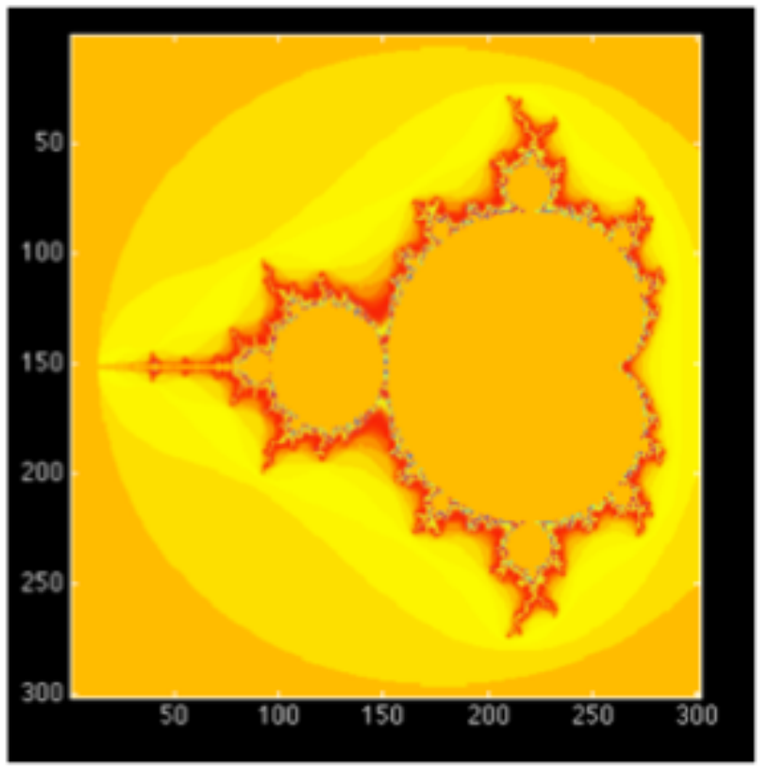
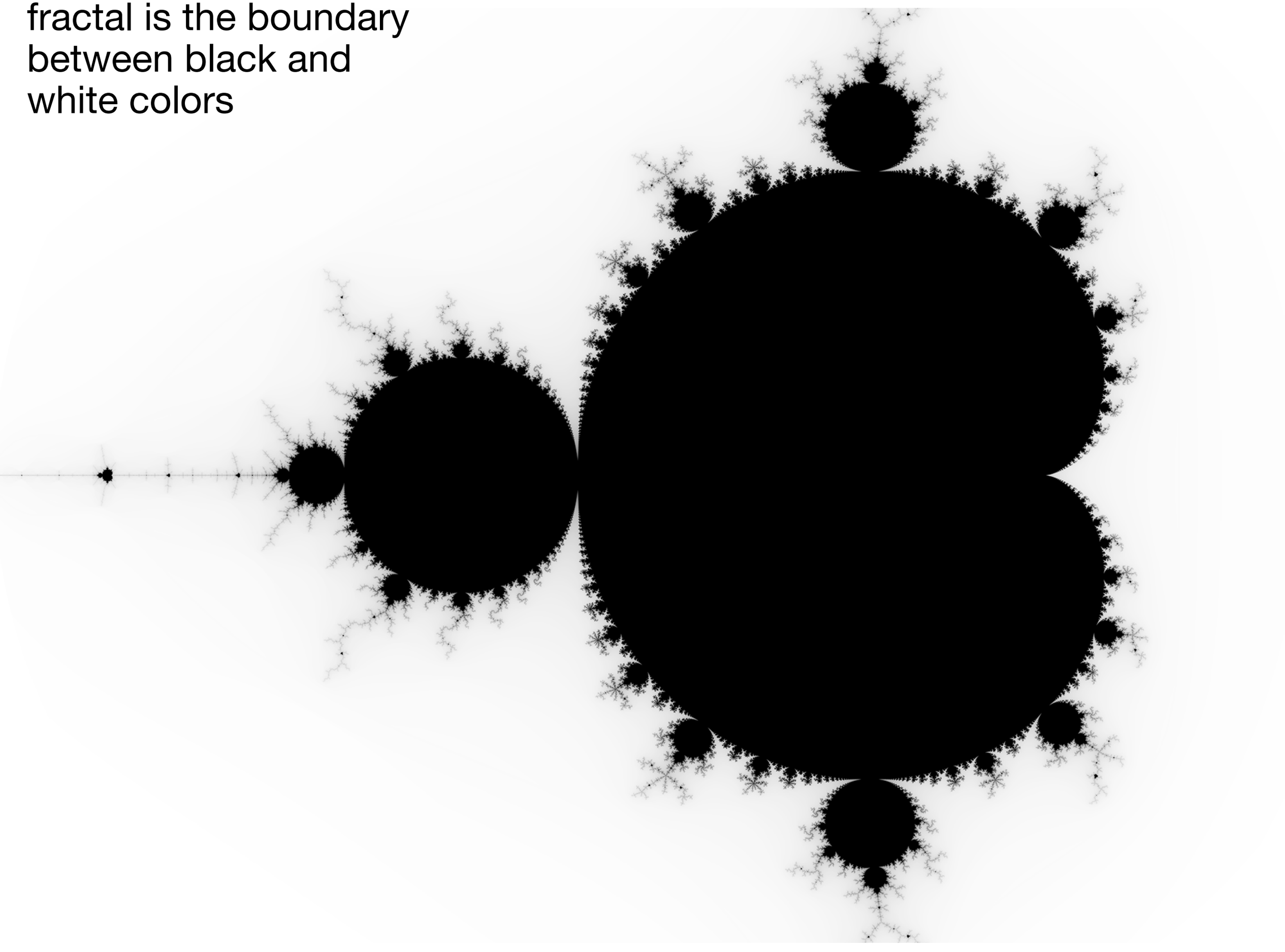


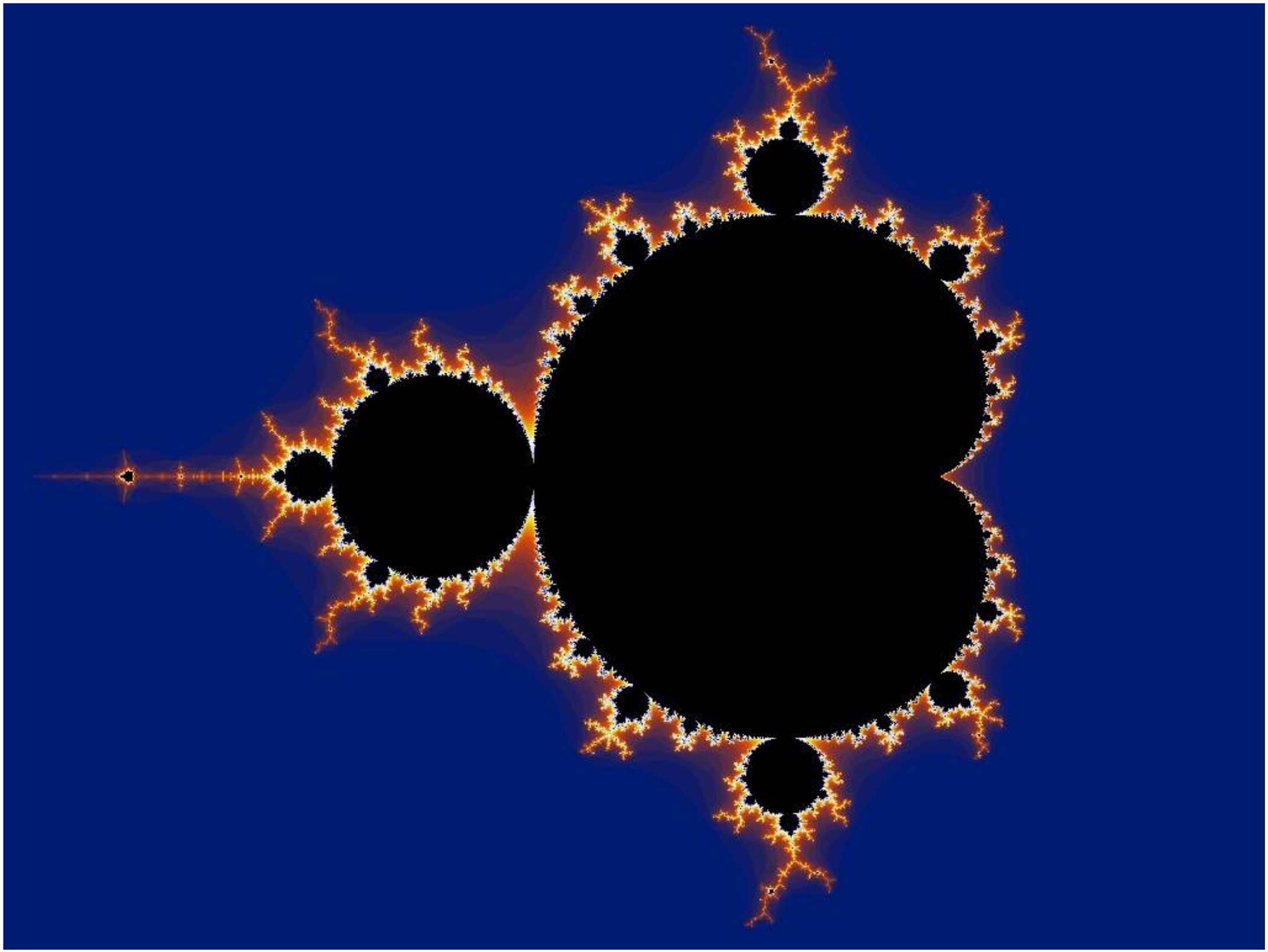
Figure 2.25: The Mandelbrot Fractal Boundary - Different Coloring Scheme

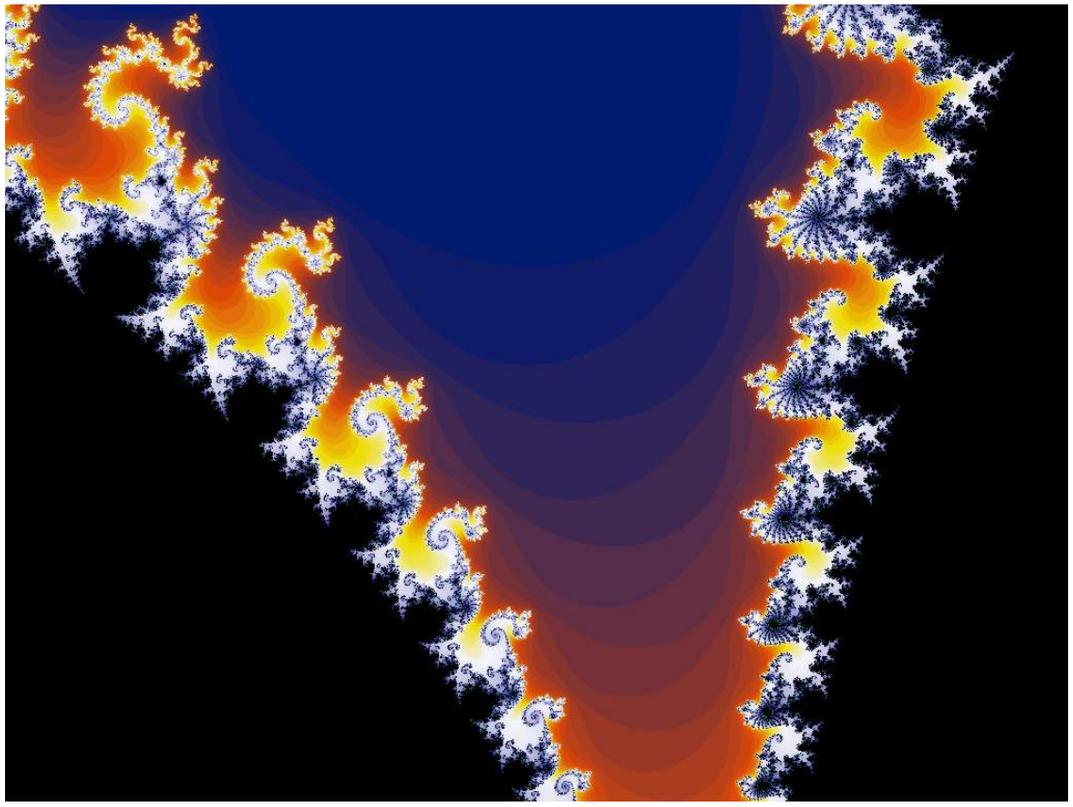
fractal is the boundary between black and white colors

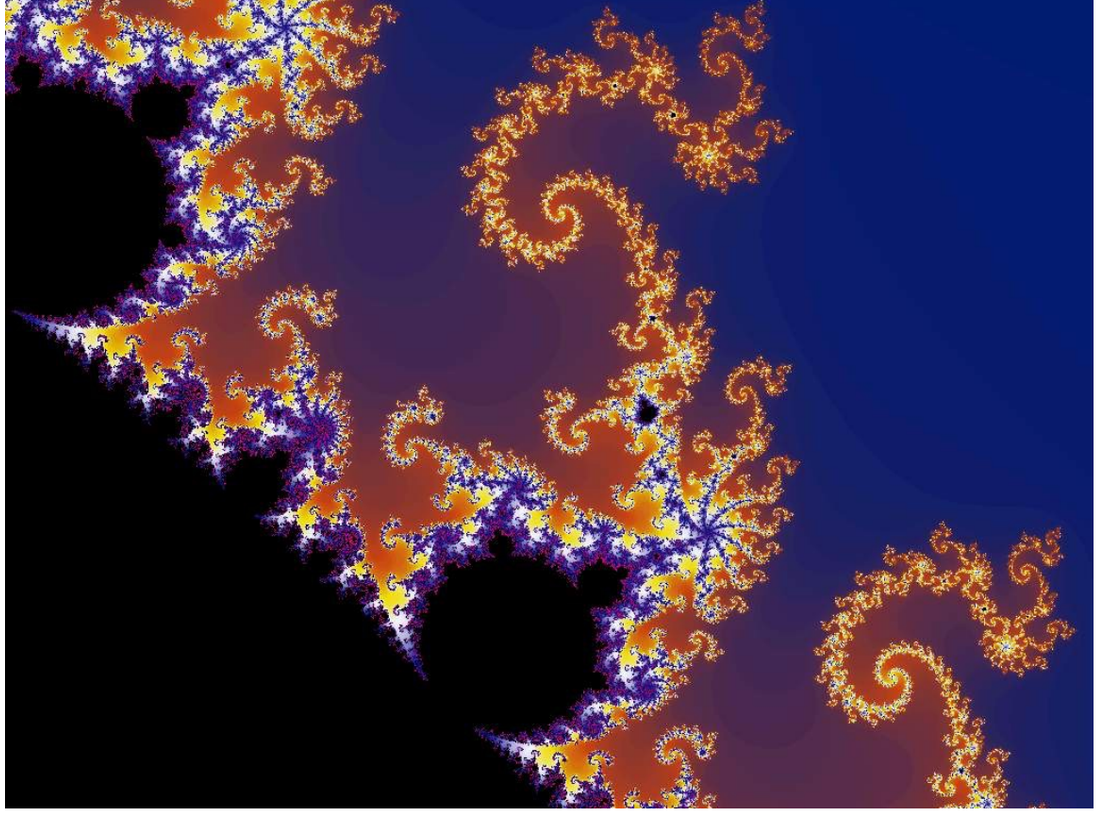


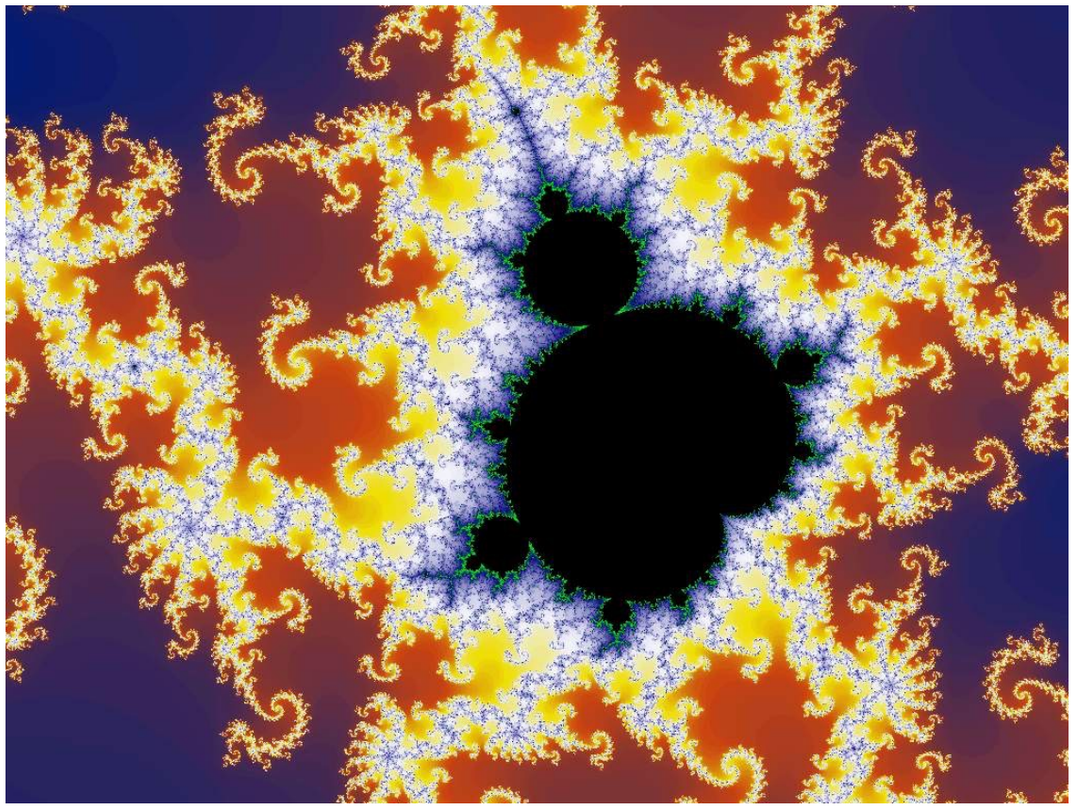
Run programs for colormaps and zooming. Mandelbrot demos - XaoS, etc

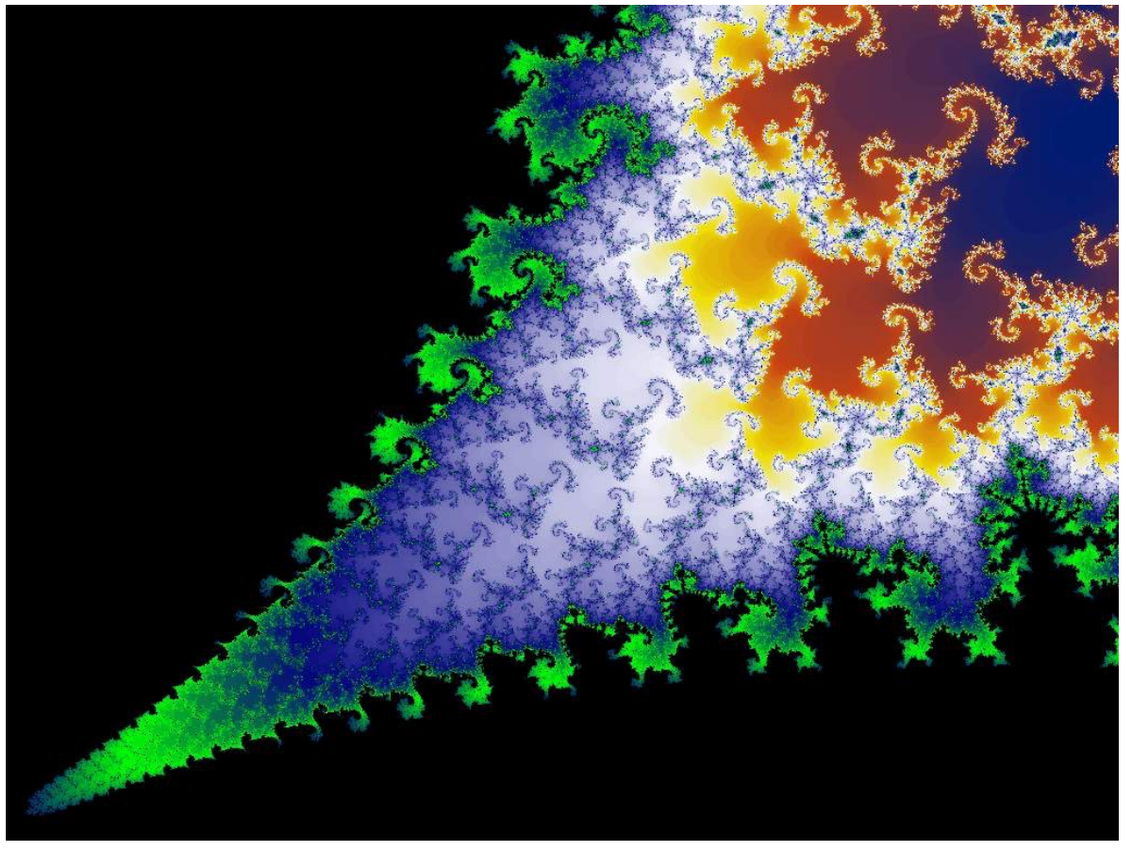
If zoom far enough see entire repeating itself as all fractals do since they are **self-similar**.

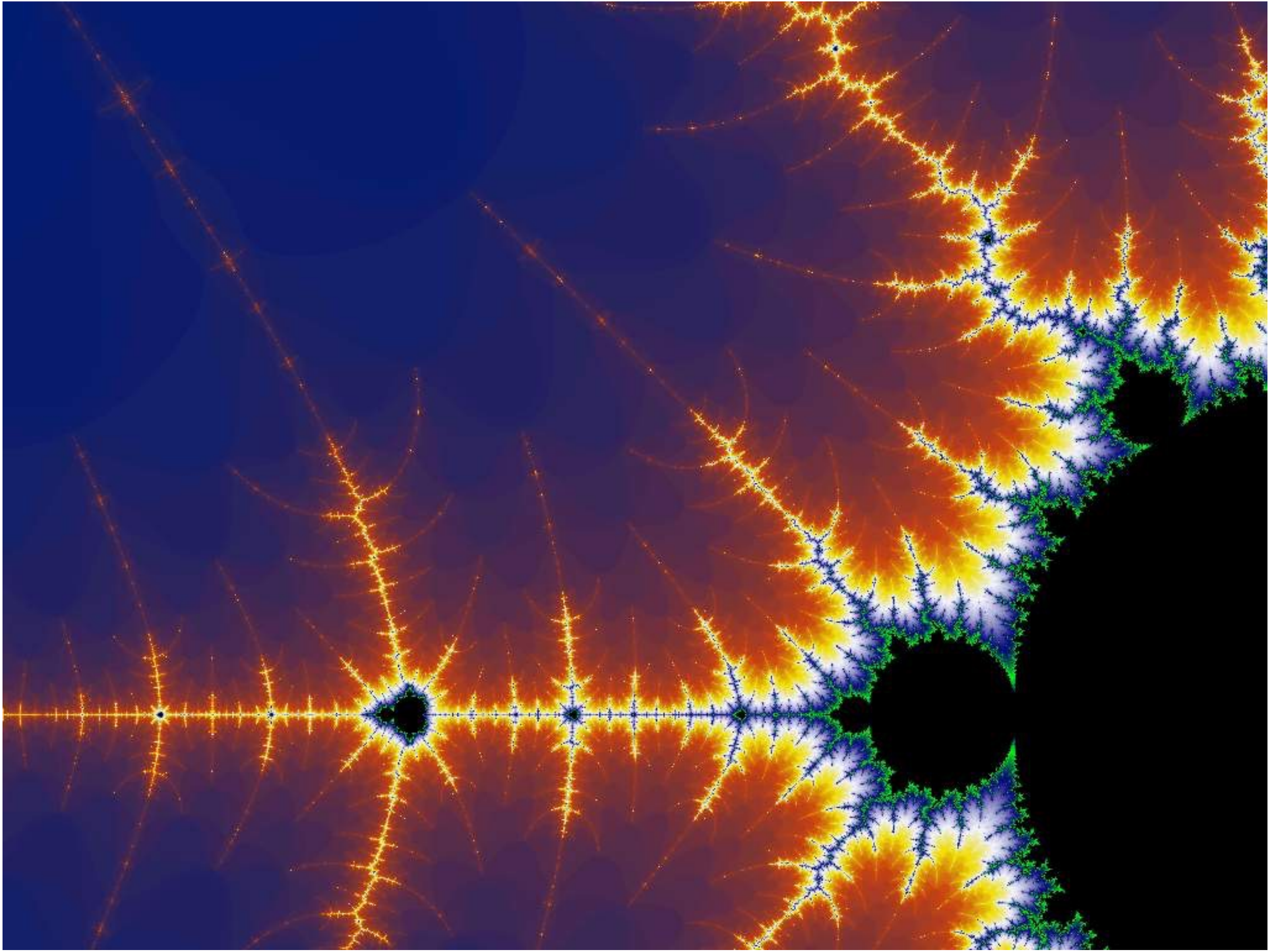


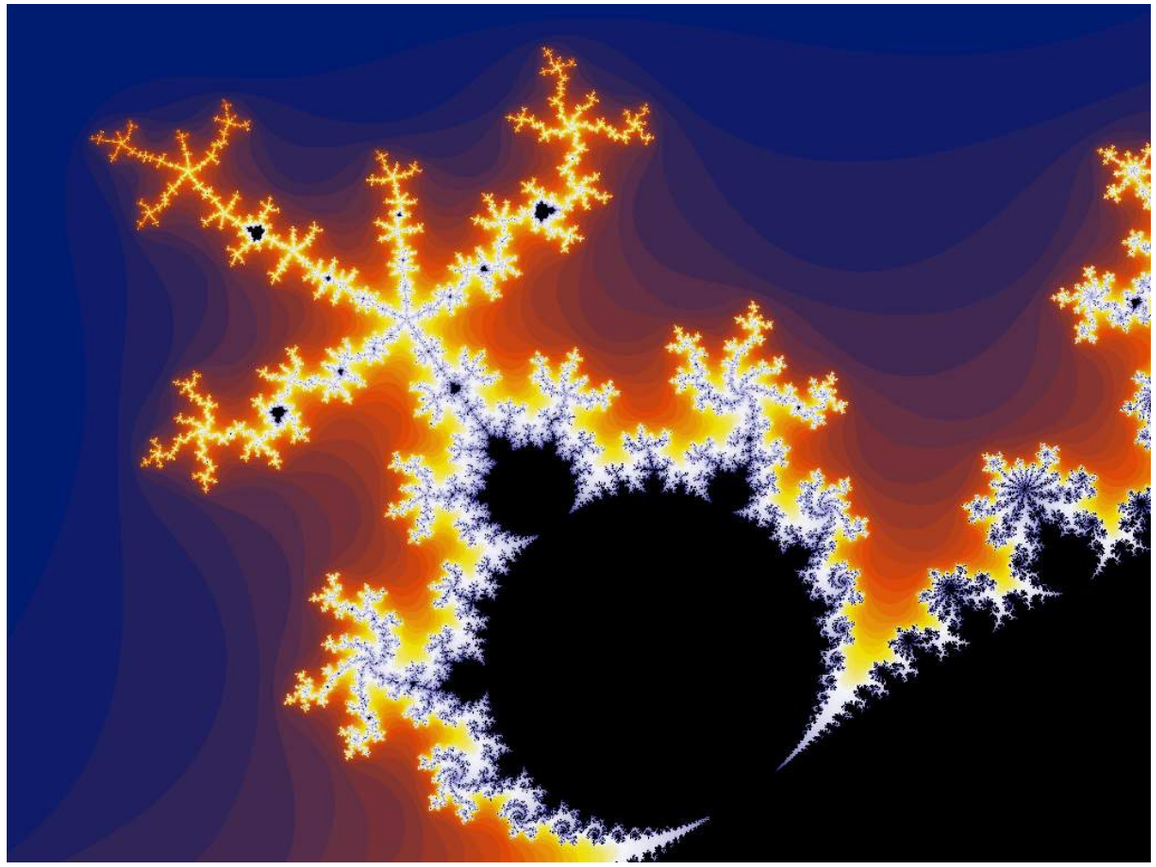


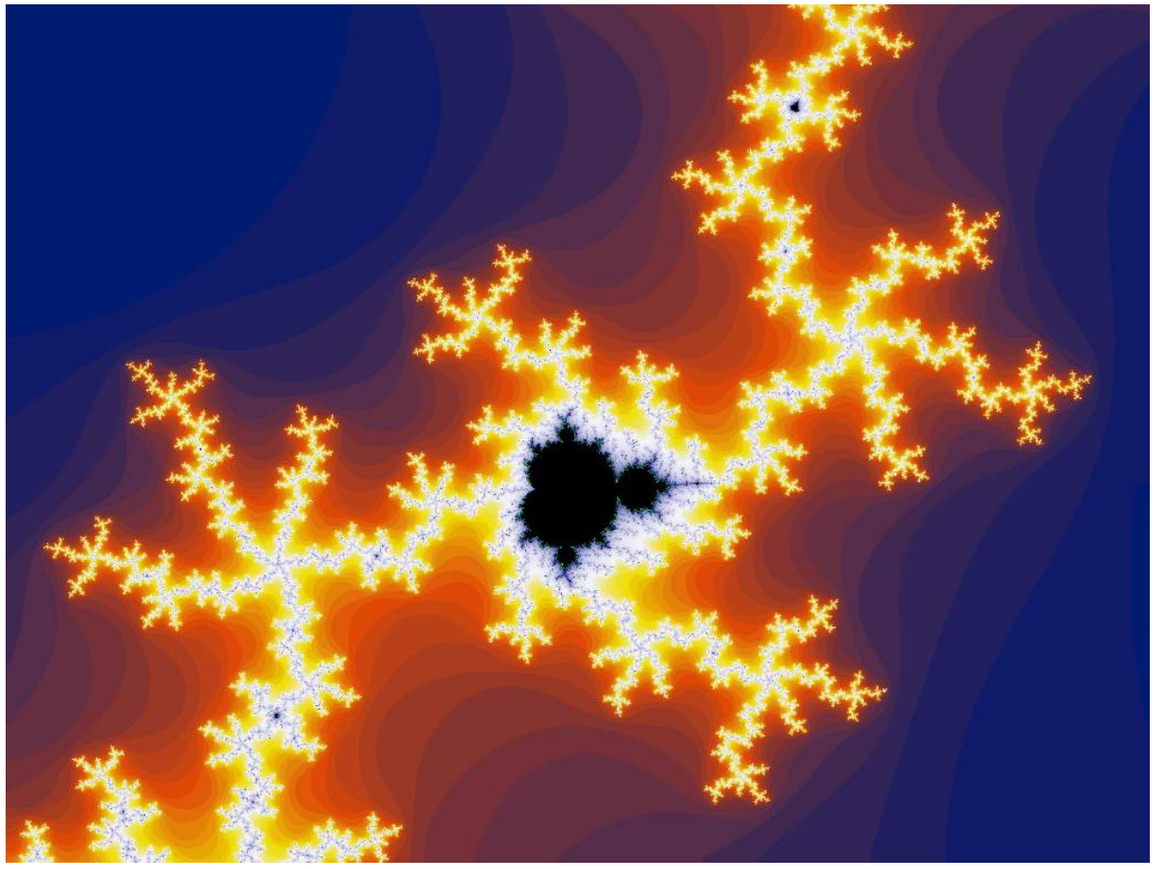


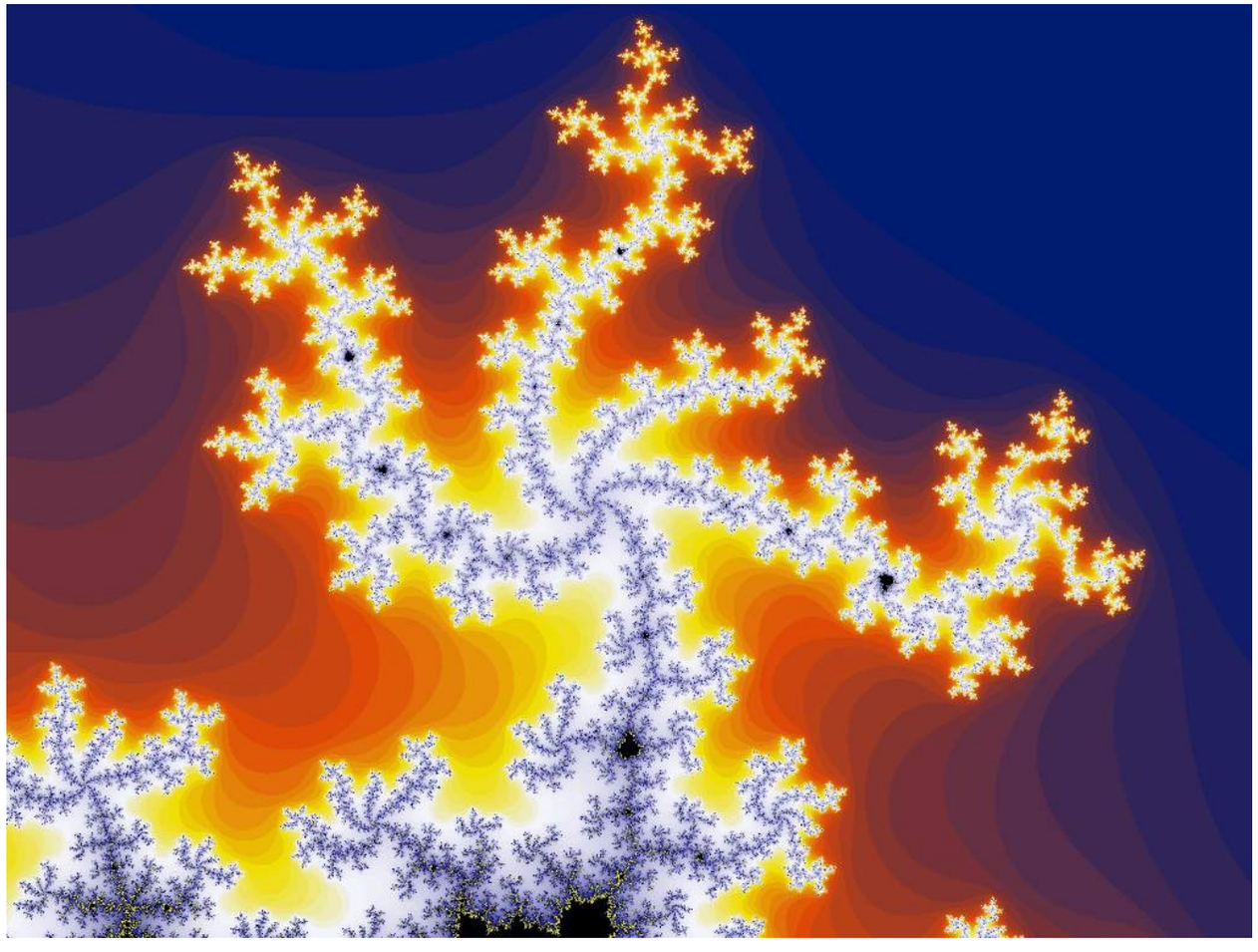












So what is a fractal?

It is an irregular geometric object with an infinite nesting of self-similar structures at all scales.

Why do we care about fractals?

natural objects are fractals

chaotic trajectories (strange attractors) are fractals

complex systems that self-organize will have self-similar or fractal behavior at all scales

and **weather** is a chaotic system…….

Initial Thoughts on Weather

Although other fields that deal with complex, non-linear systems

have accepted the implications of chaos theory,

some meteorologists and climatologists **remain reluctant**

to accept the primary implication of chaos theory,

namely that long-term climate forecasting is **impossible**.

According to chaos theory, all the current “initial" conditions throughout the atmosphere
must be known precisely to predict what the atmosphere
will be doing in the distant future.

In addition, one must know all the current conditions throughout the oceans as well,
since the oceans control the atmosphere.

In my view of the inevitable inaccuracy and incompleteness of weather observations,
the possibility of precise very-long-range forecasting would seem to be nonexistent.

So even if the molecules in the air all interacted nonrandomly,
in a totally cause-and-effect (deterministic) manner,
you still couldn’t predict with certainty what they would do or what the weather would be.

Now we go back over some of the ideas in chaos and fractals in order to make connections to weather systems

Some repetition but that always helps understanding

Chaos, Strange Attractors, and Weather

Introduction

Simplicity and regularity are associated with predictability.

For example, because the orbit of the earth is simple and regular
we can always predict when astronomical winter will come.

On the other hand, complexity and irregularity are almost synonymous with unpredictability.

The atmosphere, being so complex and irregular, is rather unpredictable.

Those who try to explain the world we live in always hoped
that in the realm of the complexity and irregularity observed in nature,
simplicity would be found **behind** everything,
and finally unpredictable events would become **predictable**.

That complexity and irregularity **exist** in nature is obvious.

We only need to look around us to realize
that **practically everything** is random in appearance.

Or is it?

Clouds, like many other structures in nature, come in an **infinite** number of shapes.

Every cloud is different, yet **everybody** will recognize a cloud.

Clouds, though complex and irregular, must on the whole possess a **uniqueness**
that **distinguishes** them from other structures in nature.

The question **remains**: is their irregularity completely **random**
or is there some **order** behind their irregularity?

Over the last decades physicists, astronomers, biologists, and scientists
from many other disciplines have developed
a **new** way of looking at complexity in nature.

This way has been termed **chaos theory**.

Chaos theory, which mathematically **redefines** the “randomness”

generated by simple deterministic dynamical systems,

allows us to **see** order in processes that were **thought** to be completely random.

Apparently, the founders of chaos theory had a very good sense of humor,

since chaos is the Greek word for the complete absence of order.

It is the **purpose** of this discussion to introduce you to some chaos theory concepts

and their **implications** in weather and climate.

Simple examples and definitions from the theory of dynamical systems

On last slide, the **term** "dynamical systems" was used.

What is a dynamical system?

In simple terms a dynamical system is a system whose evolution

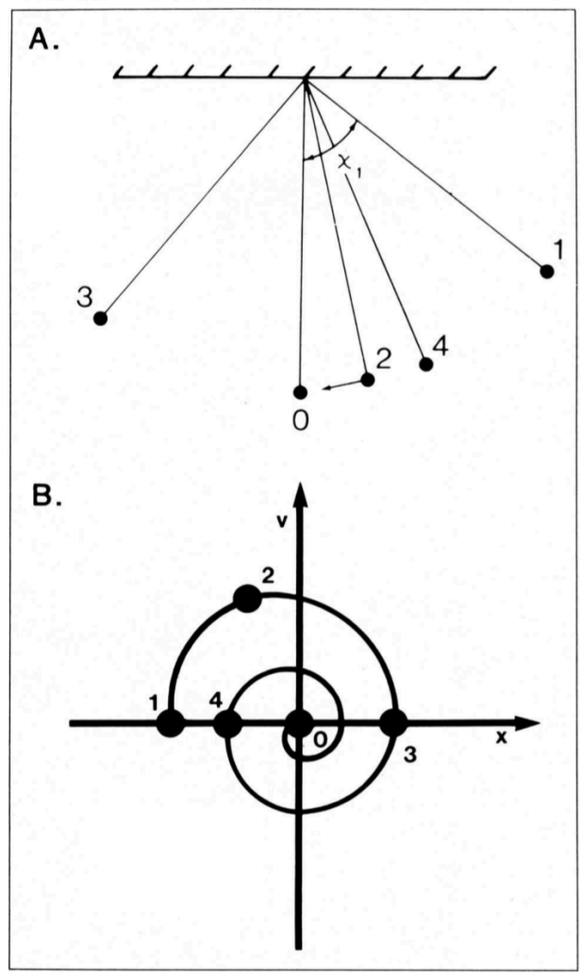
from some initial state (which we know) can be described by a **set of rules**.

These rules may be conveniently **expressed** as mathematical equations.

The evolution of such a system is best described by the so-called **“state space.**"

An example of a simple dynamical system, a **pendulum**, and its state space, is given below.

Consider a pendulum that is **allowed**
to swing back and forth from some initial state,
as shown in Figure A.



The initial state can be **completely** described by the
velocity, v, and the position of the pendulum,
where the position of the pendulum at any time
can be given by the angle x.

Under such an arrangement,
Newtonian physics **provides** the equations (rules)
that describe the system's **evolution** from the initial state.

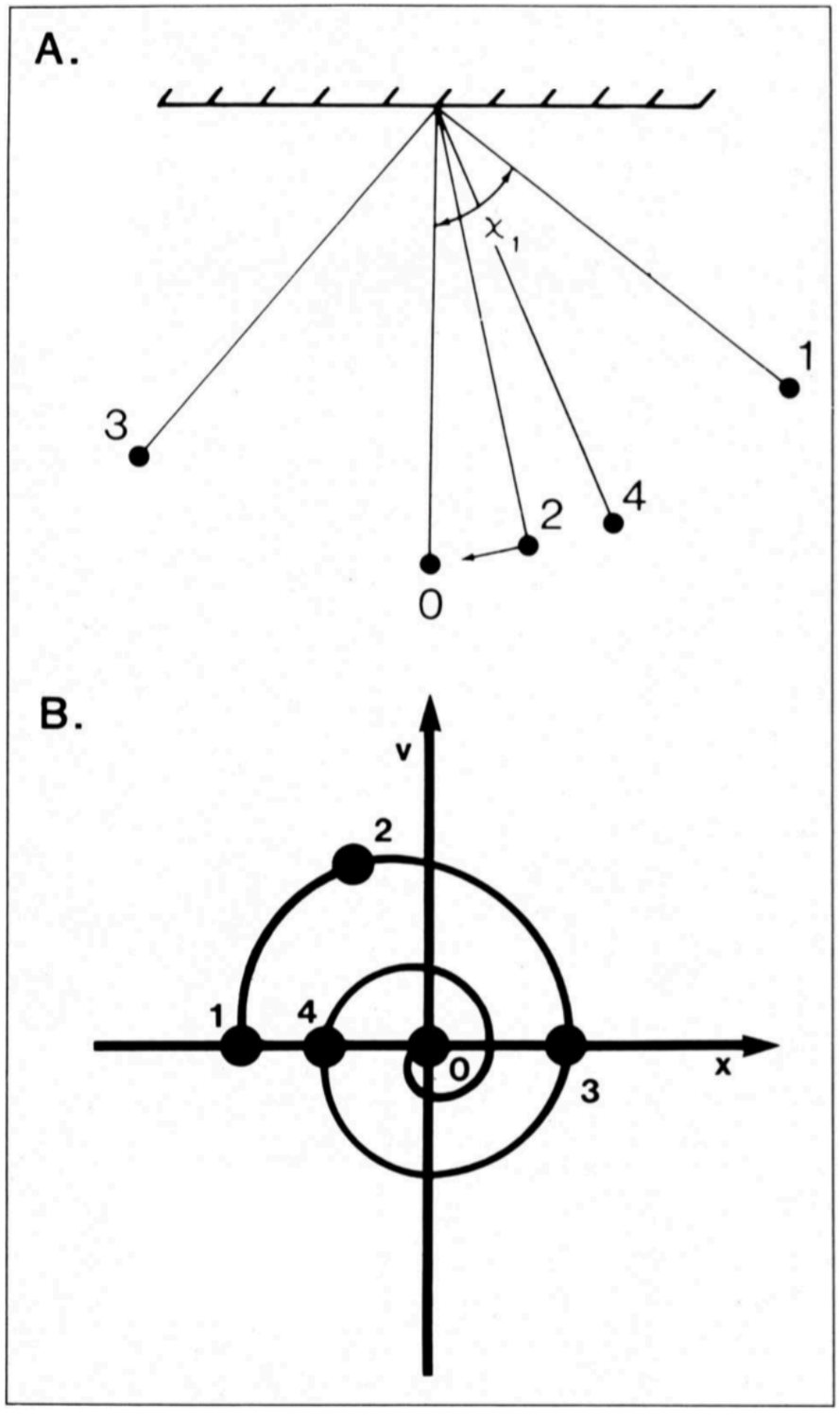


Figure 1

As I said a **dynamical system** is a system
whose evolution from some initial state
can be **determined** by some rules.

In the figure the motion of the pendulum
can be **completely described** by the laws of physics
if its initial position and velocity are **known**,

—> Example of a dynamical system
whose coordinates are the velocity
and the angle(position) of the pendulum.

As the pendulum swings back and forth
it follows a **trajectory** in the state space
which converges to a **fixed point**,
or **attractor** of the dynamical system,
as we discussed earlier.

Assume that the pendulum starts at position 1.

At position 1 its initial state will be $x = x_1$ and velocity $v = 0$.

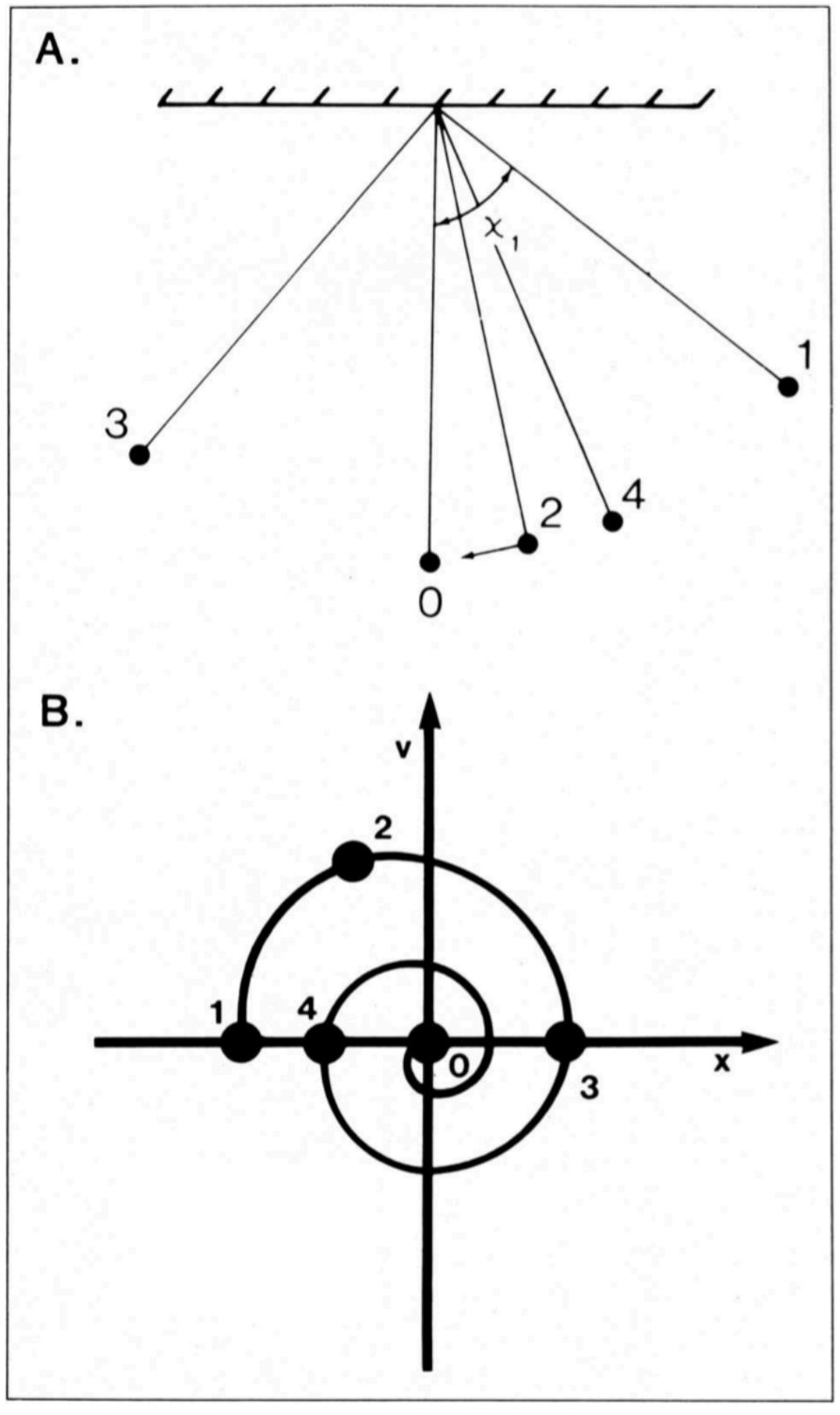
The pendulum is then let free.

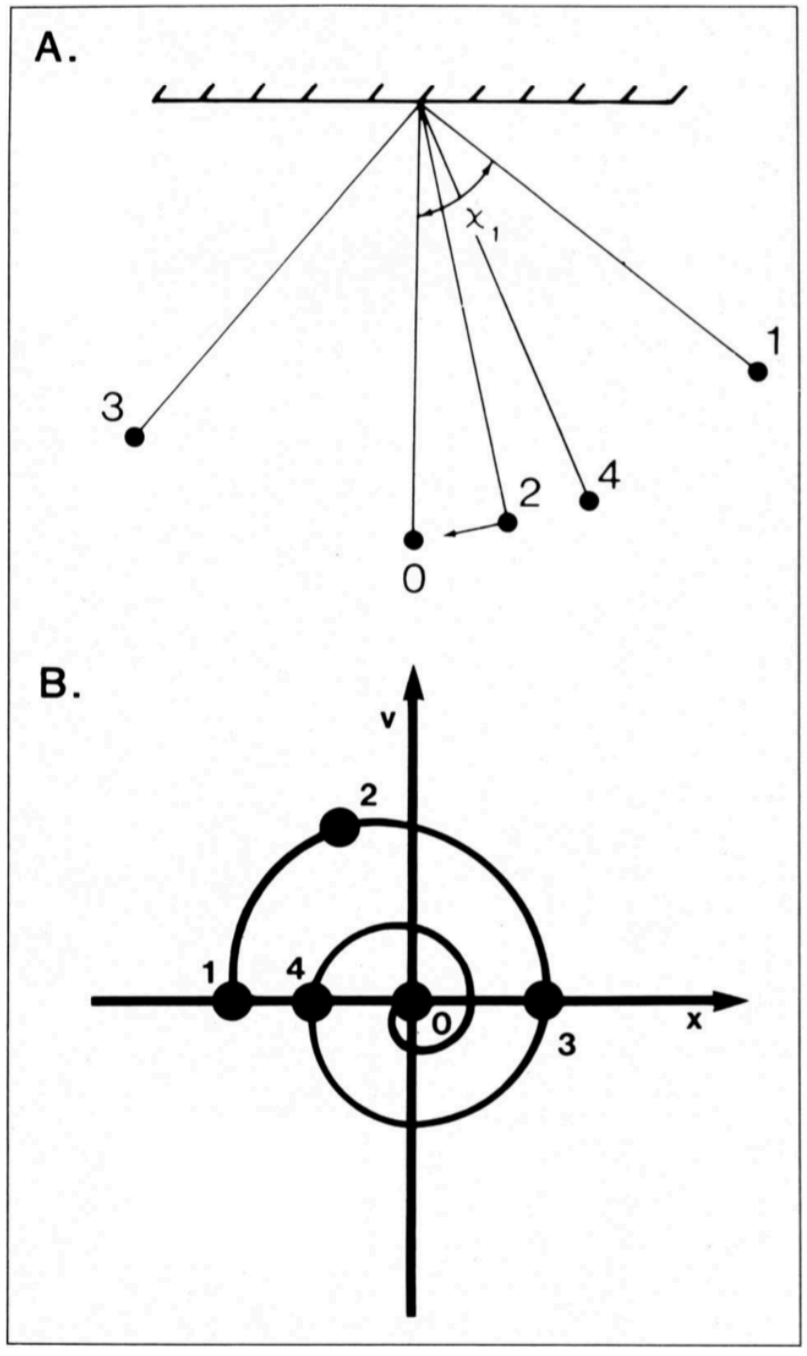
As it moves towards point 0, its speed **increases** due to gravity.

After a while (position 2), the pendulum will be closer to point 0 and will have a higher speed.

Once the pendulum crosses point 0 its speed decreases, since now gravity acts in a direction opposite to its motion.

At some point (position 3), the pendulum's speed will become zero again.





Immediately the pendulum will begin to swing back.

After it crosses point 0 it will once again attain,
at some point, a zero speed (position 4).

Because there is always some **friction**, however,
the points at which the speed becomes zero
(to the right and left of point 0)
are **not** fixed but are found **closer** and closer to point 0.

Finally, pendulum will come to **rest** at point 0 as in this phase space picture.

Apparently, the time evolution of the pendulum can be **completely described** by two variables, namely velocity and angle.

These two variables **define** the coordinates of the state space.

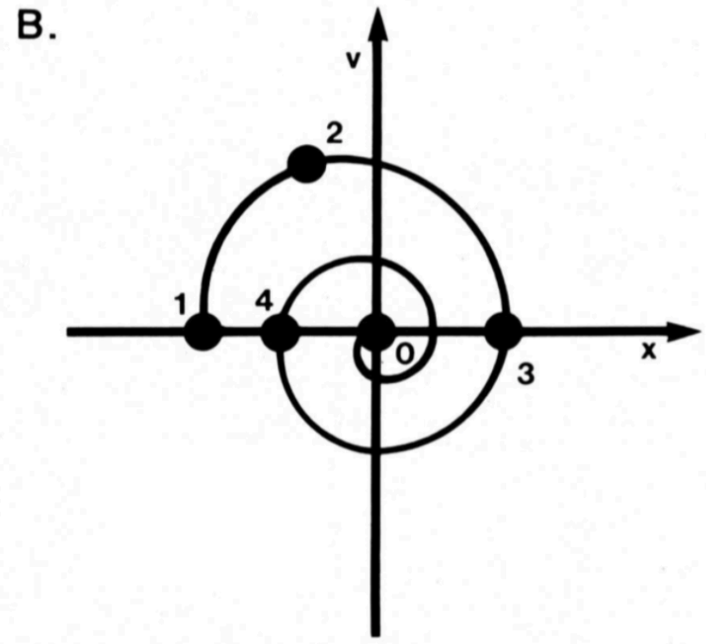
If one plots the velocity (v) as a function of the angle (x) of the pendulum,

for the times corresponding to positions 1, 2, 3, 4, one will **arrive** at figure B.

Each point **represents** the state of the system at a given moment,

therefore a **trajectory** that connects all points

gives a **visualization** of the evolution of the system.



These were the curves we drew **earlier** for the damped-driven oscilla

As shown, the trajectory converges, that is stops, at point 0.

As a matter of fact, **any** other trajectory that corresponds

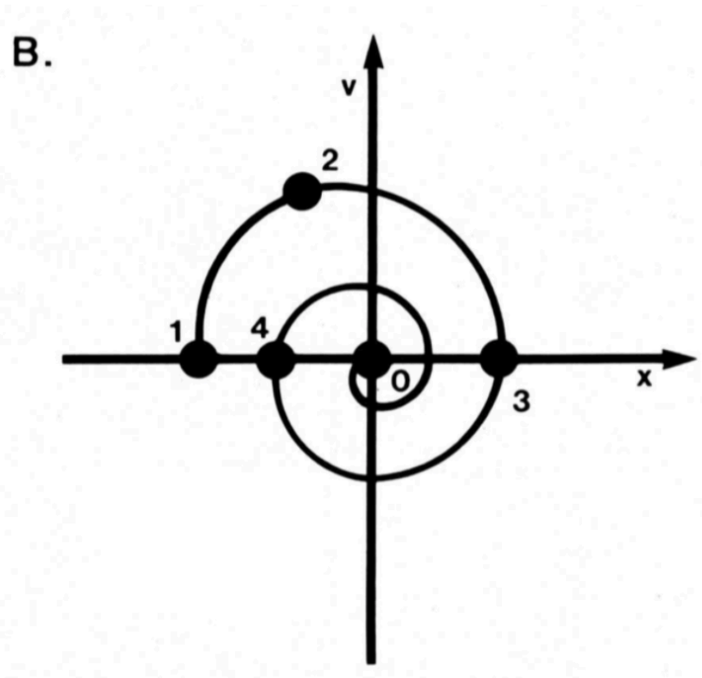
to an evolution of this dynamical system from a different initial state

(velocity and position) will **converge** at point 0 (i.e., no matter what the initial state,

the pendulum will always come to rest at point 0).

The point 0 in the state space is called an **attractor** as we mentioned earlier.

It **"attracts"** all the trajectories in the state space.



Apparently, the behavior of the dynamical system in question can be **completely** understood.

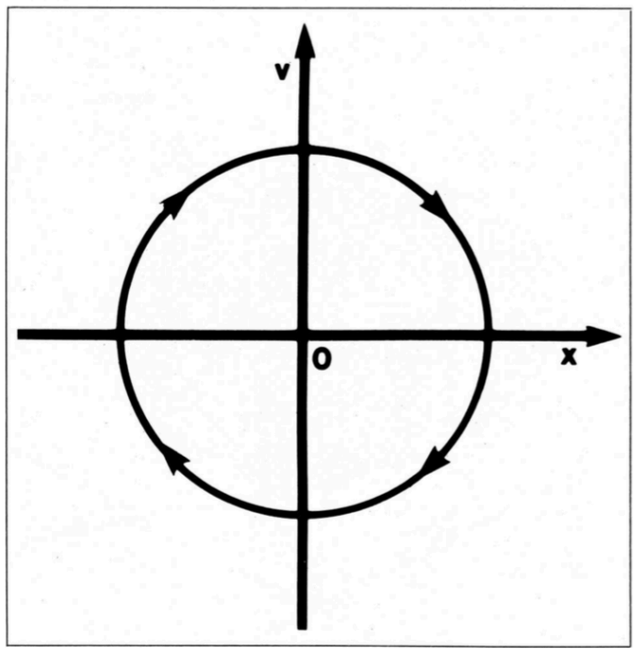
Long-term predictability is **guaranteed**.

The pendulum will always come to rest at point 0.

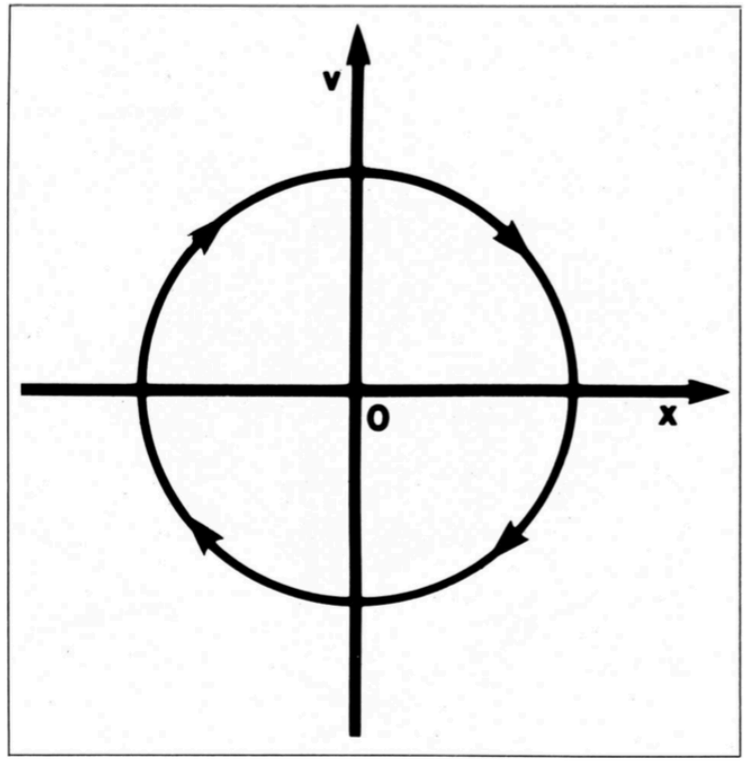
Point attractors therefore **correspond** to systems that reach a state of no motion.

The **next** simplest form of attractor is the **limit cycle** (figure).

A limit cycle in the state space indicates a **periodic** motion.



An **example** of a system whose attractor is a limit cycle
is the grandfather clock, in which the loss of kinetic energy
due to friction is compensated mechanically via a mainspring.



No matter how the pendulum clock is set swinging,
a perpetual, periodic motion will be **achieved**.

This periodic motion **manifests** itself in the state space as a limit c

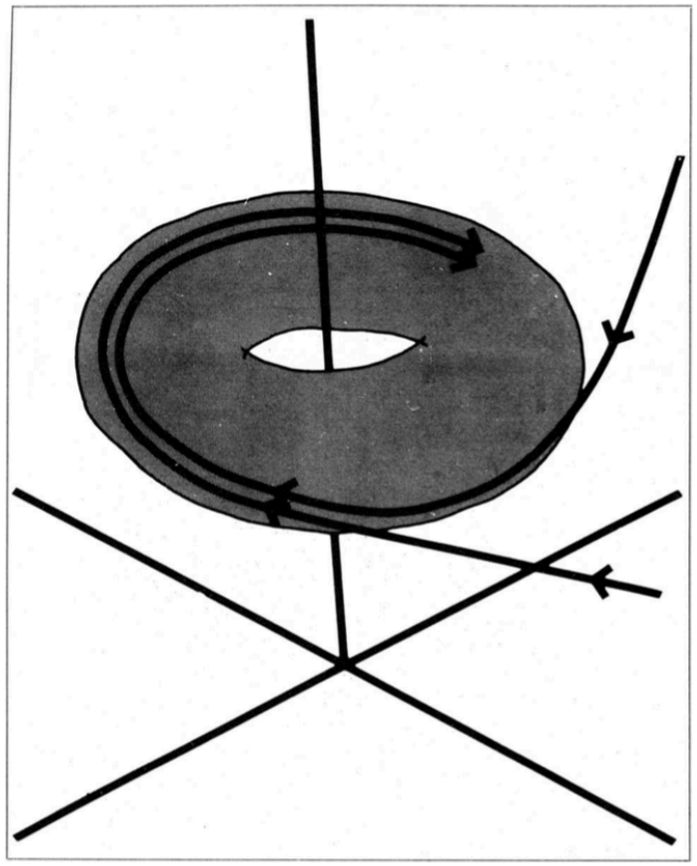
Again, in the case of systems that have a limit cycle as an **attractor**,
long-term predictability is **guaranteed**.

In this case all trajectories are **attracted** by the limit cycle,
which represents a periodic evolution.

The pendulum of a grandfather clock is a system
that **possesses** a limit cycle as an attractor.

Another familiar system with a limit cycle as its attractor is the **heart**.

Another form of attractor is the **torus**.



The torus looks like the surface of a doughnut (figure).

In this case, all the trajectories in the state space

are **attracted** to and **remain** on the surface.

Systems that possess a torus as an attractor are **quasi-periodic**.

In a quasi-periodic evolution a periodic motion

is **modulated** by a second motion, itself periodic,

but with another (usually different) frequency.

The **combination** of frequencies will

produce a time series whose regularity is **not clear**.

The power spectrum(energy versus frequency), however, should consist of sharp **peaks**

at each of the **basic** frequencies with all its other prominent features

being **combinations** of the basic frequencies.

In this case the **evolution** of the corresponding dynamical system from any initial condition

will follow a **trajectory** in the state space that will eventually

be **attracted** and **remain** forever on the torus.

The most important characteristic of a system that exhibits such an attractor

is usually that two initially nearby trajectories on the attractor remain nearby forever.

Geometrically, a quasi-periodic trajectory **fills** the surface of a torus,

in the appropriate state space.

An important **characteristic** of such an attractor is that

when the two frequencies have **no** common divisor,

any two trajectories which represent the evolution of the system

from **different** initial conditions,

and which are **close** to each other

when they **approach** the attracting surface,

will **remain** close to each other forever.

This characteristic can be **translated** as follows.

The two points in the state space where the trajectories
enter the attractor can be two measurements (**initial states**)
which **differ** by some amount.

Since these trajectories **remain** close to each other,
the states of the system at a **later** time are going to differ
to the **same** extent that they differed initially.

Thus, if we **know** the evolution of such a system from an initial condition,
we can **predict** accurately the evolution of the system from some **other** initial condition.

Again, in this case long-term predictability is **guaranteed**.

The above mentioned forms of attractors are "**well-behaved**" attractors
and usually correspond to systems whose evolution is **predictable**.

Often they are called **non-chaotic attractors**.

In mathematical terms, the above mentioned attractors

are **smooth** copies of the original topology

of the available state space.

These attractors are, therefore, **characterized** by an

integer dimension that is **equal** to the

dimension of the original state space.

A very important **characteristic** of these attractors is that

the long-term evolution of the systems they describe

is **not sensitive to initial conditions**.

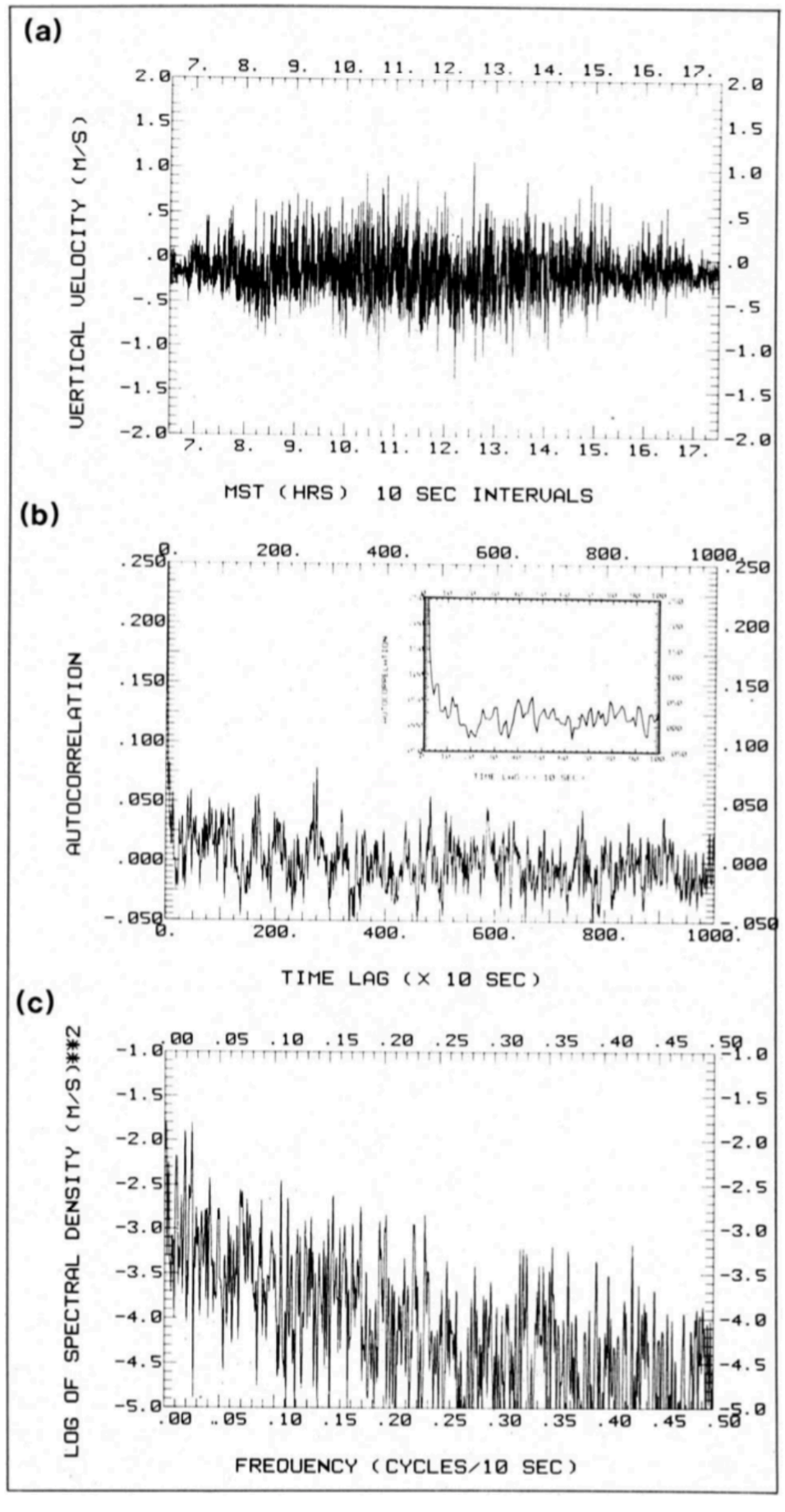
Strange attractors

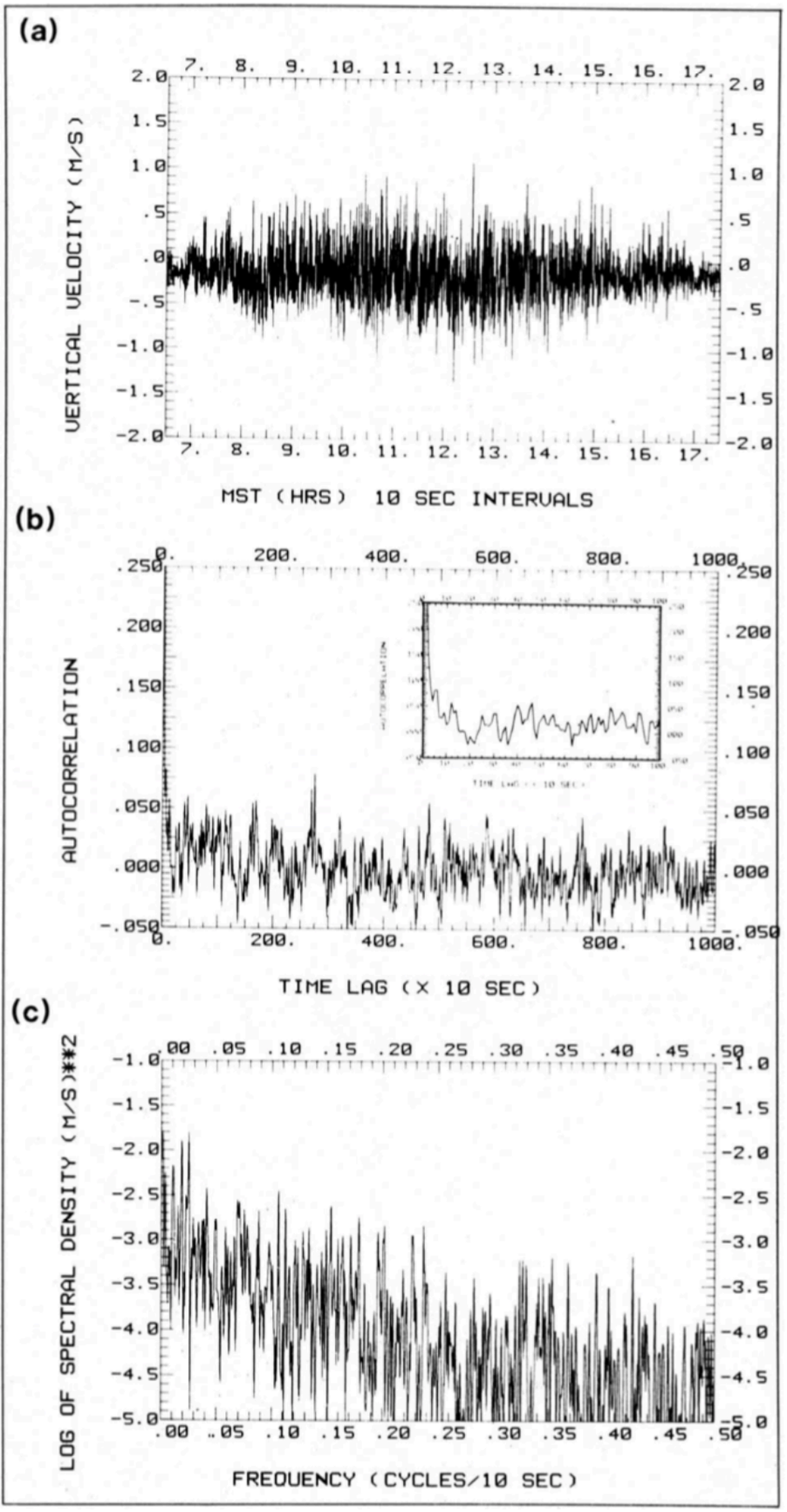
When one observes the spectra

(energy versus frequency) of turbulent motion,

one realizes that there is motion at all frequencies

with no preferred frequencies (see data in figure).





Details about the data…..

(a) The data represent 10-second averages of the vertical wind velocity over 11 hours.

The data were recorded from 6:30 to 17:30 MST (Mountain Standard Time) (12:30 to 23:30 UTC) on 26 September 1986 at Boulder, Colorado.

At about 6:30 (12:30) the sun rises.

The air close to the ground is heated and rises creating strong convection.

Positive values indicate updrafts and negative values indicate downdrafts.

(b) The autocorrelation function for the above data.

The inset graph is a magnification of the region close to the origin.

(c) The logarithm of the spectral density as a function of the frequency for the above data.

The spectra shows various peaks on a background of continuous frequency spectrum, suggesting a non-periodic evolution.

What is Autocorrelation?

Autocorrelation is a mathematical representation of the degree of similarity between a given time series and a **lagged** version of itself over successive time intervals.

It is the same as calculating the correlation between two different time series, except autocorrelation uses the same time series twice: once in its original form and once lagged one or more time periods.

An autocorrelation of +1 represents a perfect **positive** correlation (an increase seen in one time series leads to a proportionate increase in the other time series).

An autocorrelation of negative 1, on the other hand, represents perfect **negative** correlation (an increase seen in one time series results in a proportionate decrease in the other time series).

As I said, autocorrelation measures the relationship between a variable's current value and its past values.

An autocorrelation of +1 represents a perfect positive correlation,
while an autocorrelation of negative 1 represents a perfect negative correlation.

While zero means NO CORRELATION - look back at wind graphs.

This broad-band structure of the spectrum indicates

that the motion is nonperiodic

(or strictly speaking is periodic with an infinite period).

Could such a “non-periodic” motion be due

to a simple dynamical system?

Let us assume that the answer to this question is yes.

In such a case the trajectory in the state space

would be nonperiodic (never repeat itself)

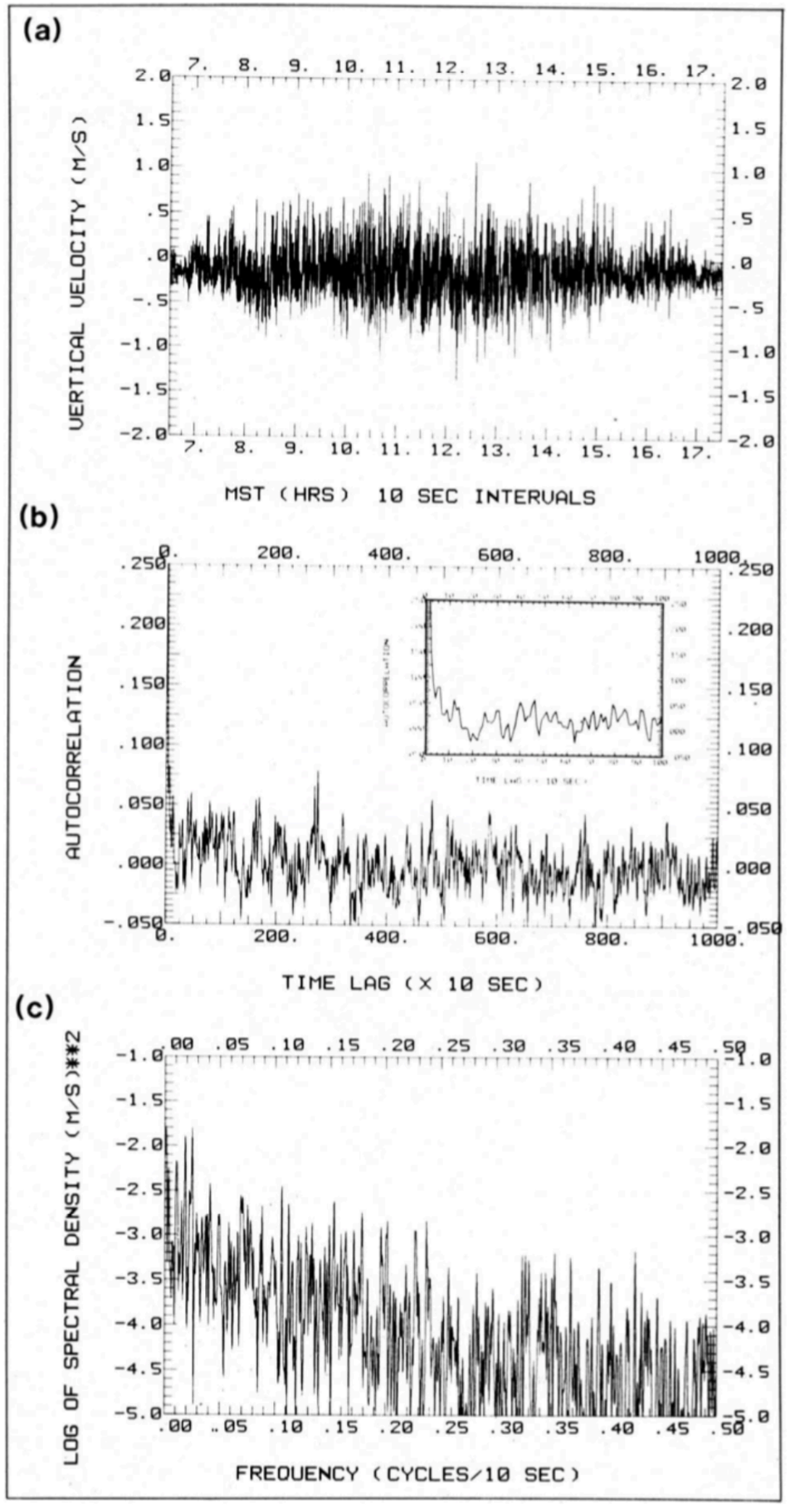
and never cross itself

(since once a system returns to a state

it was in some time in the past it

then has to follow the same path).

Thus the trajectory should be of infinite length but confined to a finite area in the state space.



This can only be the case if the attractor is **not** a simple topological structure as earlier

but rather is a **fractal** set (as we have discussed).

The first such system was discovered in 1963 by Edward Lorenz (Lorenz 1963).

More details now.

This system, described by the following three differential equations,

gives an approximate description of a horizontal fluid layer heated from below.

$$
\begin{aligned}
dx/dt &= -ax + ay \\
dy/dt &= -xz + bx - y \\
dz/dt &= -xy - cz
\end{aligned}
$$

The fluid at the bottom gets warmer and rises, creating convection.

For a choice of the constants that correspond to sufficient heating,

the convection may take place in an irregular and turbulent manner:

$$
\begin{aligned}
dx/dt &= -ax + ay \\
dy/dt &= -xz + bx - y \\
dz/dt &= -xy - cz
\end{aligned}
$$

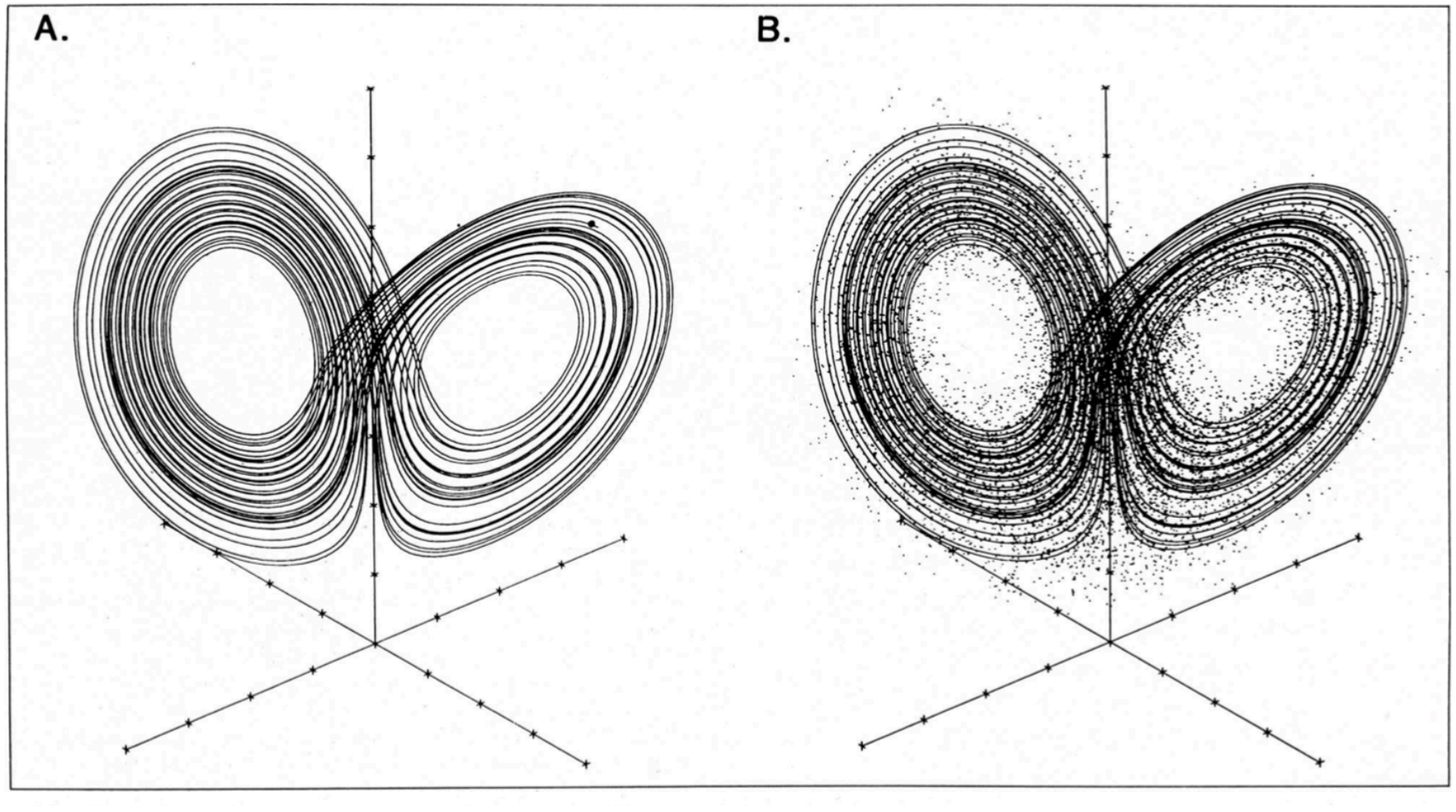
where x is proportional to the intensity of the convective motion,

y is proportional to the horizontal temperature variation,

z is proportional to the vertical temperature variation,

and a, b, and c are constants.

Figure A below depicts the path of a trajectory in the state space **(x, y, z)**.



The Lorenz attractor itself does not look like the well-behaved attractors previously described.

The trajectory is **deterministic** (since it is the result of the solution of the above system of equations), but is **strictly** nonperiodic.

The trajectory loops to the left and then to the right **irregularly**.

Extensive studies have shown that the **fine structure** of the Lorenz attractor

is made up of **infinitely nested** layers (infinite area) that occupy zero volume.

One may think of it as a Cantor-like set in a higher dimension.

Its fractal dimension has been estimated to be about 2.06.

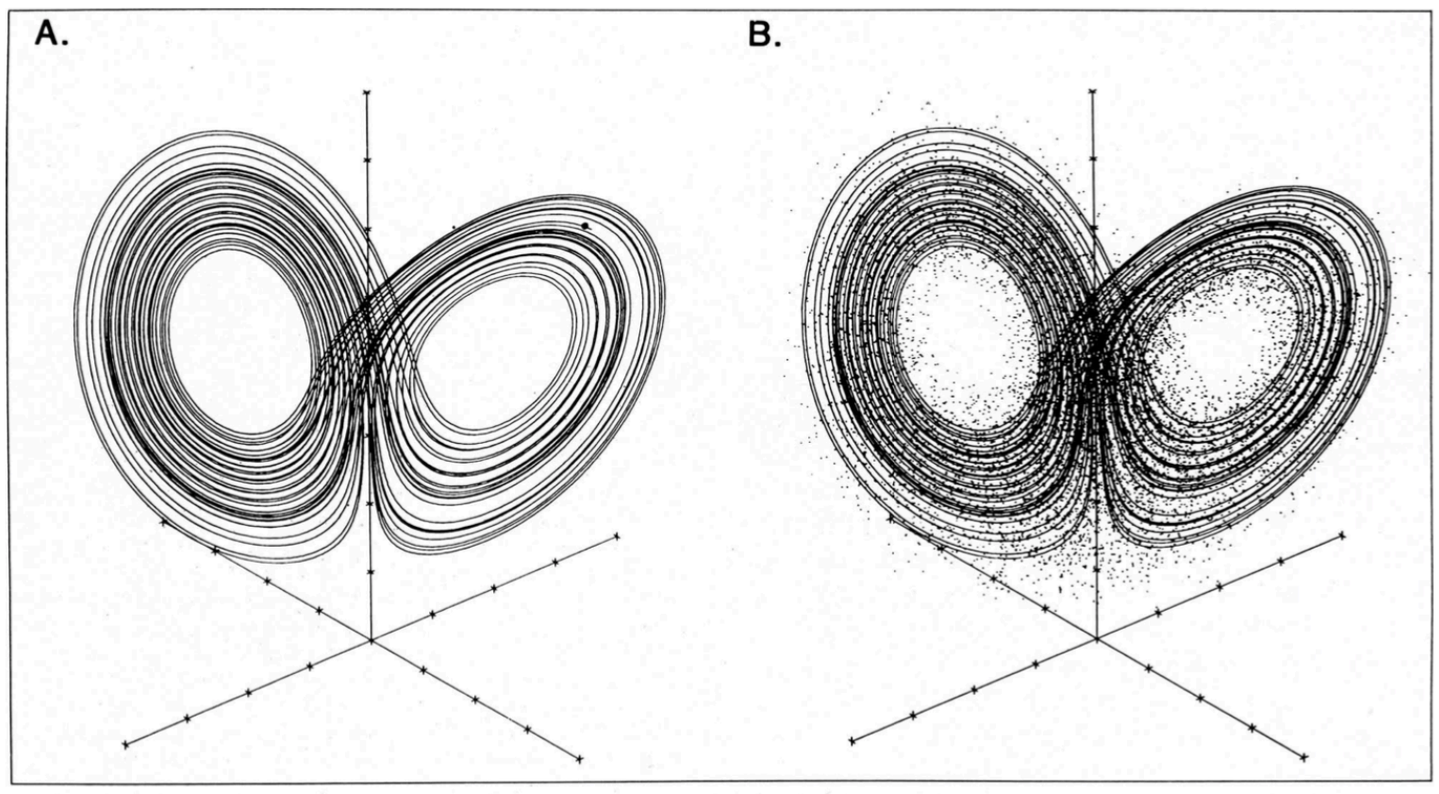


Figure 5

This is an example of a strange attractor with **implications** in the weather forecasting problem.

All **trajectories** (which will represent the evolution of that system for different initial conditions) will **eventually** converge and **remain** on that structure.

However, any two initially **nearby** trajectories in the attractor **do not remain** nearby, but **diverge**.

The fractal nature of an attractor does not only **imply** nonperiodic orbits.

As I said, it also causes nearby trajectories to **diverge**.

As with all attractors, trajectories that are **initiated** from different initial conditions

soon reach the attractor, but two nearby trajectories

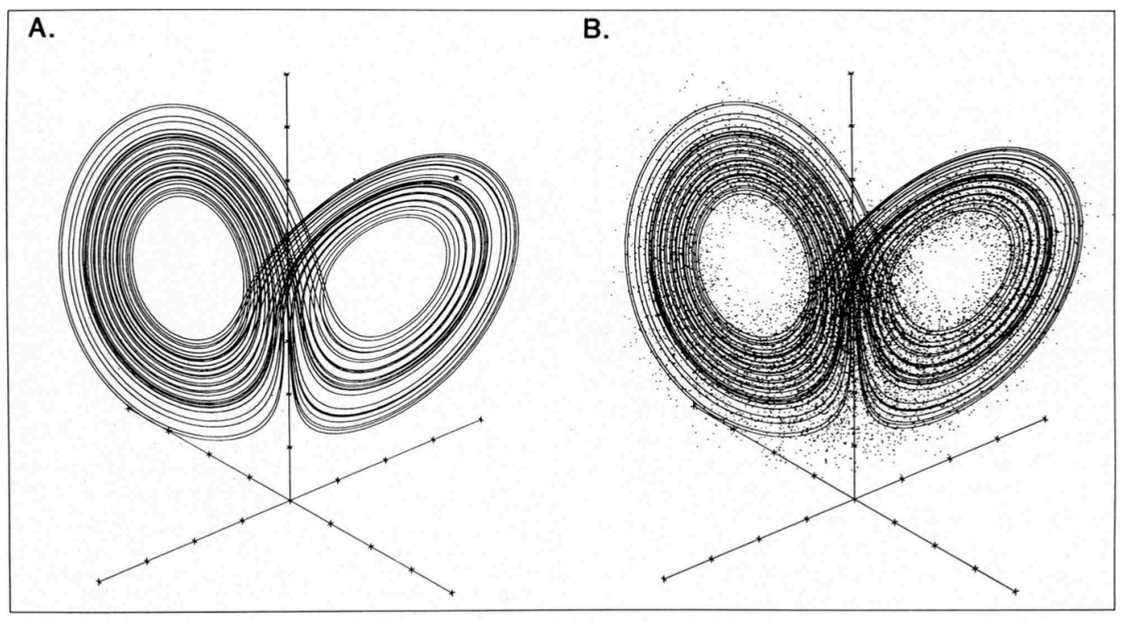
do not stay **close** to each other

(as was the case with the torus).

They soon **diverge** and

follow **totally** different paths in the attractor.

Remember Lorenz movie.



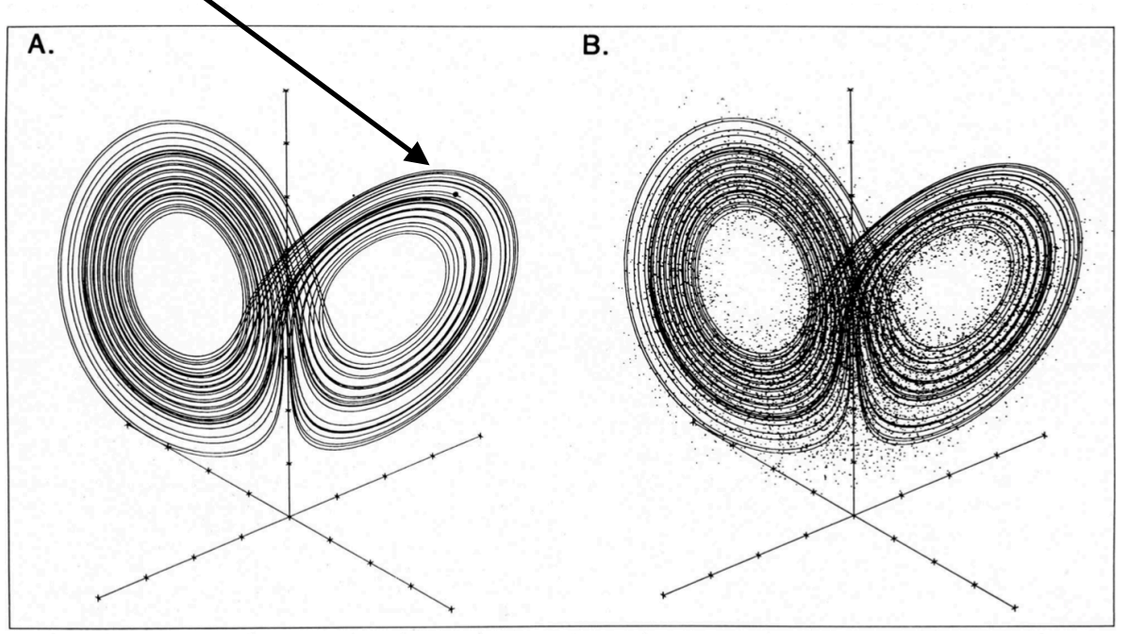
The divergence **means** that the evolution of the

system from two slightly different initial conditions

will be **completely** different, as may be seen in figures A and B.

The dot in figure A represents **10,000** initial conditions that are so close to each other

in the attractor that they are **indistinguishable**.



They may be **viewed** as 10,000 initial situations that differ only slightly from each other.

If we **allow** these initial conditions to evolve according to the rules (equations)

that describe the system, we **see** (figure B) that after some time

the 10,000 dots can be **anywhere** in the attractor.

In other words, the state of the system **after** some time can be **anything**

despite the fact that the initial conditions were **very** close to each other.

Apparently, the evolution of the system is very **sensitive to initial conditions**.

In this case we **say** that the system has **generated** randomness.

We can now **see** that there **exist** systems that, even though

they can be **described** by simple deterministic rules, can **generate** randomness.

Randomness generated **this way** has been termed **chaos**.

These systems are **called** chaotic dynamical systems

and their attractors are often **called strange** or **chaotic** attractors.

The implications of such findings are profound.

If one knows **exactly** the initial conditions, one can follow the trajectory
that corresponds to the evolution of the system from those initial conditions
and basically **predict** the evolution forever.

The **problem**, however, is that we **cannot** have a perfect knowledge of initial conditions.

Our instruments can only measure **approximately** the various parameters
(temperature, pressure, etc.) that will be used as initial conditions.

There will **always** be some deviation of the measured from the actual initial conditions.

They may be very close to each other, but they will **not** be the same.

In such a case, **even** if we completely know the physical laws that govern our system,
due to the nature of the **underlying** attractor the state of the system
at a later time can be **totally** different from the one predicted.

Simply, **due** to the nature of the system,

initial errors are **amplified**

and therefore **prediction** is **limited**.

Ideas from the theory of chaotic dynamical systems

have been **applied** to simple models

that describe **climatic** fluctuations and transitions between ice ages and today's climate.

These studies **attribute**

the broad-band structure of the spectrum of observed climatic data

to the **presence** of nonperiodic chaotic attractors.

The attractors are very **sensitive** to the initial conditions,

in accordance with the **intrinsic** unpredictability of the climatic system .

For something specific…. Now let me go back to Lorenz and look at his weather model in more detail….

As I said Edward Lorenz was a **meteorologist** who discovered by accident that weather is chaotic.

In the 1960s, he repeated a calculation to **predict** a weather trend, but rounded an initial value from six digits after the break point to only three digits.

Despite the **tiny** difference in the initial value, he got wildly different results.

That's **chaos**, and it gave rise to the idea of the "butterfly effect" - that the flap of a butterfly in China might cause a tornado in Texas two weeks later as discussed earlier.

To understand better what was happening, Lorenz took his rather complicated set of equations and **simplified** it to a set of only three equations that nevertheless captures the strange behavior he had noticed.

These three equations are now **commonly** known as the **"Lorenz Model"**.

In the Lorenz model, we have **three** variables, x, y, and z and they are functions of time, t.

$$\frac{\mathrm{d}x}{\mathrm{d}t} = \sigma(y - x),$$

$$\frac{\mathrm{d}y}{\mathrm{d}t} = x(\rho - z) - y,$$

$$\frac{\mathrm{d}z}{\mathrm{d}t} = xy - \beta z.$$

This model can be **interpreted** as a simplified description of convection in gases or fluids, but just what it describes does not really matter for our purposes.

The nice thing about the Lorenz model is that you can **integrate** the equations on a laptop.

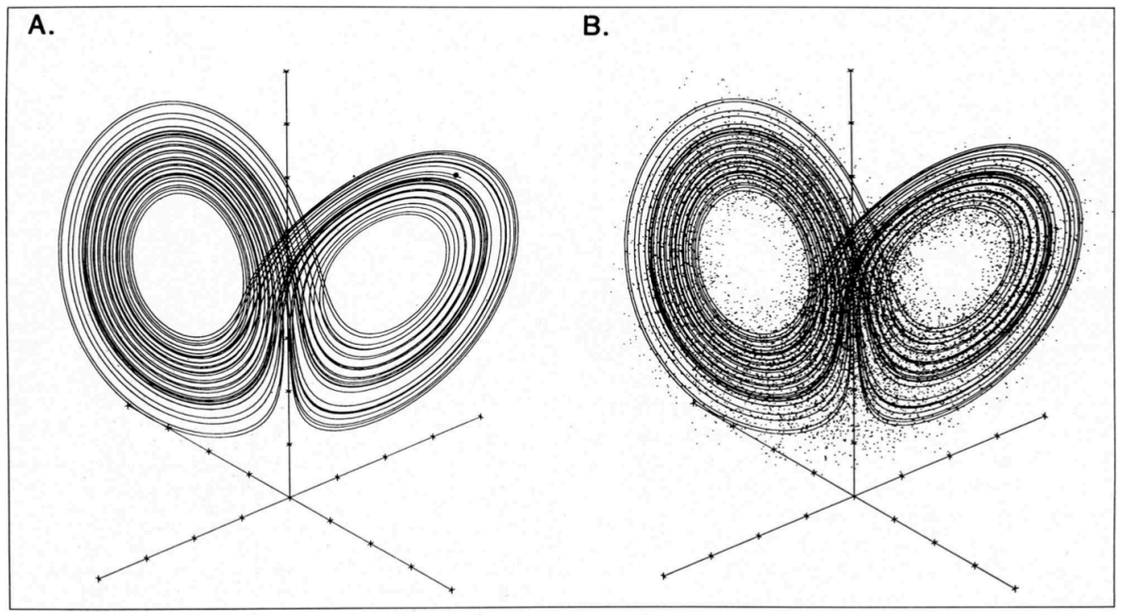
Showed you **one** of the solutions earlier —— **Lorenz** curves……

Each of the axes in the graph is one of the directions x, y, z, so the solution to the Lorenz model will be a curve in these three dimensions.

As you saw, it circles around two different locations, **back and forth**.

It's not only this one solution which does that, actually all the solutions will end up doing circles close by these two places in the middle, which is called the **“attractor”**.

The attractor has an interesting shape, and coincidentally happens to look somewhat like a butterfly with two parts you could call “wings”.



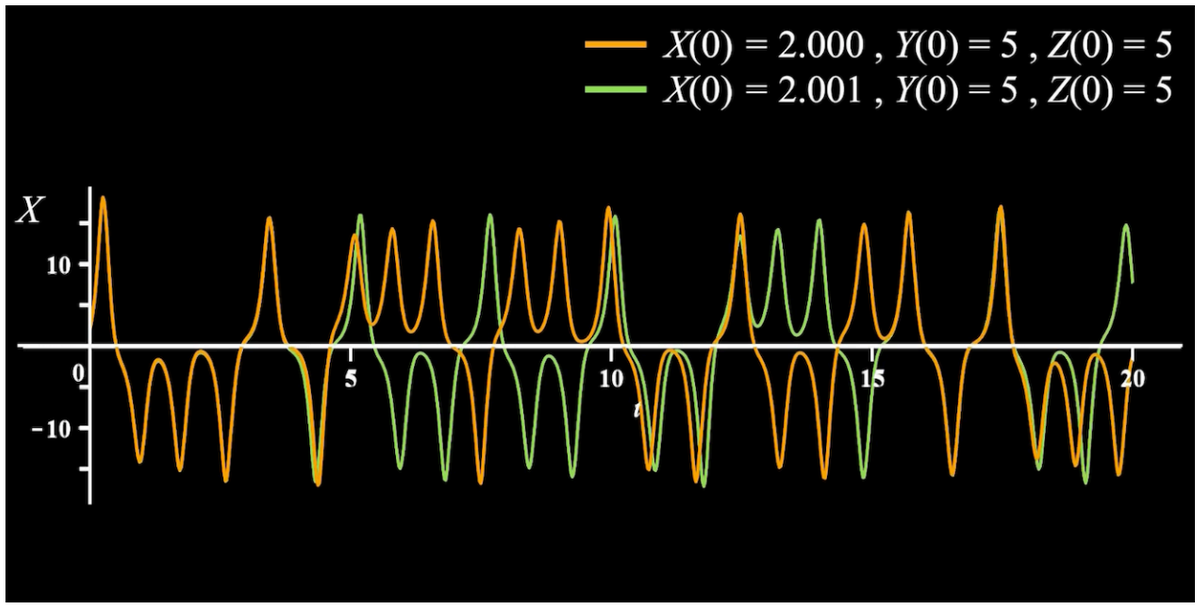
But more **relevant** for us is that the model is **chaotic**.

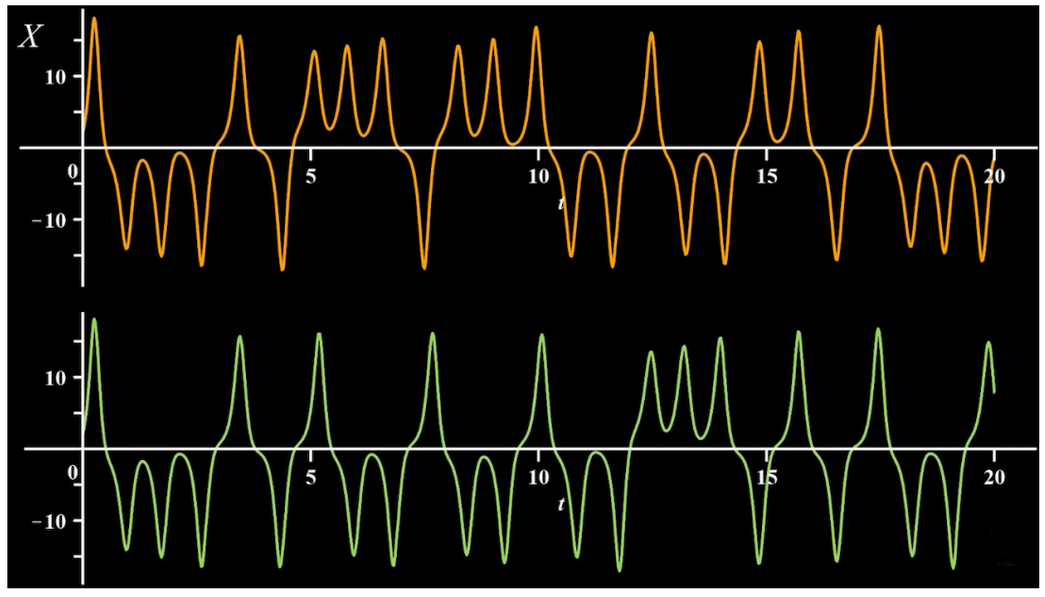
If we take two initial values that are very **similar**, but not exactly identical, then the curves at **first** look very similar, but then they run **apart**, and after some while they are entirely **uncorrelated** as we saw.

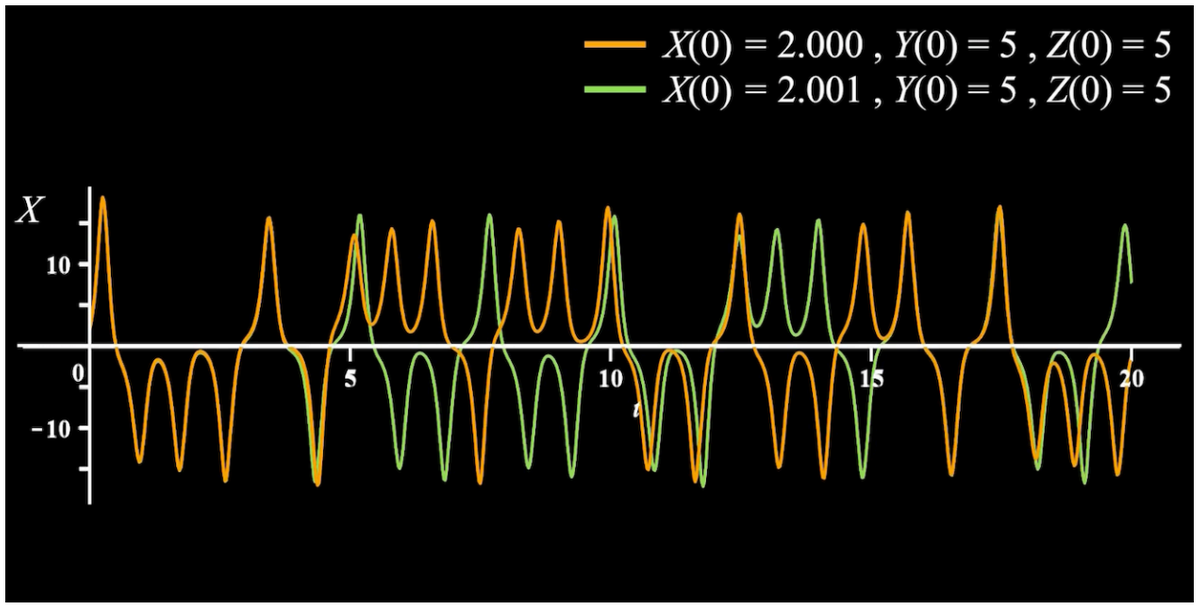
These three dimensional plots are **pretty**, but it's **somewhat** hard to see just what is going on, so in the following will merely look at **one** of these coordinates (one of the variables), that is the x-direction.

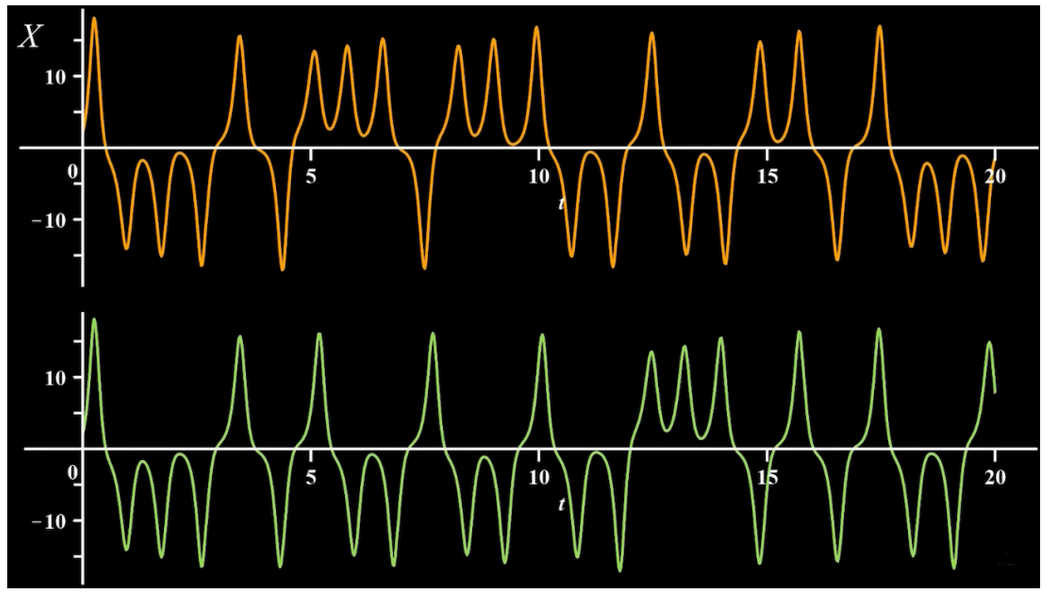
From the three dimensional plot, you **expect** that the value in x-direction will go back and forth between two numbers, and indeed that's what **happens**.

Here you see again the curves previously shown for two initial values that **differ** by a tiny amount.









At **first** the two curves look pretty much identical, but **then** they diverge and after some time they become entirely uncorrelated.

As you see, the curves **flip** back and forth between positive and negative values, which **correspond** to the two wings of the attractor.

In this early range, maybe up to t = 5, you **would** be able to make a decent weather forecast.

But **after** that, the outcome depends very sensitively on exactly what initial value you used, and then measurement error makes a good prediction impossible.

That's chaos.

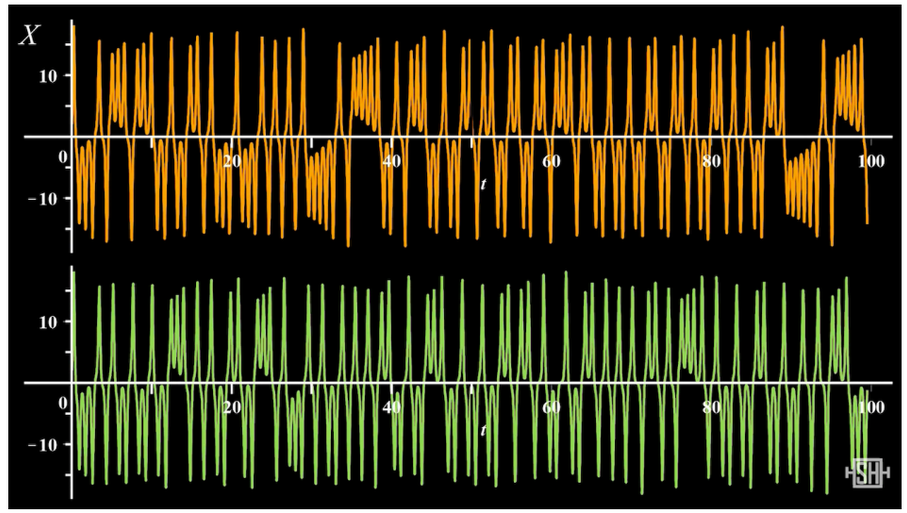
Now, we want to **pretend** that these curves say something about the weather(a model).

Maybe they describes the weather on a strange planet where it either doesn't rain at all or it pours and the weather just flips back and forth between these two extremes.

Besides making the short-term weather **forecast** you could then also ask what's the average rainfall in a certain period, say, a year.

To calculate this **average**, you would integrate(sum) the curve over some period of time, and then divide by the duration of that period.

So let us **look** at these curves again.



Just by eyeballing these curves
you'd **expect** the average to be approximately zero.

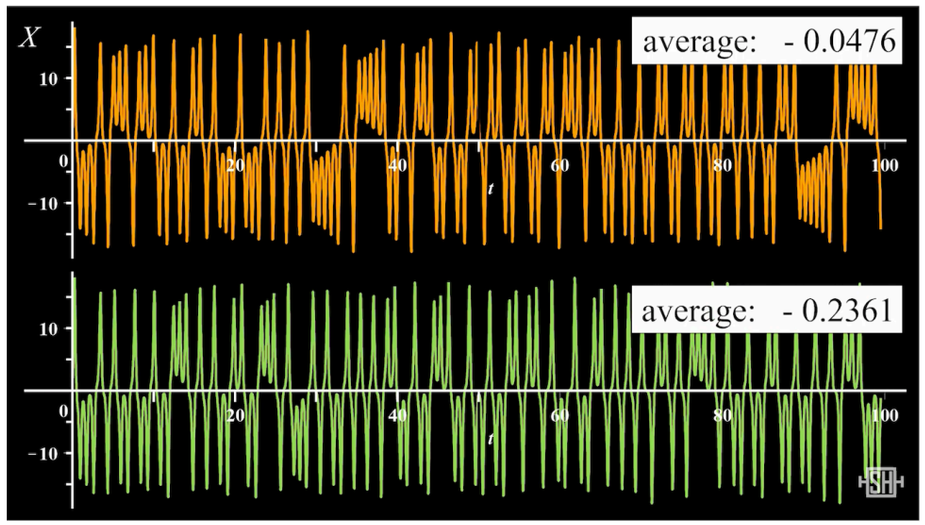
Indeed, if calculate the average from t equals zero to
t equals one hundred, it comes out to be **approximately** zero.

What this means is that the system spends about **equal** amounts of time on each wing of the attractor.

To stick with our story of rainfall on the weird planet, you can imagine that the curve shows **deviations** from a reference value that you set to zero.

The average value **depends** on the initial value and will **fluctuate** around zero because I am only integrating over a finite period of time, so one arbitrarily cuts off the curve somewhere.

If you'd **average** over longer periods of time, the average would **inch** closer and closer to zero.



What one does now is add a **constant** to the equations of the Lorenz model.

Call this constant "f" and mimics what climate scientists call "**radiative forcing**"(will discuss in great detail later).

The radiative forcing **could be** the excess power per area that Earth captures due to increasing carbon dioxide levels.

Again that's **relative** to a reference value.

$$\frac{\mathrm{d}x}{\mathrm{d}t} = \sigma(y - x) \ \texttt{+f}$$

$$\frac{\mathrm{d}y}{\mathrm{d}t} = x(\rho - z) - y \ \texttt{+f}$$

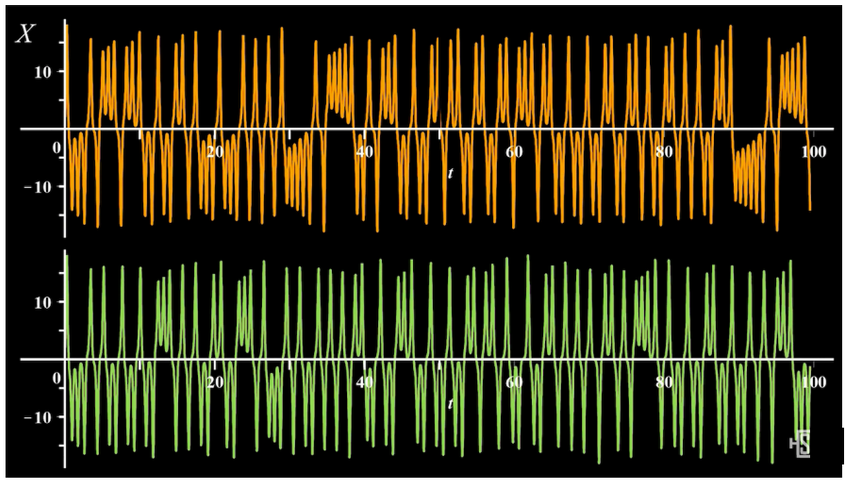
$$\frac{\mathrm{d}z}{\mathrm{d}t} = xy - \beta z.$$

Must emphasize again that am using this model only as an analogy.

It does not actually describe the real climate.

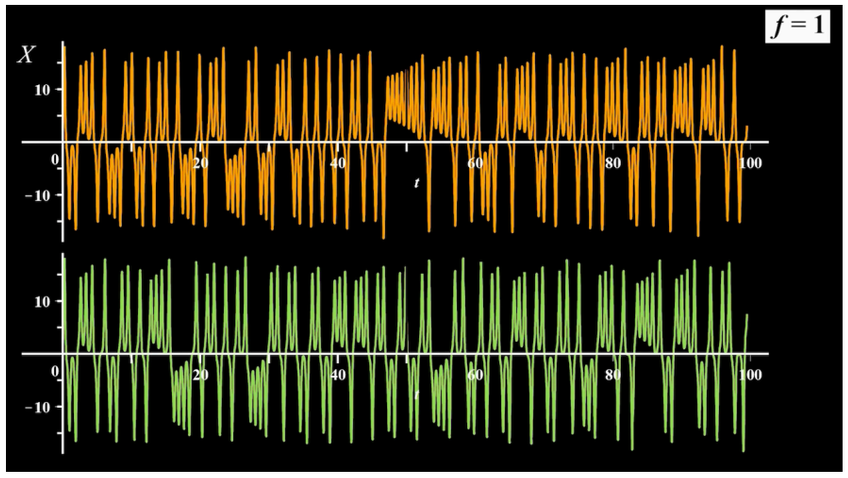
But it does make a good example for how to make predictions in chaotic systems.

Having said that, let us look again at what the curves look like with the added forcing.



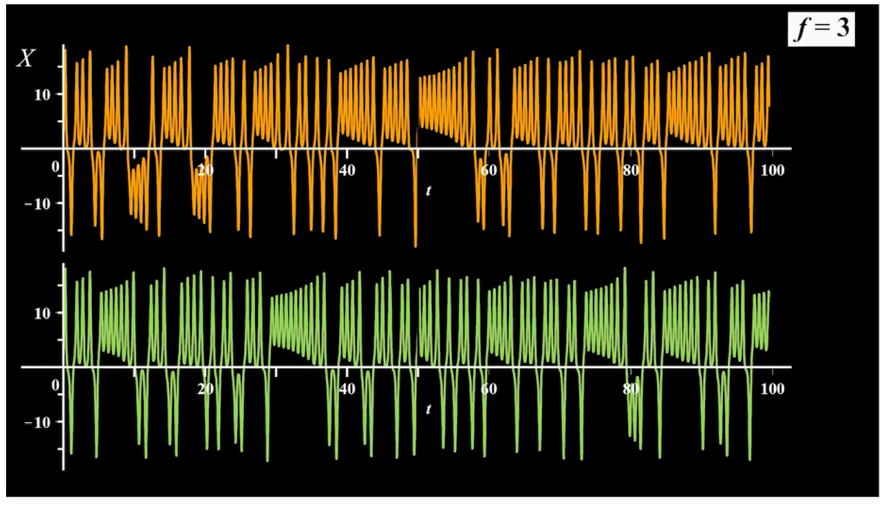
f=0

original data



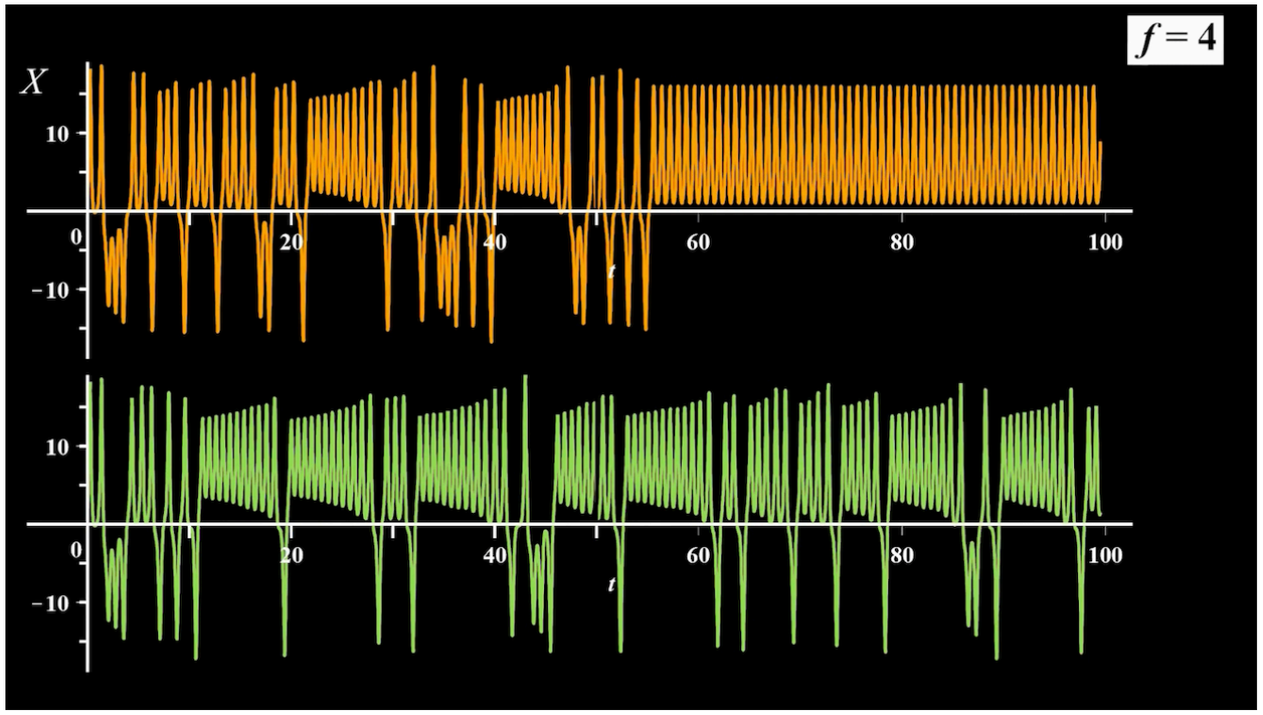
f=1

looks pretty much the **same** as previously.



f=3

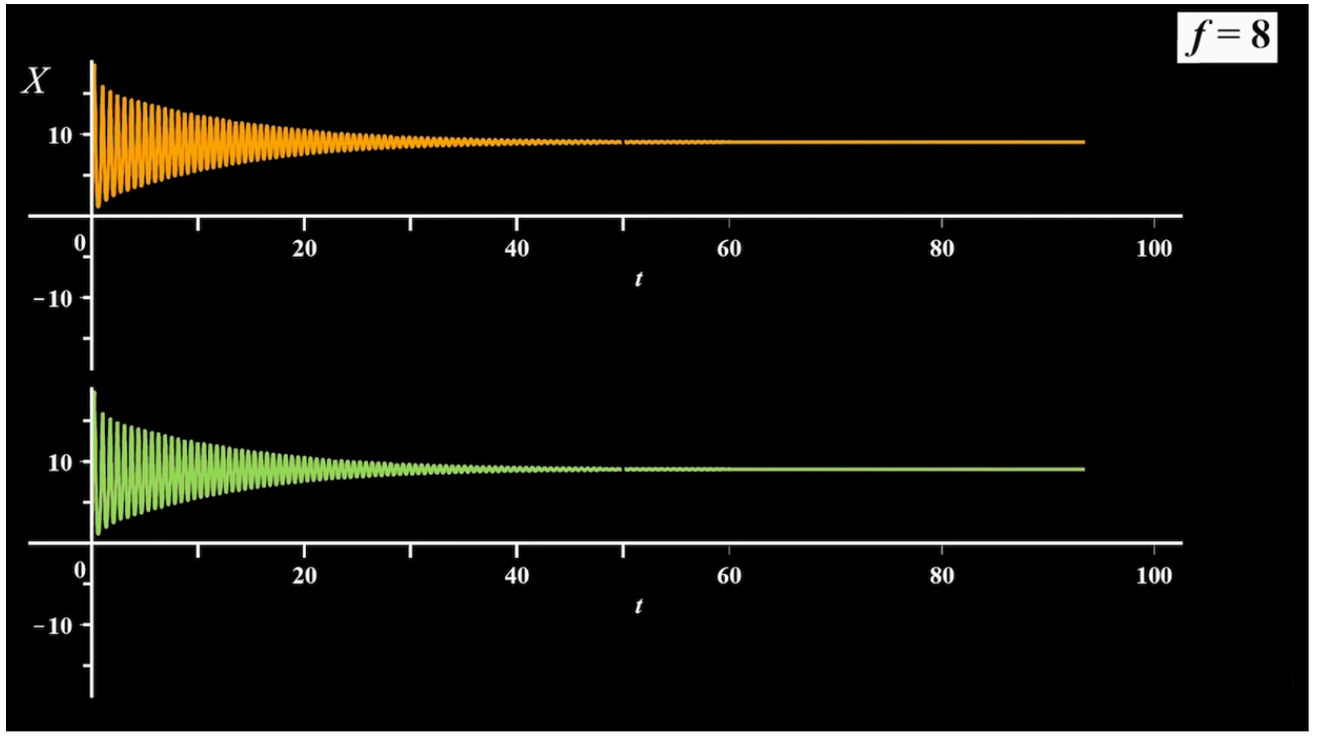
looks like the system is spending a **little** more time in the upper range



f=4

it **clearly** changes

f=8

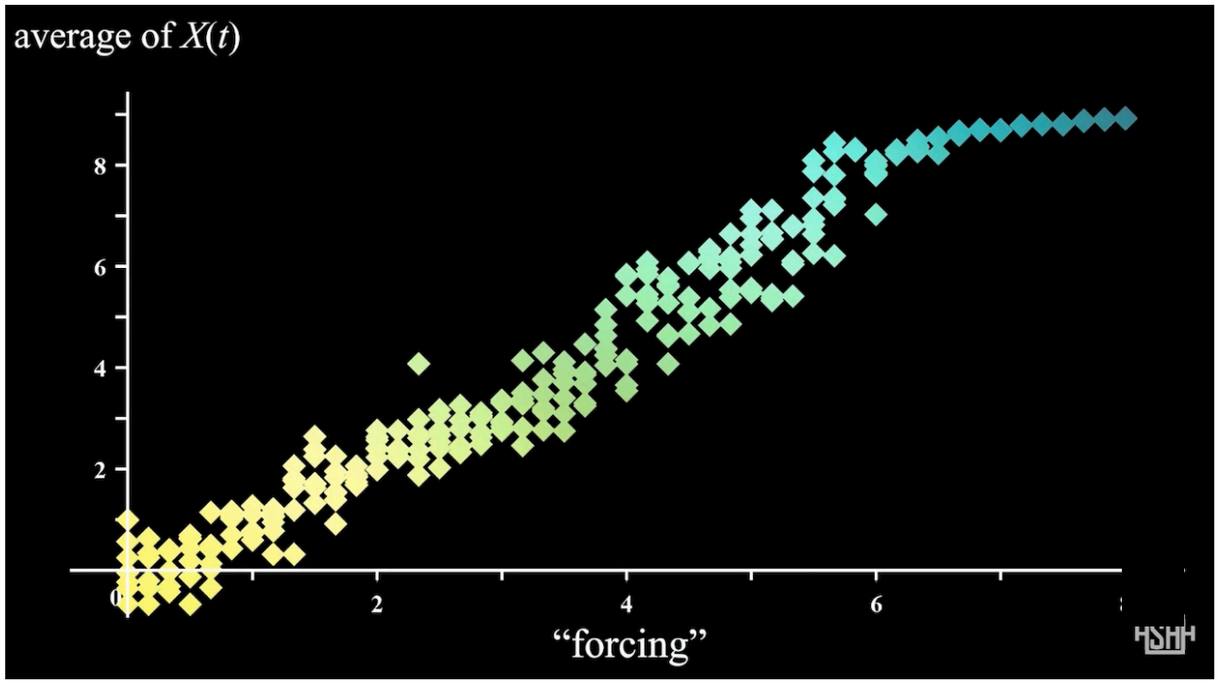


And just for **fun**, if you turn **f** up beyond seven or so, the system will get **stuck** on one side of the attractor **immediately.**

The relevant point is now that this happens for all initial values.

Even though the system is **chaotic**, one **clearly** sees that the **response** of the system **does** have a **predictable** dependence on the input parameter.

To see this **better**, one **calculates** the average of these curves as a function of the “radiative forcing”, for a sample of initial values.



One can see that the average value is **strongly correlated** with the radiative forcing.

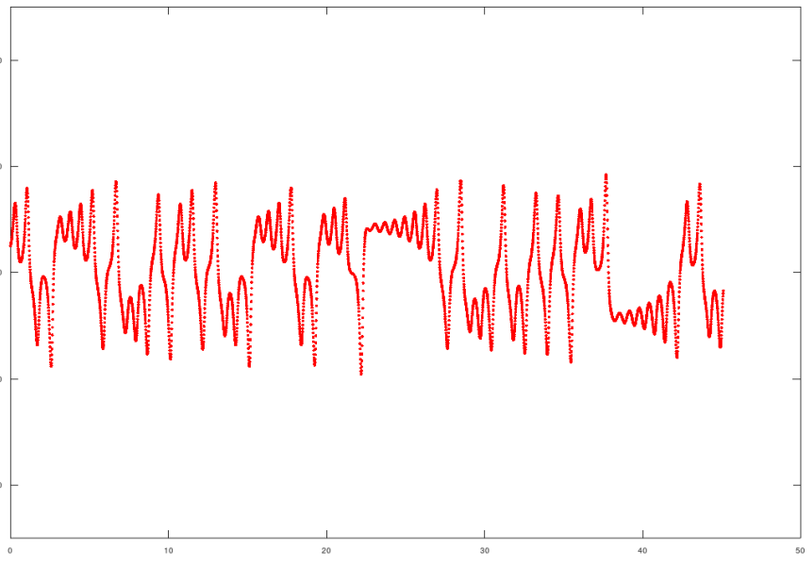
Any **scatter** that appears is **because** the averaging is over a rather arbitrarily chosen **finite** period.

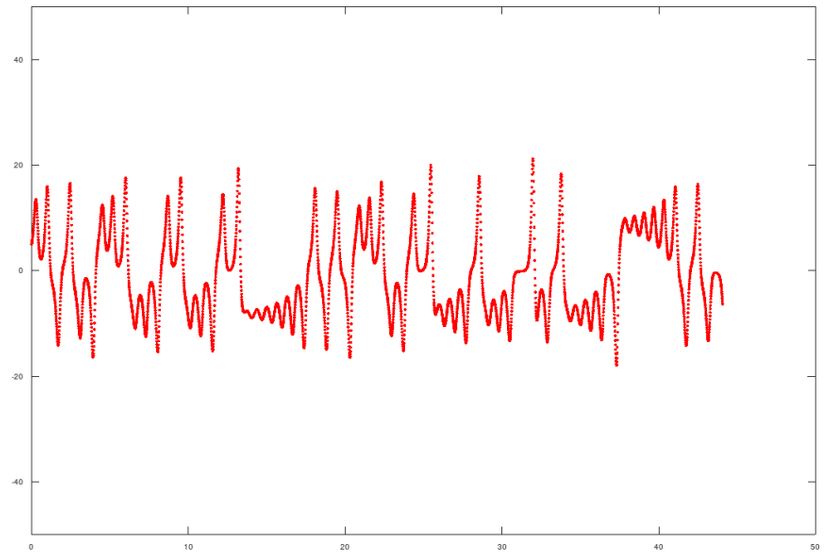
What this means is that in a chaotic system, the trends of average values can be predictable, even though you cannot predict the exact state of the system beyond a short period of time.

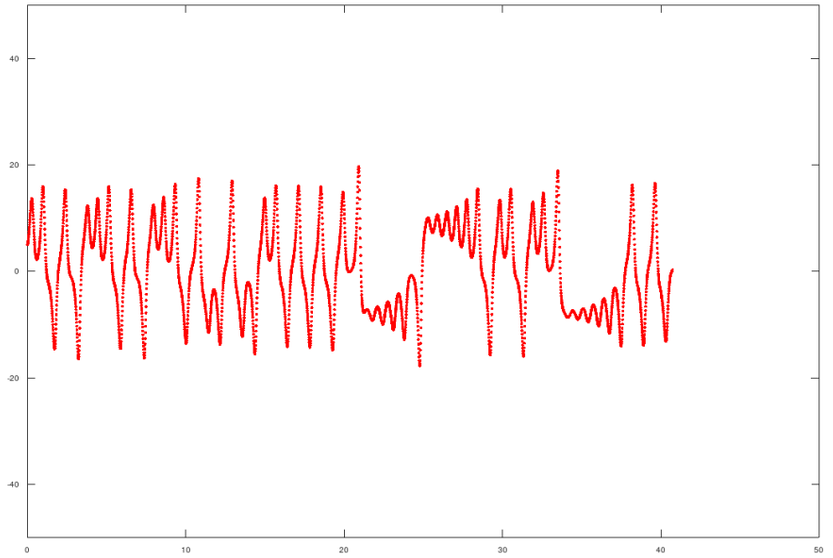
And this is exactly what is happening in climate models.

Scientists **cannot** predict whether it will rain on June 15th, 2079, but they can very well predict the **average rainfall** in 2079 as a function of **increasing** carbon dioxide levels(f).

A better simulation …….



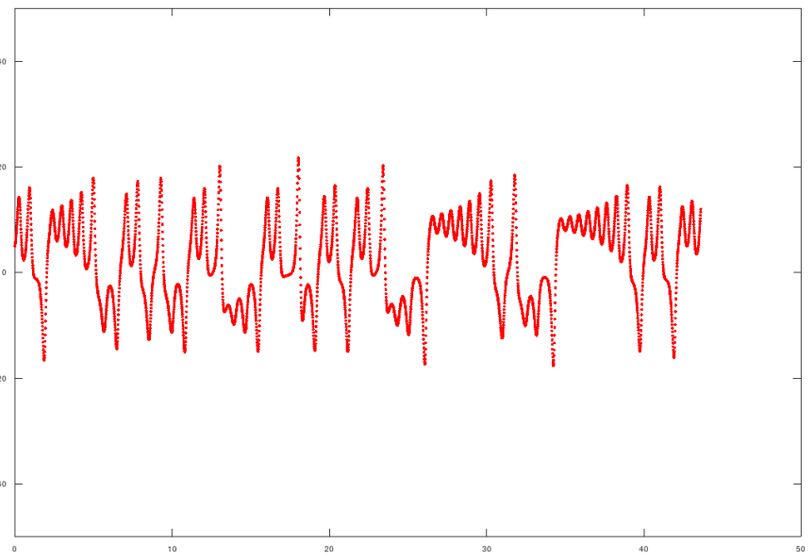


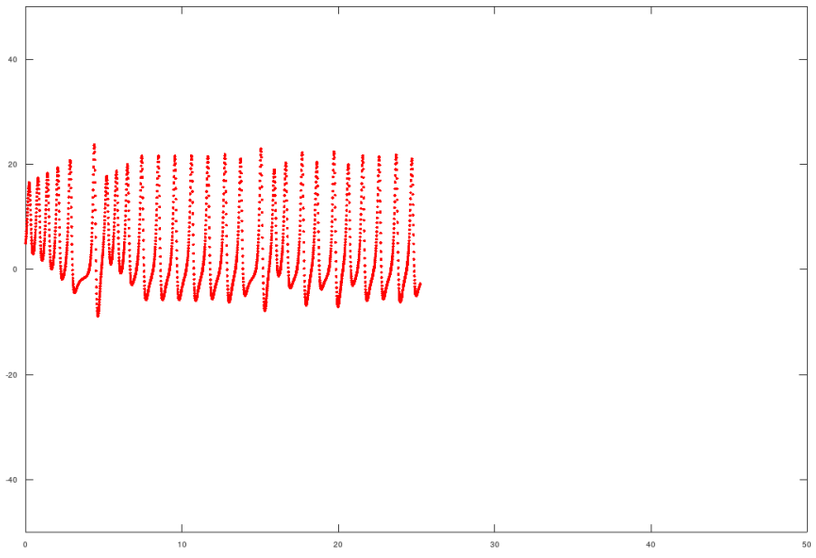


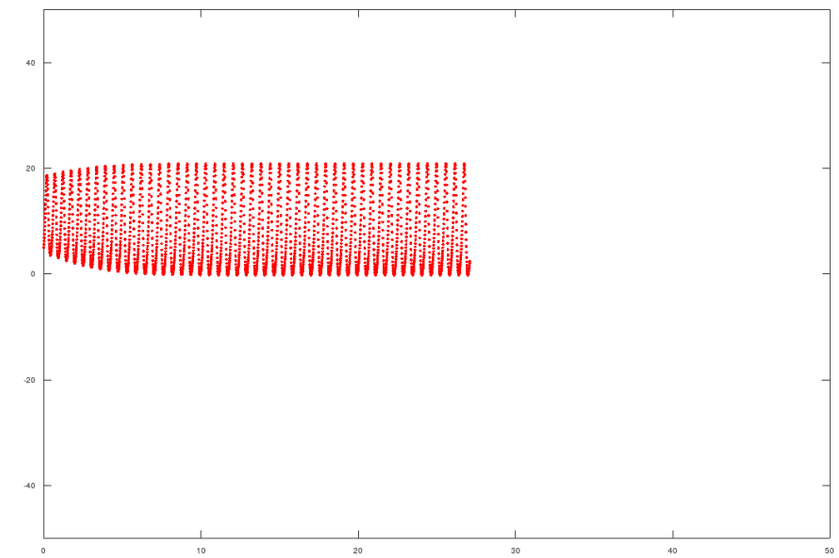
f=0 aver=0.40

f=3 aver=-1.02

f=5 aver=0.32







f=10 aver=2.17

f=30 aver=3.42

f=50 aver=8.23

The shift upwards due to “radiative forcing” is clear …

The search for attractors in weather and climate

Most **important** consequence of knowing the **fractal dimension** of an attractor

is that the **dimensionality** of an attractor

indicates the **minimum** number of variables

present in the evolution of the corresponding dynamical system

Therefore, the **determination** of the fractal dimension of an attractor

sets a number of constraints that must be satisfied

by any model used to **predict** the evolution of a system.

If the mathematical description(equations) of a dynamical system is **given**,

the number of variables is **known,**

—> generation of the state space and the attractor is **straightforward** (we did it earlier).

If the mathematical formulation of a system is **not available** (only have data),

the state space can be **replaced** by a phase space

produced using a **single** record of some observable variable x(t) from that system.

The physics **behind** such an approach is that a single record from a dynamical system
is the **outcome** of all interacting variables and
thus information about the dynamics of that system should, in principle,
be **included** in any observable variable.

The mathematical **procedure** used to prove the existence and nature (chaotic or not)
is as follows (for **completeness**- **do not worry** if the mathematics is too high level).

It is **assumed** that variables present in the evolution of the system in question
satisfy a set of n **first-order** differential equations (remember Lorenz):

$$\frac{dx}{dt} = \sigma(y - x),$$

$$\frac{dy}{dt} = x(\rho - z) - y,$$

$$\frac{dz}{dt} = xy - \beta z.$$

$$\dot{x}_1 = f_1(x_1, x_2, \ldots\ldots\ldots x_n)$$
$$\dot{x}_2 = f_2(x_1, x_2, \ldots\ldots\ldots x_n)$$
$$\vdots$$
$$\dot{x}_n = f_n(x_1, x_2 \ldots\ldots\ldots x_n)$$

the coordinates of the state space are (x_1, x_2 , …. x_n).

System of differential equations can be **reduced** to 1 nonlinear equation of **nth order**
by **successive** differentiation of one of the equations describing the system.

In this way, one may **obtain** an nth order differential equation.

$$x_1^{(n)} = f(x_1, \dot{x}_1, \ldots\ldots\ldots, x^{(n-1)})$$

and **replace** the coordinates of the state space with

$$(x_1, \dot{x}_1, \ddot{x}_1, \ldots\ldots\ldots x_1^{(n-1)}) \text{ or } (\dot{x}_1, \ddot{x}_1, \ldots\ldots\ldots x_1^{(n)})$$

without any loss of information about the evolution or the dynamics of the system.

This way we can **define** the coordinates of an **appropriate** phase space,
which should **approximate** the dynamics of the system (original state space) from
which the observable x(t) was **sampled** (or in other words the **unknown** state space).

The **search** for attractors in weather and climate is driven
by the **desire** to investigate the observed **complexity** in the atmosphere.

The theory of chaos **provides** new tools to do just that. .

Observable weather variables are **used** in the search for attractors in weather and climate

Various numerical studies **found** climatic attractors of dimension 3.1 and between 6 and 7.

Conclusions

Many systems in nature are **chaotic**.

The developments in the study of chaotic dynamical systems

have **suggested** that nature thus imposes **limits** on prediction.

At the **same** time, however, it has been **realized** that the very **existence**

of the attractors **implies** that **randomness** is **restricted** to the attractors.

The **atmosphere** may be **chaotic**, but its **evolution** is **confined**

to a **specific** area in the state space that is **occupied** by the attractor.

No states outside this attractor are **allowed**.

FOR EXAMPLE:

The **winds** associated with a high pressure system, for example,

can **never** blow counterclockwise.

Even though it has provided an **excuse** for the unpredictability of weather,
the theory of dynamical systems is **slowly shaping** our way
of **investigating** the weather and its prediction.

Together with some **pessimism**, the study of chaotic dynamical systems
provides some **optimism**.

We may **never** be able to predict the weather exactly,
but improvements in weather forecasting are **feasible**
if we **improve** the completeness and accuracy with which
we **measure** the initial condition of the atmosphere,
and, most important, if we **understand** predictability at **different time scales**.

Chaos and Climate

the equations used in all models

The chaotic **nature** of atmospheric solutions of the Navier-Stokes equations for fluid flow
have great **impact** on weather forecasting,
but the evidence **suggests** that it has much less importance for climate prediction.

Chaos is usually **associated** with the **sensitivity** of a deterministic system to **infinitesimal** perturbations in initial conditions.

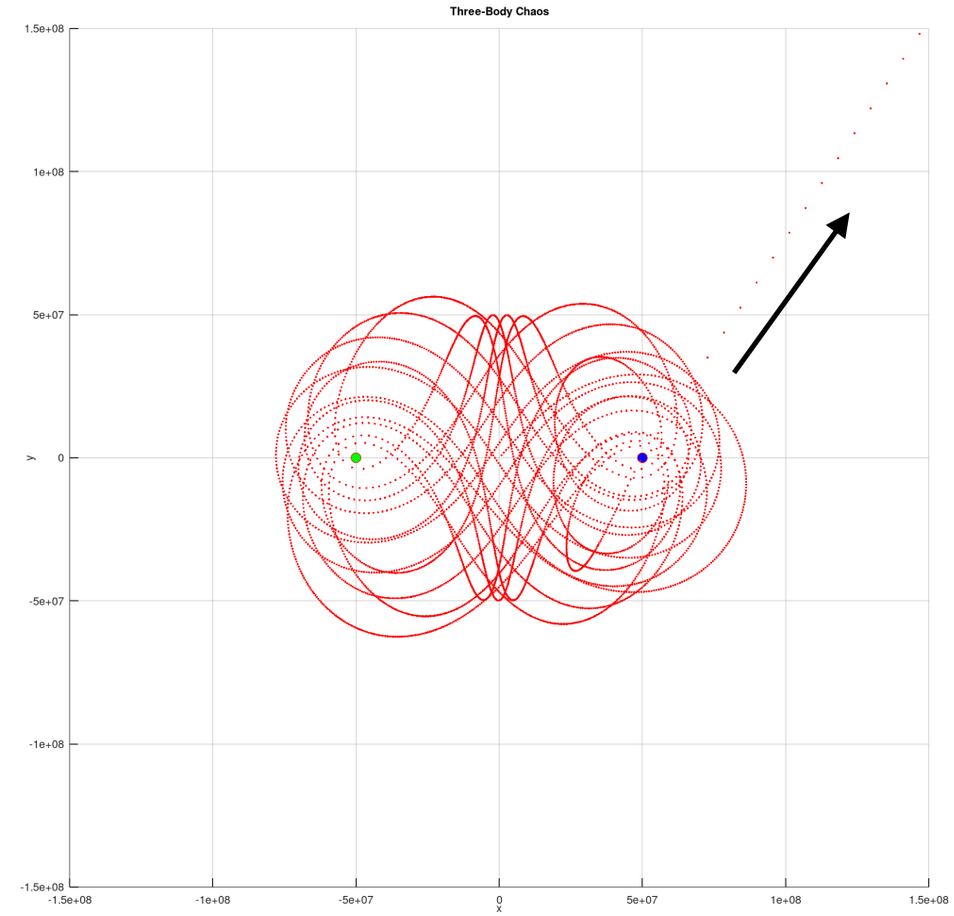
The **identification** of chaos in atmospheric systems is **due**, as we saw, to an **accidental** discovery by Lorenz in 1961.

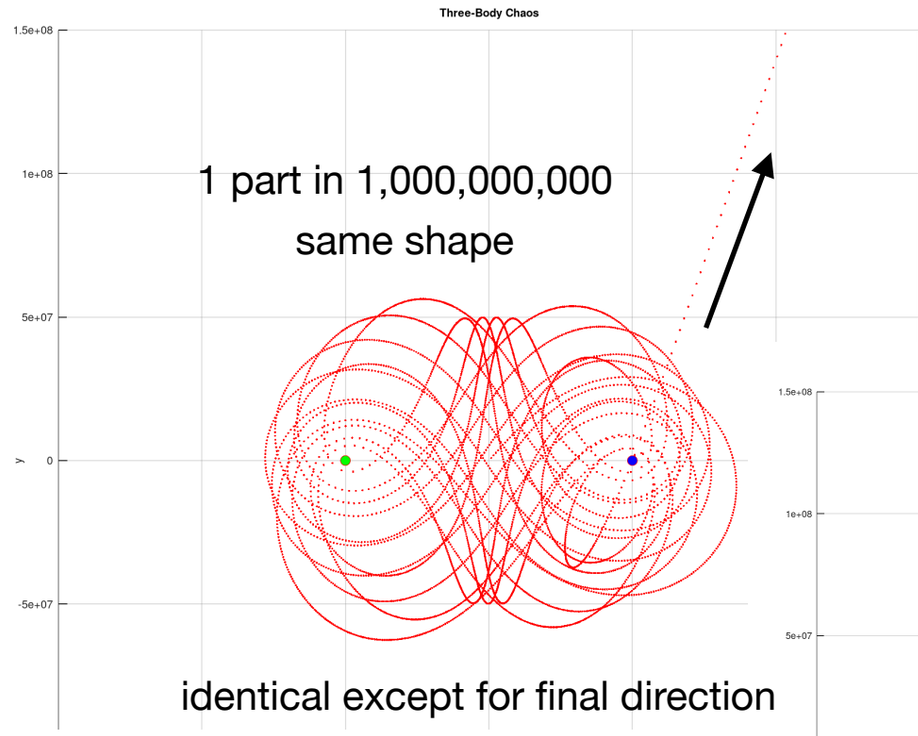
So it was an **early atmospheric model** which provided some of the **first insight** into the `chaos effect', thus **teaching** us something quite profound about nature

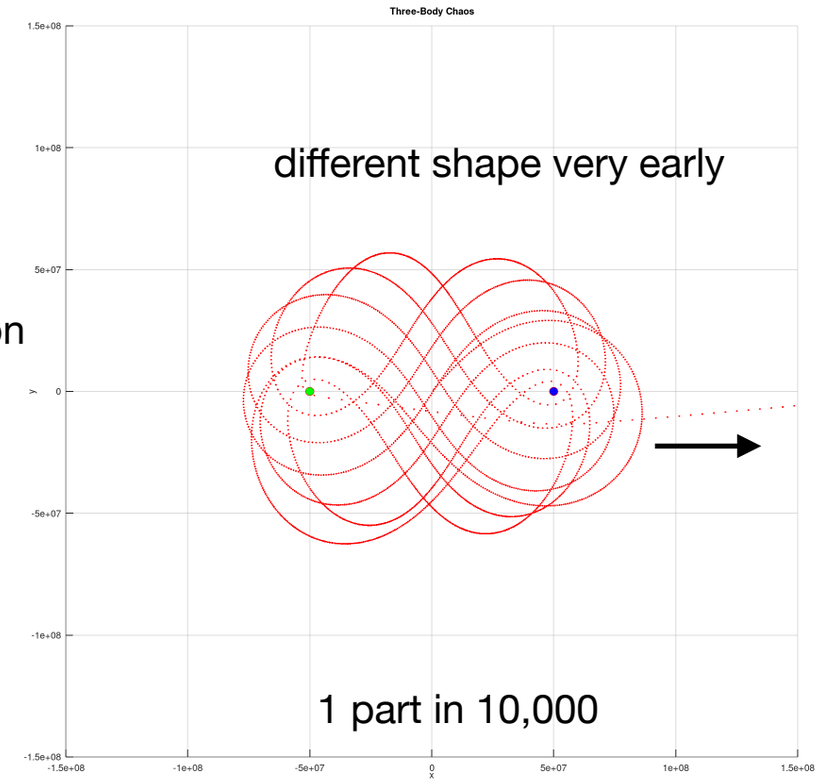
In fact, this type of behavior had **already** been **identified** and studied more than 60 years earlier by Poincare, in the form of the ``**3-body problem**'' of celestial dynamics.

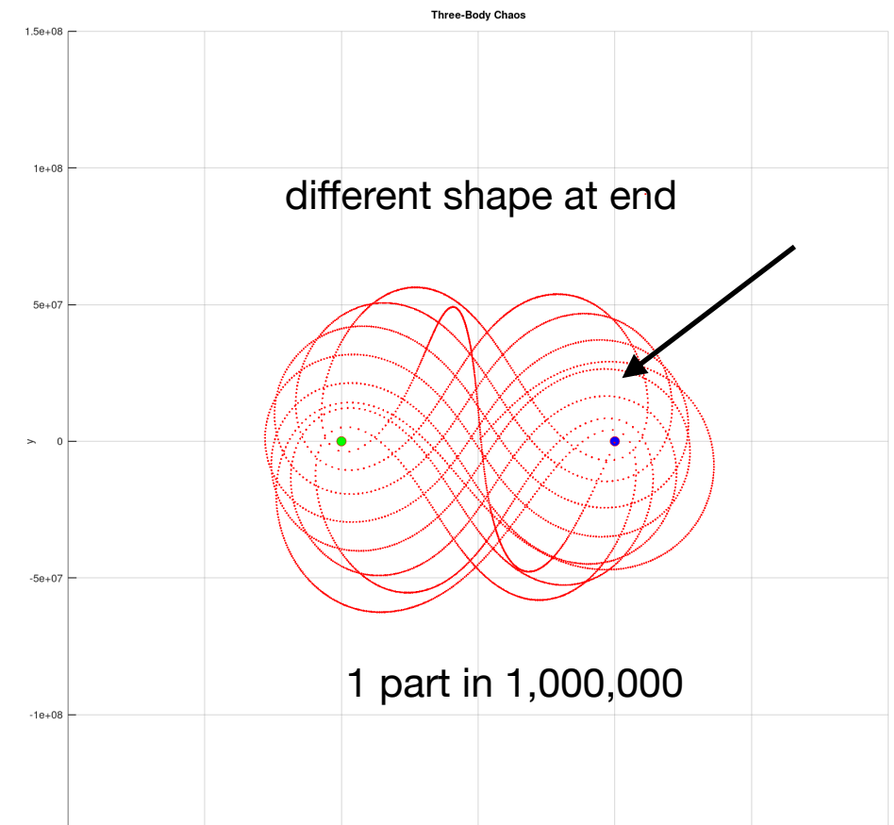
2 stars (or planets etc) in **orbit** around each other will each **follow** a **regular** ellipsoidal trajectory around their joint center of mass.

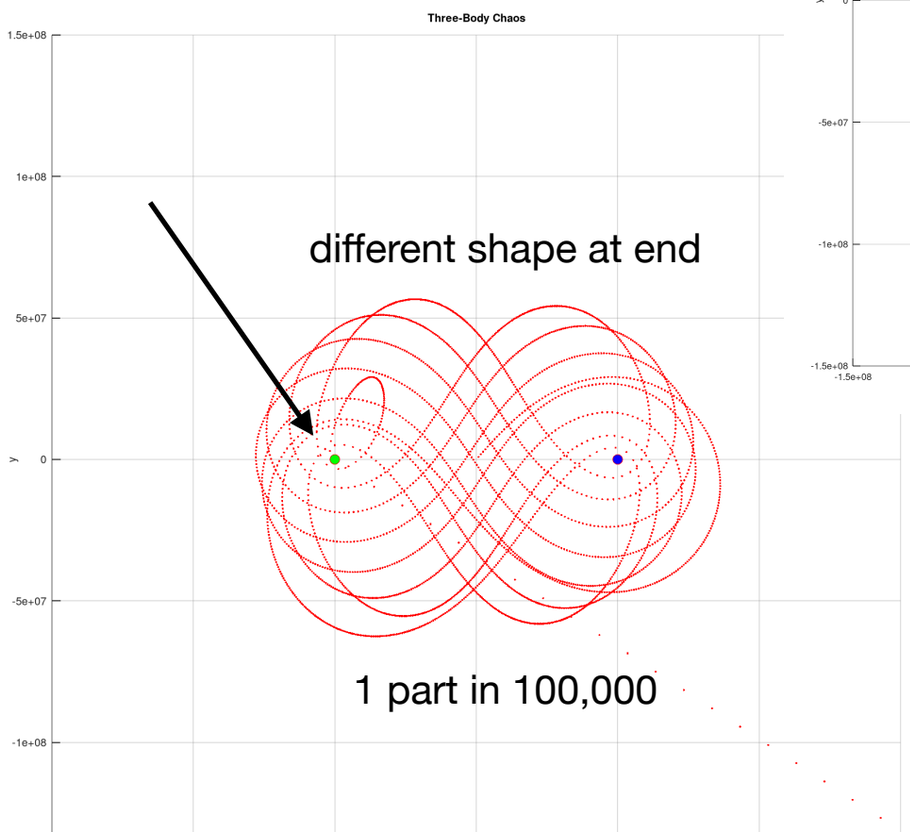
However, when a **3rd** (or more) body is thrown into the mix, their **future** trajectories are highly **sensitive** to the precise initial conditions.


Three-Body Chaos
1.5e+08
1e+08
5e+07
0
-5e+07
-1e+08
-1.5e+08
y
-1.5e+08
-1e+08
-5e+07
0
5e+07
1e+08
1.5e+08
x


Three-Body Chaos
1.5e+08
1e+08
5e+07
0
-5e+07
y
1 part in 1,000,000,000
same shape
identical except for final direction


Three-Body Chaos
1.5e+08
1e+08
5e+07
0
-5e+07
-1e+08
-1.5e+08
y
different shape very early
1 part in 10,000
-1.5e+08
-1e+08
-5e+07
0
5e+07
1e+08
1.5e+08
x


Three-Body Chaos
1.5e+08
1e+08
5e+07
0
-5e+07
-1e+08
y
different shape at end
1 part in 1,000,000


Three-Body Chaos
1.5e+08
1e+08
5e+07
0
-5e+07
-1e+08
y
different shape at end
1 part in 100,000

Back to atmospheric physics.

Chaotic **behavior** is a highly-studied and well-understood phenomenon
of all **realistic** global models, arising directly from
the **nonlinearity** of the Navier-Stokes equations for fluid flow used in the models.

So any uncertainty in the current atmospheric state, **however**, small,
will **ultimately** grow and **prevent** accurate weather forecasts in the long term.

This is the sort of thing that is **easy to show** with numerical models of the atmosphere.

Simply perturb a reference run, and see what happens.

So long as the perturbation is
not rounded out by the limited numerical precision of the model, it will **invariably** grow.

Describe…….

Of **course** the existence of an unknown butterfly flapping its wings

has no direct bearing on weather forecasts,

since it will take far too **long** for such a small perturbation to grow to a significant size,

and we have many more **immediate** uncertainties to worry about.

Chaos is **defined** with respect to infinitesimal perturbations and infinite integration times,

but our **uncertainties** in the current atmospheric state

are far too **large** to be treated as infinitesimal,

and furthermore, all of our models have **errors**

which means that they will inevitably **fail** to track reality anyway

within a few days **irrespective** of how well they are initialized.

Nevertheless, chaos theory and associated methods continue to play a major role

in the research and development of ensemble weather prediction methods.

Although **ultimately** chaos will kill a weather forecast, this **does not necessarily** prevent long-term prediction of the climate.

By climate, we mean the statistics of weather, averaged over suitable time and perhaps space scales.

We **cannot** hope to accurately predict the temperature

in Dallas at 9am on the 23rd July 2050,

but we **can** be highly confident that the average temperature in the USA in that year

will be substantially higher in July than in January.

Of course, we **don't** need a model to work that out

historical observations(data) already give strong evidence for this prediction.

The models are useful for a different purpose!

These models, which are based on physical principles, **seem to be effective**

in reproducing the **response** to changes due to radiative forcing.

This **strongly supports** their use in the prediction of responses

to anthropogenic (originating in human activity) forcing.

Fortunately, the calculation of climatic variables (i.e., long-term averages)

is much **easier** than weather forecasting,

since weather is **ruled** by the vagaries of random fluctuations,

while climate is **not**.

Imagine a pot of boiling water.

A weather forecast is **like** the attempt to predict

where the **next** bubble is going to rise (physically this is an **initial value** problem).

A **climate** statement would be that

the **average** temperature of the boiling water is 100^{o}C at normal pressure,

while it is **only** 90^{o}C at 2,500 meters altitude in the mountains,

due to the lower pressure (that is a **boundary** value problem).

We can **demonstrate** this sort of climate response clearly in the Lorenz model,

or any more complex climate model.

Perturbing the initial conditions gives a **completely** different trajectory (weather),

but this **averages** out over time,

and the **statistics** of different long-term runs are **indistinguishable**.

However, a **steady** perturbation to the system which **leads** to unpredictability

can generate a significant change to the long-term statistics.

Remember in the model I was able to **completely change** future weather with strong forcing.

Simulations **disprove** the common(for decades) but misguided claim that chaotic weather necessarily **prevents** meaningful climate prediction.

In fact, **all** climate models do predict that the **change** in globally-averaged steady state temperature,

is almost exactly proportional to the **change** in net radiative forcing,

indicating a **near-linear** response of the climate, at **least** on the broadest scales,

and linear equations are non-chaotic. Why they work in this case!!

The only **uncertainty** is in the steepness(rate of change) of the slope of the change, which is **what** “climate sensitivity” describes.

The **climate** of a model can be easily **defined** in terms of the **limit** of the statistics of the model output as the averaging time **gets large**.

This **limit** is well-defined for all climate models.

However, it is still the case that the real world is more **messier** to deal with!

The real climate system **varies** on all time scales, from daily weather, through annual, multi-year and decadal effects, Milankovitch behavior, glacial-interglacial cycles, plate tectonics and continental configurations, **right up** to the ultimate death of the Sun, etc, etc.…

The **average** temperature, and all other details of the climate system, will **vary** substantially depending on the time scale being used.

So how can we talk meaningfully about "the climate" and "climate change"?

Well, **although** there are interesting scientific questions to **ask** across all the different time scales, the **policy-relevant** portion is on the **multi-decadal and centennial time scales**.

It is quite **clear** that the perturbation that we are currently imposing is **already** large, and will be **substantially** larger, by up to an **order of magnitude**, than any **plausible natural** variability over this time scale.

So **for** the policy-relevant issues, we **generally focus** on the physical atmosphere-ocean system, **sometimes** with coupled carbon-vegetation system, and **treat** the major ice sheets, orbital parameters and planetary topography as **fixed** (change very slowly on time scale of interest) boundary conditions.

It's an **approximation to the real world**, but it turns out to be a **good** one.

So now we attempt to talk about and quantify "the climate" and "climate change"?

Part #2

Introduction

to

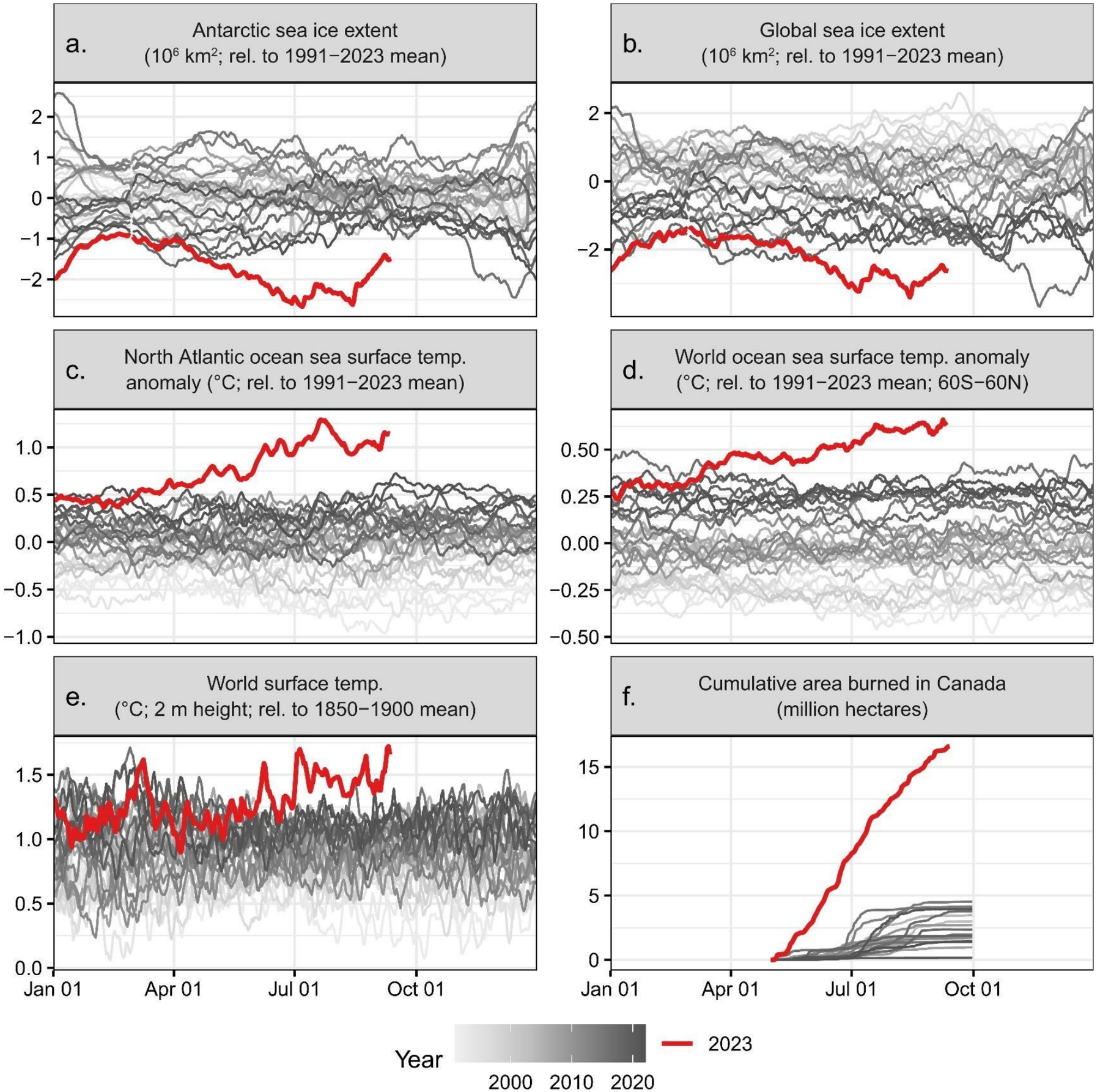
THE PHYSICS OF CLIMATE CHANGE

First Pass

with

more details to follow later in Parts #3, #4 and #5

Before we start — If you still do not think we have a problem?



YES WE DO!!!!

Now to proceed.

First let me tell you a story that I learned from a friend…..

1 - A RIVER LIKE NO OTHER

Images of the Mekong River

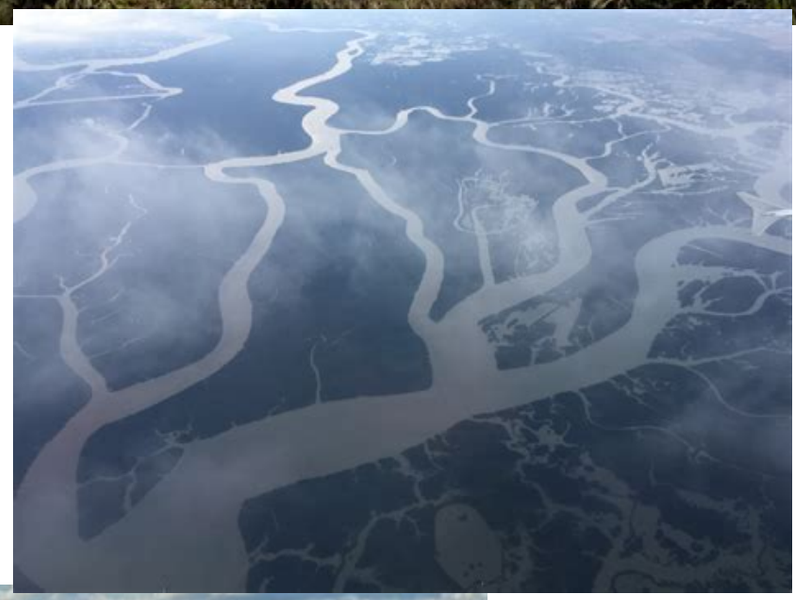
















To travel down the Mekong River near its delta is to experience a waterway like no other in the world.

Unnavigable for much of its 2,800-mile length, it is the longest river in Southeast Asia and the twelfth-longest river in the world.

By the time the river flows past Phnom Penh in Cambodia and onward into Vietnam, the low-lying and flat terrain causes it to spread out into many separate branches.

The Mekong basin covers an area the size of France and Germany combined.

On average, the river is almost a mile wide and is far wider at many points.

Khone Falls, on the Laos-Cambodian border, is the widest waterfall in the world.

Its series of rapids and falls are almost seven miles wide, with a drop of seventy feet!

All told, the river disgorges over 475 billion cubic meters of water each year into the sea, and it provides food and water for sixty million people.

In 1866, the French Mekong River Commission of twenty men embarked on a two-year journey up the Mekong, traveling a distance longer than the entire length of Africa, to map the full system.

They started in Saigon and ultimately made it all the way to the Yangtze in China.

Thirteen men survived the journey.

Because the lower Mekong Delta south of Phnom Penh lies just above sea level and is exceedingly flat, the shallowness of the river produces a striking annual variation.

At Phnom Penh, the river is joined by the Tonlé Sap river and lake system.

Depending on the season and the river's varying height, the direction of the Tonlé Sap actually changes.

At times, it becomes a tributary, flowing into the Mekong.

During flood season, the flow reverses, and the floodwaters flow along the Tonlé Sap into its large lake.

Beyond its annual ebbs and flows, the Mekong Delta experiences a daily surge that, while not unique in the world, is nevertheless rare enough to have aroused the fascination of the earliest Western visitors—and surprise all the rest of us who first learn about it.

For much of the year, the delta experiences only one high tide from the surrounding China Sea each day.

Sir Isaac Newton explained the fundamental physics behind the tides when he developed his law of universal gravitation in the seventeenth century.

As Newton described it, the gravitational force of the moon on the Earth varies as the inverse square of the distance between the moon and the Earth.

Therefore, the side of the Earth nearest the moon, being slightly closer to the moon than the center of the Earth, is pulled with a stronger gravitational force than the average force on the Earth.

Similarly, the side of the Earth opposite the position of the moon is pulled toward the moon with a smaller force.

Ignoring for a moment the motion of the moon around the Earth, as the Earth rotates, one would expect two bulges to occur in the world's oceans: one on the side facing the moon and another on the opposite side.

Roughly speaking, the first occurs because the water is pulled away from the Earth, and the second because the Earth is pulled away from the water.

Looking down on the Earth-moon system, one would expect schematically to see something like Figure 1.1 on the next slide.

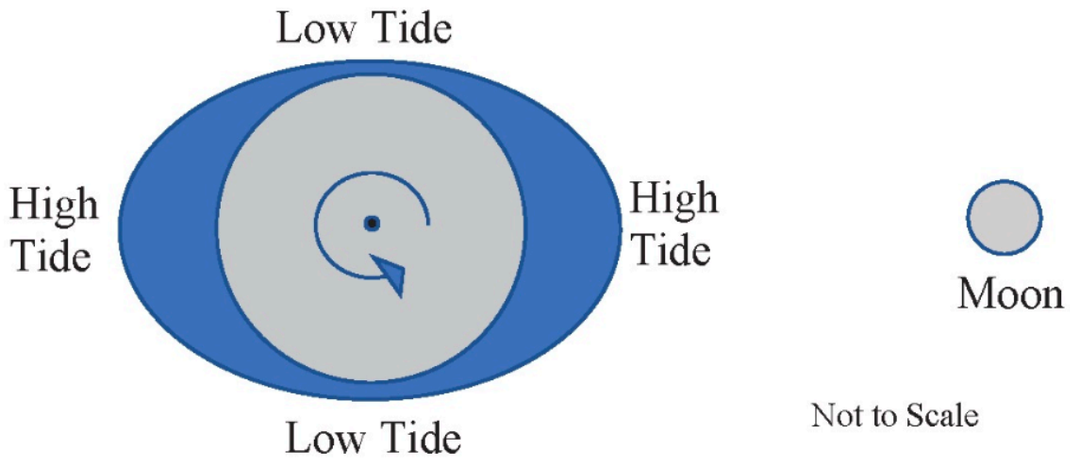


Figure 1.1

As the Earth rotates fully once each day, each spot on the Earth would therefore be expected to experience two high tides and two low tides, as Newton famously described.

In practice, understanding the tides is more complicated.

Water on Earth cannot instantaneously relocate to the equilibrium configuration shown in the figure, but must move from place to place, and flow rates depend on local conditions, including ocean depth.

The rotation of the Earth and the motion of the moon must also be considered, as must the position of the sun.

Happily, however, we don't need to take all that into account to understand how two tides a day can turn into one tide a day.

The key is to recognize that the moon does not orbit around the Earth's equator as the Earth rotates each day.

The plane of its orbit is tilted relative to the Earth's axis, varying between eighteen and twenty-eight degrees relative to the equatorial axis of the Earth over an eighteen-year period.

Consider the Earth-moon system in a frame where the Earth's axis of rotation is vertical (Figure 1.2).



Figure 1.2

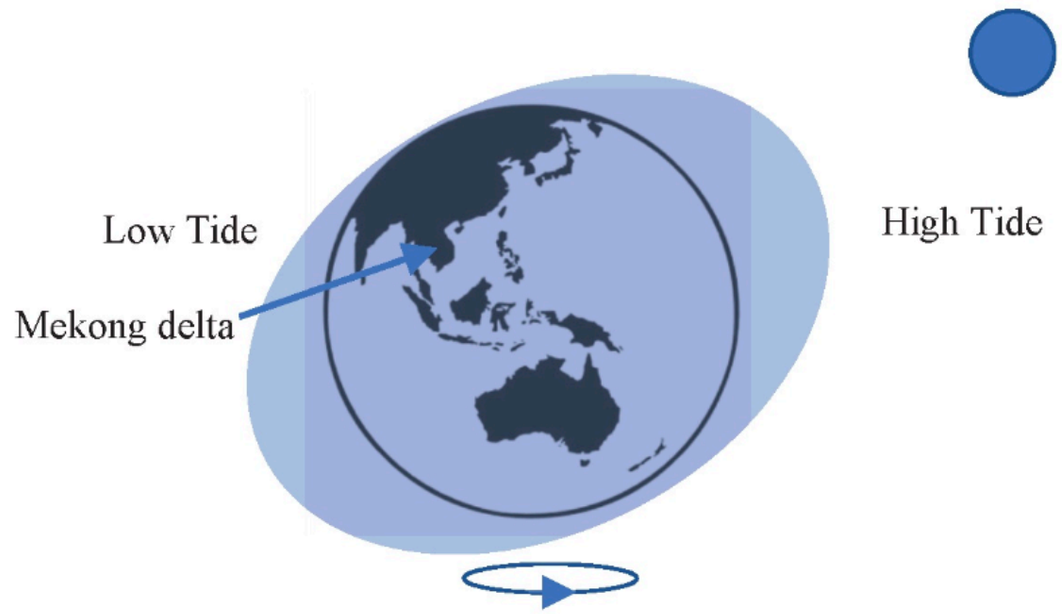


Figure 1.3

The tidal response of the ocean, using the reasoning from the figure shown earlier, is seen in Figure 1.3.

The actual magnitude of tides and the relative size of the tides depend on local conditions.

But roughly speaking, given the latitude of Vietnam and the position of the moon, as the Mekong Delta rotates around with the Earth, the seas in the region will tend to experience a high tide when southern Vietnam is on the side of the Earth facing the moon, where the bulge is big, and a low tide twelve hours later when it is on the far side, where the water has pulled away and there is no bulge.

I have belabored this issue not just because it involves a bit of physics and is a frequent source of confusion, but because it plays an important role in the future of the Mekong ecosystem that will be relevant later.

A large single daily tide impacts on the flow of the Mekong River itself.

The Mekong has the richest density of freshwater fish in the world and is home to what is estimated to be over one thousand species of fish.

Directly supporting a population of over fourteen million people, a greater population of freshwater fish is harvested from the Mekong each year than from all the lakes and rivers in the US combined.

Over the course of the year, its floods bring water and silt to nourish rice paddies, making the Mekong delta the world's most productive rice bowl.

While some of the most dramatic potential global impacts of climate change might not become manifest for many decades, centuries, or even millennia, the Mekong may be one of the first casualties in the battle to head off Earth 2.0.

The shallowness of the river, the flatness of the delta, and the flooding caused both by the seasonal weather and the delicate balance of tides and river dynamics make the Mekong Delta particularly sensitive to even small near-term changes in any of these systems.

It is not just the highly publicized dire predictions of climate change that can have a dramatic impact on many people's lives.

The Mekong Delta is a canary in a coal mine for climate change, and that is one reason I have begun this discussion by describing it and why I shall return to discuss the specific predictions and impacts for the Mekong region at the end of this 2nd part of our discussion.

While the particular circumstances of the Mekong are unique, various other locations around the world live with similar fragile balances of opposing ecological forces, from the lowlands of Bangladesh to the everglades of Florida and the mouth of the mighty Mississippi.

Climate change as a global issue may manifest itself in a thousand different ways in a thousand different places.

But just as no man is an island, in an interconnected world, no single place and no single country is likely to be completely immune from the impact of even seemingly small changes first detected on the other side of the planet.

2 - HISTORY AND NUMBERS: HALF EMPTY OR HALF FULL?

If you observe something you have never seen before, how do you determine if it is good or bad?

Or if it is anomalously big or small?

Dangerous or benign?

Perspective is everything, of course, but how can we gain it?

The problem is numbers alone can be deceiving.

The same data viewed in different contexts can appear to present a vastly different picture.

Is the glass half-full or half-empty?

This was the problem facing those who were first presented with estimates of carbon dioxide (CO_2) concentrations in the Earth's atmosphere.

In 1953, Charles David Keeling had just begun a postdoctoral position at Caltech.

Originally he was working on a project to extract uranium from granite, but then moved to another geochemical problem: investigating how carbonate achieves equilibrium in a mixture of water, limestone, and atmospheric CO_2.

To do this, he needed to construct a precision device to measure CO_2 extracted from the air and water.

To test his apparatus, he measured the CO_2 concentration in various locations in Pasadena but found significant variations, which he figured were probably due to locations of heavy industry.

So he took the equipment up to a more isolated spot: Big Sur, near Monterey, California.

During every afternoon he measured the same value of CO_2 concentration in the atmosphere, 310 parts per million (ppm).

He started taking samples during both night and day to get better estimates and discovered a diurnal pattern he hadn't anticipated.

At night, there was more CO_2 in the air than there was during the day.

Also, with great prescience, he measured the ratio of ^{13}C to ^{12}C and discovered this ratio was smaller at night than during the day.

Checking a meteorology book, Keeling discovered that the concentrations he measured were representative of uniform mixing with the "free-atmospheric" concentrations that prevailed over the continent.

However, at night there was a lower boundary layer so the air measured was more heavily influenced by concentrations near the ground, where local plant and soil respiration would be systematically more important.

This interpretation was confirmed by a decrease in the $^{13}C/^{12}C$ ratios at night, as plants preferentially respire ^{12}C.

In 1956 Roger Revelle at Scripps Institution of Oceanography and Harry Wexler at the US Weather Bureau joined with Keeling to suggest a bolder global measurement of CO_2 during the upcoming International Geophysical Year (1957–58) at a variety of remote locations presumably unaffected by local contamination, including the South Pole station and at Mauna Loa in Hawaii.

In March 1958 Keeling installed his first infrared gas analyzer at Mauna Loa, and on its first day of use, it recorded a concentration of 313 ppm.

This was the first reading in what has become one of the most significant continuous terrestrial scientific projects ever carried out.

It has been ongoing for the last sixty-six years and has given the world its first quantitative assessment of the impact of global industrial activity on the composition of the atmosphere.

Over the first few months of its installation, Keeling noticed a surprising monthly increase in CO_2 concentrations up until May.

After this it declined again until October, and this pattern repeated in 1959.

It was as if the Earth was breathing in and out once each year.

It actually is, due to the existence of life on our planet.

Life has changed the composition of the Earth's atmosphere since its inception some four billion years ago, and it continues to govern the dynamics of CO_2 on the planet today.

What Keeling was detecting was a modern annual cyclic version of the ancient processes of life that first produced the existing atmosphere on the planet today.

He was directly observing, for the first time, the seasonal impact life has on the atmosphere in the Northern Hemisphere through the process of photosynthesis in plants, which converts CO_2 and water into organic compounds with O_2 as a residue.

As Keeling later put it in a 1960 article, “We were witnessing for the first time nature’s withdrawing CO_2 from the air for plant growth during summer and returning it each succeeding winter.”

The second significant observation Keeling made between 1958 and 1960 at Mauna Loa was that the average concentration of CO_2 measurably increased during this period.

If one considers comparable months each year, March CO_2 levels went from 313.4 ppm in 1958 to 314.4 ppm in 1960.

A somewhat more dramatic effect was observed from samples taken from surface flasks collected at the South Pole, which increased from 311.1 ppm in September 1957 to 314 ppm in September 1959.

Were such small increases significant?

Noting that the increase observed at the South Pole was consistent with what one would expect from the terrestrial combustion of fossil fuels,

Keeling was nevertheless well aware that claiming a trend based on such a small time sequence was dangerous.

Here is how he put it in his first paper for the Scripps Institution of Oceanography in March of 1960:

> *Where data extend beyond one year, averages for the second year are higher than for the first year.*
>
> *At the South Pole, where the longest record exists, the concentration has increased at the rate of about 1.3 ppm per year.*
>
> *Over the northern Pacific Ocean, the increase appears to be between 0.5 and 1.2 ppm per year.*
>
> *Since measurements are still in progress, more reliable estimates of annual increase should be available in the future.*
>
> *At the South Pole the observed rate of increase is nearly that to be expected from the combustion of fossil fuel (1.4 ppm), if no removal from the atmosphere takes place.*
>
> *From this agreement one might be led to conclude that the oceans have been without effect in reducing the annual increase in concentration resulting from the combustion of fossil fuel.*
>
> *Since the seasonal variation in concentration observed in the northern hemisphere is several times larger than the annual increase, it is reasonable to suppose, however, that a small change in the factors producing this seasonal variation may also have produced an annual change counteracting an oceanic effect.*

This skepticism is the hallmark of good science.

Was the observed increase significant?

If the expected annual output of CO_2 due to human industrial activity was smaller in magnitude than the seasonal variation itself, was it possible to claim a signal amidst the noise?

What was the role of the oceans in possibly moderating this signal?

Beyond this, did a CO_2 concentration of 310 ppm itself reflect anything important about the global dynamics of the planet as it is affected by life, including human life?

These were all important questions in 1960, some of which had been anticipated a few years earlier by the pioneering oceanographer Roger Revelle and his colleagues at Scripps.

This is probably one of the reasons Revelle enlisted Keeling in his own effort.

To begin to answer these questions, accurate measuring would have to be carried out from remote locations regularly for long periods.

That, of course, is what has happened.

Every day for the past sixty-six years at Mauna Loa, CO_2 has been measured by the technique first used by Keeling, involving infrared gas analyzers that measure the absorption of infrared radiation in atmospheric samples and compare it with absorption rates in samples with known calibrated CO_2 concentrations.

The result has been one of the most famous plots in science, appropriately named the Keeling Curve.

Scripps Institution updates the curve every day and presents the data for public consumption.

Figure 2.1 shows the curve up to 2020, when the CO_2 reading was 415.19 (! !) ppm.

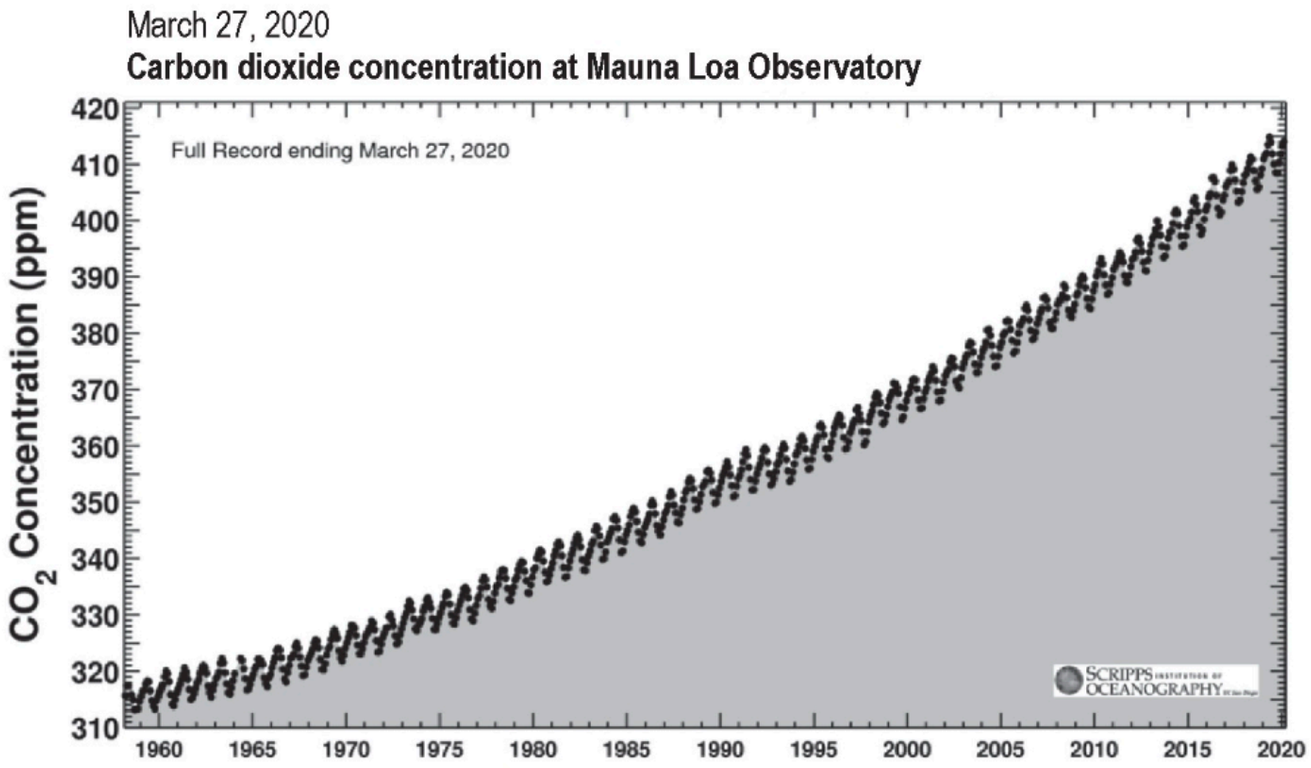


Figure 2.1

The seasonal variation first observed by Keeling is obvious, but now so is the monotonic rise in the average value of the CO_2 concentration year over year that was first tentatively inferred by Keeling in 1960.

Recall that, in 1958, the peak abundance was about 315 ppm.

The value today is therefore more than 30 percent larger than the measured atmospheric abundance at that time.

What are we to make of this?

As a physicist, the first thing I generally do is examine orders of magnitude, which usually give some perspective on any measurement.

The CO_2 abundance has increased by about 100 ppm in sixty-two years.

This is about 1.6 ppm per year.

Recall that Keeling himself estimated the CO_2 generation by fossil fuel consumption in 1960 to be about 1.4 ppm, about the same order of magnitude.

Note, however, that global annual CO_2 emissions have increased by a factor of five since 1960, while the slope of the Keeling curve has not increased by a similar factor.

Should this give us pause?

Perhaps, except Keeling himself pointed out a very plausible reason why one might expect that not all the CO_2 being produced by humanity would be reflected directly by a concomitant rise in atmospheric CO_2 content.

CO_2 can dissolve in water, creating carbonic acid, so one would expect some CO_2 to be taken up by the oceans.

So, the rough quantitative correlation between the observed rise and our human contribution is, at the very least, suggestive.

Perspective is everything, however, and we must always remember a key warning in science: **correlation does *not* imply causation**.

So, it is possible, without some underlying physical explanation and without more data, that the comparable rates of CO_2 in the atmosphere and human-generated CO_2 is just a coincidence.

To get a better idea of whether the observed CO_2 increase is correlated to human industrial activity, we can hope to explore longer-term variations to see if the current fossil fuel-generating era is anomalous or not.

But Charles Keeling wasn't making direct measurements before 1958, and before his time the few direct measurements that were made were scattershot and discordant.

Fortunately, however, nature has given us a time capsule.

In places that remain frozen all year long, ice builds up as snow falls.

It is for this reason that places like Greenland and Antarctica have ice sheets over a mile thick.

Like the growth of tree rings or sedimentary layers of rock, as one drills deeper into ice, one encounters ice layers that were deposited at ever-earlier times.

And like growth rings, ice deposition is different in summer than in winter, so a regular pattern allows one to literally count years.

Many cores are extracted from each area so better estimates can be obtained.

Figure 2.2 shows examples of core sites from a recent US International Trans-Antarctic Scientific Expedition study in Antarctica.

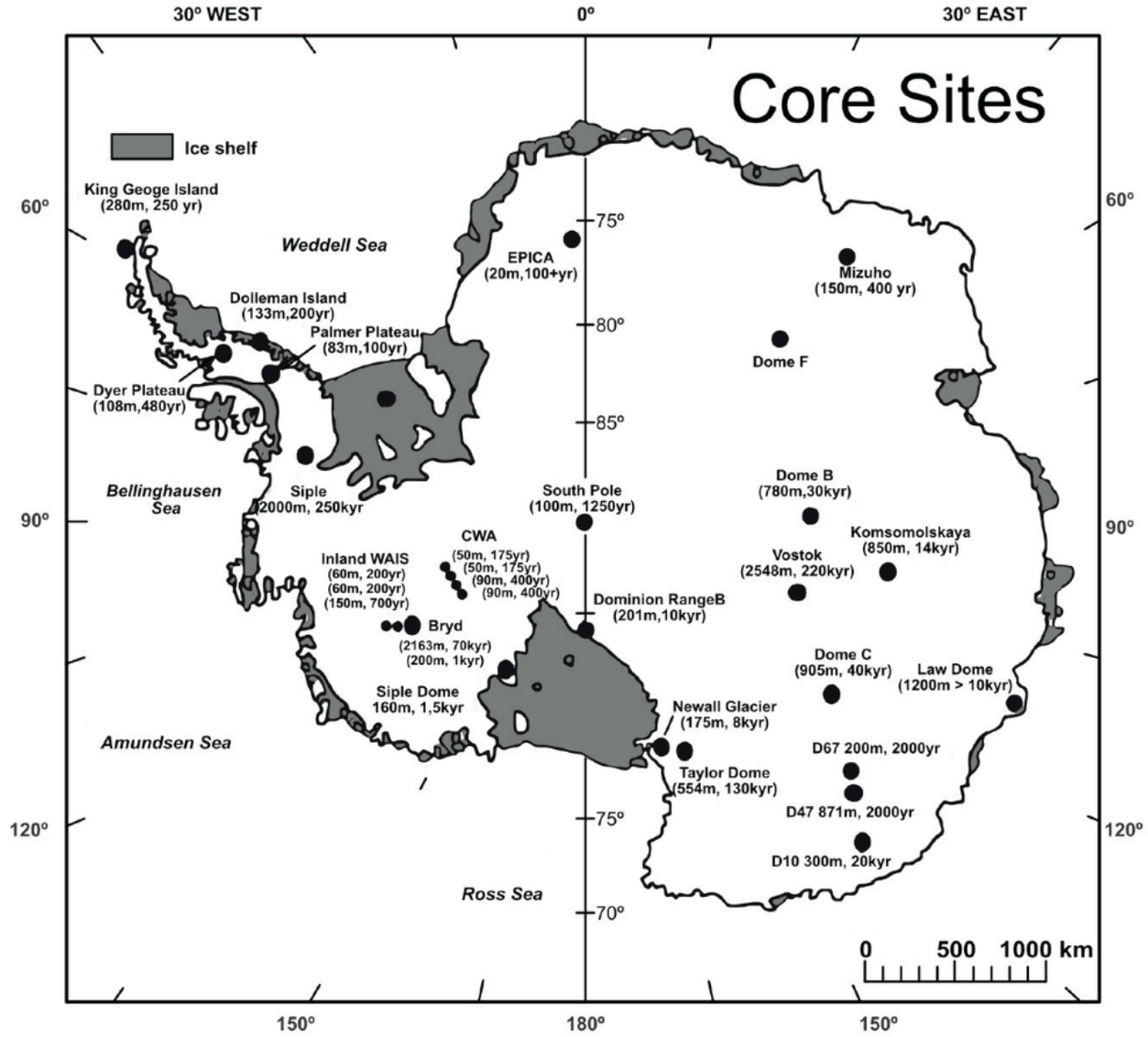


Figure 2.2

In the ice there are bubbles.

Air gets trapped in the snow as it is compressed, and the bubbles therefore reflect the atmosphere at the time and place when the ice first formed.

Measure the gas composition of the bubbles and you know the gas composition at that time and place.

Since Greenland and Antarctica have the deepest, largest, and historically the most stable ice conglomerations, most ice cores come from these two locations, and at their upper layers, can be compared with the direct measurements going back to 1958 to calibrate the ice core estimates.

Because each location gets a lot of snow each year, ice from successive years can be visually separated relatively easily so good time resolution is possible as well.

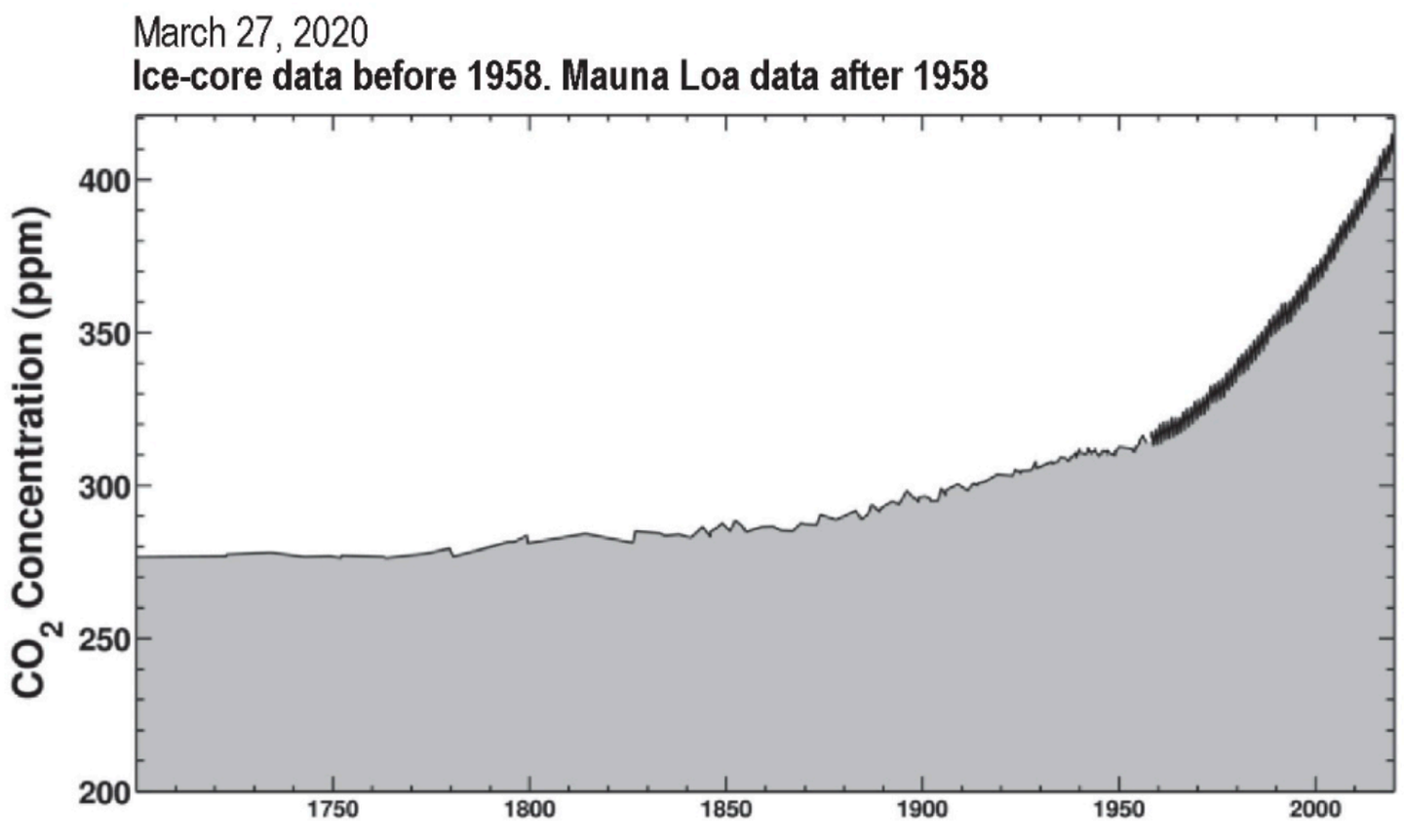
It is also fortunate that Greenland and Antarctica are in opposite hemispheres, so in some sense comparing values from ice cores in both locations gives us both a global consistency check as well as a global average for each period.

Let’s take a walk back in time using the ice core data.

Scripps Institution provides graphical examples of the data before 1958, matched to their own measurements from 1958 onward (where the black line thickens).

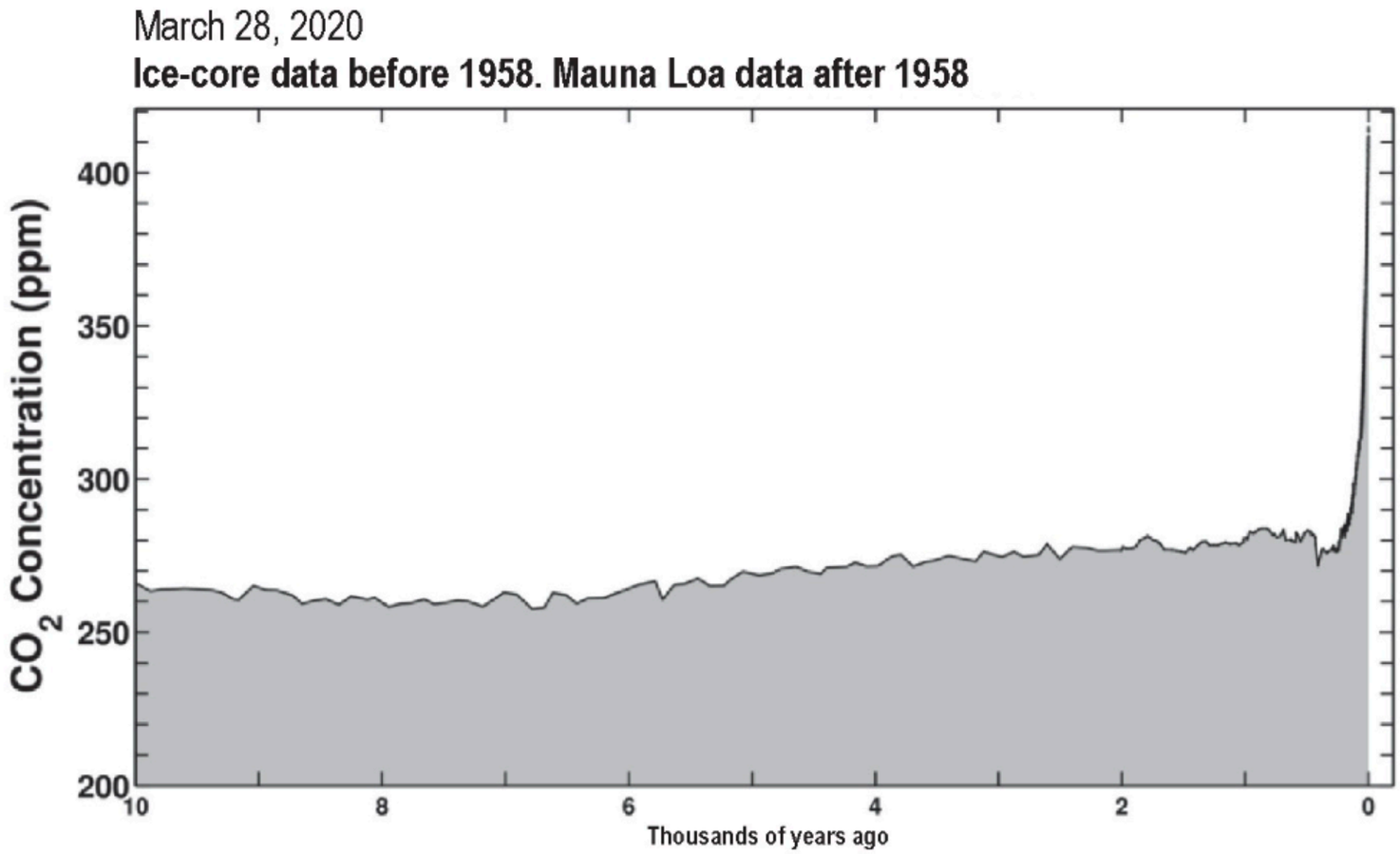
Figure 2.3 shows the data going back to 1700, before the advent of the modern industrial era.

Figure 2.3



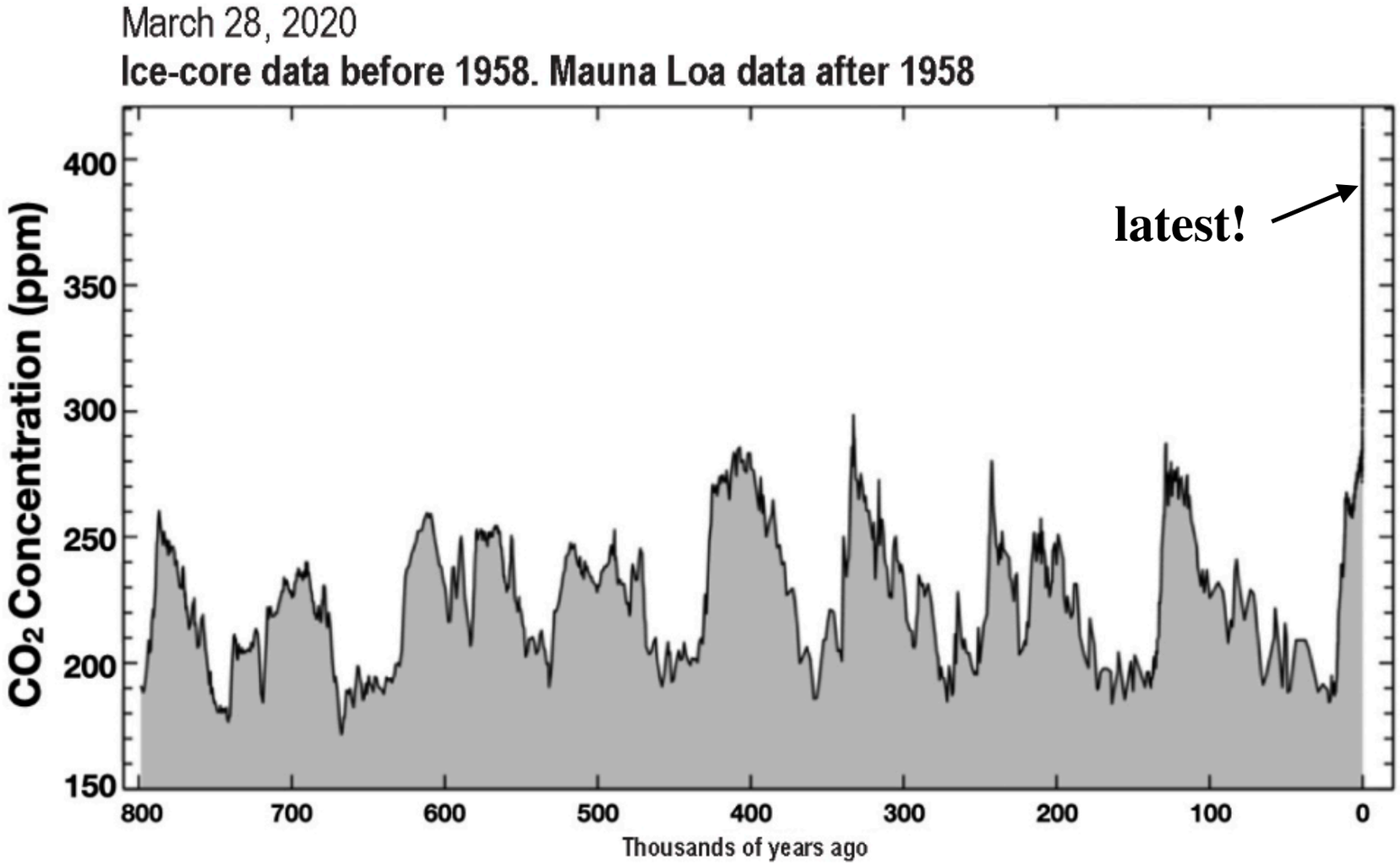
Alternatively, one can go back to the dawn of human recorded history, about ten thousand years (Figure 2.4).

Figure 2.4



The ice core record actually goes back about eight hundred thousand years. Here is the full record (Figure 2.5).

Figure 2.5

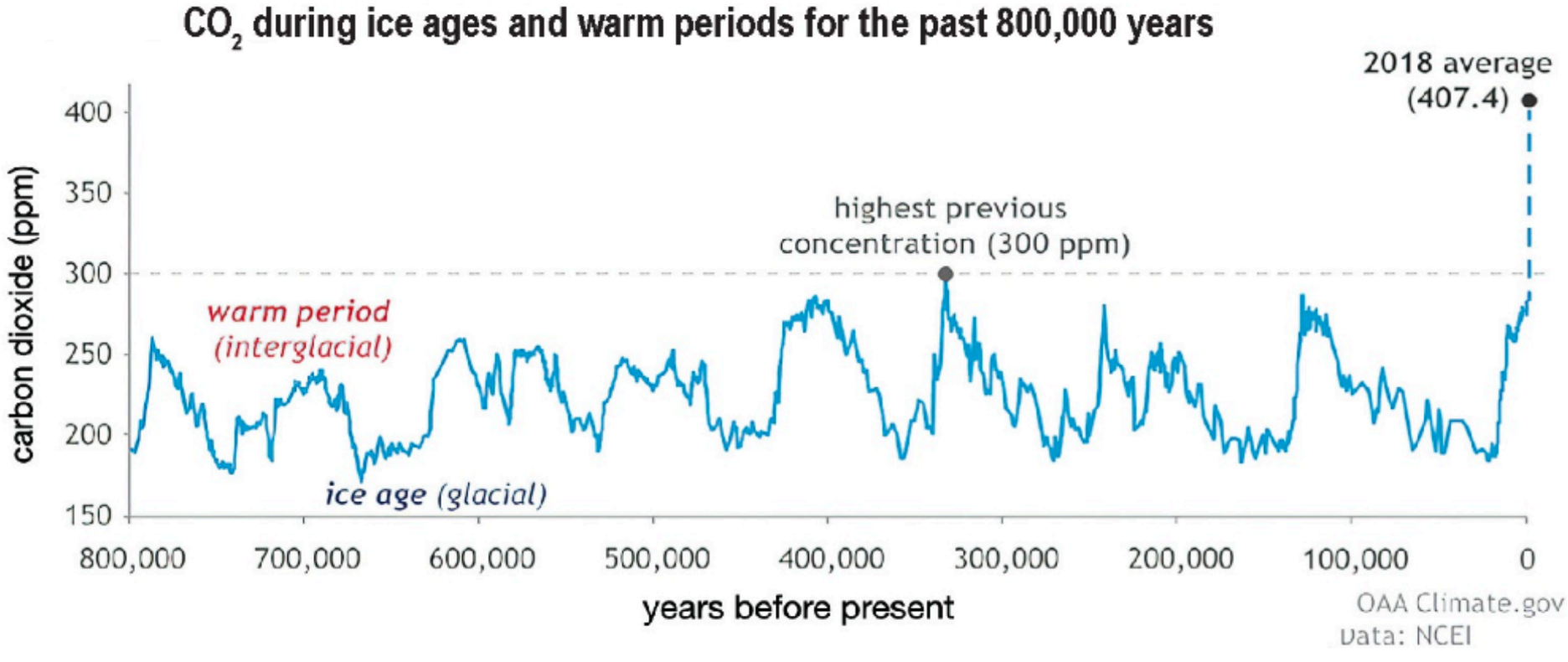


Because this last figure spans so much time, it is useful to point out some markers to guide you.

Of some relevance is the **correlation** between ice age and interglacial period, caused by periodic variations in the orbit of the Earth around the sun, with low and high CO_2 concentrations, respectively.

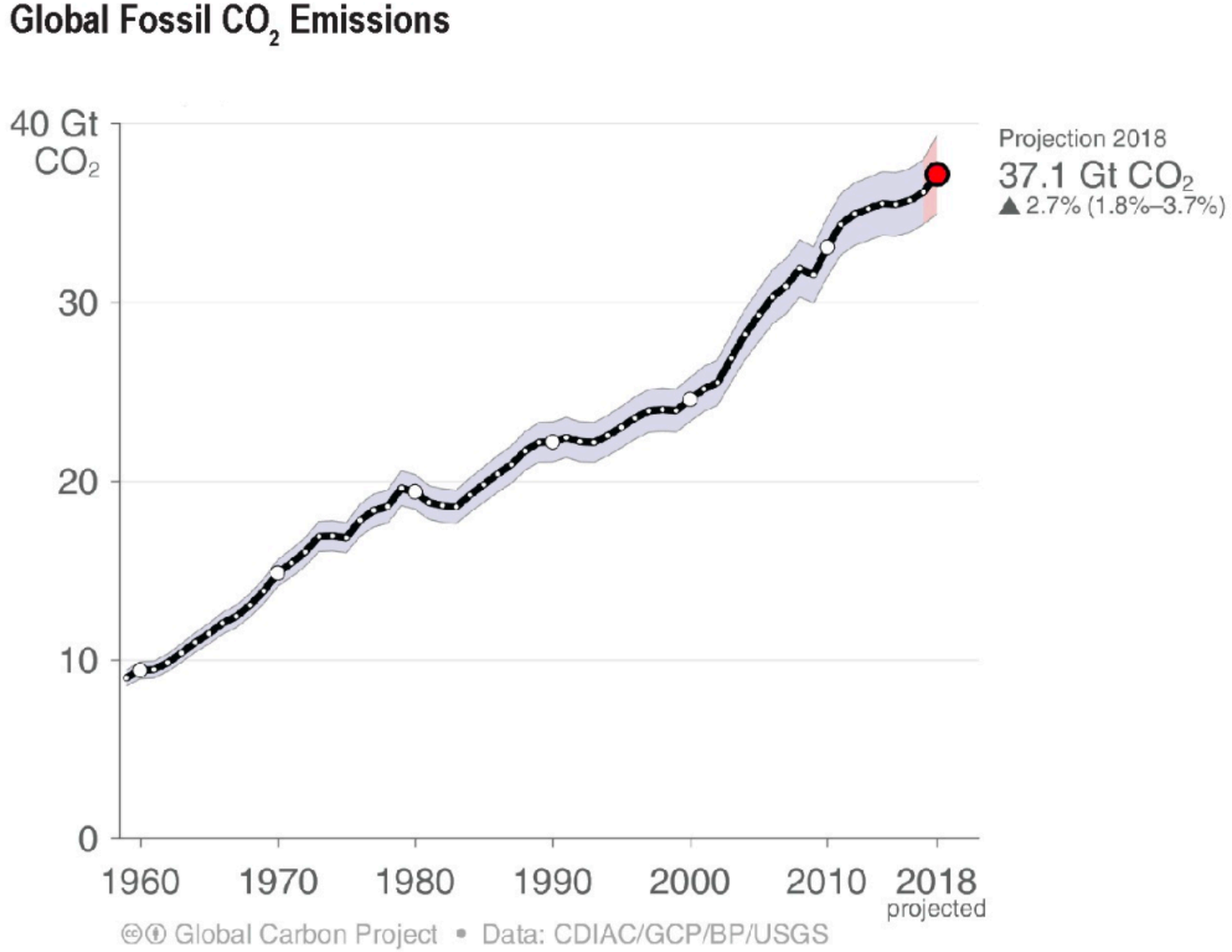
Also, an important overall baseline number for comparison is the highest concentration of CO_2 previous to the present era, which was at 300 ppm, about 350,000 years ago (Figure 2.6).

Figure 2.6



Returning to the original Keeling curve, we can compare it to the global CO_2 emission from human activity, as taken from a figure prepared by the Global Carbon Project (Figure 2.7).

Figure 2.7



The monotonic rise in CO_2 production is compatible with the Keeling data, and the dip in slope visible after 1990, when the first IPCC report was followed by the Kyoto Protocol, is also reflected in the Keeling curve.

Marginally visible is the increase in slope in the 1970s preceding 1974, followed by a dip at that time due to the first oil crisis.

What are the immediate takeaways from this data?

Some things stand out:

(a) the present era is unprecedented in the recorded history of CO_2 in the atmosphere of the planet over almost the past million years;

(b) geological variations in the CO_2 concentration have and over much longer timescales than the recently observed rise;

(c) those times with higher levels of CO_2 in the atmosphere appear correlated with warming periods, and those times with lower levels with ice ages;

(d) the rise began with the beginning of the modern industrial era, and the rate and overall magnitude of increase appears commensurate with global fossil fuel consumption by human industrial activity; and

(e) the economic and political vicissitudes of the human condition appear to be reflected at some level in the recent undulations in directly measured atmospheric CO_2 concentrations. occurred, but at much smaller levels

The connection between CO_2 concentration in the atmosphere and the growth of human industrial production and fossil fuel appears **unambiguous**, making the current era qualitatively and quantitatively new in recorded human history.

But is this quantitative change likely to be significant from the point of view of climate?

That will require a discussion of the basic dynamics of CO_2 on Earth, which we shall turn to next.

Before completing this historical tour, however, there is one more curve of CO_2 concentration that I find personally compelling.

I have shown plots over the recent era, over the industrial era, and over geological eras.

The following is a plot over the last two thousand years.

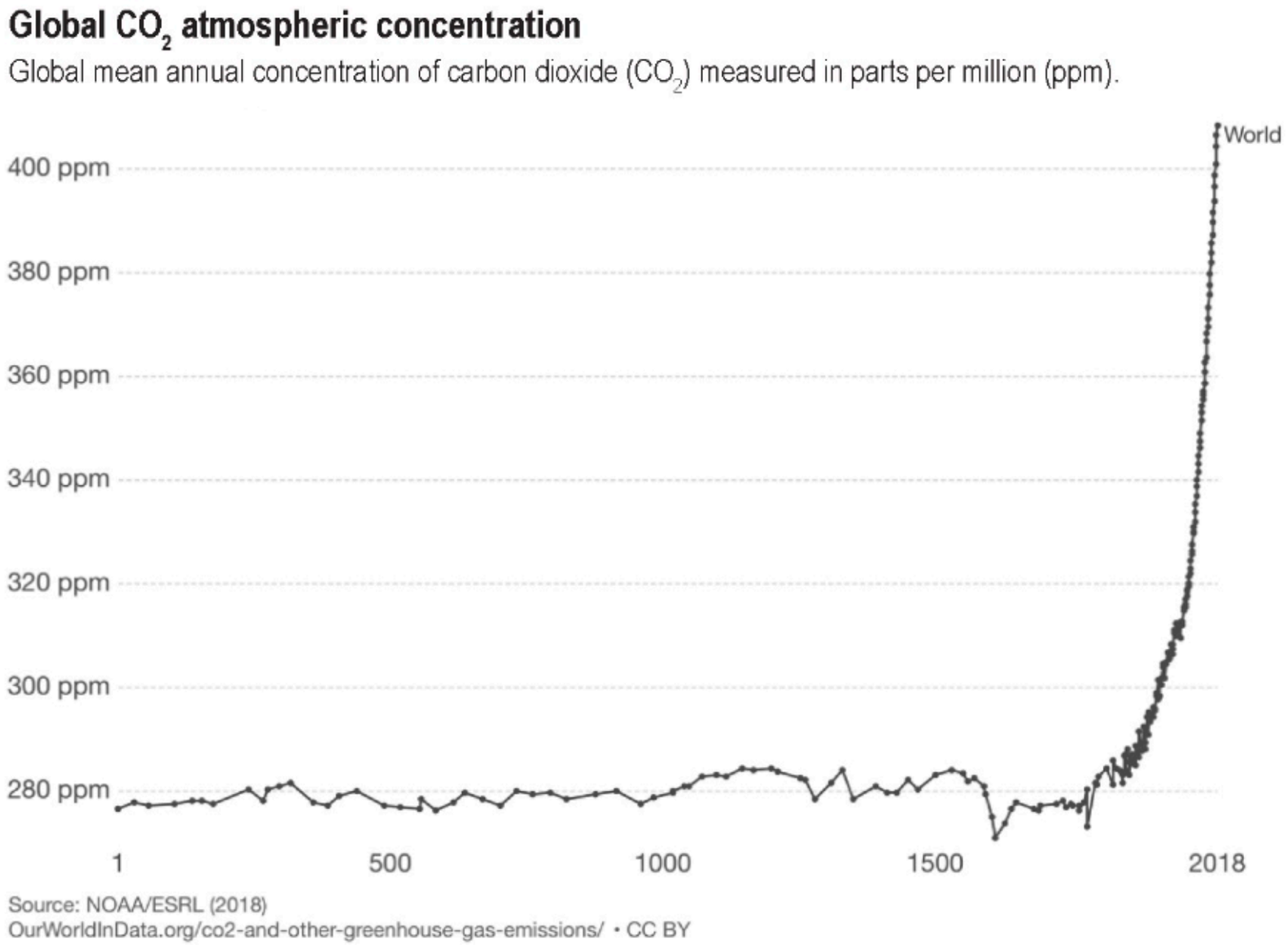
All of the modern dramas in human civilization took place over this time, from the rise of Christianity to the fall of Rome, the creation of Islam, the dynasties of China, the medieval era, imperial wars, and in the west, the Enlightenment, the Renaissance, and ultimately the modern technological world, including the atom bomb and two world wars.

During this time the human population has grown from perhaps two hundred million to almost eight billion, a fortyfold increase.

The energy use per human has increased by a far larger rate.

During almost all of these remarkable developments of modern civilization, humans had little global physical impact on the planet—until today.

Clearly, we are the culprits!!



3 - CYCLES AND CYCLES

The history of the Earth is a history of overlapping cycles, repeated with great regularity on ever-decreasing timescales.

We are relative latecomers to that history, and while we often imagine ourselves as the Masters of the Universe, we are nevertheless the slaves of chemistry.

Every atom that makes up you and me has been recycled throughout the universe and throughout Earth on its cosmic journey.

Yes, we are stardust, but we are also intimately connected to everything that has ever lived on Earth and to the very rocks we walk on, water we swim in, and air we breathe.

With the exception of hydrogen, every other element in Earth's atmosphere, and in our bodies, including carbon, oxygen, and nitrogen, was fabricated in the fiery cores of stars.

Their explosive deaths in supernova explosions seeded the galaxy for the formation of new stars and the solar systems that formed around them.

Hydrogen is the dominant element in the universe, but stars produce significant amounts of carbon, nitrogen, and oxygen, as well as silicon and iron, during their lifetimes.

This accounts for the fact that iron and silicon make up much of the interior of rocky planets like Earth, as well as the interior of asteroids and meteoroids that orbit throughout the solar system.

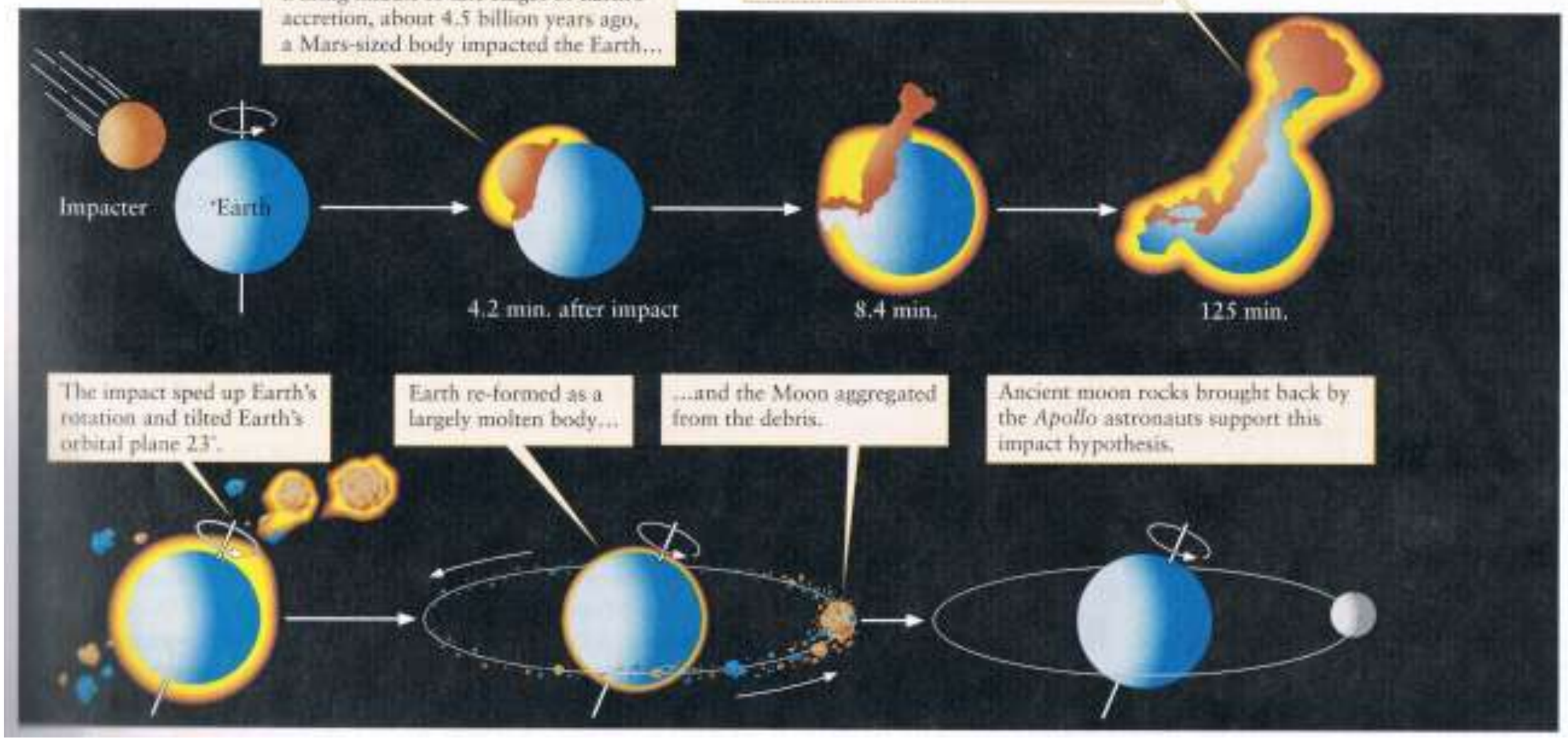
Equally important, planets large enough to have atmospheres and hold on to them for billions of years have lighter elements, like carbon, nitrogen, and oxygen, and compounds containing hydrogen, dominating their atmospheres.

From the dawn of terrestrial time, these elements have been recycled throughout the planet.

The hellish period of the Earth shortly after the gigantic impact that created the moon and liquified much of the planet is called the Hadean period, after Hades, and for good reason.

As the molten rock condensed after the impact within a few thousand years, a dense atmosphere was left behind containing mostly CO_2, hydrogen, and water vapor, with **no** free oxygen.

Estimates are this early atmosphere was almost thirty times denser than our current atmosphere, and the dominant gas was CO_2.



This means this initial concentration of CO_2 in the early atmosphere was well over ten thousand times greater than it is today.

CO_2 was reduced to its present abundance as a result of the first great carbon cycle on Earth, the geological cycle.

Before the emergence of life, it was the only carbon cycle.

The dense CO_2 layer began to decrease as the Earth cooled and liquid water oceans began to cover the Earth's surface.

This first bit of chemistry is simple.

Carbon dioxide dissolves in water and combines with it to form carbonic acid:

$$H_2O + CO_2 = H_2CO_3$$

The carbonic acid interacts in the ocean, or with the rock, forming substances called carbonates, where a CO_3^{2-} ion combines with things like calcium, magnesium, or iron.

These tend to be insoluble in water and sink to the bottom of the ocean floor (like the buildup of scale inside pipes with hard water), effectively removing carbon from the atmosphere-ocean system.

This process could not continue indefinitely if that were the end of the story, as eventually some equilibrium between the carbonates and carbonic acid and dissolved carbon dioxide would occur, including the formation of bicarbonates, which are soluble in water.

However, the continental crust of the Earth floats on a moving mantle, causing the continents to drift on large continental plates.

At the places where the plates meet, one plate is thrust upward by continental drift and the other plate downward.

This downward process is called subduction.

Subduction, beginning at deep ocean trenches, drags material, including the carbonate products of the carbonic acid interactions, down into the Earth.

This further drives the removal process of the initial atmospheric CO_2.

If this were the end of the story, the removal of CO_2 from the atmosphere could continue indefinitely.

However, as the material is dragged down into the mantle, it is heated, breaking apart the carbonate materials and releasing CO_2 into the mantle.

Eventually, after enough CO_2 has been moved from the atmosphere to the mantle, the pressure of CO_2 in the mantle builds up.

Excess CO_2 is then released in regions where the crust is thrusting upward due to continental drift.

This is where new land rises above the ocean, accompanied by violent volcanos that spew hot rock and gases back into the atmosphere.

Ultimately, over a long enough period, this would create some sort of equilibrium between CO_2 in the atmosphere and CO_2 stored in the Earth.

During this process, an average CO_2 molecule in the atmosphere would get recycled through the Earth about once every hundred million years or so.

Over geological time frames (in other words, tens or hundreds of millions of years) this cycle is self-regulating.

Roughly speaking, if the CO_2 concentration in the atmosphere increases, more CO_2 dissolves in water, producing more carbonic acid in water.

This increases the creation of carbonates that remove CO_2 from the atmosphere/ocean interface, reducing the atmospheric CO_2 ratio.

If the CO_2 concentration in the atmosphere falls, it is replenished by the CO_2 released by volcanos based on the previous buildup of carbonates in the crust and mantle.

While it gets me a bit ahead of myself, I can't resist relating here how this process dramatically affected Earth in a fascinating way during a period about six hundred to eight hundred million years ago.

Over the past twenty-five years, it has become widely accepted that Earth experienced one or more periods where, due to the merging of continents into a supercontinent and Earth's subsequent ability to reflect sunlight, Earth's surface actually froze over.

We think of Earth as a beautiful water world, and even today we classify the possibility of "habitable" planets around other stars by whether they have liquid oceans.

The idea that during at least one relatively recent period in our history, our habitable planet resembled not Mars but the frozen water moons Europa or Enceladus, is stunning

The “Snowball Earth” hypothesis, as it became known, is now well accepted in the community.

What would a possible Snowball Earth event do to the global geological carbon cycle?

With the ocean surfaces frozen, the atmosphere/ocean interface governing the flow of CO_2 from the atmosphere into the ocean would be interrupted.

During this period, volcanos would continue to emit deeply stored CO_2 back into the atmosphere and increase the atmospheric CO_2 concentration.

The way to halt this increasing concentration would be to melt the Earth’s surface so CO_2 could once again dissolve into the oceans.

Fortunately, the driving mechanism for this melting *was* the very buildup of CO_2 in the atmosphere itself.

As we shall discuss in some detail in what follows, CO_2 helps store heat that would otherwise be reflected back into space by Earth.

Thus, as the CO_2 buildup during Snowball Earth continued, more heat was stored and less radiated into space, eventually melting the oceans and allowing the geological carbon cycle to once again resume.

The emergence of life on Earth completely changed not only the atmospheric CO_2 concentration but almost everything else, transforming the planet.

While likely initially powered by inorganic reactions at the sea floor, eventually life discovered the sun as an energy source.

Not only is sunlight ubiquitous at Earth's surface, but more than ten thousand times the energy released from Earth's core to the surface impinges on the same surface each day from the sun.

Photosynthesis uses sunlight to elevate the energy levels of electrons in atoms so those electrons can later do work to power the chemical reactions of life.

In the process, as I have already briefly mentioned, it breaks apart water and uses CO_2 as a basic ingredient to build complex hydrocarbons.

The **waste** product of this process is oxygen in the form of O_2 molecules.

The earliest forms of photosynthetic life—cyanobacteria, tiny single-celled objects—provide a literal testament to the idea that even in small groups, working together can change the world.

Over time, day after day, year after year, century after century, millennium after millennium, over the first few billion years of the history of this planet, each small puff of O_2 released by every single cyanobacterium contributed to a remarkable conversion of CO_2 to O_2 in the atmosphere.

Colonies of cyanobacteria, producing structures called stromatolites—visible as extinct fossils or extant forms in places such as Australia and Mexico—photosynthesized CO_2 and H_2O into O_2.

Since O_2 was initially a poison, layers of cyanobacteria would die as these structures built up, and slowly these colonies helped remake the Earth's atmosphere to bring O_2 and CO_2 concentrations close to their present values about a billion years ago.

Why was O_2 a poison?

Because it could "oxidize" materials, from iron to organic substances, stealing the free electrons organic materials could use to store energy for later use.

It was vitally important that there was little or no free oxygen in the early atmosphere.

Only after O_2 levels in the atmosphere rose sufficiently did living systems respond to this new abundance by evolving a mechanism to exploit oxygen rather than succumb to it.

Respiration is basically "controlled burning."

Living systems release stored oxygen electrons to help produce a substance called ATP—akin to a "battery" universally used by living systems to power their physical processes—during the last phases of the electron-mediated power cycle.

As a result, living systems are able to extract over thirty times the energy they would otherwise have been able to generate by photosynthesis.

During respiration, incoming O_2 allows organic molecules to be fuels for life, just as during burning it supports the combustion of wood.

In both cases oxygen combines with carbon from organic materials to produce CO_2.

Photosynthesis utilizes CO_2, water, and sunlight as fuel sources, while respiration exploits O_2 and returns CO_2 to the atmosphere.

These two energy-generation processes of life have recycled CO_2 on Earth in ways that ultimately have governed the dynamics of CO_2, even as its geological cycling through Earth has continued.

As oxygen began its slow but inexorable rise from its near-zero concentration in the atmosphere of the initial, lifeless planet to its current value of about 20 percent of the atmosphere, the CO_2 concentration inexorably fell to its current value of about 0.04 percent of the atmosphere.

The carbon that was extracted by photosynthesis was processed into organic materials that built up over the billions of years life has been evolving on the planet.

The carbon is now stored in biota on the surface, in the soils of the world, and deep underground.

When first confronted by the claim that human civilization has recently altered the global dynamics of our atmosphere, it seems slightly outrageous.

The Earth is vast, and the atmosphere extends tens of miles upward to the borders of space.

Every day the sun rises and sets, tides go in, tides go out.

We experience the annual cycles as winter turns to spring, spring to summer, summer to fall, and fall again to winter.

We experience beautiful sunsets and clear evenings full of stars if we are lucky to be in the right place at the right time.

All this reinforces the seeming insignificance of humanity on a global scale.

To be sure, the calm regular rhythm of life is periodically interrupted by natural disasters, but even these reinforce the perception that Mother Nature is far more powerful than we are.

With hurricanes, typhoons, snowstorms, floods, droughts, and deluges, the natural world seems too powerful to be dramatically affected by anything everyday human activity can generate.

Moreover, if Earth took billions of years to reach its current state, how can we change it in the course of decades?

The key to understanding our human impact here is to realize that we are able to access in a short time, through our technology, quantities of carbon that took literally hundreds of millions of years to build up on Earth before the rise of humans.

First, rather than thinking about the percentage of CO_2 or O_2 in the atmosphere, it is perhaps more useful to think in terms of absolute amounts of each substance.

These are traditionally expressed in terms of gigatons (Gt=billions of tons) of material, either as gigatons of CO_2 or gigatons of just carbon.

Since the atomic weight of carbon is about 12 atomic mass units (AMU), and oxygen is about 16 AMU, the ratio of the mass of one CO_2 molecule to one carbon atom is about (12 + 2x16=44)/12, or about 3.67.

Because different studies quote one or the other, I will sometimes quote carbon and sometimes CO_2 abundances in what follows, and you can use the 3.67 figure to transform from one to the other.

The atmosphere contained about 600 Gt of carbon, or about 2,200 Gt of CO_2.

As illustrated here, the carbon in CO_2 gets cycled through Earth's oceans and mantle, and these locations have built up their own carbon reservoirs over time.

At this equilibrium level, 600 Gt (the same amount as in the atmosphere, it turns out) resides in the terrestrial biosphere, while about three times that amount, about 1,800 Gt, resides in terrestrial soils.

Figure 3.1 is an illustration by Donald DePaolo at Lawrence Berkeley Lab that displays the global carbon cycle during recent history but before the current industrial revolution.

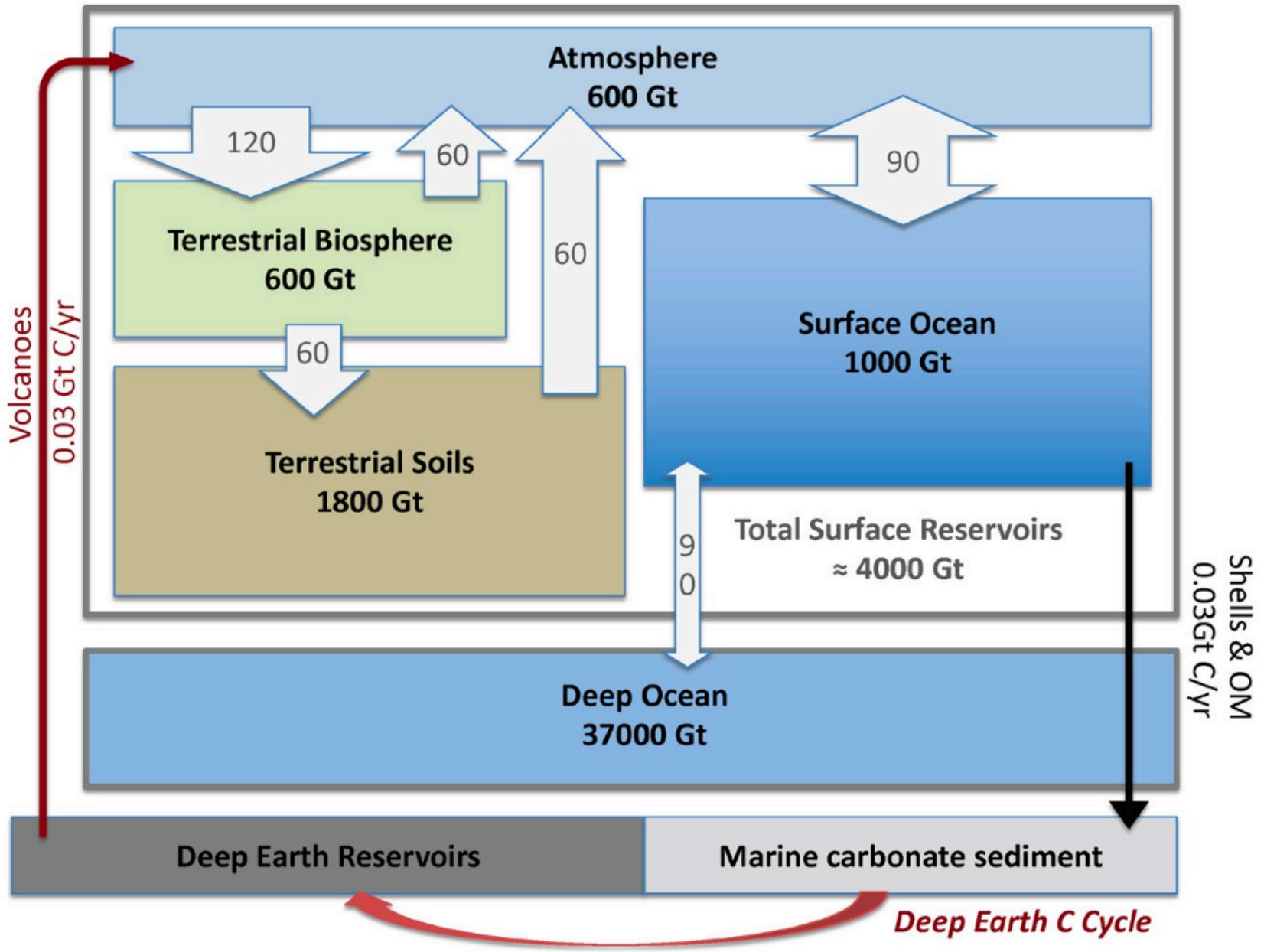


Figure 3.1

The surface levels of the ocean that are in contact with the atmosphere contain about 1,000 Gt, while the deep ocean, not in contact with the atmosphere, stores a much larger reservoir of 37,000 Gt.

During this equilibrium, CO_2 gets recycled in various ways, moving between the atmosphere, the biosphere, soils, and the ocean.

As shown in the figure, about 120 Gt/yr. of carbon is taken up by the biosphere, and about 60 Gt/yr. of that is converted into the soils as living organisms die, for example.

Similarly, through respiration, decay, fires, and the like, about 60 Gt/yr. is released back into the atmosphere from the biosphere, and an equal amount from soil.

About 90 Gt/yr. of carbon is cycled back and forth between the atmosphere and the surface layers of the ocean, and between the surface layers and deep ocean.

Altogether, this cycling keeps the carbon abundance in these reservoirs and the atmosphere roughly constant.

On top of this largely biologically inspired movement of carbon, which happens on short timescales, there is the long-term geological carbon cycle.

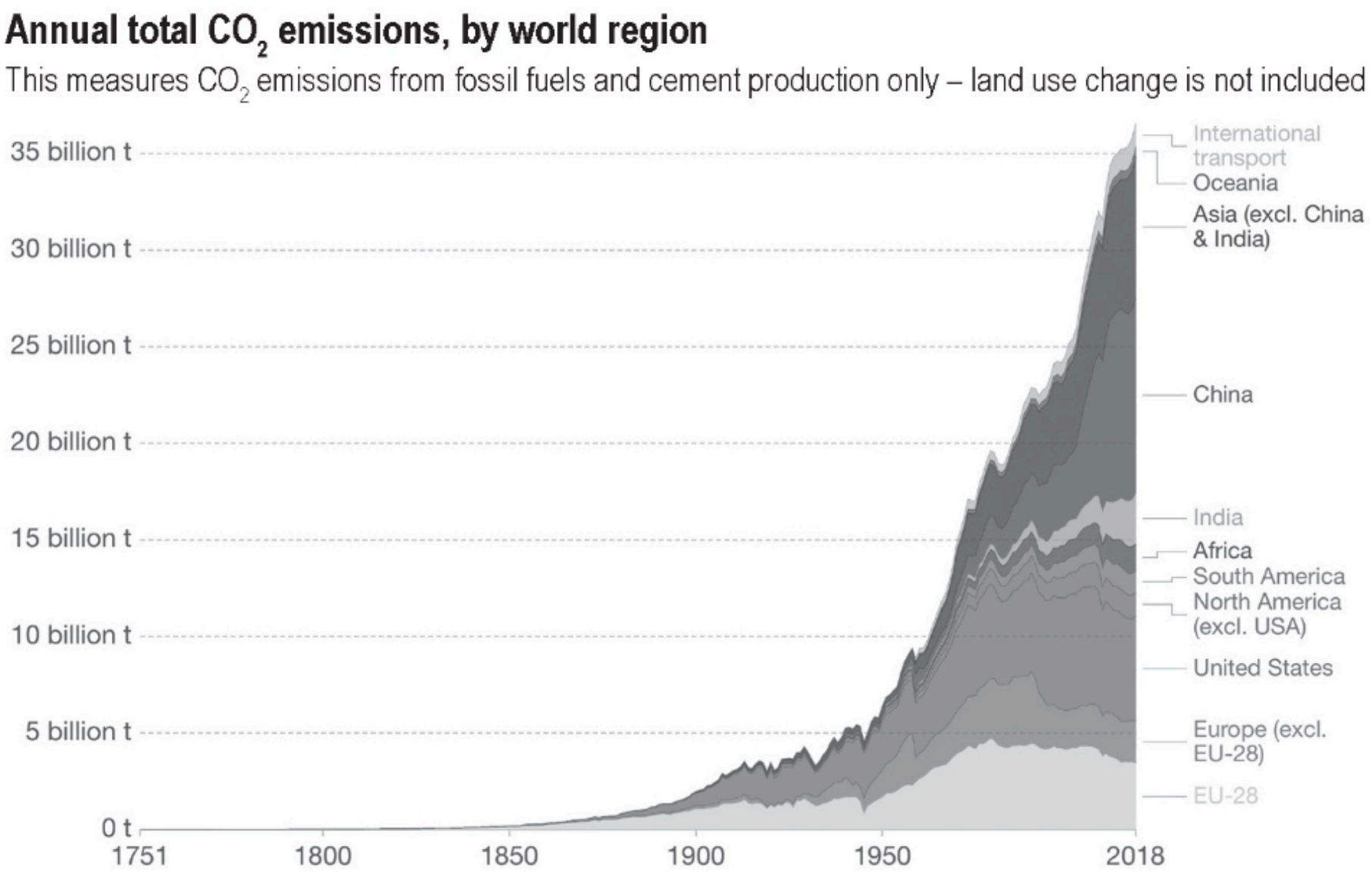
As I have described, this is eventually subducted deep into the Earth to be returned to the atmosphere through volcanos.

These spew out an equal amount of carbon as CO_2 (less than 1 percent of what humanity now does) back into the atmosphere.

This was the equilibrium that has existed on Earth over much of the recent history of the planet, since atmospheric abundances achieved their current concentration and life emerged in more or less its present form hundreds of millions of years ago.

What has happened **since**, as humans evolved modern technology, is shown in Figure 3.2.

Figure 3.2



At the current time, humanity is emitting over thirty-six billion tons of CO_2 per year into the atmosphere, or in the units used in the previous figure, about ten billion tons of carbon.

This graph also breaks down the locations where most of the emissions have taken place.

While China has the world's biggest population and is also emitting the most CO_2 at the present time, the USA has been the greatest single emitter when its cumulative contributions over time are considered.

I will discuss the impacts of the distribution of CO_2 emissions later but now want to concentrate on the totals.

Up to the present time, human industrial activity over the past sixty years has resulted in the emission of about 400 Gt of carbon into the atmosphere.

That is about two-thirds as much CO_2 as previously existed in the atmosphere over the past million years or so.

Not all of this has remained in the atmosphere, however, as the Keeling curve demonstrates.

The CO_2 abundance in the atmosphere has increased by only about 30 percent since 1960, meaning about 50 percent of the emitted CO_2 has been taken up by a combination of the surface ocean layers and terrestrial soils.

Humanity has efficiently tapped into a much larger reservoir of carbon than has been available at the Earth/ocean-atmosphere interface up to now.

By digging into the deep Earth, we have efficiently extracted amounts of carbon that would otherwise have required eons for any previous carbon cycle to generate.

To get a sense of this, we can add the human contribution to the global CO_2 balance in an updated block figure (Figure 3.3), along with the net amount of extra CO_2 that gets taken up by ocean and soil.

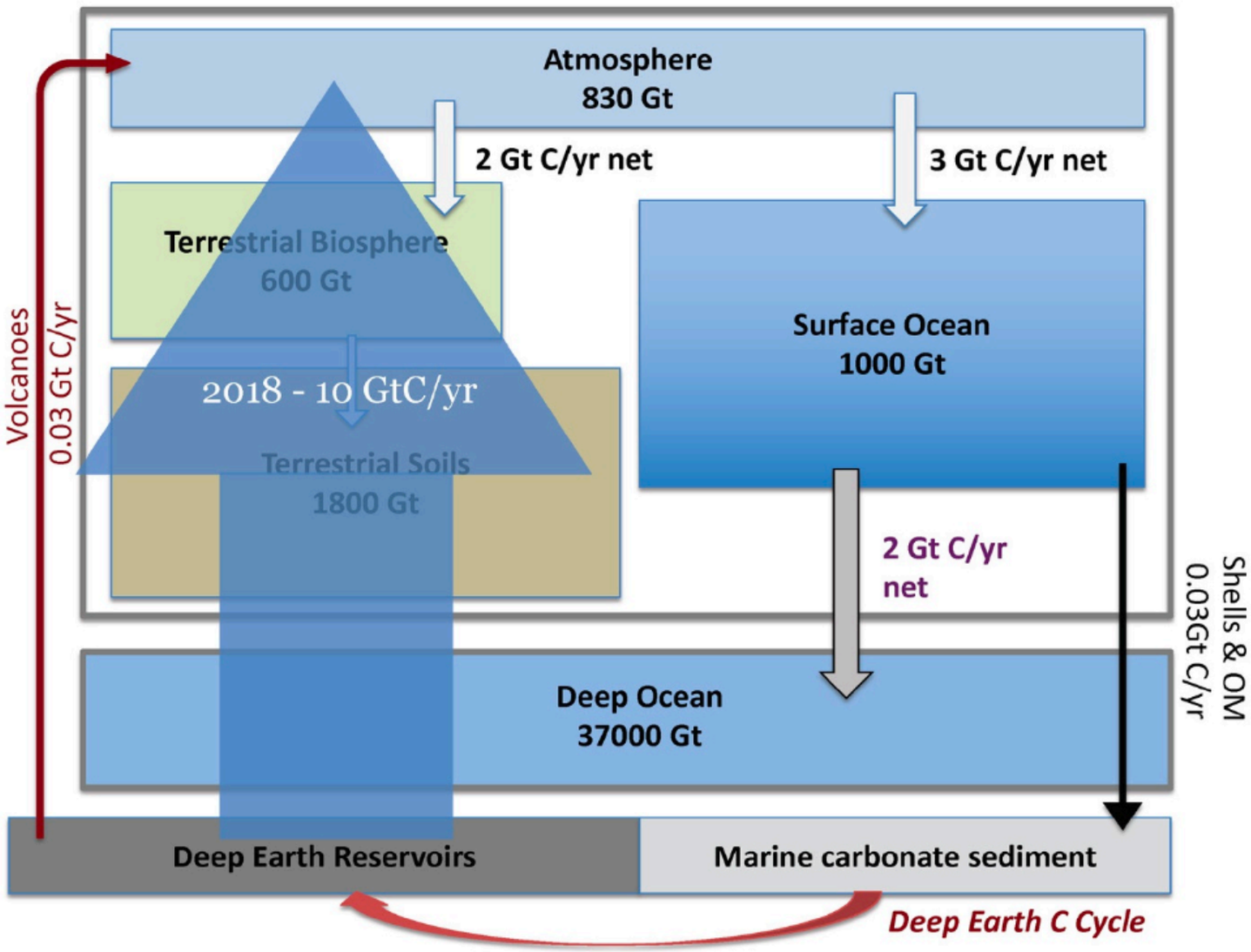


Figure 3.3

Unless we drastically alter the way we generate energy, which today is primarily by burning fossil fuels, during the next 150 years, humanity will emit an additional 1,000 to 4,000 Gt of carbon into the atmosphere, or up to seven times the total amount of CO_2 that existed in the atmosphere before the advent of modern civilization.

If this seems staggering, consider an estimate I shall later describe that was made over one hundred years ago: If all of the carbon in the atmosphere in 1900 were compressed into solid form, it would comprise a global layer less than 1 mm thick on the Earth's surface.

It does not seem so daunting, therefore, to imagine that humanity's carbon contribution to the atmosphere from industrial activity over the past two hundred years might exceed this amount.

As non-intuitive as it may seem, when it comes to the global terrestrial carbon budget at the planet's surface, the human footprint is anything but negligible

4 - EARTH'S BLANKET

Why isn't the Earth as cold as the space around it?

While not something you may wonder about every day, a version of this question was what the famous mathematician and physicist Joseph Fourier, whose work on what is now called Fourier analysis forms the mathematical basis of much of modern theoretical and experimental physics work today, asked himself in 1824.

More specifically, Fourier reasoned that Earth would radiate much of its heat out into space if something were not trapping that heat—something like a blanket, which keeps us warm at night by stopping us from radiating our own heat out into the room when we are sleeping.

Fourier had been studying the mathematics of heat flow for some time and, in the course of that study, discovered the mathematics that now bears his name.

It was in the context of his studies of heat flow that he began to study Earth's temperature.

He explored diurnal temperature variations as well as seasonal ones and determined that the planet was warmer than one would expect if the only source of warming came from the sun's radiation, at least if much of that radiation could then be reradiated by Earth out into space.

The Earth needed an insulating blanket, and he reasoned that Earth's blanket of air, which shields us from the harshness of space in many ways, acts like an insulator.

He considered the example of a box with a glass cover.

The air inside the box heats up when exposed to sunlight.

Anyone who has sat in a car with the windows closed in the summertime has experienced this phenomenon.

The specific example that Fourier quoted came from an experiment designed by Horace Bénédict de Saussure, involving a vase with the interior containing black cork and covered by layers of glass separated by air.

Thermometers put into the vase and between the layers showed that, when exposed to sunlight, the interior of the vase reached the highest temperature and subsequent layers of air between the glass plates had successively lower temperatures.

His example was a somewhat unfortunate choice, as it became later known as the Greenhouse Effect, coined by Swedish meteorologist Nils Gustaf Ekholm, after greenhouses, which are essentially glass boxes that store heat from the sun to produce a temperate environment for the plants inside.

The reason the choice was unfortunate was that greenhouse glass keeps the inside air warm in large part because it traps the warm air heated by the sun inside of it, whereas on Earth warm air is free to rise by convection or transporting heat upward by conduction.

Absorbing heat radiation before it escapes is important for both greenhouses and the atmosphere, but that effect is less important in greenhouses than the fact that they trap air.

Fourier recognized this fact but concentrated on the issue of heat radiation when he discussed this experiment.

In his (translated) words:

> *The theory of this instrument is easy to formulate.*
>
> *It suffices to remark that: (1) the heat acquired is concentrated because it is not dissipated immediately by exchange of air with the surroundings; (2) the heat emanated by the Sun has properties different from those of dark heat.*
>
> *The rays of this star are for the most part transmitted through the glass without attenuation and reach the bottom of the box.*
>
> *They heat the air and the surfaces which contain it: the heat communicated in this way ceases to be luminous, and takes on the properties of dark radiant heat.*
>
> *In this state, the heat cannot freely traverse the layers of glass which cover the vessel; it accumulates more and more in the cavity enclosed by materials which conduct heat poorly, and the temperature rises to the point at which the incident heat is exactly balanced by the dissipated heat.*
>
> *One could verify this explanation, and render the consequences more evident, if one were to vary the conditions of the experiment by employing colored or darkened glass, and by making the cavities containing the thermometers empty of air.*
>
> *When one examines this effect by quantitative calculations, one finds results which conform entirely to those which the observations have yielded.*
>
> *It is necessary to consider this range of observations and the results of the calculations very carefully if one is to understand the influence of the atmosphere and the waters on the thermometric state of the Earth.*

The term “dark heat,” as translated here, refers to infrared radiation, discovered by Sir William Herschel in 1800.

In any case, what Fourier described in his antique language is more or less what is understood today to account for the fact that Earth is warmer than it would be without its atmosphere.

Actually, two factors come to play: Earth’s overall reflectivity and a greenhouse effect in the atmosphere.

I will consider each in turn, approaching the subject as a physicist does, by starting with the simplest picture then adding some additional complexity after exposing the key ideas.

Every second of every day, about 1361 watts per square meter crosses every point on a sphere located at Earth’s radius, centered on the sun.

Since Earth is a sphere and not a disk, not all of the radiation impinges directly on every part of Earth facing the sun.

So to estimate the total radiation hitting Earth, we can replace it by the area of a disk πR^2 facing the sun with a radius equal to Earth’s radius, R (Figure 4.1).

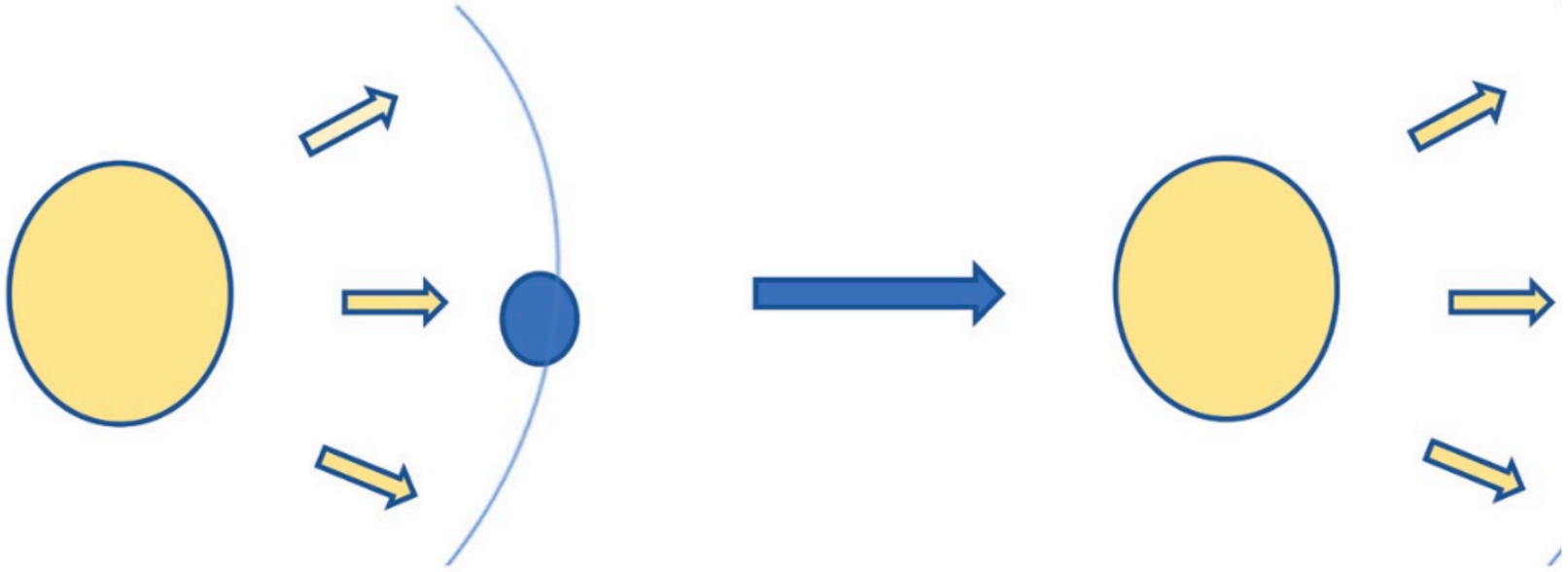


Figure 4.1

Dividing this total by actual Earth's surface area, which is four times as large, the average solar radiation power is therefore reduced by a factor of four to about 340 W/m^2.

In equilibrium, Earth will radiate away into space as much radiation energy as impinges on it from the sun.

Otherwise it would continue to heat up.

Earth also directly reflects some fraction, *a* (about 30 percent), of the sun's radiation.

So the total rate of energy flowing into Earth from the sun is given by $P_{solar} = 1361 \times (1-a)\pi R^2$.

This turns out to be about 10^{17} watts, or about 100,000 terawatts of power.

By comparison, humanity uses about 40 terawatts on average.

Therefore, more than a thousand times the energy used by all of humanity rains down from the sun each day.

If we could just harness even a small fraction efficiently, humanity's energy needs would be more than satisfied.

In any case, to not heat up any further, the Earth has to emit all of this power back into space, which it does by radiating the heat out as a thermal body.

It is one of the fundamental laws of physics, as we now understand them, that bodies at some temperature, T, radiate energy out at a rate that goes as the fourth power of temperature, multiplied by a constant that physicists label as the Stefan-Boltzmann constant.

I simply label it as k here.

The Earth radiates over its entire area, which is given by $4\pi R^2$, so the total power radiated by the Earth at temperature T is given by $P_{\text{Earth}} = (4\pi R^2) \times (kT^4)$.

To figure out the equilibrium temperature of the Earth, we just equate this quantity to 10^{17} watts and solve for T.

This is a high school-level physics problem, and when we plug in the numbers, we get that $T \approx -18°C$, about 251°K (K=Kelvin, or degrees Celsius above absolute zero) or about 0°F.

Earth should be frozen solid!

This was the quandary Fourier faced, and the solution he suggested used the fact that a body at this temperature radiates most of its energy in the form of infrared radiation.

The wavelength of radiation emitted by hot bodies depends on their temperature.

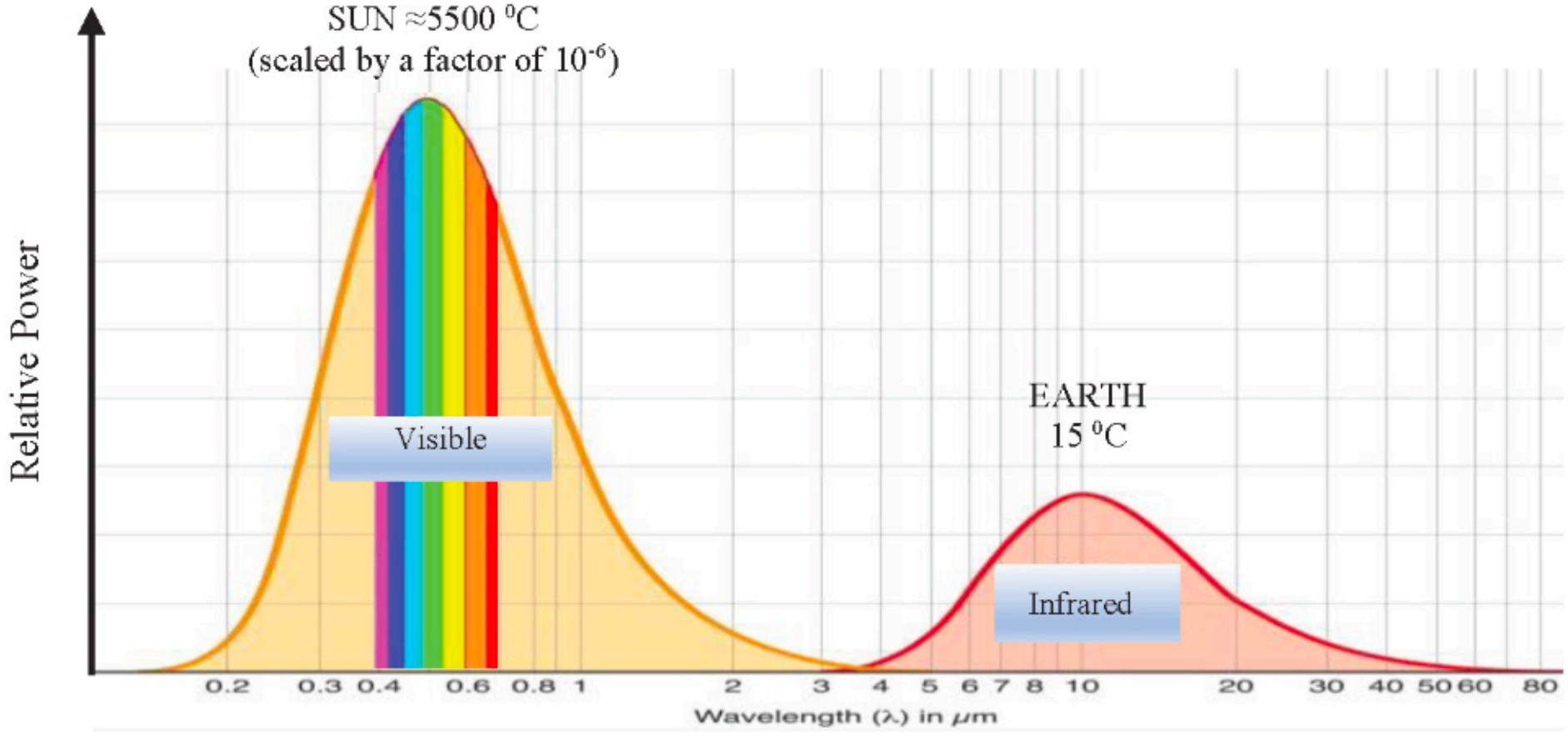
That is why heating elements go from red-hot to blue- or white-hot as their temperature increases.

An object with the average temperature of the Earth's surface (about 15°C) radiates in the infrared band, with wavelengths longer than visible light.

But, as you can literally see by looking up, the wavelengths of light from the sun are centered in the visible band.

Shown in Figure 4.2 are the spectra of light (as predicted by modern physics) from an object at the sun's surface temperature and a "black body" (one that absorbs and re-emits all incoming radiation thermally) at the Earth's surface temperature.

Figure 4.2



The difference in the frequency/wabelength of the two different types of radiation is important.

From the fact that we can see the sun, we know the atmosphere is largely transparent to radiation in the visible band.

But it may not be transparent to radiation in the infrared band.

In this case it would absorb radiation emitted upward by the surface of the Earth, but not incident radiation coming down from the sun.

That would change everything, and this possibility is central to the terrestrial greenhouse effect.

To see this more explicitly, assume that the Earth's atmosphere absorbs and stores some fraction, say b, of that energy.

Then instead of radiating a power P_{Earth} into space as before, the Earth would radiate

$$(1-b)\, P_{\text{Earth}} = (1-b) 4\pi R^2\, kT^4$$

Then, to radiate as much power into space as it receives from the sun,

$$(1-b)\, P_{Earth} + P_A = P_{solar}$$

Since *(1-b) < 1*, this means the power radiated at the surface of the Earth would have to be greater than it would be if *b* were 0.

Since the power radiated is proportional to the fourth power of its temperature, this in turn means the Earth's temperature would have to be greater than it would otherwise be if some of the energy it would otherwise radiate into space didn't get trapped in the atmosphere first.

Absorption of IR radiation is therefore at the heart of the greenhouse effect, although the actual details are more complex.

Even though the model I have just presented is simplistic, as physicists are wont to do on a first try, simple models often reveal the key physical factors that are at play.

A joke in which a physicist first pictures a cow as a sphere when trying to analyze it is often told.

But even that simplistic picture can reveal a lot—about not just cows, but life in general.

We can be a little more realistic by recognizing that the atmosphere itself, if it absorbs heat energy, cannot simply just store it, but it must also radiate energy if it is not to continue heating up.

Approximating the entire atmosphere as having a single temperature, it will radiate with a power

$$P_A = 4\pi R^2 \, kT_A^4$$

But it will then emit radiation in two directions, both out into space and back down again to Earth (Figure 4.3).

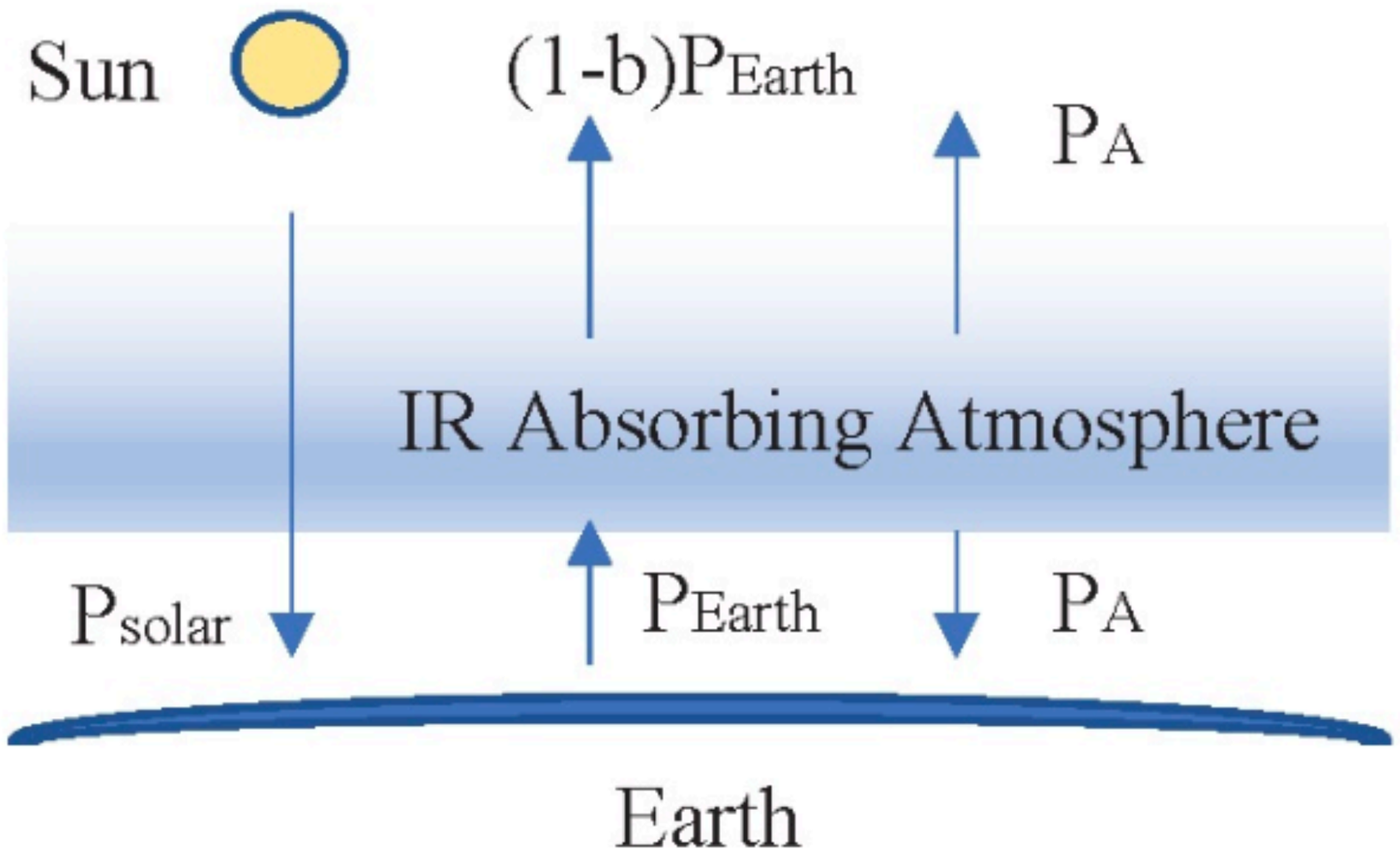


Figure 4.3

For Earth to be in equilibrium with the incoming solar radiation, the sum of the emission from the Earth into space plus the emission from the atmosphere into space must equal the incoming power coming from the sun:

$$(1\text{-b})\, P_{Earth} + P_A = P_{solar}$$

For the atmosphere to be in equilibrium with the radiation impinging upon it from the Earth, its total emission, that is, its emission back toward the Earth plus its emission into space, must equal the amount it absorbs.

This is a fraction *b* of the radiation emitted from the Earth.

$$bP_{Earth} = 2\, P_A$$

Using this relation for P_A in the previous expression yields:

$$(1\text{-b/2})\, P_{Earth} = P_{solar}$$

This somewhat more realistic model implies that even if subsequent emission from the atmosphere into space is included in the estimate following absorption of IR radiation from the Earth, the power emitted from the Earth still increases compared to what it would be if the atmosphere absorbed no radiation.

The quantitative amount by which it increases is now different, however.

More importantly, this model is only consistent if we assume the temperature of the atmosphere radiating into space is lower than the temperature of the Earth.

If it were the same, then the top of the atmosphere would be emitting radiation with the same power as the surface of Earth.

This is consistent with $b = 0$, or no absorption of IR radiation by the atmosphere.

As we shall see, this requirement is satisfied precisely because the atmosphere is not homogeneous and the top of the atmosphere is both less dense and cooler.

Expecting precise and accurate quantitative estimates on the basis of such a simple model is unrealistic.

But using it as a rough guide demonstrates the significant role of absorption in determining the overall magnitude of the observed greenhouse effect on Earth.

The average temperature at Earth's surface is observed to be 15°C and not the value of -18°C we derived for an Earth without any absorbing atmosphere.

Using this simplistic model, one finds that most of the IR radiation emanating from the Earth's surface would have to be absorbed by the atmosphere—indeed almost 80 percent of it.

Perhaps the main reason detailed quantitative estimates using this model are approximate at best is because the model as I have presented ignores a number of physical effects besides absorption of solar radiation by the Earth's surface and subsequent thermal radiation by the surface and the atmosphere.

First off, while the atmosphere is mostly transparent to the sun's radiation, thankfully it is not fully so, or we might all die of skin cancer from ultraviolet radiation.

In total the atmosphere absorbs about 23 percent of the incoming solar radiation, meaning between reflection and atmospheric absorption, only about 49 percent of the incident solar radiation actually makes it down to the surface of the Earth.

Next, the Earth's atmosphere and surface are far from uniform, so temperatures vary greatly from the equator to the poles, and temperatures and absorption rates vary vertically through the atmosphere.

Thus, it is a rough approximation to treat the atmosphere as a single homogeneous entity.

Also, wind and ocean currents transfer heat laterally around the Earth.

In addition, the Earth's thermal response time (in large part due to the huge heat capacity of the oceans) is such that it can take centuries for heat buildup to manifest itself as a rise in temperature, so a radiation imbalance can occur for some time before equilibrium temperatures are reached.

Equally important in changing the overall energy balance is the fact that radiation is not the only way heat is transferred from the Earth to the atmosphere since the Earth's surface is in direct physical contact with the atmosphere.

Slightly more than 50 percent of the incoming solar energy that reaches the Earth's surface (or about 25 percent of the total incoming solar flux) leaves the surface through evaporation.

The heat goes into evaporating water, which is then stored and can be released later in storms.

As I will elaborate on later, this allows heat energy to be transferred from warmer to colder places on Earth and drives an atmospheric "heat engine."

Another 10 percent or so of the solar energy reaching the Earth's surface gets transferred back to the atmosphere by convection.

Warm air close to the Earth's surface rises up into the atmosphere, where it releases its energy by expanding and cooling.

This leaves only 35–40 percent of the incoming radiation absorbed at the Earth's surface to be released in the form of thermal radiation.

The atmosphere naturally absorbs about 1/3 of this radiation, meaning about 24 percent of the radiation that is absorbed by the Earth's surface ends up being reradiated out into space from the surface.

This in turn means only about 12 percent of the total incident radiation from the sun (remember, only 48 percent of that total incident radiation makes it down to the Earth's surface in the first place) gets radiated back into space from the Earth's surface.

Since about 30 percent of the total incident radiation is reflected, that leaves almost 60 percent of the incident radiation that must be reradiated into space *by the atmosphere* for the Earth-sun system to remain in thermal equilibrium.

This illustrates a feature that is central to the terrestrial greenhouse effect.

Most of the incident solar radiation is absorbed by the Earth's surface, while most of the outgoing radiation is radiated into space at the top of the Earth's atmosphere.

This is a consequence of the heat energy stored in the atmosphere, as captured in both of the toy models I described earlier.

Note also that the top of the atmosphere is in thermal radiative equilibrium with about 60 percent of the incoming solar energy flux.

As a result, one would expect, using the original estimate we made for the Earth's temperature without an atmosphere (-18°C or 253° Kelvin), that the average temperature in the top layer of the atmosphere where the radiation is emitted would be something like -48°C ($(0.6)^{1/4}$ x 253°K).

Happily, this is in accord with observations of the temperature near the top of the troposphere, about 10 km or so above the Earth's surface, below which most of the absorbing part of the atmosphere lies.

With the top of the atmosphere radiating energy upward, at least an equal amount (60 percent of the incident solar flux) must be radiating downward.

But the atmosphere actually radiates much more energy down to the surface than it does to outer space.

This so-called “back radiation” may seem surprising, as it has been a source of confusion, especially among those who are reluctant to accept the basic physics of the greenhouse effect.

First, note that the atmosphere is warmer at lower altitudes, where it is radiating down to the surface, and one might expect more power emitted from this region.

But more importantly, recall that the surface is transferring roughly an additional 35 percent of the solar heat it absorbs up to the atmosphere via evaporation and convection, as well as depositing about 5 percent of its radiant energy output directly to the atmosphere.

When you add these up, it turns out that the atmosphere needs to radiate back down to the Earth an additional amount of radiation equivalent to about 40 percent of the solar energy originally incident upon the Earth to not continue heating up.

This means it radiates downward an energy equivalent of almost 100 percent of the energy originally incident on the Earth from the sun.

The surface, which is far warmer than the atmosphere in general, ends up radiating upward about 116 percent of the solar energy incident on the Earth, which balances the fraction of the original solar flux that makes its way down to the surface plus the energy radiated downward by the atmosphere.

Of this upward radiation, about 12 percent or so makes its way back up through the atmosphere and outward directly to space, and the remaining 104 percent gets absorbed by the atmosphere.

If the percentages seem confusing, here is a visualization of the Earth-atmosphere energy balance when the two are in equilibrium with incoming solar radiation, provided by the National Oceanic and Atmospheric Administration (NOAA) in Figure 4.4.

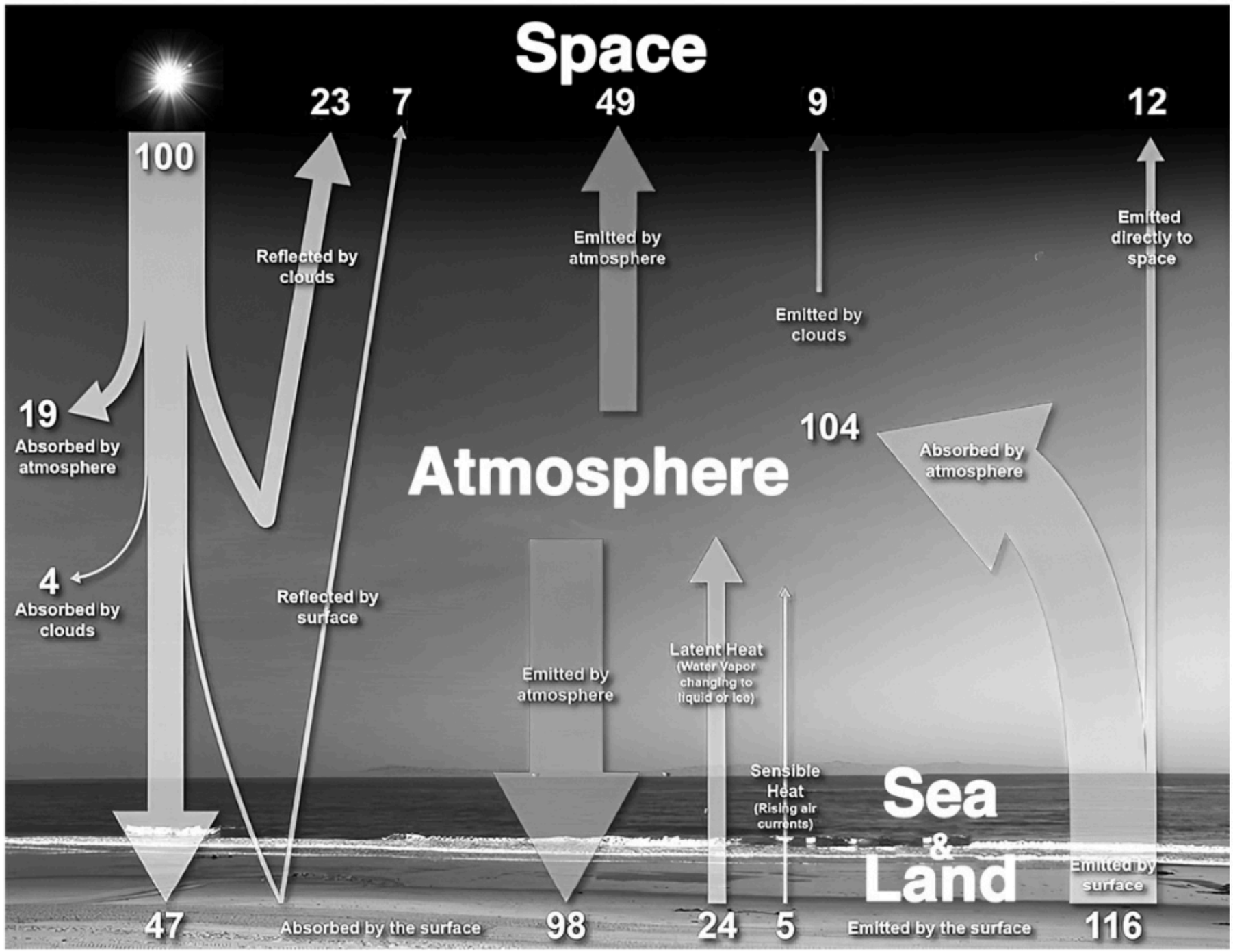


Figure 4.4

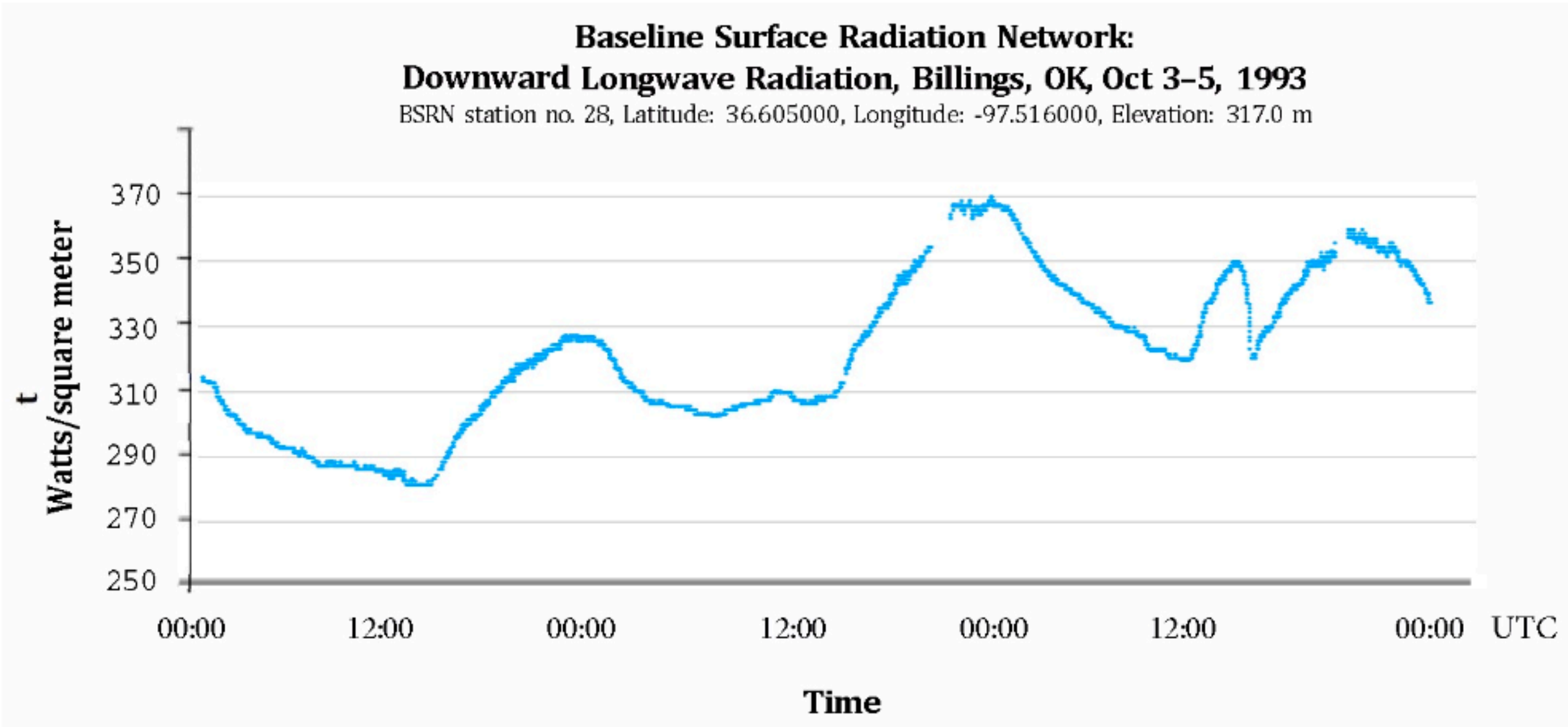
It is worth stressing that we can actually measure all of these quantities, so you don't have to take them on faith, even if the general features are already predicted by the simplest models.

Downward-facing satellites can measure the upward-going radiated power emanating from the atmosphere and the Earth's surface.

Also, because the spectrum of light incoming from the sun is different than the thermal spectrum of radiation emitted by the atmosphere, upward-facing detectors at the Earth's surface can distinguish between them.

Here, for example, are the data from three days of "downward longwave radiation" (the colder atmospheric—or long wavelength—component of downward radiation) measurement in one mid-latitude location in the US, with the power in W/m^2 shown in Figure 4.5.

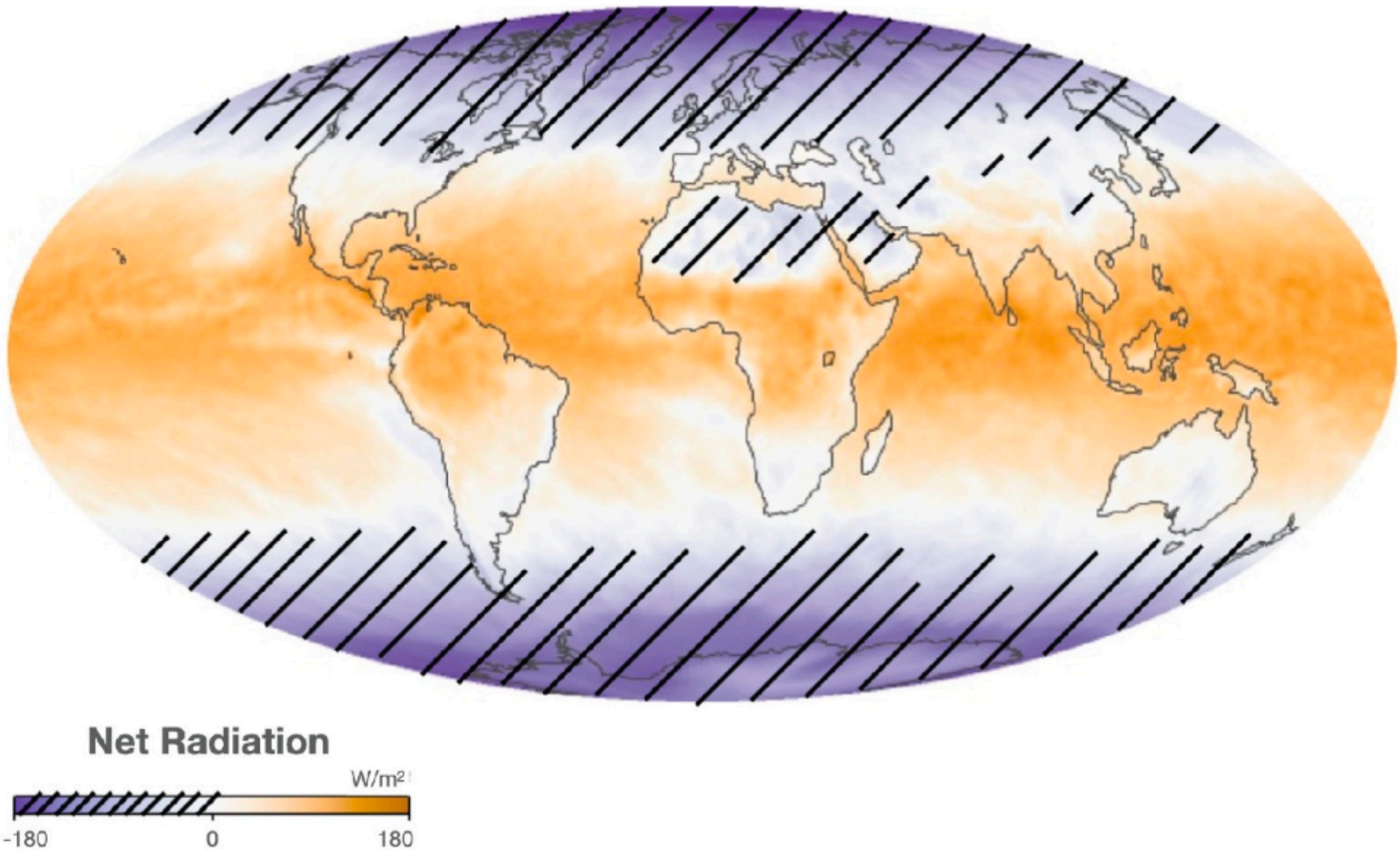
Figure 4.5



While this component varies just as the incident solar flux does, with the highest magnitude at low latitudes and the lowest magnitude at high latitudes, at mid-latitudes it achieves a value close to its average value, which one can see here is indeed comparable in power to the total average incoming solar power impinging on the Earth, as predicted.

Also, for reference, here is satellite data from Sept. 2019 showing the net radiation (incoming sunlight minus reflected light and outgoing heat) flowing into the Earth (Figure 4.6).

Figure 4.6



Note again that near the equator more radiation is absorbed than emitted or reflected, and the opposite is true near the poles.

Mid-latitudes are roughly in balance.

To conclude, then, just so we don't lose the forest for the trees, here are the key takeaways from the physical picture I have presented thus far.

1. The heart of the greenhouse effect, first proposed by Fourier and confirmed by the fact that the surface temperature of Earth is warmer than it would otherwise be, is as follows:

If the atmosphere absorbs heat energy that would otherwise be radiating out into space by Earth, the resulting surface temperature of Earth will be warmer than the atmosphere.

As a consequence, it will be warmer than it would otherwise be in the absence of an atmosphere.

The atmosphere indeed acts like a blanket.

2. Observations of the temperature of the surface and the atmosphere and measurements of radiant energy flows upward and downward are consistent with this fundamental understanding of the greenhouse effect.

This understanding derives from longstanding and well-understood physics arguments.

Basic model predictions and observational data combine to establish the basis of this physical picture of the greenhouse effect.

3. While a precise determination of how effective a blanket the atmosphere will provide requires detailed modeling of the atmosphere itself, the generic features and consistent quantitative estimates of the greenhouse effect derive from the following basic principles:

a. In thermal equilibrium: energy in = energy out.

b. Thermal bodies radiate with power that is proportional to the fourth power of their temperature.

This basic picture is not controversial.

What comes next depends on the details....

5 - THE BIRTH OF CLIMATE CHANGE

While Fourier realized that the absorption of outward-going infrared heat energy is the key to what became known as the greenhouse effect, he did not propose what the source of this absorption was.

Nor did he provide quantitative estimates of what it might be.

The quantitative science underlying the greenhouse effect began with John Tyndall and Eunice Foote. Tyndall, who was born just around the time Fourier, was beginning to ponder the question of what governed the Earth's temperature.

In the late 1840s he and a colleague, who both taught at an English boarding school, had the good sense to move to Germany to study. German universities were far ahead of Britain in experimental science at the time.

By the time Tyndall returned in 1851, he was well prepared to begin a career in experimental chemistry and physics, where his early work focused on magnetism, one of the subjects of greatest interest at the time.

In 1853 he obtained a professorship at the Royal Institution, led by Michael Faraday, one of the greatest experimental physicists of the nineteenth century, whose work established the modern theory of electromagnetism.

A decade later, upon Faraday's retirement, Tyndall was appointed his successor.

A lifelong interest in mountain climbing and glaciology led Tyndall to explore the heating effects of sunlight and study the earlier work of Fourier and others proposing that the absorption of infrared light by the atmosphere might ultimately heat the Earth's surface.

Ever the experimentalist, in 1859 Tyndall began a decade-long set of experiments on the absorption of thermal radiation by different gases.

As for Eunice Foote - her work predated discoveries made by John Tyndall, who had been recognized by scientists as the first person to experimentally show the mechanism of the greenhouse effect involving infrared radiation.

Detailed examination of her work by modern scientists has confirmed that **three years** before Tyndall published his paper in 1859, Foote discovered that water vapor and CO_2 absorb heat from sunlight.

Furthermore, her view that variances in the atmospheric levels of water vapor and CO_2 would result in climate change preceded Tyndall's 1861 publication by five years.

Because of the limits of her experimental design, and possibly a lack of knowledge of infrared radiation, Foote did not examine or detect the absorption and emission of radiant energy within the thermal infrared range, which is the cause of the greenhouse effect.

As usual, she was not recognized until recently and very little is known about her, her methods and reasoning.

Thus, I will talk about Tyndall’s work but remembering Foote probably did everything earlier!!!!

Tyndall began experiments in May of that year with a setup that consisted of a known heat source, a tube containing the gas he wanted to study, and at the end of that an object called a thermopile, which converts thermal energy into an electrical signal he could measure.

By measuring the thermal energy emanating from the tube, he could determine how much energy was absorbed by the gas.

Incidentally, this was an early application of what is called absorption spectroscopy, which, turned on its head, was ironically essentially the same tool used by Keeling to first accurately measure the fraction of CO_2 in the atmosphere.

Tyndall helped first determine the absorption properties of CO_2, and his work was followed by a century of more refined studies.

When Keeling did his work, he could compare the absorption of infrared radiation by atmospheric samples to the known absorption properties of test samples with predetermined CO_2 fractions to determine the CO_2 fraction in the atmosphere.

A hundred years earlier, Tyndall actually used his own method to invent a system for measuring the CO_2 fraction in human breath, a method that is still in use today.

Returning to Tyndall, by the end of May, he enthusiastically reported his results to the Royal Society:

> *With the exception of the celebrated memoir of M. Pouillet on Solar Radiation through the atmosphere, nothing, so far as I am aware, has been published on the transmission of radiant heat through gaseous bodies.*
>
> *We know nothing of the effect even of air upon heat radiated from terrestrial sources.*

By June 10 he had firmly established the experimental physics of the greenhouse effect.

As he put it in a Royal Society lecture, regarding solar heat:

> *When the heat is absorbed by the planet, it is so changed in quality that the rays emanating from the planet cannot get with the same freedom back into space.*
>
> *Thus the atmosphere admits of the entrance of solar heat; but checks its exit, and the result is a tendency to accumulate heat at the surface of the planet.*

Tyndall studied the infrared absorption by nitrogen, oxygen, water vapor, carbon dioxide, ozone, and methane, among other gases in the atmosphere, and was the first to get quantitative data on this absorption.

His chief result was that the main gases, oxygen and nitrogen, did not have any significant absorptions, but that water vapor is the strongest absorber of infrared radiation in the atmosphere.

He also demonstrated that the gases that absorb infrared radiation are also infrared emitters, establishing, as we have seen, another key component of the greenhouse effect.

There is a somewhat simple reason why O_2 and N_2 molecules don't absorb much infrared radiation, while CO_2 and H_2O do.

Absorption of light can occur when the incident radiation excites vibrations of the molecules.

For O_2 or N_2, this looks like Figure 5.1.

- where the atoms are vibrating back and forth.

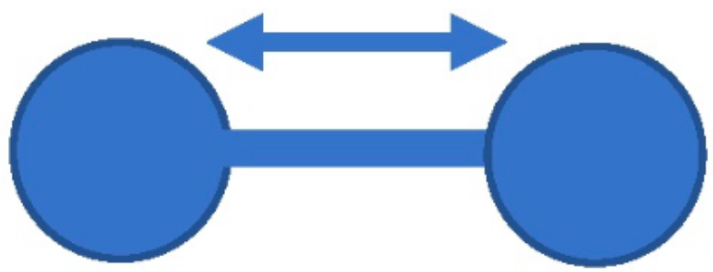


Figure 5.1

These vibrations happen rapidly, so only higher-frequency radiation (like ultraviolet radiation, for example) will excite the vibrations, so these gases absorb much of the harmful ultraviolet radiation from the sun.

H_2O and CO_2 look more like Figure 5.2, however.

This bending vibration is slower, so lower-frequency infrared radiation can excite these vibrations.

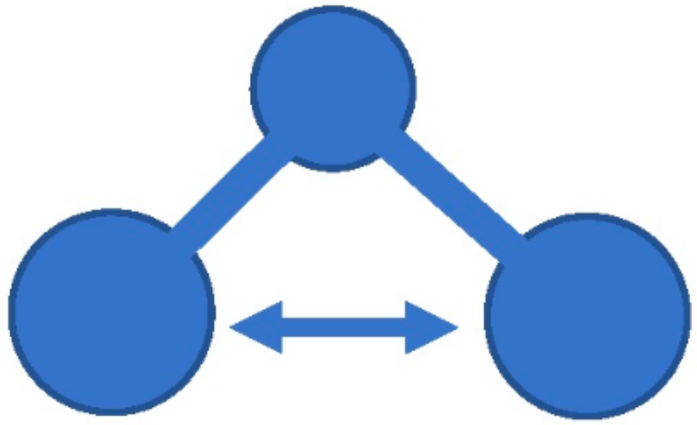


Figure 5.2

As such, CO_2 and H_2O preferentially absorb in the infrared.

With Tyndall's results, by the 1860s the emerging experimental basis of the greenhouse effect was largely demonstrated.

You will notice, however, that CO_2 had not yet played a prominent role.

Rather, Tyndall demonstrated that it was water vapor that dominates the warming of the Earth, although small amounts of hydrocarbons and CO_2 could have a significant effect.

Incidentally, water vapor was initially responsible for the hellish heat on Venus and will be responsible for a runaway greenhouse effect on Earth in about two billion years or so, by which time the solar luminosity will have increased by about 15 percent.

The increased temperatures this will produce will cause the world's oceans to begin to evaporate, putting more infrared-absorbing water vapor in the atmosphere, which will further heat up the Earth, causing more evaporation, and so on.

After that time, the surface temperature of the Earth will resemble that on Venus now, in excess of 450°C, or hot enough to melt lead.)

Enter Svante August Arrhenius, a Swedish physical chemist and Sweden's first Nobel laureate in 1903.

The work that earned him the prize involved demonstrating that when salts dissolve in water, they dissociate into separate charged particles, which Faraday had called ions.

Later on, he demonstrated that for most chemical reactions to proceed, additional heat energy is required to overcome an energy barrier that otherwise stops separate molecules from reacting and developed an equation that still bears his name for a relationship between this "activation" energy and reaction rates.

Still later he did seminal work in the early development of biochemistry and even proposed the theory now known as *panspermia*, which suggests that life first arrived on Earth through the transport of spores from earlier life-harboring planets.

So, he was no slouch.

But he seems to have irritated his colleagues from his student days onward.

His PhD thesis, which contained the germ of the idea for which he would win the Nobel Prize, was rated as “third-class” by his professors, and even that was done only after his defense that the original grade of “fourth-class” should be raised.

He became a professor of physics at Stockholm University, but only after overcoming the opposition of his colleagues, and was elected to the Swedish Academy of Science in 1901, but only after further opposition from his colleagues.

He weathered the opposition well, however.

He became rector of Stockholm University in 1896 and later on rector of the Nobel Institute itself.

I find it intriguing that he actually helped set up the Institute and the Nobel Prizes, and was a member of the Nobel Committees on Physics and Chemistry, then two years later was the first Swede to be awarded the Prize.

Happily, even if there may have been a bit of nepotism involved, his contributions to the fields of chemistry and physics were nevertheless important and lasting.

In 1896 Arrhenius turned his considerable skill in physical chemistry to the question of climate change.

He knew of the work of Fourier and Tyndall (and also Claude Pouillet) demonstrating that a greenhouse effect could warm the Earth.

The physical question that drove him appeared to be his interest in the appearance and disappearance of ice ages.

As he tellingly put it in his groundbreaking 1896 paper, “On the Influence of Carbonic Acid in the Air upon the Temperature of the Ground,” published in the *Philosophical Magazine and Journal of Science*,

> *I should certainly not have undertaken these tedious calculations if an extraordinary interest had not been connected with them. In the Physical Society of Stockholm there have been occasionally very lively discussions on the probable causes of the Ice Age; and these discussions have, in my opinion, led to the conclusion that there exists as yet no satisfactory hypothesis that could explain how the climatic conditions for an ice age could be realized in so short a time as that which has elapsed from the days of the glacial period.*

He therefore became interested in exploring how possible periodic variations in the atmosphere could change the greenhouse effect and therefore also the Earth's temperature over "short" time intervals.

Naturally he was concerned with whether variations large enough to affect the climate might be possible due to natural processes on the Earth's surface.

Here, he translated into English the lesser-known work of his colleague Arvid Högbom, who estimated how the CO_2 (referred to as carbonic acid at the time in deference to the substance formed when CO_2 dissolves in water) concentration then in the atmosphere might compare to the carbon stored in living systems, and I quoted his result earlier in this discussion.

I find Högbom's estimate to be a wonderful example of the type of "order of magnitude" arguments that are so important and powerful in physics.

They allow us to quickly discern what are the crucial physical factors we should focus on and which are likely to be quantitatively irrelevant—namely the type of thinking that motivates this discussion.

For that reason, I present his discussion verbatim here:

On the supposition that the mean quantity of carbonic acid in the air reaches 0.03 vol per cent. [Note by me: this agrees extremely well with the current estimate that CO_2 comprised .03 percent of the atmosphere in 1900 as can be seen by the graphs in chapter 2], this number represents 0.045 per cent. by weight, or 0.342 millim. partial pressure, or 0.466 grams of carbonic acid for every cm^2 of the Earth's surface.

Reduced to carbon this quantity would give a layer about 1 millimeter thickness over the Earth's surface.

The quantity of carbon that is fixed in the living organic world can certainly not be estimated with the same degree of exactness; but it is evident that the numbers that might express this quantity ought to be of the same order of magnitude, so that the carbon in the air can neither be conceived of as very great nor as very little, in comparison with the quantity of carbon occurring in organisms.

With regard to the great rapidity with which the transformation in organic nature proceeds, this disposable quantity of carbonic acid is not so excessive that changes caused by climatological or other reasons in the velocity and value of that transformation might not be able to cause displacements of the equilibrium.

It may be over one hundred years old, but this still stands as a beautifully simple and powerful order-of-magnitude argument.

It demonstrates that global atmospheric CO_2 might be affected, in principle, at significant levels by living systems on biological, and not geological, timescales.

I was pleased to learn of it while researching for this class.

The first thing Arrhenius did was ingeniously use infrared observations of the moon taken by Samuel Pierpont Langley to calculate how much infrared radiation is absorbed by CO_2 and water vapor in the atmosphere.

He focused on CO_2 as well as water vapor in part because Tyndall had shown that while water vapor was the dominant greenhouse gas, CO_2 and some other trace gases could also have some impact.

Also, researchers who followed after Tyndall argued that CO_2 played a more significant role than Tyndall had accorded it.

Arrhenius came up with the absorption table (Figure 5.3), which I also reproduce here because the amount of effort he devoted to developing it was considerable.

Figure 5.3

TABLE III.—*The Transparency of a given Atmosphere for Heat from a body of* 15° C.

→ H_2O ↓ CO_2	0·3.	0·5.	1·0.	1·5.	2·0.	3·0.	4·0.	6·0.	10·0.
1	*37·2*	*35·0*	*30·7*	*26·9*	*23·9*	*19·3*	*16·0*	*10·7*	*8·9*
1·2	*34·7*	32·7	28·6	*25·1*	22·2	17·8	14·7	9·7	8·0
1·5	*31·5*	29·6	25·9	*22·6*	19·9	*15·9*	13·0	8·4	6·9
2	*27·0*	*25·3*	*21·9*	*19·1*	16·7	13·1	10·5	6·6	*5·3*
2·5	*23·5*	22·0	19·0	16·6	14·4	*11·0*	8·7	5·3	4·2
3	*20·1*	18·8	16·3	*14·2*	12·3	9·3	*7·4*	4·2	*3·3*
4	*15·8*	14·7	12·7	10·8	9·3	*7·1*	5·6	3·1	2·0
6	*10·9*	10·2	*8·7*	*7·3*	6·3	4·8	3·7	*1·9*	*0·93*
10	*6·6*	6·1	5·2	4·3	3·5	*2·4*	1·8	1·0	*0·26*
20	*2·9*	2·5	2·2	*1·8*	1·5	*1·0*	0·75	0·39	*0·07*
40	*0·88*	0·81	0·67	*0·56*	0·46	*0·32*	0·24	*0·12*	*0·02*

His units took me some time to figure out.

For CO_2, the units are relative: "1" means the amount of CO_2 that would be encountered by a light ray then traversing vertically through the atmosphere, where it passes through the smallest mass of air.

The units for water vapor are a bit more standard: "1" represents the amount of water vapor traversed by a vertical ray of light for a humidity of 10 grams/cubic meter at the Earth's surface, which he argued was the mean humidity of the air.

Note that Arrhenius therefore estimated that about 70 percent of infrared radiation traveling vertically upward was absorbed by the atmosphere at that time.

He used this data and combined that with radiation equilibrium arguments similar to those I described earlier and calculated how the expected equilibrium surface temperature might be expected to vary as one changed the CO_2 concentration.

In several ways, his models were somewhat more complex than I previously described.

First, Arrhenius didn't assume a uniform temperature in the atmosphere and explicitly demonstrated a fact that I alluded to before, namely that a colder upper atmosphere and warmer lower atmosphere will increase the surface temperature because less radiation will be emitted from the top of the atmosphere and heat will be more effectively concentrated near the ground.

He also attempted to incorporate the effects of clouds on temperature by recourse to data.

Finally, as far as I can tell, Arrhenius explicitly accounted for the increase in absolute humidity (and thus, the increase in absorption by water vapor) that occurs for latitudes with higher average temperatures when determining how a variation in the CO_2 abundance would change surface temperatures by using measured humidity and temperature at various locations in his calculations.

He didn't include, however, an effect that can amplify the impact of rising CO_2 on temperature.

Namely, as rising CO_2 concentrations produce rising temperatures, there can be more evaporation of water, increasing the vapor content of the air, and thus further increasing infrared absorption by the atmosphere.

But it was pretty impressive for a first pass.

Figure 5.4 shows his results, which, though overestimates, nevertheless hold up roughly under modern scrutiny.

TABLE VII.—*Variation of Temperature caused by a given Variation of Carbonic Acid.*

Latitude.	Carbonic Acid = 0·67.					Carbonic Acid = 1·5.					Carbonic Acid = 2·0.					Carbonic Acid = 2·5.					Carbonic Acid = 3·0.				
	Dec.–Feb.	March–May.	June–Aug.	Sept.–Nov.	Mean of the year.	Dec.–Feb.	March–May.	June–Aug.	Sept.–Nov.	Mean of the year.	Dec.–Feb.	March–May.	June–Aug.	Sept.–Nov.	Mean of the year.	Dec.–Feb.	March–May.	June–Aug.	Sept.–Nov.	Mean of the year.	Dec.–Feb.	March–May.	June–Aug.	Sept.–Nov.	Mean of the year.
70	−2·9	−3·0	−3·4	−3·1	−3·1	3·3	3·4	3·8	3·6	3·52	6·0	6·1	6·0	6·1	6·05	7·9	8·0	7·9	8·0	7·95	9·1	9·3	9·4	9·4	9·3
60	−3·0	−3·2	−3·4	−3·3	−3·22	3·4	3·7	3·6	3·8	3·62	6·1	6·1	5·8	6·1	6·02	8·0	8·0	7·6	7·9	7·87	9·3	9·5	8·9	9·5	9·3
50	−3·2	−3·3	−3·3	−3·4	−3·3	3·7	3·8	3·4	3·7	3·65	6·1	6·1	5·5	6·0	5·92	8·0	7·9	7·0	7·9	7·7	9·5	9·4	8·6	9·2	9·17
40	−3·4	−3·4	−3·2	−3·3	−3·32	3·7	3·6	3·3	3·5	3·52	6·0	5·8	5·4	5·6	5·7	7·9	7·6	6·9	7·3	7·42	9·3	9·0	8·2	8·8	8·82
30	−3·3	−3·2	−3·1	−3·1	−3·17	3·5	3·3	3·2	3·5	3·47	5·6	5·4	5·0	5·2	5·3	7·2	7·0	6·6	6·7	6·87	8·7	8·3	7·5	7·9	8·1
20	−3·1	−3·1	−3·0	−3·1	−3·07	3·5	3·2	3·1	3·2	3·25	5·2	5·0	4·9	5·0	5·02	6·7	6·6	6·3	6·6	6·52	7·9	7·5	7·2	7·5	7·52
10	−3·1	−3·0	−3·0	−3·0	−3·02	3·2	3·2	3·1	3·1	3·15	5·0	5·0	4·9	4·9	4·95	6·6	6·4	6·3	6·4	6·42	7·4	7·3	7·2	7·3	7·3
0	−3·0	−3·0	−3·1	−3·0	−3·02	3·1	3·1	3·2	3·2	3·15	4·9	4·9	5·0	5·0	4·95	6·4	6·4	6·6	6·6	6·5	7·3	7·3	7·4	7·4	7·35
−10	−3·1	−3·1	−3·2	−3·1	−3·12	3·2	3·2	3·2	3·2	3·2	5·0	5·0	5·2	5·1	5·07	6·6	6·6	6·7	6·7	6·65	7·4	7·5	8·0	7·6	7·62
−20	−3·1	−3·2	−3·3	−3·2	−3·2	3·2	3·2	3·4	3·3	3·27	5·2	5·3	5·5	5·4	5·35	6·7	6·8	7·0	7·0	6·87	7·9	8·1	8·6	8·3	8·22
−30	−3·3	−3·3	−3·4	−3·4	−3·35	3·4	3·5	3·7	3·5	3·52	5·5	5·6	5·8	5·6	5·62	7·0	7·2	7·7	7·4	7·32	8·6	8·7	9·1	8·8	8·8
−40	−3·4	−3·4	−3·3	−3·4	−3·37	3·6	3·7	3·8	3·7	3·7	5·8	6·0	6·0	6·0	5·95	7·7	7·9	7·9	7·9	7·85	9·1	9·2	9·4	9·3	9·25
−50	−3·2	−3·3	—	—	—	3·8	3·7	—	—	—	6·0	6·1	—	—	—	7·9	8·0	—	—	—	9·4	9·5	—	—	—
−60																									

Figure 5.4

As you can see from the table, for an increase of 50 percent in the CO_2 concentration in the atmosphere, he estimated a relatively uniform increase of about 3–3.5°C, whereas for a doubling of CO_2 he predicted an increase of 5– 6°C.

In all cases he predicted the temperature increase would be larger at the poles than at the equator.

Being the consummate physicist, however, Arrhenius wasn't just satisfied with a table.

He preferred a more general rule, or law, he could derive.

He did that by interpolating the results in his table.

It is also worthwhile to note that the last three columns for highest values of CO_2 were actually obtained by extrapolating the original data of Langley beyond the domain in which the data were taken.

That said, his result was clear from the table.

A regular change in temperature of about 3°C would be expected each time the value of the concentration of CO_2 increased by a factor of 3/2.

As Arrhenius stated,

“if the quantity of carbonic acid increases in geometric progression, the augmentation of temperature will increase nearly in arithmetic progression.”

This relationship, where one quantity increases linearly while the other quantity increases as a power law (that is, by 3/2 between 0.67 and 1, by $(3/2)^2$ between 0.67 and 1.5, and by $(3/2)^3$ between 0.67 and 2.25), can be mathematically shown to imply that the first quantity (such as surface temperature) varies as the logarithm of the second quantity (CO_2 concentration).

Arrhenius not only pushed the limits of the validity of Langley’s data in his extrapolation, he ignored Langley’s warning that his data should be taken with a grain of salt.

Nevertheless, after a previous century of scientific effort devoted to figuring out how to merely understand the average temperature at the Earth’s surface, Arrhenius’s goal of exploring how possible changes in that temperature might be induced by a change in atmospheric content was particularly novel and ambitious.

And to be fair, he used what was then the best data and theory available to him.

As we shall see, the basic framework of his results still stands today.

Not bad for work done in 1896.

Interestingly, while Arrhenius is usually credited for being the first person to conclude that CO_2 emissions by human industrial activity were potentially large enough to heat up the planet, his fundamental work was neither motivated by this, nor did he make this claim in his landmark scientific paper.

As he stated there, he was motivated by an interest in understanding both the origin of ice ages and how the ice ages ended, and suspected that varying CO_2 concentration might explain these variations in the Earth's temperature.

Ultimately, however, he was not able to suggest a precise mechanism for how CO_2 might vary by the requisite amounts over geological time.

Nevertheless, he was aware, from his colleague Arvid Högbom, that the consumption of coal at the time —about five hundred million tons of coal per year—would emit an amount equal to about 1/1000th of the CO_2 then present in the atmosphere.

Högbom argued that this additional contribution was easily compensated by the formation of carbonate compounds by weathering processes, though as I described earlier, we now know this latter process is balanced geologically by CO_2 emissions of volcanos.

Interestingly, if you take Högbom's estimate and assume roughly a century of coal burning at the rate he suggested, this would increase the atmospheric CO_2 concentration by about 10 percent, were it to all remain in the atmosphere.

This is more or less the increase that did occur between 1750 and 1900 (from 275 to 300 ppm) as seen in the extended Keeling curves I displayed earlier.

In a public lecture Arrhenius gave a few years after his scientific paper was presented, he didn't hesitate to extrapolate Högbom's estimate into the future.

When he accounted for absorption of CO_2 by the oceans, he estimated that, at the then-current rate of fossil fuel burning, the CO_2 concentration in the atmosphere would double in three thousand years, and global temperatures would increase by 3–4°C.

He extolled this possibility as a boon for humanity.

Not only would we avoid another ice age, but it would "allow our descendants, even if they only be those of a distant future, to live under a warmer sky in a less harsh environment than we were granted."

As a resident of Scandinavia, global warming seemed to him a most welcome prospect!

Arrhenius expanded on these ideas in two books written over the next decade, earning him a reputation today as the first scientist to predict human-induced global warming.

Poor Mr. Högbom, whose estimates provided the basis of Arrhenius's claims, has been largely buried in the dustbin of history.

Credit aside, Arrhenius was the first to construct an empirically derived model that allowed for the possibility of understanding how sensitive our climate may be to changes in the abundance of a molecule whose concentration in the atmosphere was less than one part in a thousand, something that without sufficient intellectual preparation, may seem absurd.

For that, and the fact that the basic features of his model and the overall magnitude of his predictions remain relevant today, he certainly deserves to be remembered.

6 - FORCING THE ISSUE

So far, we have established:

a) CO_2 levels have been measured to be increasing yearly at a level consistent with emissions produced by human industrial activity. Current levels are unprecedented over the past one million years or so and more than 30 percent higher than at any other period during that time.

b) CO_2 is measured to be a greenhouse gas, in that it absorbs infrared radiation emitted by the Earth's surface. Along with water vapor, it is quantitatively responsible for keeping the Earth's surface temperate over its history, right up to the present.

c) Increasing the absorption in the atmosphere of outgoing radiation from the Earth's surface increases the surface temperature of the Earth. This is the original greenhouse effect, misnamed perhaps, but nevertheless the physics of which is well established both theoretically and experimentally.

These facts led to the conclusion that the increasing CO_2 concentration in our atmosphere should lead to further warming of the Earth's surface.

The key question is how much?

Was Svante Arrhenius correct in estimating 3°C of warming if the CO_2 concentration hits 450 ppm, which, at current rates, will occur by 2050?

And even if this were true, what impact would a 3°C rise have on the globe?

These two questions have driven much of the internationally coordinated climate research program that has been carried out over the past sixty years.

They have also led to a furor of debate among the public and governments about how to respond to this research since Keeling made his first measurements in 1960.

They are not easy questions to give precise answers to.

The general principles are clear and unambiguous, but like much of science, when you want to go beyond a rough understanding to get things right at the percent level, the devil is in the details.

Even in 1900, after Arrhenius first presented his predictions, there was almost an immediately significant academic backlash.

The most forceful response came from Knut Ångström, a physics colleague at Uppsala University. Ångström's name carried weight, both due to his genealogy and his scientific expertise.

His father was a famous Swedish physicist, after whom the unit of atomic spacing, (10^{-10} m) is named.

More important, his area of expertise was in the measurement of solar radiation.

Ångström based his objections both on theoretical and experimental grounds.

He argued that the absorption of infrared radiation was already so great—'saturated' is the technical term—that increasing the abundance of any greenhouse gas, be it water or CO_2, would make little difference.

After all, if you are designing a sunshade, and a sheet of wood a millimeter thick already blocks all the sunlight, you don't gain anything by machining the wood any thicker.

Ångström's specific arguments were twofold.

First, he had observed the frequency range of absorption by water vapor, and when he measured the same thing for CO_2, he found that the absorption frequency bands appeared to essentially overlap.

Since water vapor is far more abundant in the atmosphere, and is also a stronger greenhouse gas, it essentially completely blocks certain wavelengths of IR radiation in the atmosphere.

What then is to be gained by increasing the CO_2 abundance?

Second, he added experimental insult to theoretical injury.

In 1900 his laboratory assistant, John Koch, performed an experiment with a tube of pure CO_2 gas 30 cm long.

This, he claimed, would have contained the same amount of CO_2 as a column of air from the ground up to the top of the atmosphere (remember that even now CO_2 is only 410 parts per million of the total atmosphere).

Koch found that when he reduced the amount of CO_2 in the tube by a factor of three, there was no perceptible change in the percentage of IR radiation absorbed by the tube.

These results both turned out to be wrong yet nevertheless carried great weight with the scientific community at the time.

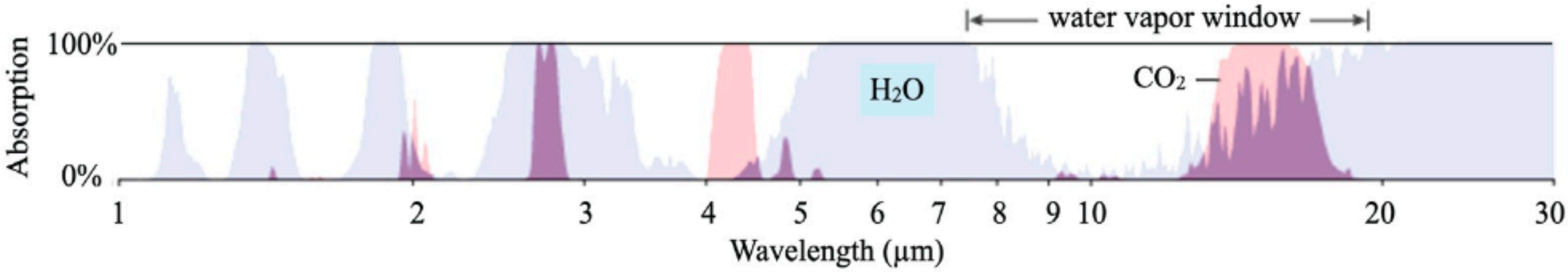
Almost fifty years passed before Arrhenius's ideas resurfaced and were vindicated.

Let's deal with the H_2O/CO_2 question first.

Precise measurements of the spectrum of infrared absorption by H_2O, which were simply not possible at the time Ångström made measurements, show that what were interpreted at the time as wide bands of H_2O absorption overlapping with CO_2 are in fact a set of narrow spikes separated by valleys.

In the following illustration of the absorption spectra (Figure 6.1), you can see that at wavelengths greater than ten micrometers, the CO_2 absorption overlaps with water vapor absorption, and the CO_2 absorption essentially fills in the gaps between the H_2O peaks:

Figure 6.1



This effect gets even more pronounced at high altitudes, where, as I will remind you, the radiation of heat energy from the atmosphere into space occurs.

At these altitudes, where the density of air is thin, there are fewer collisions between molecules.

Collisions broaden the absorption lines of compounds, so at high altitude the peaks become narrower and the valleys deeper.

All of this implies that even in more humid air, where IR absorption in various frequency ranges by H_2O may be saturated, there is still plenty of frequency real estate where CO_2 enhances absorption.

And at high altitudes not only do the H_2O absorption lines become narrower, the H_2O abundance itself also falls off to almost zero, whereas CO_2 is well mixed throughout the atmosphere, meaning its absorption becomes even more important at these altitudes.

Now, what about Koch's experiment, and in particular the claim that once CO_2 absorption is saturated, changing the amount of CO_2 won't further increase the absorption?

It was incorrect on two counts, but in any case it turns out to be largely irrelevant to the process that drives what has become known as "radiative forcing"—the central parameter that today enters into estimates of global climate change.

The experimental problems with Koch's experiment involved the source of radiation and the length of the tube used.

Koch used as his source of radiation essentially boiling water at a temperature of 100°C, well above that of either the Earth or the atmosphere, so the spectrum of the radiation being absorbed would be different.

He also used a single 30 cm long tube, and the way he searched for different absorption as a function of CO_2 abundance was to change the CO_2 pressure in the tube.

As it happens, the 30 cm tube also underestimated the actual amount of CO_2 that radiation would traverse in the atmosphere—by a factor of five.

Countering this is the fact that using CO_2 at ground level atmospheric pressure broadens the absorption lines of CO_2 compared to absorption rates through much of our much lower-pressure atmosphere.

Also, using pure CO_2 further broadens the CO_2 absorption lines compared to what would occur in an atmosphere in which trace amounts of CO_2 are mixed with N_2 or O_2.

Putting these two competing factors together implies that the appropriate length of tube would have been about one meter to recreate the absorption in our atmosphere.

But neither of these factors would have significantly affected what should have been the correct interpretation of the experiment.

The faulty conclusion lied less in execution than in the two assumptions that went into the experimental design.

What Ångström didn't take into account in his notion of saturation was the actual frequency dependence of CO_2 absorption, and the fact that absorption is not an all-or-nothing process.

At any frequency, the fraction of transmitted light falls exponentially with the number of CO_2 molecules encountered on the trajectory through the atmosphere.

Effectively, saturation occurs when some large fraction of the incoming light is absorbed, say 90 percent.

This value can be reached either by a large absorption coefficient of CO_2 at the frequency in question or by increasing the number of CO_2 molecules encountered on the trajectory of light

Next, the frequency dependence of the CO_2 absorption affects saturation.

We can approximate a broad peak in absorption spectrum for CO_2 in some frequency band as looking like Figure 6.2.

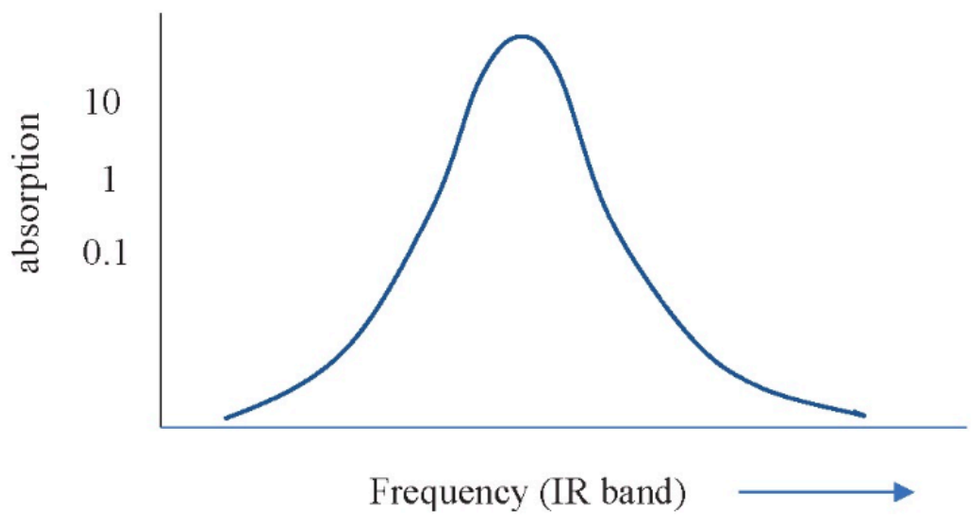


Figure 6.2

Say we consider absorption for all frequencies for which the absorption coefficient is greater than two as being sufficiently saturated so almost all the radiation is absorbed (Figure 6.3).

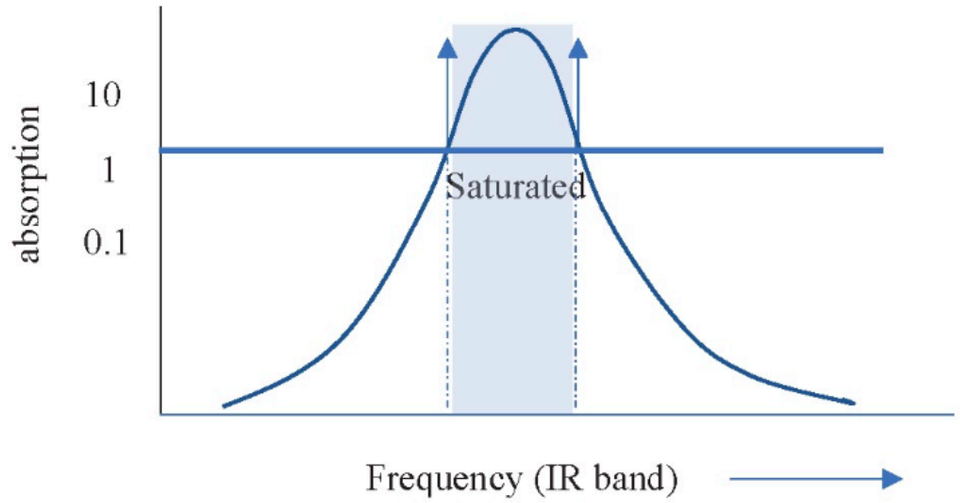


Figure 6.3

Now, if we increase the CO_2 concentration by a factor of, say, four, then all frequencies for which the absorption coefficient is now greater than a value four times smaller, or 1/2, will be saturated (Figure 6.4).

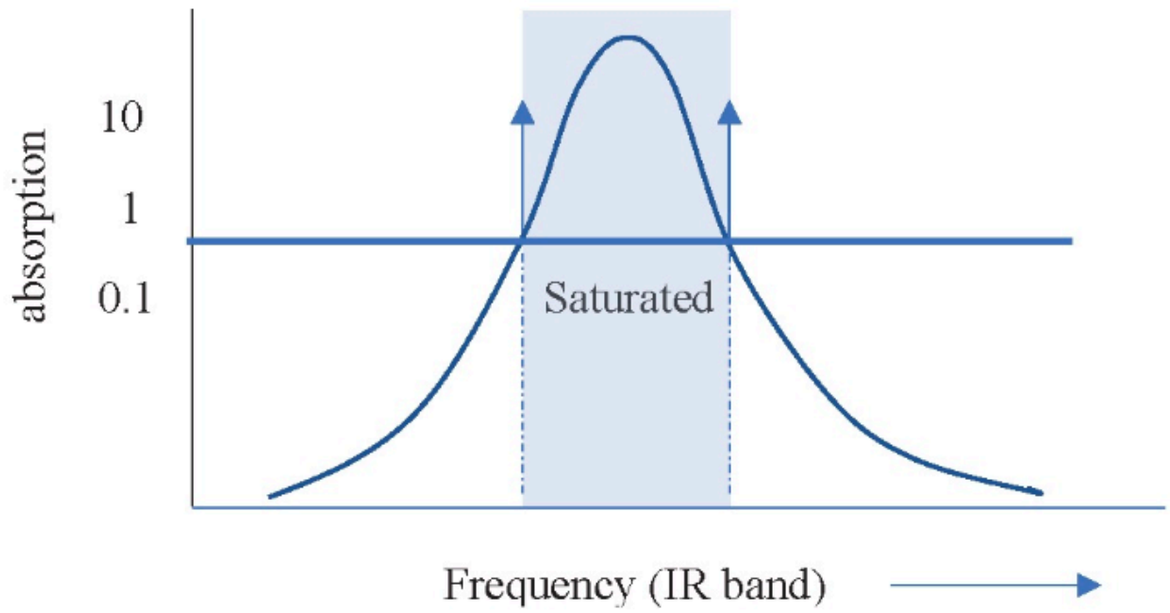


Figure 6.4

What increasing the CO_2 concentration does, therefore, is increase the *range* of frequencies that can have saturated absorption, and this increases the total rate of IR absorption by CO_2 in the atmosphere.

So Ångström's concern about saturation was a red herring.

Absorption can increase with increasing CO_2 concentration *even if absorption at some frequencies is already* saturated.

What is the effect of this increase?

For regions of the atmosphere where the CO_2 density might be high enough, so that effectively CO_2 absorption would be saturated at all frequencies, there wouldn't be any effect.

In fact, modern measurements have since shown there aren't such regions.

However, as one moves up higher in the atmosphere, the atmosphere gets less dense, and with that density reduction, the number of CO_2 molecules encountered by outgoing radiation falls off.

At some point, the atmosphere will become transparent to upward-traveling radiation, and that is the point where the radiation will be emitted out into space.

The key effect then is this: if there is more CO_2 in the atmosphere, the point at which the atmosphere becomes transparent to thermal radiation will move to a progressively higher altitude.

But since the temperature at higher altitudes is colder, the thermal radiation emission power at this altitude will decrease (recall that the emission rate goes as the fourth power of the temperature at the point the atmosphere becomes transparent).

Less radiation emitted outward means there will be an imbalance in outgoing radiation versus incoming radiation, and Earth will heat up.

Eventually, a new temperature equilibrium will be established.

Thus, Ångström's concern about the possible saturation of absorption by CO_2 was misplaced, independent of the fact that the details were also incorrect.

What is important is that the atmosphere is not a single slab, but that the temperature and density vary with altitude.

My use of the term "blanket" earlier was more than merely poetic.

The temperature at the outer surface of an insulating blanket is colder than at the inside because of the insulating material.

Thus, the blanket radiates away less of your body heat.

The inner surface radiates back downward at a higher temperature, keeping you warm.

That is precisely what happens with the atmosphere.

The major difference between this analogy and the actual process for the atmosphere is that for the blanket, the source of heat resides inside the blanket (your body), while for the atmosphere, the source of heat is external—the sun, whose rays make it successfully downward through the atmosphere to heat up the Earth and allow it to radiate upward, like your body, into the atmospheric blanket.

But, of course, the other major difference is the fact that the blanket traps the air inside, like a greenhouse does, so convection and conduction cannot play a role as they do on Earth.

Because this aspect of CO_2 absorption is both qualitatively and quantitatively important, it is worth putting some meat on the above discussion by showing actual data and the results of some detailed atmospheric modeling predictions.

Shown here are two coincident measurements of infrared radiation emission by the atmosphere in a cloud-free evening at the arctic ice sheet, compiled by researchers at the University of Wisconsin.

The upper image is the infrared emission received by a detector located 20 km above the ice sheet, looking *downward*, while the lower image displays the radiation spectrum received by a detector on the ground looking *upward*.

To guide the eye, superimposed on the actual data are dashed lines that display the radiation spectrum one would predict for a perfect black body radiating at the labeled temperatures (Figure 6.5).

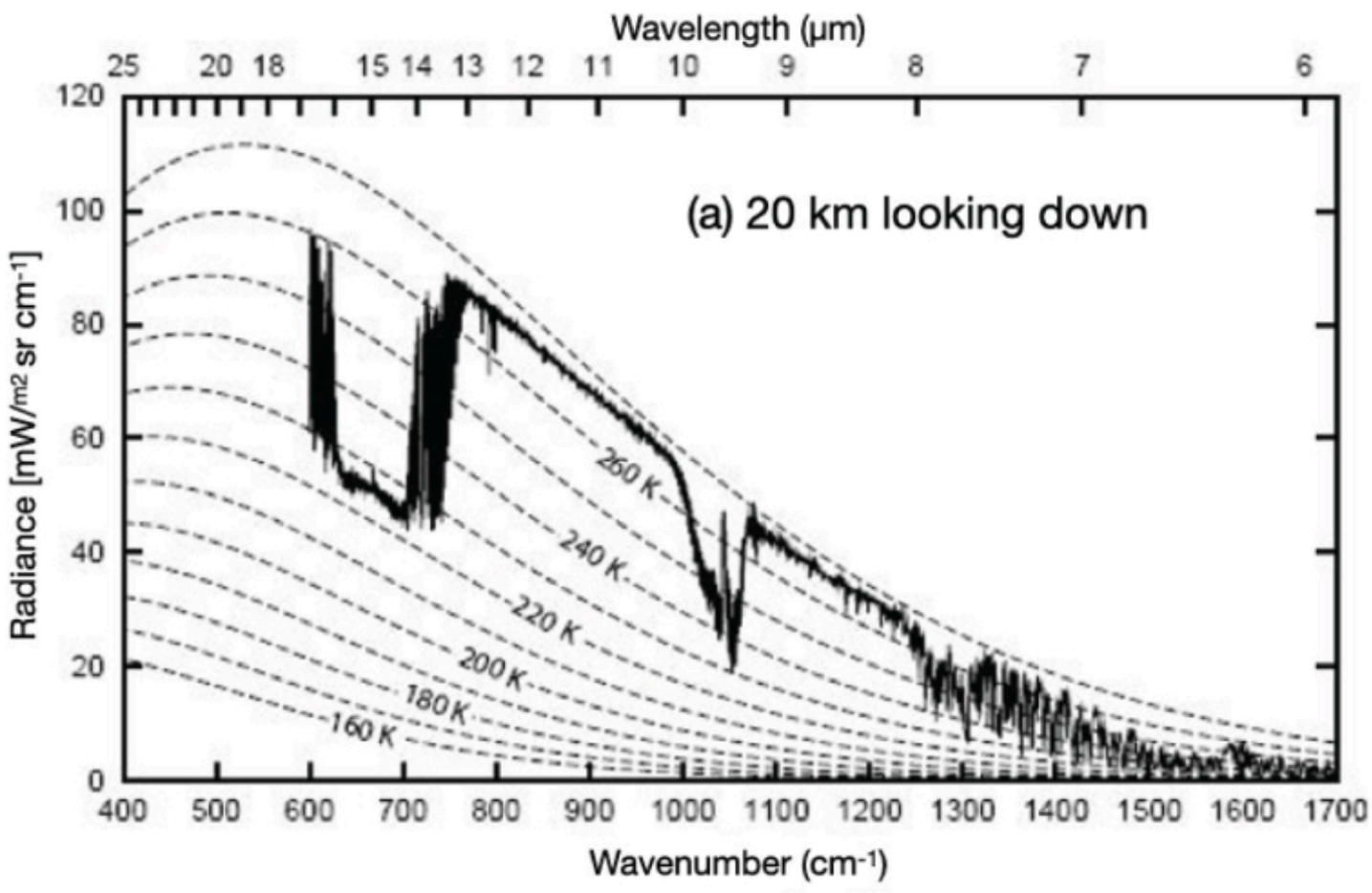
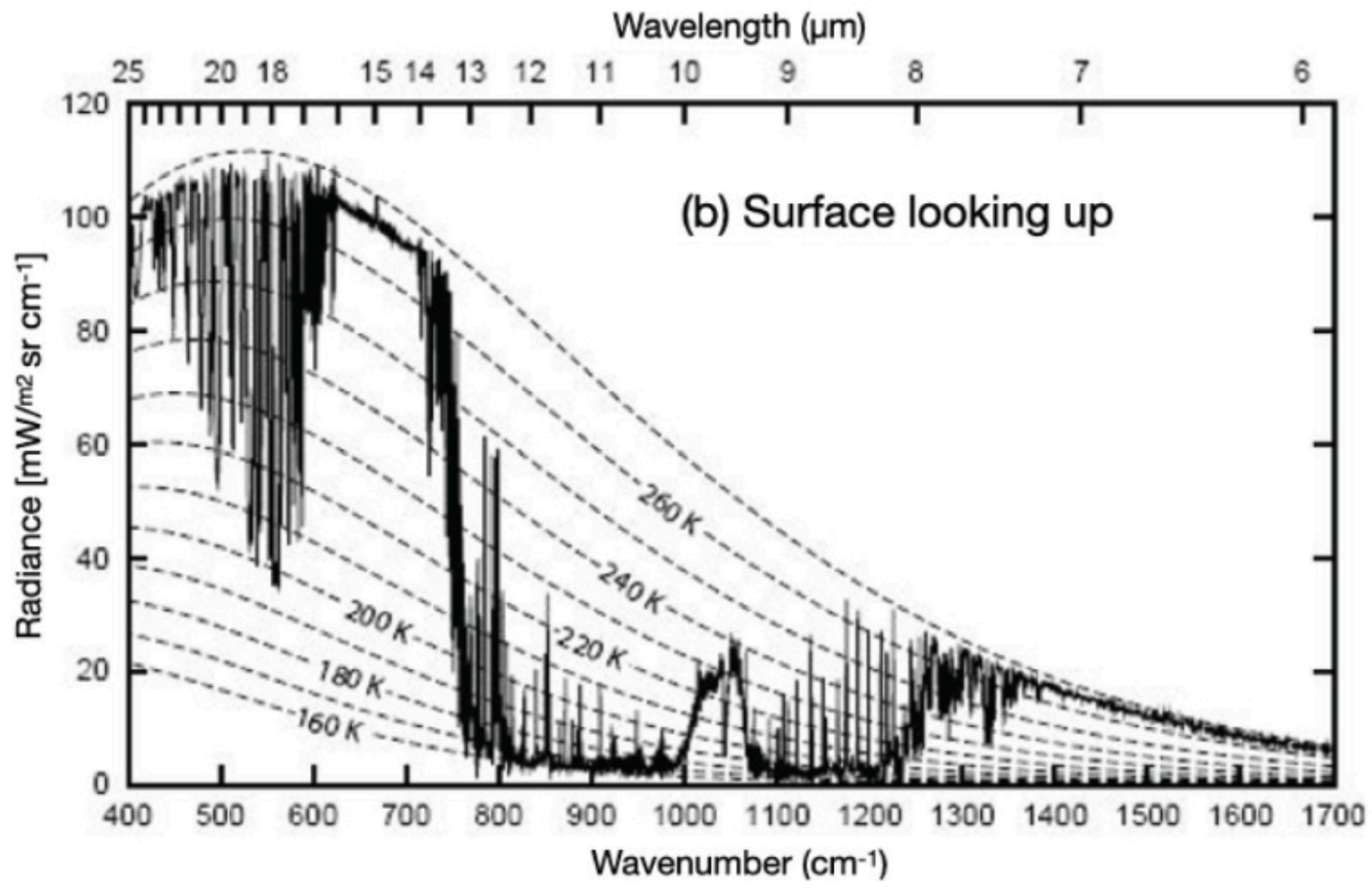

Wavelength (μm)
25 20 18 15 14 13 12 11 10 9 8 7 6
120
100
80
60
40
20
0
Radiance [mW/m2 sr cm-1]
(a) 20 km looking down
260 K
240 K
220 K
200 K
180 K
160 K
400 500 600 700 800 900 1000 1100 1200 1300 1400 1500 1600 1700
Wavenumber (cm-1)

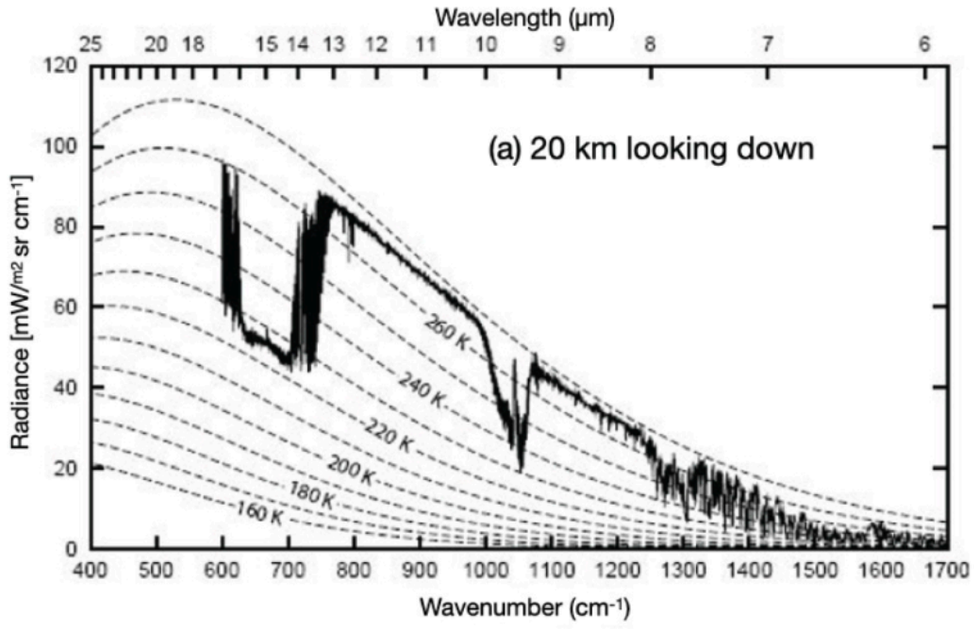
Figure 6.5


Wavelength (μm)
25 20 18 15 14 13 12 11 10 9 8 7 6
120
100
80
60
40
20
0
Radiance [mW/m2 sr cm-1]
(b) Surface looking up
260 K
240 K
220 K
200 K
180 K
160 K
400 500 600 700 800 900 1000 1100 1200 1300 1400 1500 1600 1700
Wavenumber (cm-1)

What the detector looking downward observes in the frequency band between about 600–700 (in the units used here, which are 1 over the wavelength of the radiation, in cm) is a dip in the spectrum.

This is due to absorption by CO_2 in the atmosphere of the radiation that would otherwise come from the surface of the Earth, which in the temperature units here is about 270° Kelvin, or just below freezing.

Notice that above this dip, where the atmosphere is largely transparent to infrared radiation, the emission curve follows precisely the radiation pattern one would expect from a hot body at the temperature of the Earth's surface.



In the dip, however, the radiation emission is characteristic of a body *at a much colder temperature*, around 225°K, or about -46°C (-51°F).

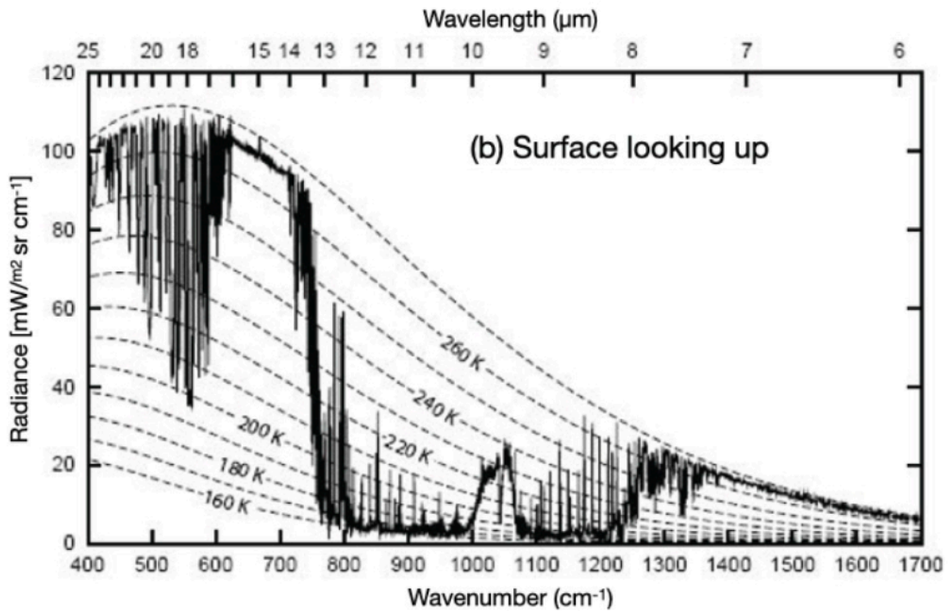
This is because this radiation is coming from a higher altitude, at the point where the atmosphere becomes transparent to upward-going infrared radiation.

Because the characteristic temperature at this altitude is much colder, the intensity of radiation emitted in this frequency range is much smaller. In other words, the radiative power emitted by Earth in this frequency band is much less than it would be if the atmosphere were transparent in this range.

This reduction in emission causes the blanketing effect I described.

Radiation in this region is trapped and reradiated down toward the Earth's surface, keeping it warm, and not out into space.

To demonstrate that this is the case, look at the lower curve.



Radiated downward toward the surface in this same frequency band, there is much more radiation received than received above the band, where the atmosphere is transparent to the infrared.

In this band the atmosphere is opaque to infrared radiation, so the radiation that makes it down to the Earth emanates from a region much closer to the surface, which is at a much higher temperature (like the inside of the blanket in my blanket analogy).

This keeps the surface warmer than it would otherwise be.

The data therefore validates the theoretical prediction.

Now to the models, which themselves are empirically based.

Using the best available laboratory data on the infrared absorption properties of CO_2 as a function of frequency, numerical codes incorporate this data with models of the atmosphere to predict radiative emission from the Earth.

Comparing the predictions and the data tells us how good the models are likely to be.

One can do more with a good model, however.

One can obtain a prediction for what might be expected for a hypothetical world with a CO_2 concentration that differs from the present concentration.

Figure 6.6 shows a prediction for what a hypothetical satellite located 20 km above a warmer spot (corresponding to the Earth's surface average temperature of about 15°C or about 60°F) at the Earth's surface, looking downward.

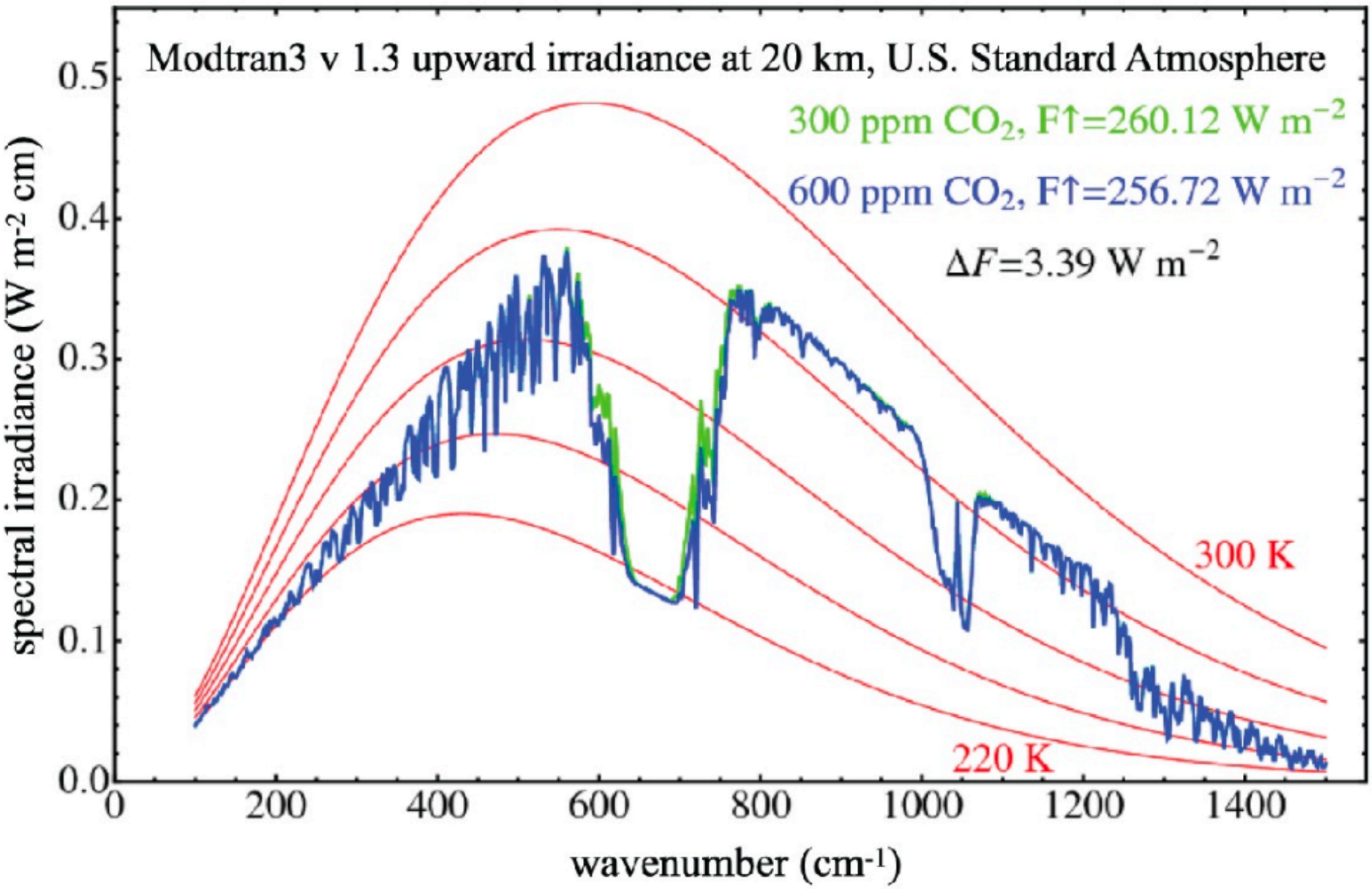
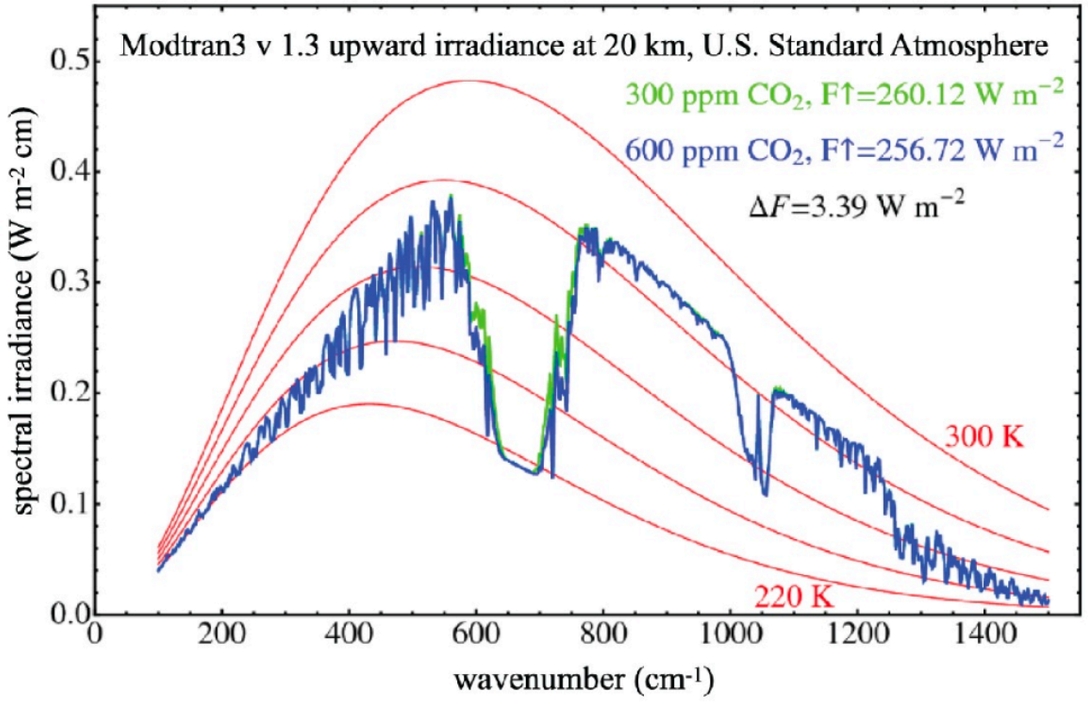


Figure 6.6

The lighter curve represents the prediction for a CO_2 abundance of about 300 ppm, just at the time when Arrhenius and Ångström had their debate in 1900.

The darker curve shows the prediction for a CO_2 abundance of double that value.

Notice that as predicted in our earlier discussion, the wings of the dip in the emission spectrum are wider for the higher CO_2 abundance, resulting in increased CO_2 absorption in the atmosphere.



As a result, the atmosphere radiates into space at a slightly higher and colder elevation.

The effect is less dramatic than that showed in my earlier sketches describing the impact of saturation, however.

Indeed, the difference in radiative power emitted upward for a CO_2 abundance of 600 ppm is only about 1 percent less than that emitted for 300 ppm.

The difference, about 3.4 watts per square meter, represents a decrease in power radiated away by the Earth compared to the power received by the sun.

This decrease in emission will ultimately heat both the ground and the atmosphere until both achieve new higher temperatures.

This will emit more radiation back out into space and a new equilibrium balance will once again be obtained.

This is shown schematically in Figure 6.7 showing temperature as a function of altitude, and the intensity and altitude at which the atmosphere is radiating into space.

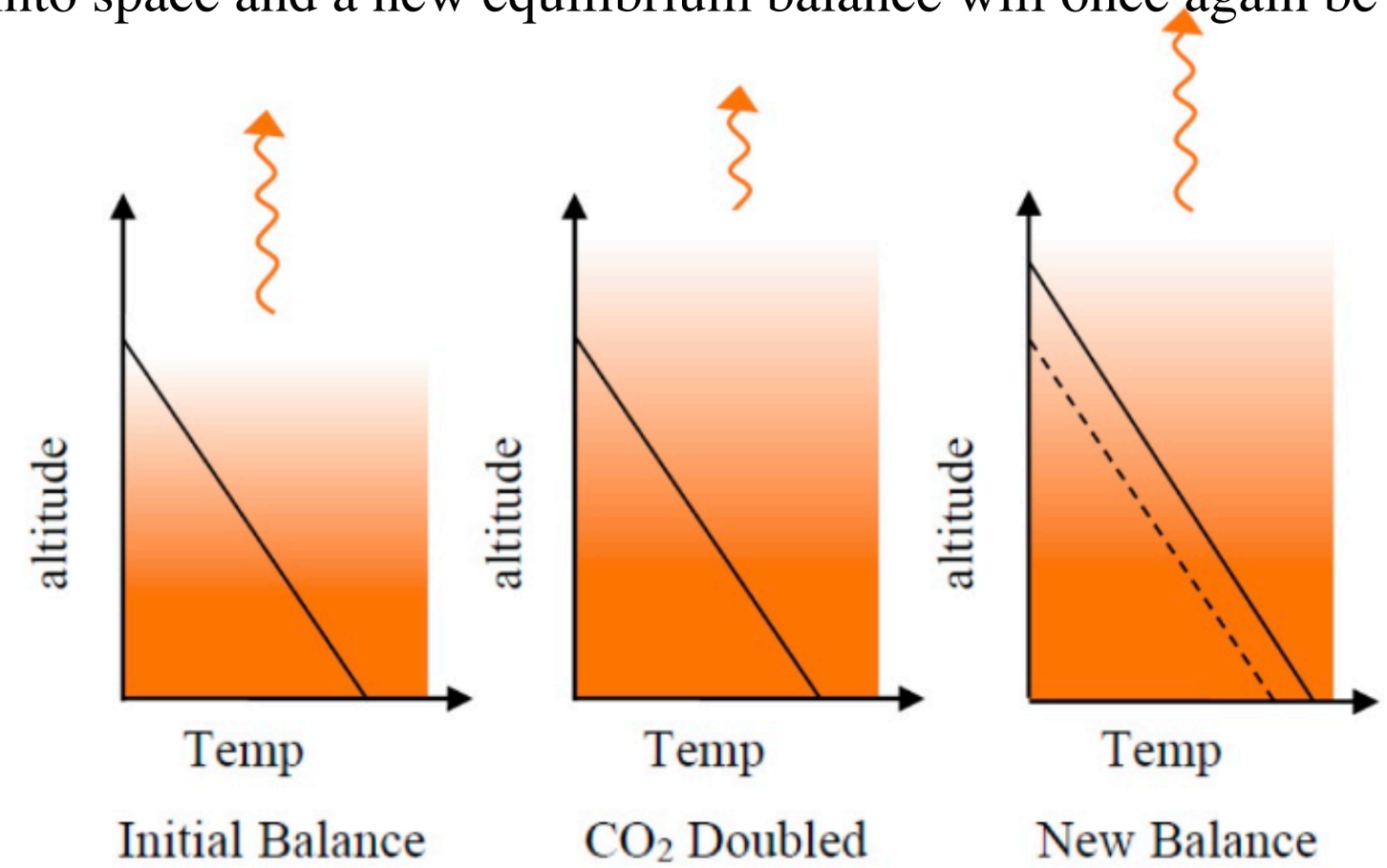


Figure 6.7

When the CO_2 concentration is doubled, the altitude of emission is increased, and the intensity of emission (length of the wavy line) is decreased.

Ultimately, a new energy balance is reached with the outgoing emission being the same as it was with the original CO_2 concentration.

In the process, the temperature of both Earth and atmosphere is increased as a function of altitude (solid vs. dashed line).

This 1 percent difference of 3.4 W/m^2 in emitted radiation due to increased CO_2 atmospheric concentration corresponds to the additional amount the atmosphere will radiate back down toward Earth, in what has become known as "**radiative forcing**."

Calculating radiative forcing has become the standard tool for quantifying the changing IR absorption in the atmosphere and its effect on changing the energy balance between incoming radiation from the sun and outgoing radiation from Earth.

A positive value of radiative forcing means less energy escapes from the atmosphere to space, and as a result there is a positive imbalance in incoming solar radiation versus outgoing IR radiation, which will cause the Earth's surface temperature to rise.

Radiative forcing is the most important single quantity parameterizing the response of Earth's climate to the presence of humanity.

Essentially all of the predicted impact on Earth's climate due to an increase in CO_2 and other greenhouse gas abundances comes down to determining a detailed, precise, and accurate estimate of forcing, expressed in terms of additional heat power deposited in the atmosphere-Earth system compared to that radiated to space, in Watts/m^2.

Before turning to our modern understanding of this quantity and its expected impact on Earth's temperature and climate, it is interesting to return to Ångström and Koch's original experimental claim that changing CO_2 abundance wouldn't affect infrared absorption.

Ironically, Koch's 1900 experimental result turns out to have been roughly consistent with the picture I have presented here.

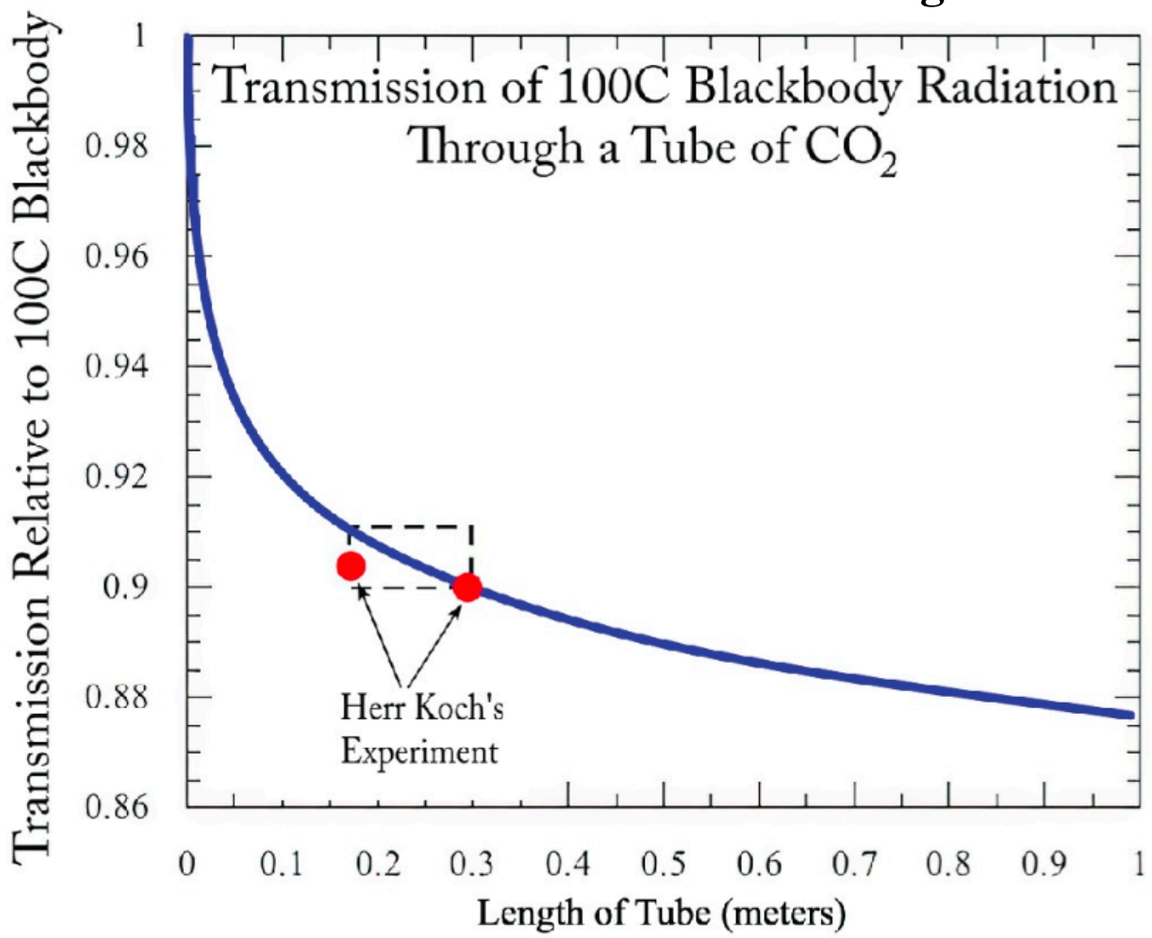
We have seen that doubling CO_2 abundance is expected to produce slightly more than a 1 percent change in the infrared heat power absorbed by CO_2 in the atmosphere.

Koch varied the net CO_2 abundance in his apparatus by only a factor of 1/3.

A recent reanalysis of the predicted change in CO_2 absorption for his configuration versus the actual values he determined is shown in Figure 6.8, for an equivalent experiment where instead of reducing the CO_2 pressure, the length of the CO_2-filled tube traversed by the IR radiation is reduced from 30 to 20 cm.

As can be seen, the absorption measured by Koch for the 30 cm tube is in good agreement with the predicted absorption in CO_2.

Figure 6.8



Moreover, the predicted change in absorption when the length of the tube is reduced to 20 cm (or the pressure is reduced by 2/3) is less than 1 percent.

Koch reported an increase in transmission for the lower pressure of “not more than 0.4 percent.”

If he overestimated the accuracy of his claim by a small amount, this would make his lower datapoint consistent with the correct value.

Since his boss was convinced there would be no effect, observing a change that might have been as large as 1 percent was, for Koch, more likely to be consistent with no change at all.

The moral of this story is that laboratory assistants have to be very careful not to be influenced by the force of their bosses’ biases!

7 - BELIEVE IT OR NOT

The sign of a good theory is when models based on it get more detailed and more precise theoretical predictions based on the improved models then agree better with the data.

This is the case with the original greenhouse effect, whose existence and theoretical interpretation are by now essentially unimpeachable.

The physical principles behind the calculation of radiative forcing—the chief driver of climate change—are the same as those that underlie the two-hundred-year-old explanation of why the Earth's surface is 33°C warmer than it would be without its current atmosphere.

When discussions turn to the issue of global warming, it is easy to lose the forest for the trees.

Debates can focus on the detailed model assumptions made to calculate radiative forcing terms and their subsequent impact on climate.

Each successive complexity that is included, from the detailed structure of the atmosphere, both vertically and across the Earth's surface, to the role of convection, the importance of clouds, ocean mixing and currents, pollution, Earth's albedo, and so on, is sometimes portrayed as a possible new Achilles heel.

But what should be the most striking takeaway, beyond the fact that ever-more-detailed models generally fit the data better, is the fact that the basic physics captured in even the simplest models gives answers that are consistent with the data.

This suggests that they capture the fundamental physics, which is sound.

One could, however, argue that this is somehow an accident—that a conspiracy of competing alternate causes and effects ends up yielding a fortuitous agreement with measurements so our interpretation of the current agreement is both qualitatively and quantitatively wrong.

But this seems increasingly without foundation.

The evidence, both theoretical and experimental, that I am aware of suggests that details now being debated among experts in climate modeling will not dramatically alter our understanding of the key scientific principles driving climate change in the current era.

I liken the situation to the design of ultra-performance sports cars and the physics of internal combustion engines.

The fundamental physics behind internal combustion engines is straightforward.

Using this, one can understand roughly how much fuel is needed to make an internal combustion engine with a certain number of cylinders to power a car of a known weight up to a given speed.

This is something we could work out in principle in an undergraduate physics class.

At the same time, huge amounts of money and time are spent by the world's best engineers to develop more efficient and faster cars, planes, trains, and so on.

Complex modeling is necessary to tweak automobile performance so such things as turbulence within the engine and the air surrounding a speeding vehicle, expansion and feedback between different parts of the engine mechanism and the vehicle chassis, and other detailed interactions are taken into account.

The modeling is so complex that it would be impossible to explain it accurately without explicitly going through all these intricacies.

But in spite of the fact that these details and more may be necessary to optimize the behavior of the vehicle, it is irrational to suppose that a better understanding of these details is likely to demonstrate that the fundamental picture, including the original physics of internal combustion engines, needs to be revised.

In any area of physics where the fundamental principles are well established, and many independent measurements confirm the same picture, it is improbable that more sophisticated models, which may nevertheless illuminate new features, will force us to throw out the baby with the bathwater.

Let's return then to Arrhenius, who, while having ingeniously accessed rough and incomplete data, built a model based on well-understood radiation physics and derived a general "rule" for associating CO_2 abundance in the atmosphere with the global surface temperature of Earth.

His rule, based on the available data and the physics at hand, implied a "logarithmic" relationship between Earth's surface temperature and the atmospheric CO_2 concentration.

As I have described, this is the more modern way of framing his statement that the surface temperature will vary by regular linear increments if the CO_2 concentration is increased by constant multiplicative factors.

Over a hundred years later, using a much more sophisticated understanding of the theory behind infrared absorption and the best data on CO_2 spectral absorption, as well as new numerical models of the atmosphere, the functional relationship that climate modelers derive is essentially the same, up to a proportionality factor that is about two times smaller.

That Arrhenius got that close using very rudimentary data may seem surprising.

But since the fundamental physics associated with thermal radiation has not changed significantly, perhaps it is not as unexpected as it might otherwise seem.

Given that absorption by CO_2 directly affects the radiation balance at the top of the atmosphere, it is simpler and more direct to derive a functional relationship not between CO_2 abundance and temperature, as Arrhenius attempted, but rather between CO_2 abundance and radiative forcing.

That is how things are generally expressed today.

Recall that radiative forcing is defined as the change in total radiation power (in watts/m^2) emitted by Earth into space as some quantity, like the CO_2 abundance, is changed.

Once one determines the radiative forcing terms, one can then use atmospheric and climate models to determine the predicted change in temperature at the Earth's surface, as well as other climate changes.

The sophisticated computer models that calculate radiative transfer through many layers of atmosphere from the ground up to produce the figures earlier in our discussion, which agreed well with observations, yield the following relationship between CO_2 and radiative forcing:

$$\text{Radiative Forcing } (\Delta P) = 5.35\ (\pm 0.5)\ \ln(C/C_0)\ \text{W/m}^2$$

Since it is fundamental to the basis of modern climate science, I felt it useful to present it explicitly.

Here, the radiative forcing (change in power radiated into space) is with a CO_2 concentration, C, calculated compared to a baseline value, which by convention is the value in 1750, at the beginning of the industrial revolution, when $C_0 \approx 278$ ppm.

If one were to take C_0 = 300 ppm, the value appropriate to the time when Arrhenius made his predictions, then the range of predicted forcing when C = 600 ppm would be 3.36–4.05 W/m^2.

How does this compare with Arrhenius's prediction?

To answer this question, we need to be able to convert radiative forcing into a surface temperature change.

This is not so straightforward, and it is not clear that the same conversion factor would prevail for different CO_2 and temperature values due to a host of other effects that can enter in as temperatures increase, from increased water vapor pressure in the atmosphere to effects of convection, clouds, surface albedo, heat storage in the oceans, and temperature variation across the globe.

Nevertheless, one can make a simple approximate estimate based on the fact that a reduction in power radiated into space leads to that power being stored in radiant heat in the atmosphere.

Assume that this radiant heat makes its way down to the Earth, which then heats up and radiates it back.

Since radiated power goes as the fourth power of temperature, for small changes in the total radiated power one can then estimate that the change in temperature compared to the pre-existing temperature would be approximately 1/4 of the ratio of the radiative forcing compared to the original total radiated power.

Assuming a radiation forcing of 3.7 W/m^2, in the middle of the predicted range, and that the original radiated power from the Earth and atmosphere into space at equilibrium is about 245 W/m^2, then we would find that the predicted change in surface temperature for a doubling of CO_2 would be about 1°C.

Another way of expressing this is that for small changes in radiated power by the Earth, the factor relating radiative forcing in W/m^2 to change in surface temperature in degrees Celsius, using this very simple approximation, is about 0.3.

It turns out that this is a conservative underestimate.

When the other factors mentioned above are incorporated into more sophisticated models, the roughly linear factor relating the predicted change in average surface temperature to the predicted radiative forcing is about 0.7–0.8, depending on the models used.

This suggests the mean temperature change for the surface of Earth, if the CO_2 concentration doubles compared to the 1900 value, would be in the range of 2.5–3.5°C, about 2.5–3 times larger than the naïve prediction based simply on scaling radiative power versus temperature, and about 2 times smaller than the temperature change predicted by Arrhenius.

While the variation between these estimates is not negligible, nevertheless at the order-of-magnitude level, these estimates are all relatively close, predicting a temperature change on the order of 1 percent, and not 0.1 percent or 10 percent of the original surface temperature.

Fortunately, or perhaps unfortunately, we can actually empirically check this estimate with at least one data point.

We can compare the mean surface temperature of the Earth today versus that in 1900, given that the CO_2 abundance in the atmosphere has changed from 300 ppm to about 415 ppm.

Using the modern estimates above, the mean prediction for the temperature change over that period, ignoring other effects for the moment, is about 1.3°C. Figure 7.1 shows the data from climate.gov.

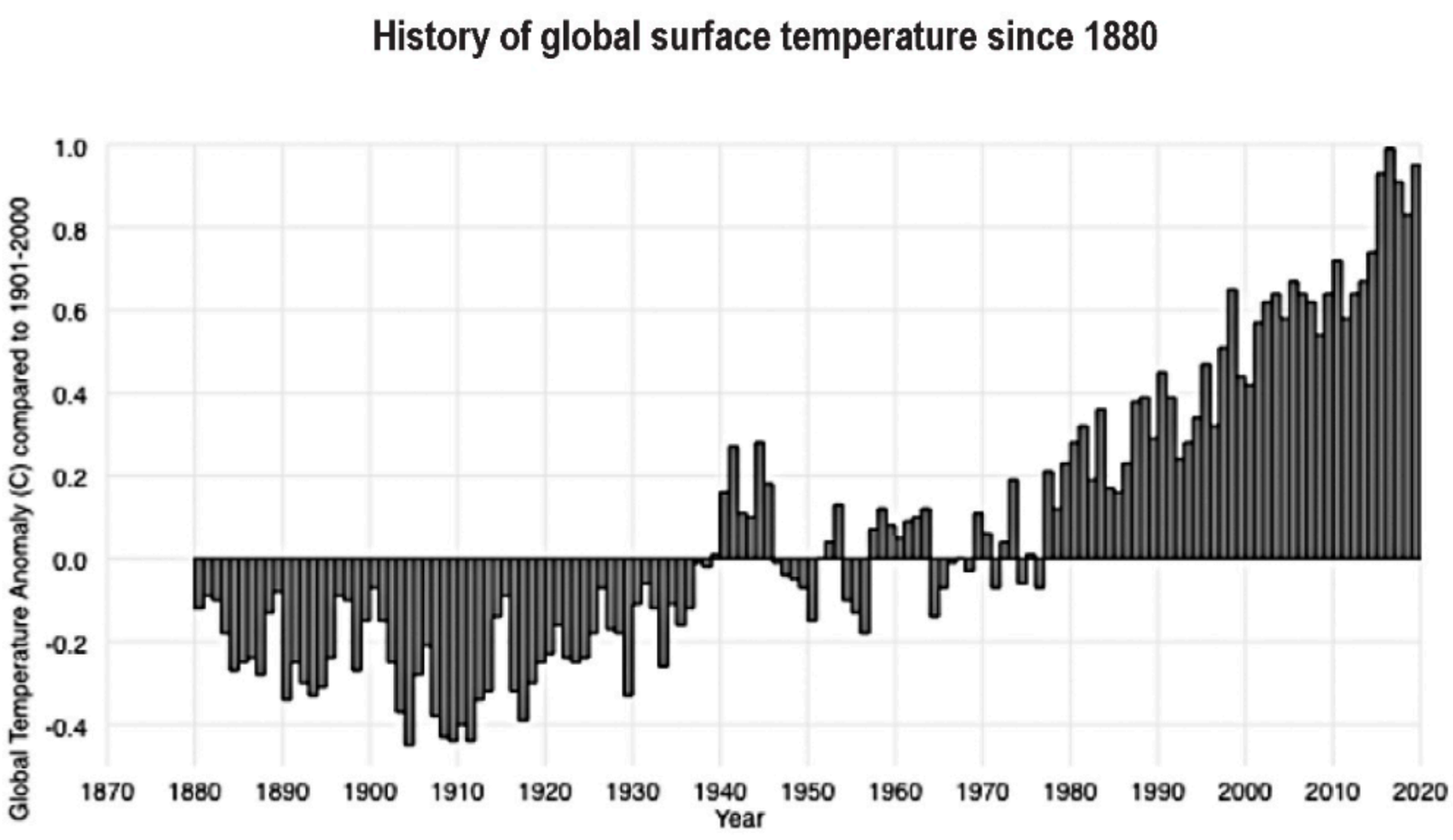


Figure 7.1

As can be seen, the actual change, accounting for the year-to-year fluctuations is in the range 1.15–1.35°C.

That is pretty good agreement with the prediction.

In fact, the agreement is actually better than one might have expected because, as I indicated, the prediction I presented assumes just a change in CO_2 abundance without considering other possible radiative forcing factors present, such as additional greenhouse gases, surface albedo, clouds, solar irradiance, or pollution.

Some of these, such as increased cloud cover or conversion of forests into urban areas, produce a negative forcing, whereas others, including contributions from other greenhouse gases such as methane and other hydrocarbons, produce a positive forcing.

In 2007 the UN Intergovernmental Panel on Climate Change (IPCC) released a chart showing the best estimates and the predicted uncertainties for various radiative forcing factors for 2005 compared to 1750.

The largest relative uncertainties involved the effect of clouds and aerosols in the atmosphere, while the predicted uncertainty associated with CO_2 abundance was relatively small.

When combined, the average predicted radiative forcing at that time was about 1.6 W/m^2.

This is about 5 percent smaller than that predicted for CO_2 alone, confirming that the estimate we obtained by just considering CO_2 was a reasonably good approximation to the full situation.

Alas, humanity has not stood still since 2005.

Greenhouse gas emission has continued to rise.

We remain on track to double the CO_2 concentration since 1750 in this century.

Figure 7.2 is a chart showing the forcing due to greenhouse gases since 1980.

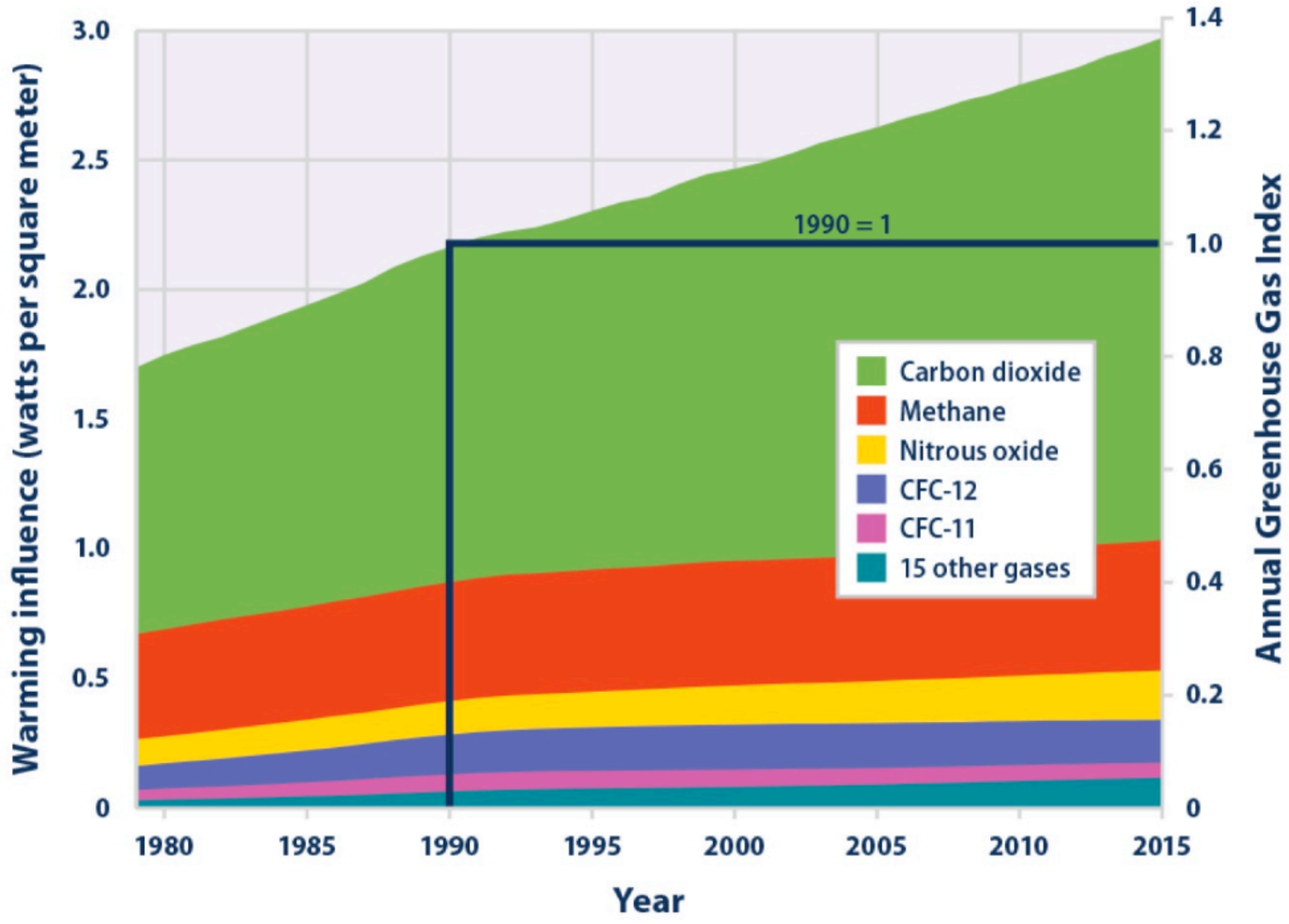


Figure 7.2

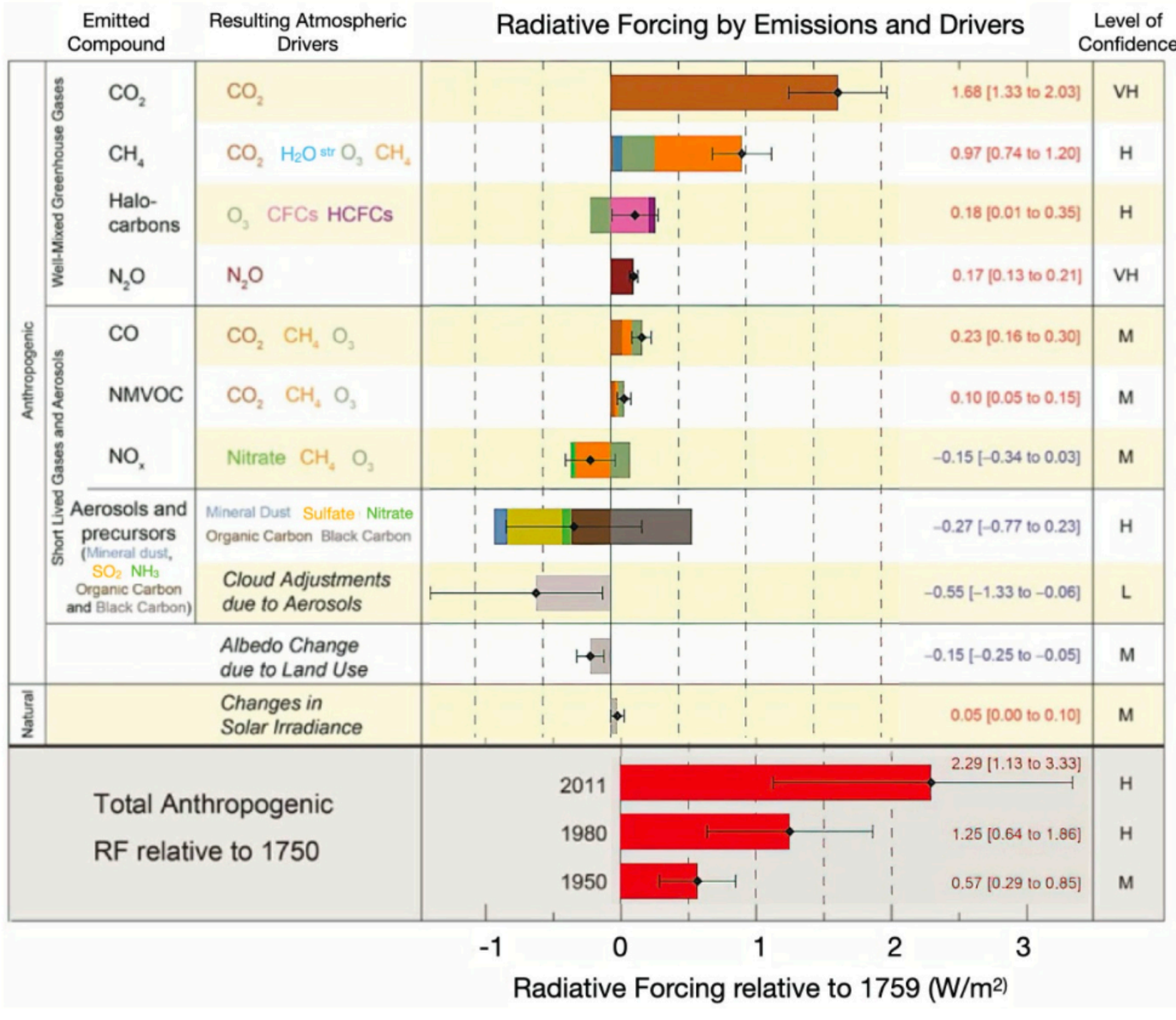
At the same time, uncertainties in various other factors, including impact of aerosols and clouds, have decreased.

In 2013, the IPCC updated their chart, also listing the increase in net anthropogenic radiative forcing over the years.

Thus, the net forcing in 2011 was estimated to have a mean value of 2.29 W/m^2, a 40 percent increase compared to the 2005 estimate, and almost double that estimated from 1980.

The chart is shown in Figure 7.3.

Figure 7.3



For some of you, the the discussion so far may have been something of a slog, so I applaud you for making it this far.

My reason for presenting the fundamental theoretical framework and the experimental tests of that theory in some detail here has been twofold.

First, the basic physics is relatively straightforward and can and should be explained comprehensibly to interested bystanders—which should include basically most of humanity—following to some extent the scientific developments that led scientists to where we are today.

Second, the predictions presented here represent the baseline perspective from which the discussions of the rest of this part of the class should be viewed and from which each of us can ultimately draw conclusions about the possible risks, impacts, and necessity for global action or inaction on this issue.

I therefore felt I owed it to you to give you the most complete and coherent general perspective I could.

The physics is clear.

If it isn’t to you, or still seems sketchy, blame my presentation, not the science.

Before concluding this preparation for what follows,

I will add one more bit of data that helps confirm that estimates of greenhouse-induced temperature change are empirically reliable.

I have discussed essentially one datapoint so far: the predicted change in temperature between 1900 and today versus the measured change.

We can do better than that, however.

Recall that ice cores have allowed us to measure the CO_2 content of the atmosphere going back at least eight hundred thousand years by looking at the composition of the gas bubbles trapped in the ice.

Ice cores allow us to do more than this.

Ice is made of water, H_2O.

However, there are two stable isotopes of both hydrogen and oxygen.

Heavy hydrogen, called deuterium, exists with a ratio of about one part in ten thousand in the oceans.

^{18}O, a heavy isotope of oxygen containing ten neutrons instead of eight neutrons in the oxygen nucleus, makes up about 0.2 percent of the oxygen in the world's oceans on average.

Now, because deuterium and ^{18}O are heavier than the more common versions of hydrogen and oxygen, it takes more energy to evaporate them out of the oceans.

Also, as water vapor in the atmosphere heads toward the poles, the heavier isotopes will preferentially fall when precipitation occurs.

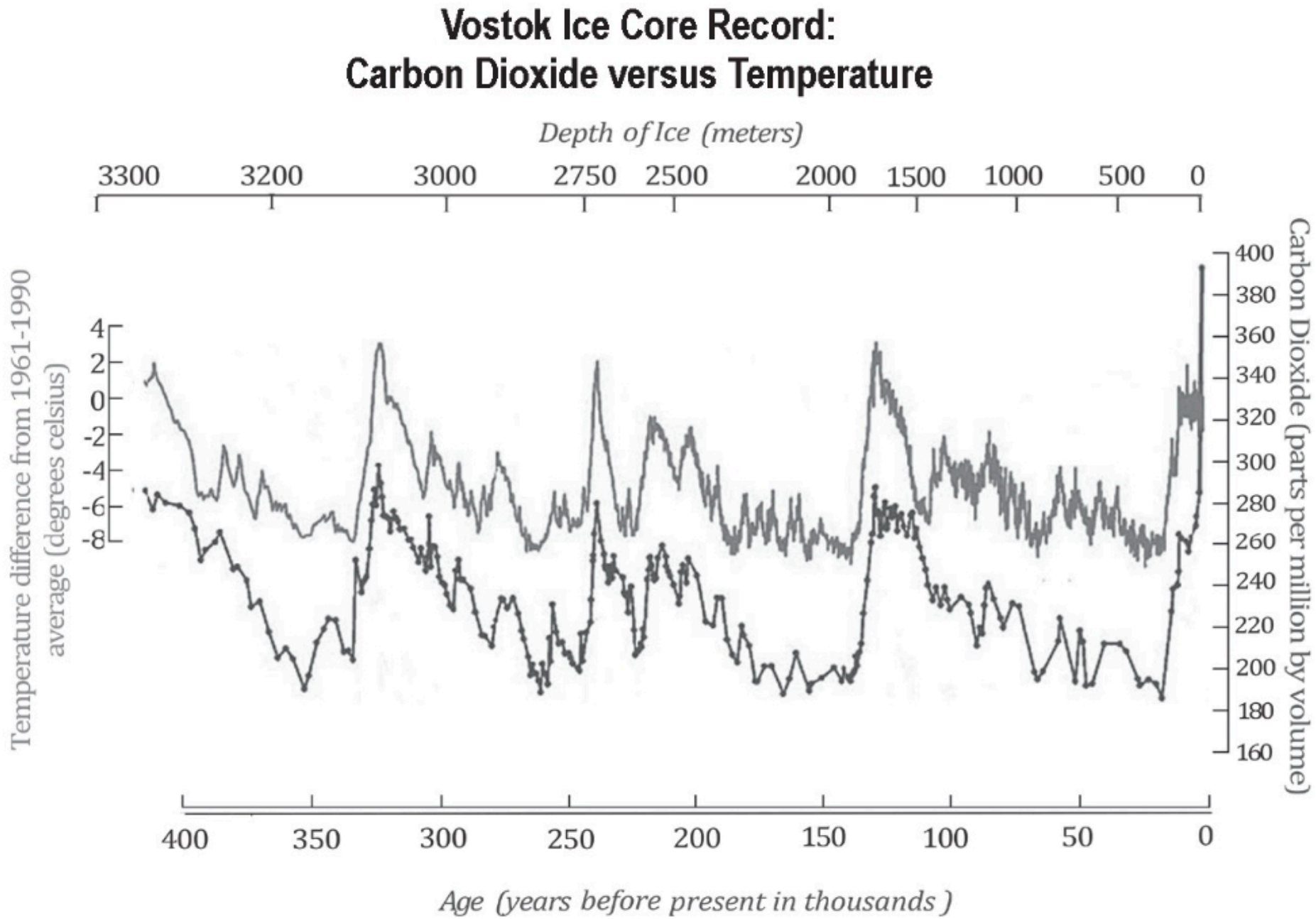
These two factors contribute to what is called "fractionation," and both are temperature dependent, with more fractionation during colder periods.

Thus, if one measures the ratio of deuterium to hydrogen, and ^{18}O to ^{16}O in ice cores, during colder times we will find systematically greater depletion of these heavy isotopes in the ice core samples, and during warmer times systematically less depletion.

Thus, ice cores give us geological record not just of CO_2 abundances, but also atmospheric temperatures!

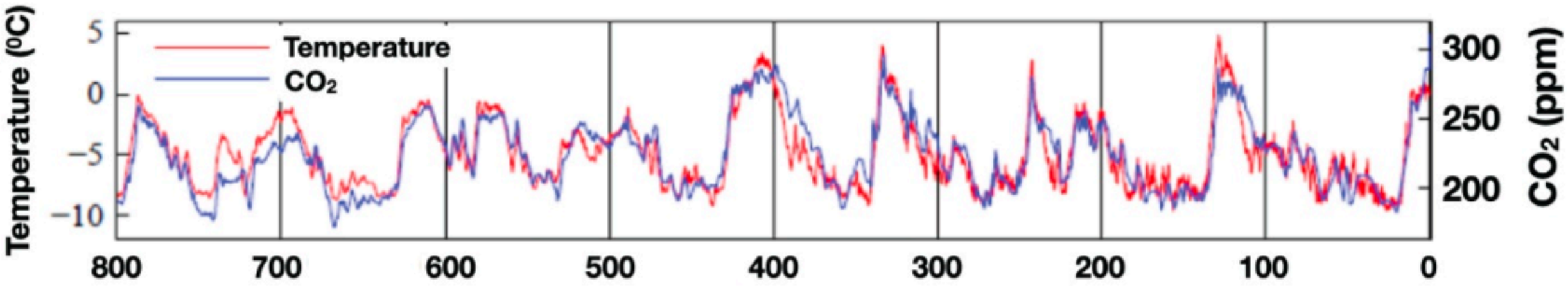
Figure 7.4 is a plot from the Vostok ice core in Antarctica, going back over four hundred thousand years, showing CO_2 abundances and inferred arctic temperatures compared to the average temperature from 1961 to 1990.

Figure 7.4



It is hard not to be struck by the correlation, which continues back at least four hundred thousand more years using data from the Dome C ice core in Antarctica (Figure 7.5).

Figure 7.5



During this period, long-term periodic changes in the Earth's orbit are what initiated the climate change.

But without the feedback due to the radiative forcing induced by the resulting changing CO_2 concentration and the changing surface albedo as ice sheet size changed, the actual temperature changes would have been far smaller.

Using the relation derived earlier for the predicted change in temperature (°C) as ≈ 0.75 times the total radiative forcing, along with the predicted relationship between CO_2 abundance and its contribution to forcing (about one third of the total forcing in this case), Jim Hansen compared the predicted and observed temperature change for the Vostok 440,000-year core (temperature changes divided by two to fit on the plot) in Figure 7.6.

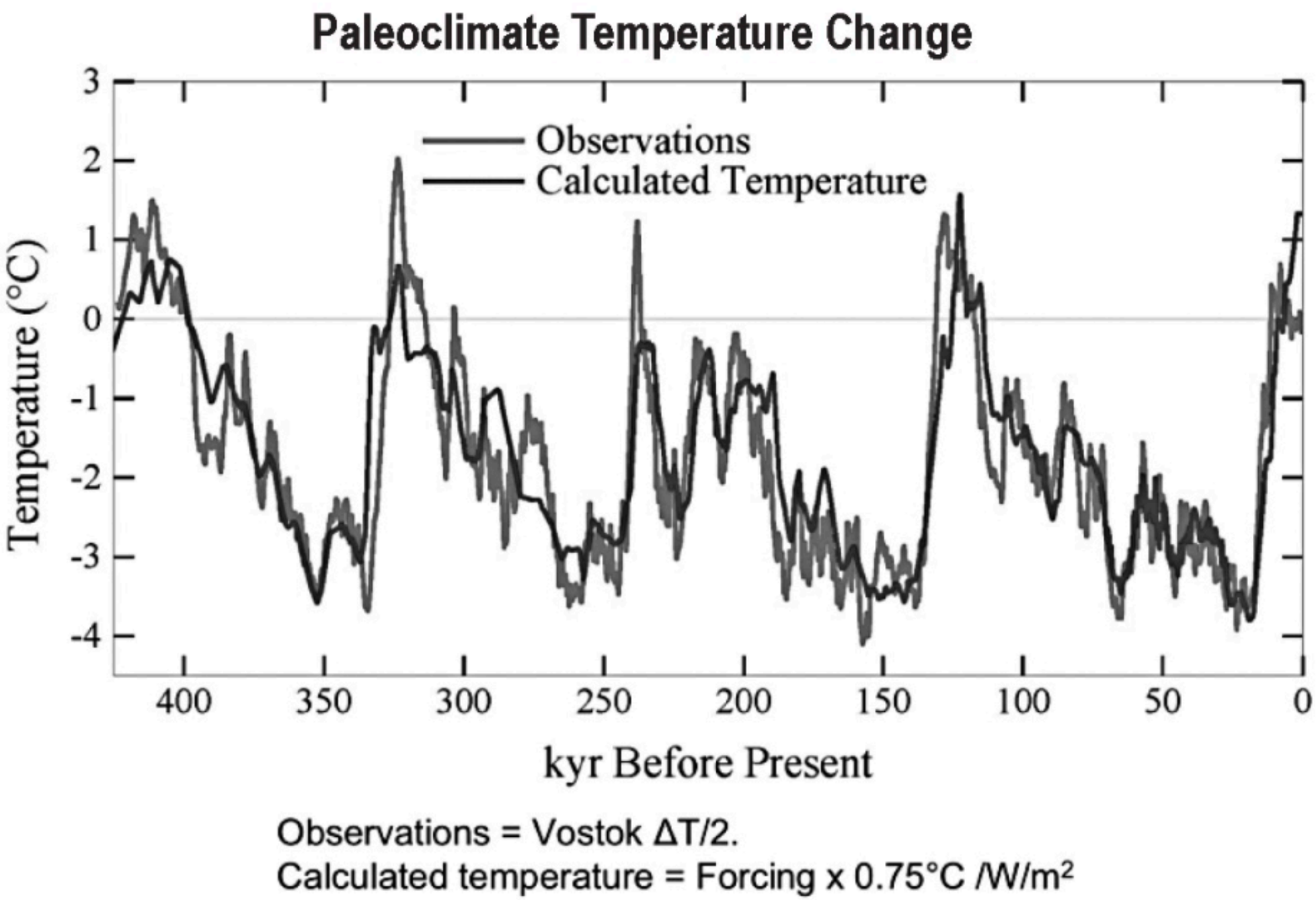


Figure 7.6

That this relatively simple set of approximations to the detailed model results provides as good a fit as this provides, at least to me, strong further confirmation that the physical arguments and predictions associated with radiative forcing and CO_2 abundance are sound.

8 - SOME LIKE IT HOT

It is often said that one can only make predictions about two trillion years *into* the future.

First, the cosmos will be colder, darker, and simpler.

Second, no one will be around to check if we are wrong.

Predicting instead what might happen in the coming century is fraught with inevitable uncertainties, as is humanity.

However, the laws of physics do put some restrictions on tomorrow based on what we know about today.

Now, I want to summarize some future predictions oddly reminiscent of Dickens's *A Christmas Carol*. Christmas Past couldn't be changed, and we cannot change our past, neither can we change some of the present and future consequences of humanity's past actions.

Christmas Present invited Scrooge to begin to imagine how actions he took that day might pan out.

And Christmas Future, the most terrifying of all, invited Scrooge to look at the consequences of the future as it *might be* with an invitation to ensure that future never happens.

We often hear or read claims about our future with global warming, but these predictions range from the guaranteed to the likely, the plausible, and the possible, often with few distinctions made.

There can be a world of difference, as Dickens devoted a whole story to demonstrate, between the future that will be and the future that might be.

I want to keep track of those differences here.

I will begin with the guaranteed developments because they are already built into the system, and simple physics arguments constrain what will happen next.

First, what happens if we stop emitting tomorrow?

I am tempted to suggest a COVID-19 pandemic-related cause, as that seemed to be central on everyone's minds over the past few years, but I sincerely hope that for this class and into the future the time, there might be other things to talk about.

Be that as it may, Earth's surface temperature has already changed by about a full degree in the past century as a result of the emission of about 450 gigatons of carbon into the atmosphere, as well as other forcings.

If carbon production stops, will the temperature return to its earlier value?

Not anytime soon.

As demonstrated in 2008 by Susan Solomon and colleagues, the time period over which it will drop by half, due to varying rates in ocean uptake, exceeds one thousand years.

Much of what we have put up there will stay there till about the year 3000.

That is, unless we figure out some new technology to remove it, or an effort at geoengineering that manipulates the atmosphere on a global scale to otherwise remove the radiative forcing that already exists.

Solomon *et al*. considered numerical climate models where CO_2 emissions continued at a 2 percent yearly increase after 2008, reaching CO_2 atmospheric concentrations ranging between 450–1200 ppm, before emissions abruptly ceased.

Figure 8.1 was their result.

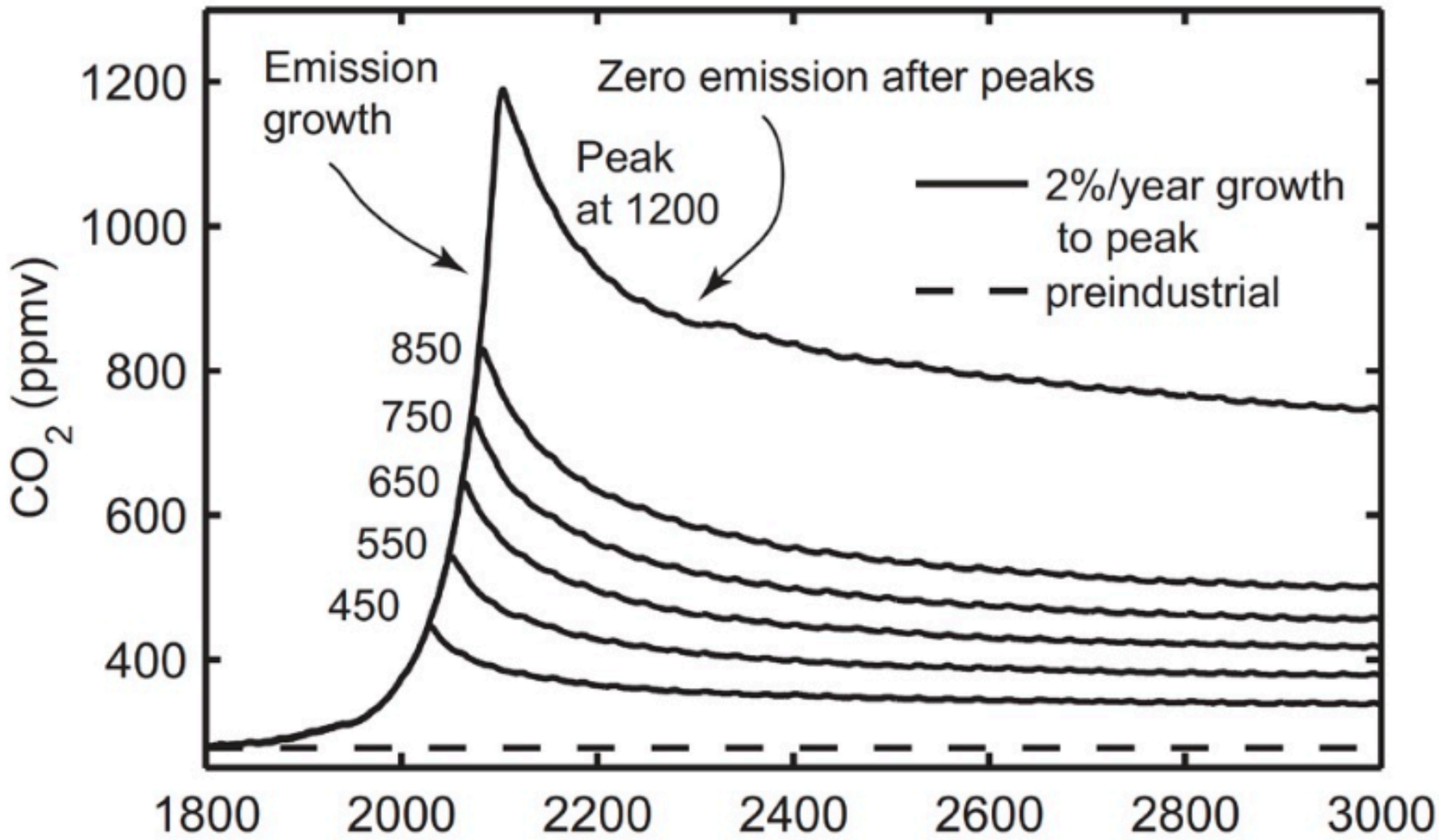


Figure 8.1

The CO_2 abundance "quickly" drops by about 20 percent over a century or so, due largely to some rapid CO_2 uptake by the terrestrial biosphere, which it then begins to give back a century later.

After that, the abundance falls much more slowly, and a millennium later they find that 40 percent of the original CO_2 peak abundance on top of the original value of 280 ppm is still in the atmosphere.

Similar 2010 calculations by a Canadian group who considered two cases—an abrupt end to CO_2 emission in 2010 or an abrupt end in 2100 with a peak concentration of almost 800 ppm—found similar results, with 55 percent of the peak CO_2 concentration remaining after a millennium.

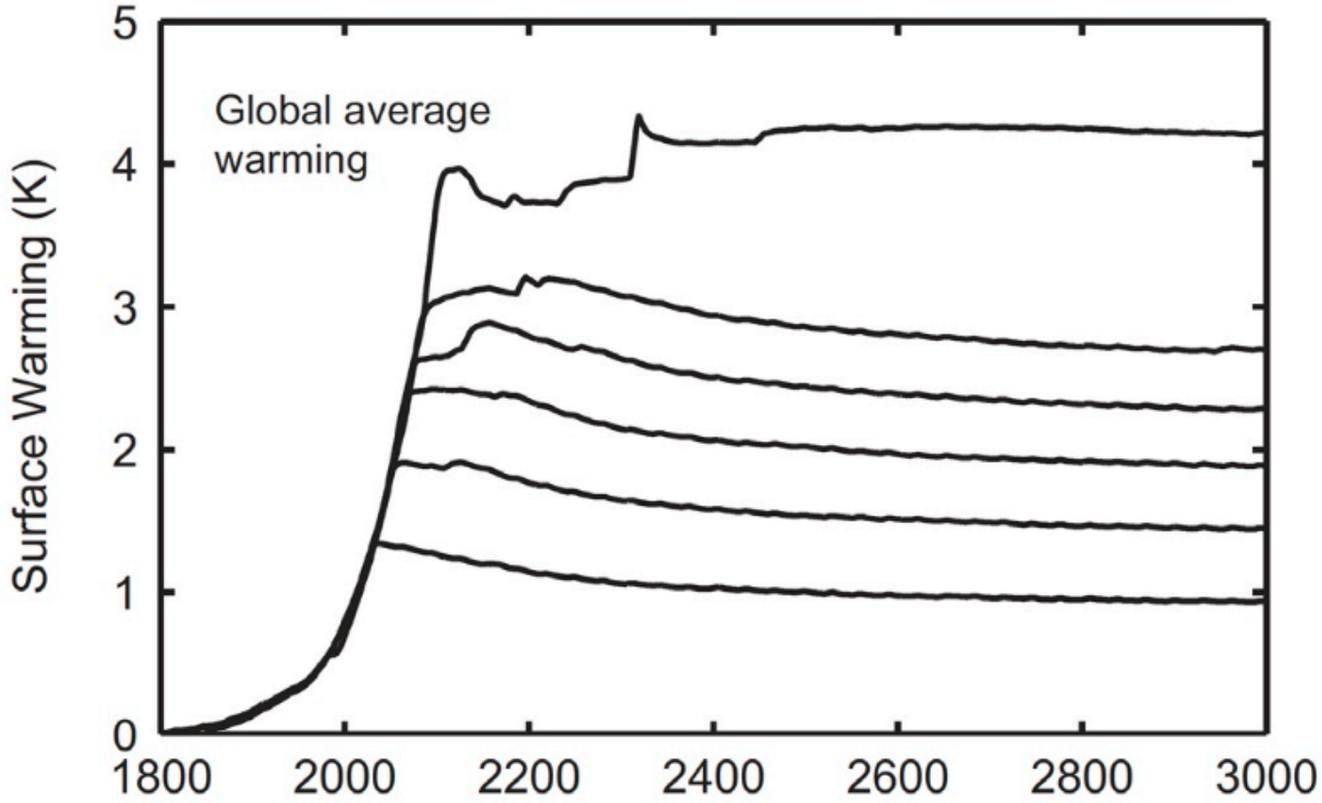
While calculating CO_2 abundance evolution is more direct, converting this into a global surface temperature change involves more model-dependent assumptions.

Nevertheless, the models allow one to numerically estimate the mean global temperature evolution, including fluctuations from year to year, as well as the distribution of temperatures across the globe.

As we saw, the global temperature predictions appear to be accurate thus far to about 0.1–0.2 degrees Celsius.

Figure 8.2 is a result for global average surface warming from the Solomon study.

Figure 8.2



The key takeaway here is that the relatively recent rapid temperature rise that has occurred as long as CO_2 emissions have continued unabated over the past fifty years *ends very quickly upon cessation*, but the temperature itself does not noticeably decrease even as CO_2 levels themselves slowly decrease over the next thousand years.

This is due in large part to a factor that will play a key role in much of what follows and in much of the mid- and long-term future of the planet.

The deep oceans take a long time to mix and equilibrate the additional heat that has been pouring into the Earth due to the greenhouse effect.

Hence the oceans will continue to warm for many hundreds of years after the source of excess heat has disappeared.

Land masses, on the other hand, store less heat and therefore cool more readily as the radiative forcing decreases.

For this reason, since the Southern Hemisphere has much more ocean and less land than the Northern Hemisphere, there is expected to be a differential temperature change across the planet's surface, and that change will itself change with time.

Figure 8.3, for example, displays the relative temperature changes across the planet between 1850 and 2105 in the Canadian study where carbon emission continues unabated until 2100, and below it the subsequent temperature change between 2105 and 2995.

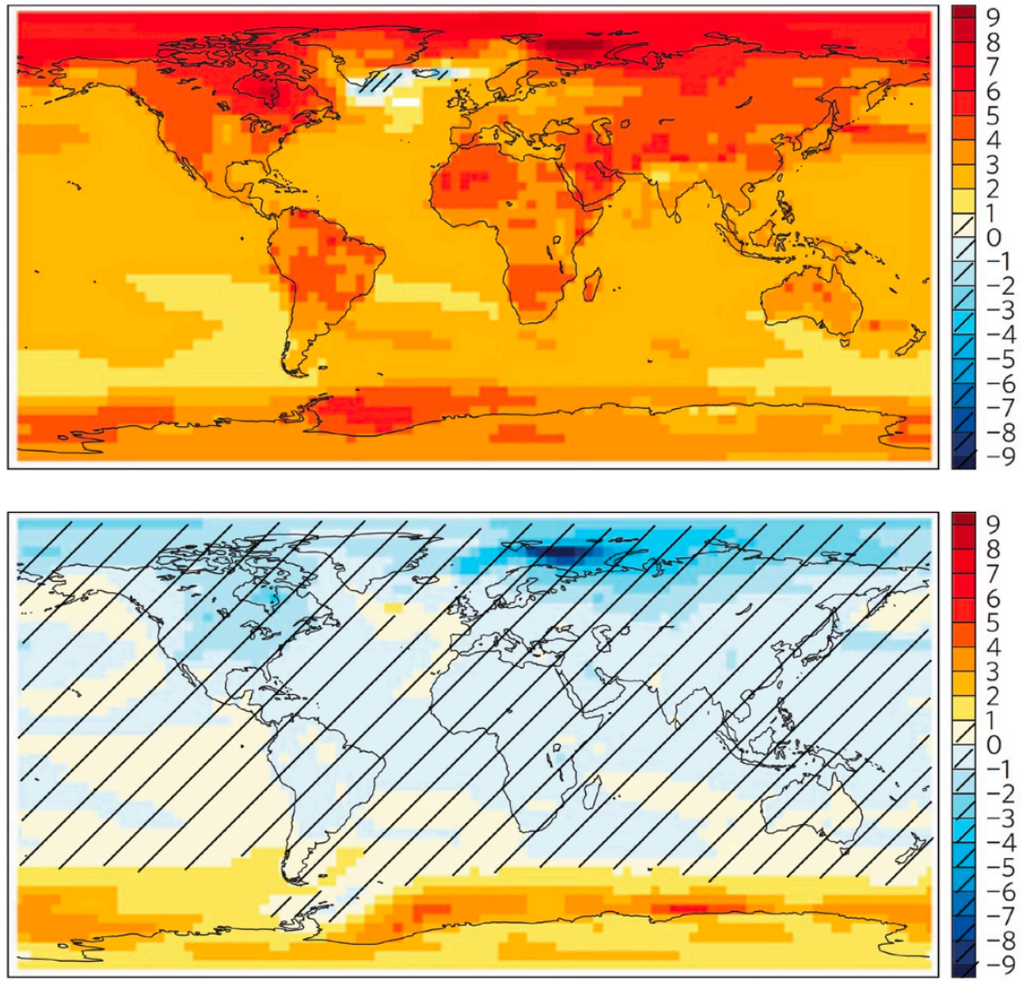


Figure 8.3

As can be seen in the top figure, the highest 'immediate' rise is in the Northern Arctic, where a shallow ocean is surrounded by large land masses.

Hundreds of years later, in the lower figure, the Arctic will have already begun to cool significantly whereas the Southern Ocean and Antarctica continue to warm well after CO_2 stopped rising.

Similar but smaller changes occur in the model in which emission stops in 2010.

We are somewhere in between, having had one additional decade of CO_2 emission.

This can be understood in large part because of the huge latent heat stored in the deep oceans, which store over 90 percent of the heat that has been added thus far to the climate system, and which, as I have indicated, takes some time to equilibrate.

In the Canadian study, an estimate of the temperature of the Southern Ocean at a high latitude of 66°S found ocean temperatures continuing to increase throughout the next millennium by as much as an additional 3°C in the pessimistic case where carbon emissions don't cease until 2100, ultimately slightly exceeding the average global surface temperature rise throughout this period.

Their overly optimistic scenario, in which carbon emissions ceased in 2010, is worth focusing on because it describes the future as it *will be* (actually it will be worse because as I have just noted, we have continued to emit CO_2 in this decade).

It yields a further rise in ocean temperature through the next millennium by almost 0.5°C, an amount that is essentially already written in stone because that heat has already been deposited over the past century.

This will have inevitable consequences for the planet, some of which I will discuss in more detail in the later.

To get an idea of the heat capacity of the world's oceans compared to the heat dumped in the oceans by humanity so far, consider the measured rise in average ocean temperature in 2019 of 0.075°C compared to the 1981–2010 average temperature, as recently compiled by an international group of researchers.

Just this measured temperature difference—which I remind you does not yet reflect the full ocean response to the additional heating by CO_2 radiative forcing thus far—corresponds to the additional heat that would been produced by setting off 3.6 billion Hiroshima atom-bomb explosions in the ocean or about five Hiroshima bombs of heat every second, day and night, 365 days a year for the past twenty-five years.

Besides the overall change in ocean temperatures, there are two other long-term impacts that have already been written in stone: one direct and one indirect.

The first impact is due to the simple thermal expansion of water as its temperature increases, a fundamental basic physics result.

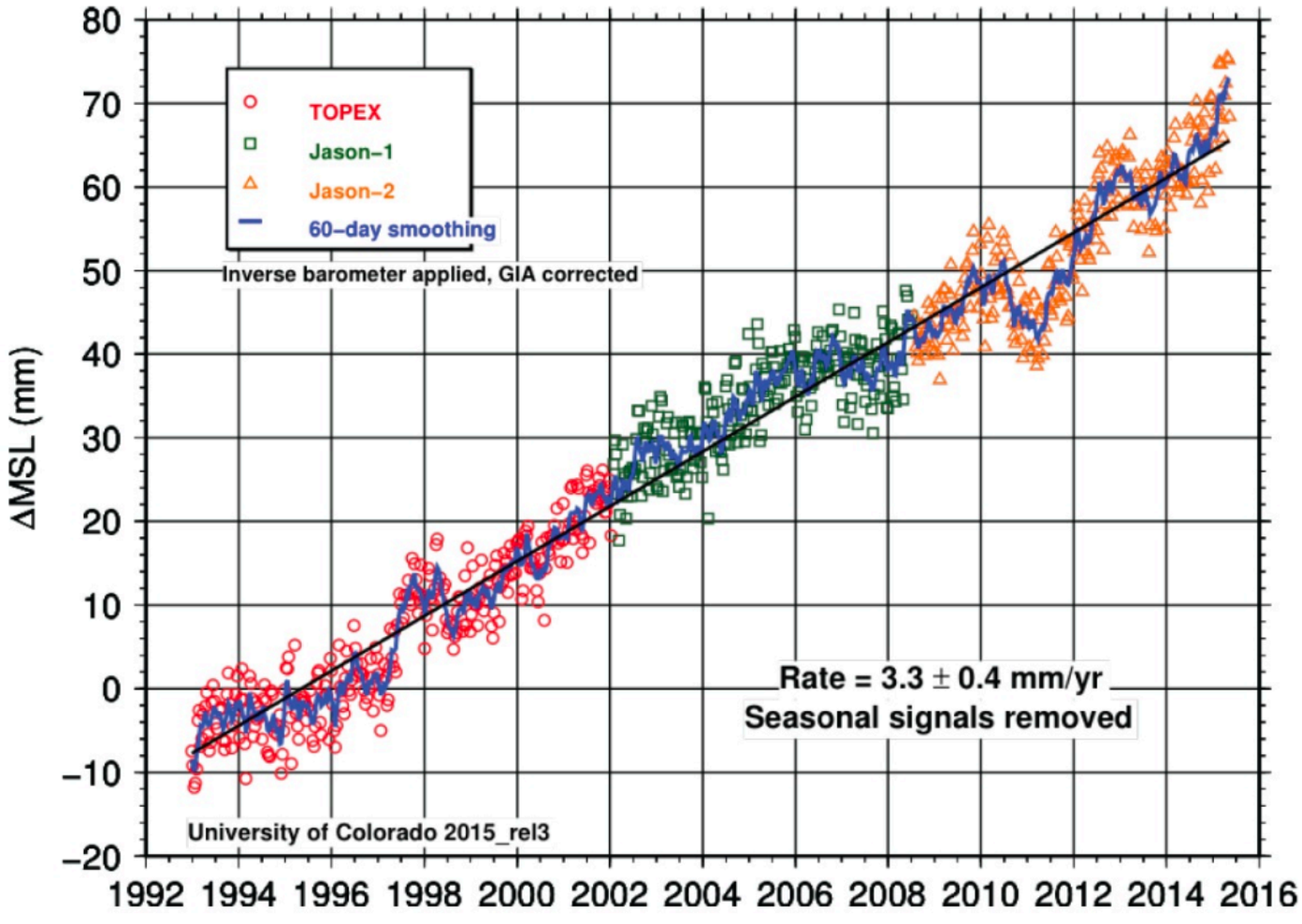
When global warming is discussed, most often reference is made to the predicted rise in sea levels due to the possible continued melting of glaciers and ice shelves in Greenland or Antarctica.

However, as I shall describe later, the dominant single source of sea level rise in recent times has not been due to this effect but rather to thermal expansion due to the average ocean temperature rise thus far.

Figure 8.4 is a graph of the measured sea-level rise, using satellite data between 1992 and 2015, of about 3.3 mm per year.

Fully half of this effect is due to thermal expansion of water and is independent of any uncertainties associated with glacial melting.

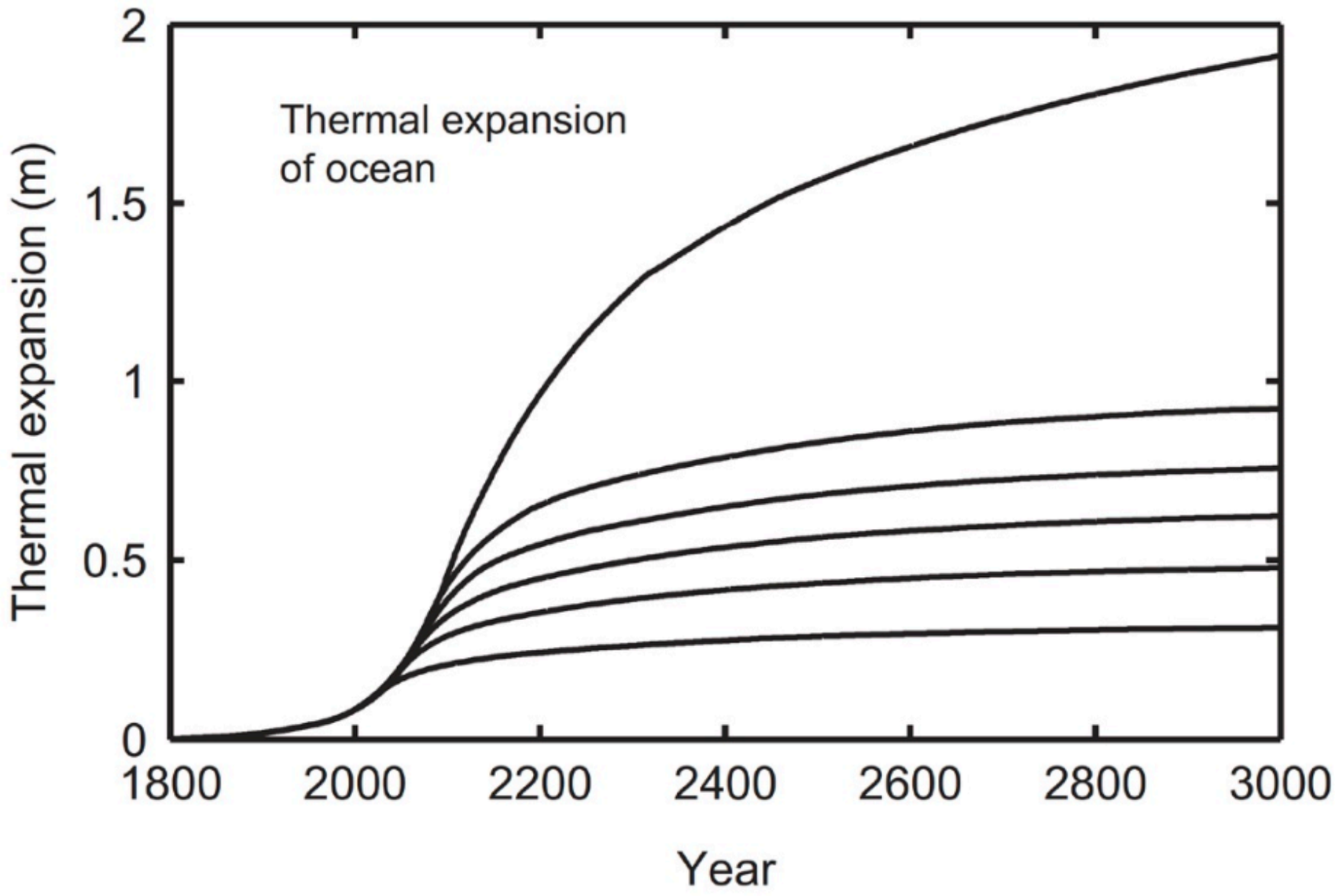
Figure 8.4



As the world's oceans continue to increase in temperature, *even long after global industrial CO_2 emission may have ceased*, the ocean level will continue to rise, again independent of glacial melting effects.

Figure 8.5 is the prediction from the Solomon simulation discussed above. Once again, note that the lower curve represents what is inevitable, even if we had stop emitting CO_2 in 2010:

Figure 8.5



Note that the lowest curve occurs with a peak concentration of 450 ppm, close to the current value of 417 ppm.

This means a long-term sea level rise of about 1⁄4 of a meter is essentially inevitable given the world's cumulative CO_2 emissions to date, independent of other possible sources of sea level rise, including melting of major ice sheets.

If CO_2 emissions continue at more or less their current rate for the rest of this century, the minimum sea-level rise we can expect is about one meter.

For this case, the more recent 2010 Canadian model analysis predicts a sea level rise slightly in excess of one meter.

The other more indirect impact that is likely to continue well after our current CO_2 emission rate declines or ceases involves a predicted change in the precipitation patterns around the world with significant potential regional impacts.

These are due to predicted changes in global ocean currents affected in part by differential global temperature shifts, deep mixing of heat and increased fresh-water content due to glacial melting.

A detailed discussion of models of ocean dynamics and currents is beyond the scope of this class, but I will mention one predicted long-term impact because it follows almost directly from a direct prediction I have already discussed: the differential long-term heating of the Southern Hemisphere versus a cooling of the Northern Hemisphere after emissions cease.

The net impact of this temperature shift is predicted to produce a consequent precipitation change: continued drying in the low latitudes of North Africa, South America, and Australia.

As for the global temperature pattern predictions displayed earlier, and from the same numerical simulations, the top image shown is the predicted change in precipitation between 1855 and 2105 if emissions continue till 2100.

The lower plate displays the predicted change between 2105 and 2995 in this case (Figure 8.6).

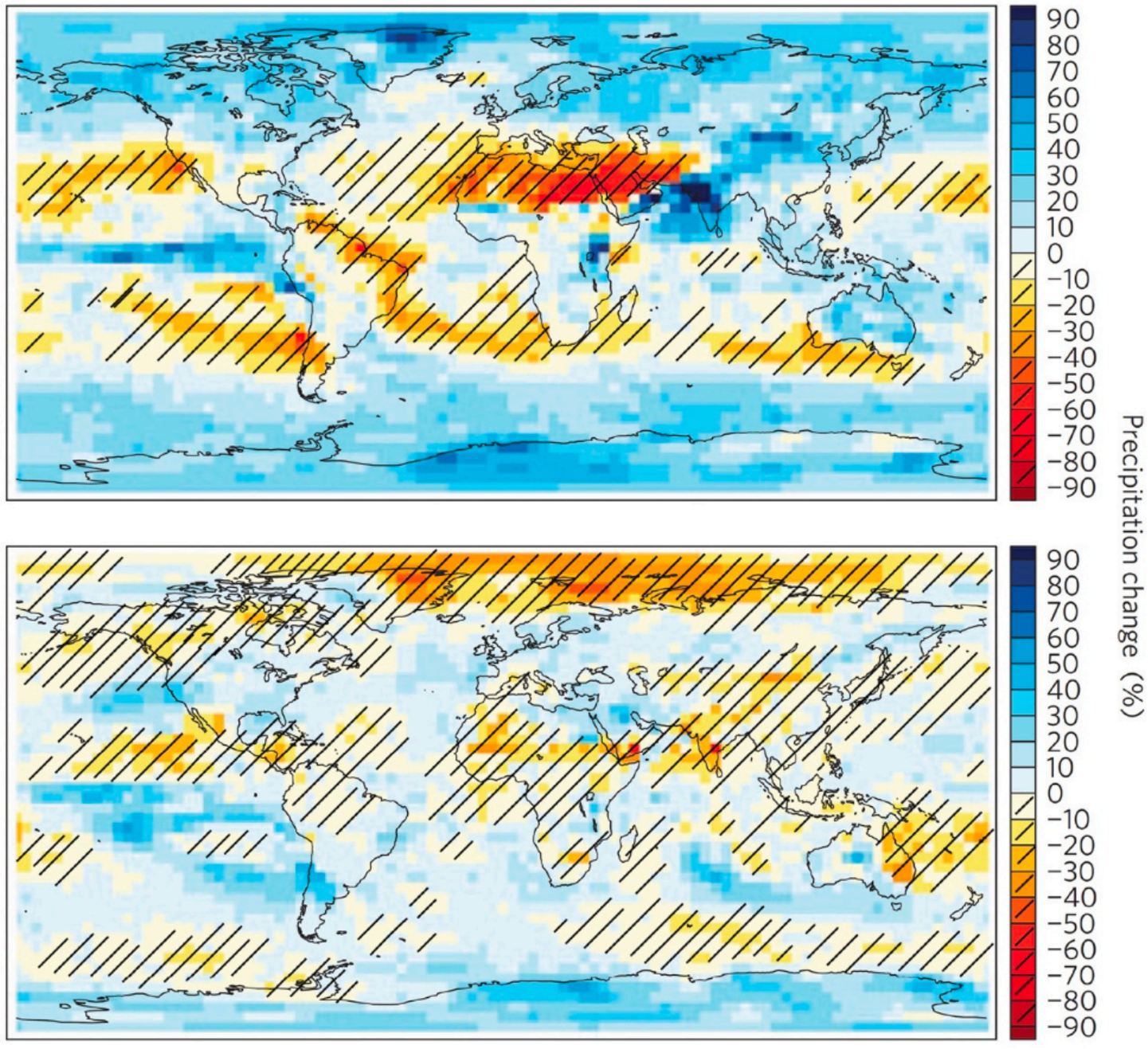
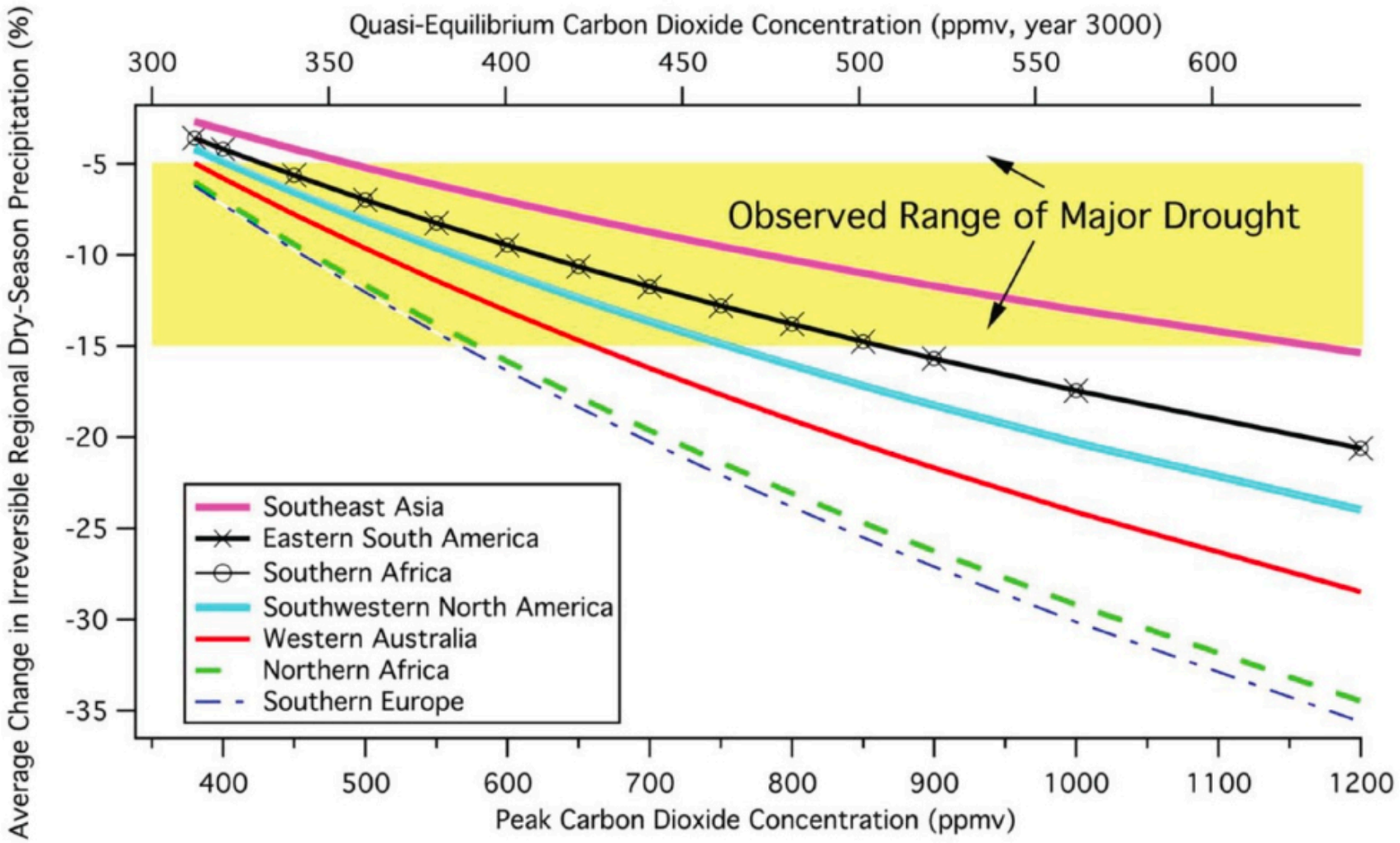


Figure 8.6

Solomon and colleagues demonstrated just how severe the impact could be in various locations during dry season in 3000 as a function of the peak CO_2 concentration before cessation (Figure 8.7).

The shaded region represents the range of precipitation experienced during major past droughts, such as the "dust bowl."

Figure 8.7



The fact that CO_2 will remain in the atmosphere long after humanity stops or reduces its global emissions, combined with the fact that the potentially negative observable consequences of its current abundance will in some cases take centuries to fully manifest themselves, should affect our thinking about the consequences of continuing business as usual and the potential need for global action sooner rather than later.

Once again, I remind you that pre-industrial carbon abundance in the atmosphere was about 600 billion tons.

We have added about 500 billion tons to that amount to date.

If emissions continue at the current rate till 2100, we will have added by that time an additional 2.2 trillion tons to the atmosphere.

Say then, for example, we wish to limit global temperature rise to some specific value, perhaps 1.5°C.

Then, because the cumulative CO_2 abundance in the atmosphere simply increases each year we add more CO_2 and doesn't significantly decrease if we stop emitting, we simply have to ask how many more years we can keep emitting at the current rate before the net radiative forcing by CO_2 produces the predicted temperature change.

For 1.5°C we are already 90 percent of the way there, meaning in ten years we will have already passed that threshold if current emission rates remain constant.

Let's say we relax our goal and accept a temperature rise of at most 2°C, the target of the 2015 Paris Agreement.

One of the arguments that has been used against trying to moderate emissions to meet this goal is based on the claimed economic costs of trying to achieve it.

I won't debate that issue here, even though I think it is spurious, but rather talk about the science.

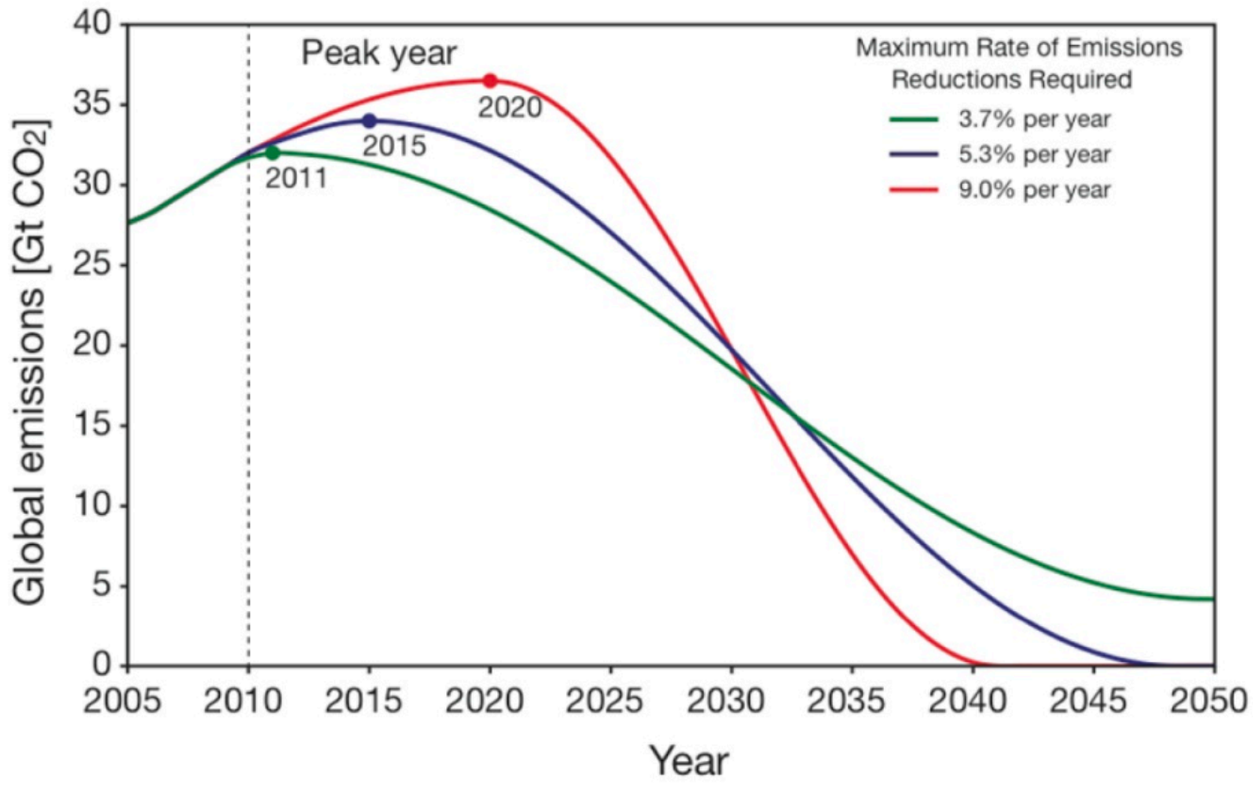
And here the science presents a dilemma.

Each year we delay trying to limit the greenhouse temperature rise to be less than some target, it becomes harder and more costly to achieve that goal.

How much would one have to reduce emissions each year so that the total cumulative emission is less than an amount required to yield a two-thirds probability of global temperature rise being less than 2°C?

Figure 8.8 is a chart from 2009, called the "ski slope" diagram, showing what would have been required if we had started this process globally in 2011 (beginner slope), 2015 (intermediate slope), or in 2020 (expert slope):

Figure 8.8

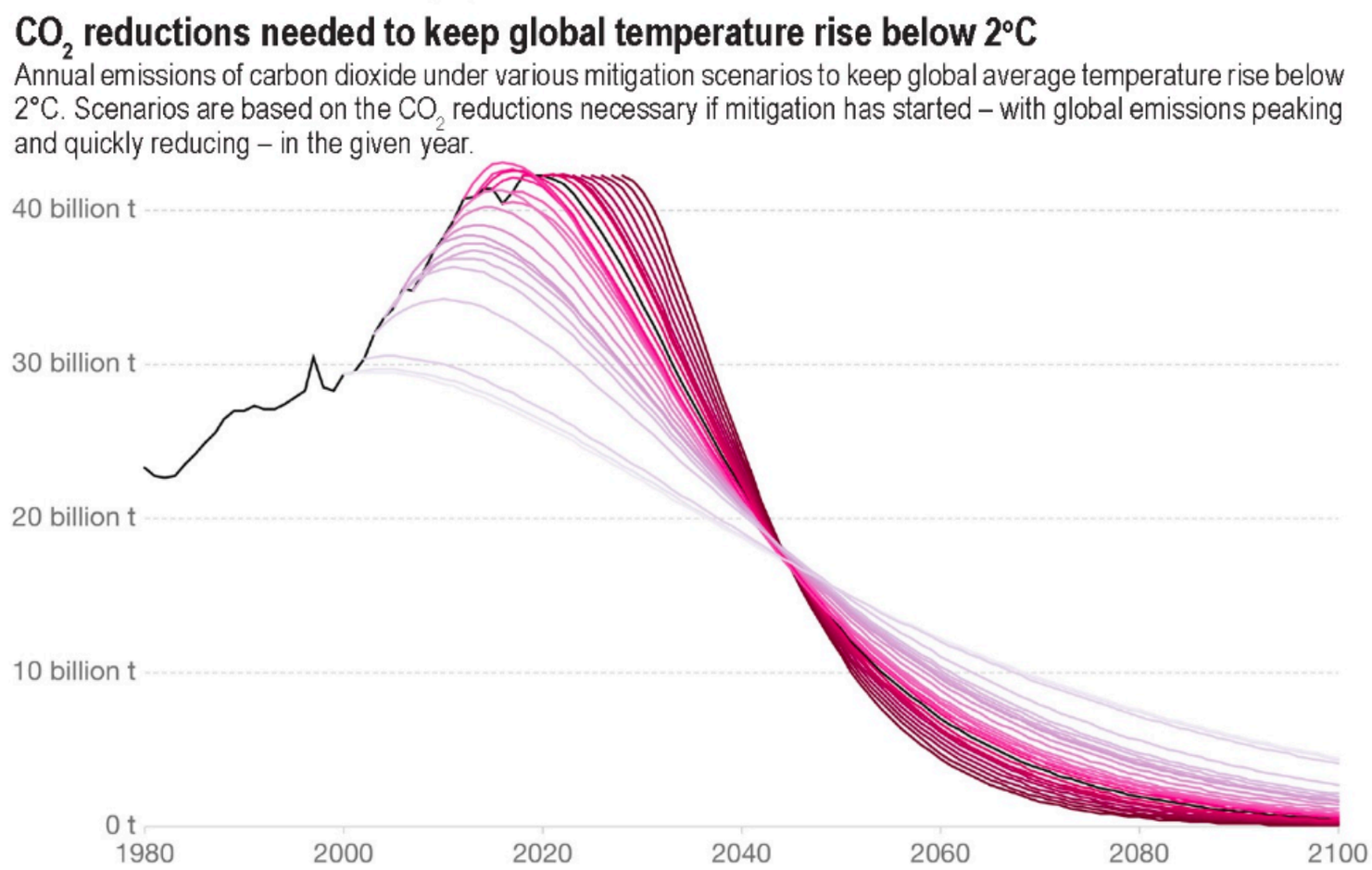


Observe that in 2011 we would have only had to consider a 3.7 percent annual reduction in emissions.

Now we would have to somehow arrange a 9.0 percent annual reduction.

Figure 8.9

Figure 8.9 is a similar graph, projecting things further into the future, also showing actual emission rates over the past forty years.



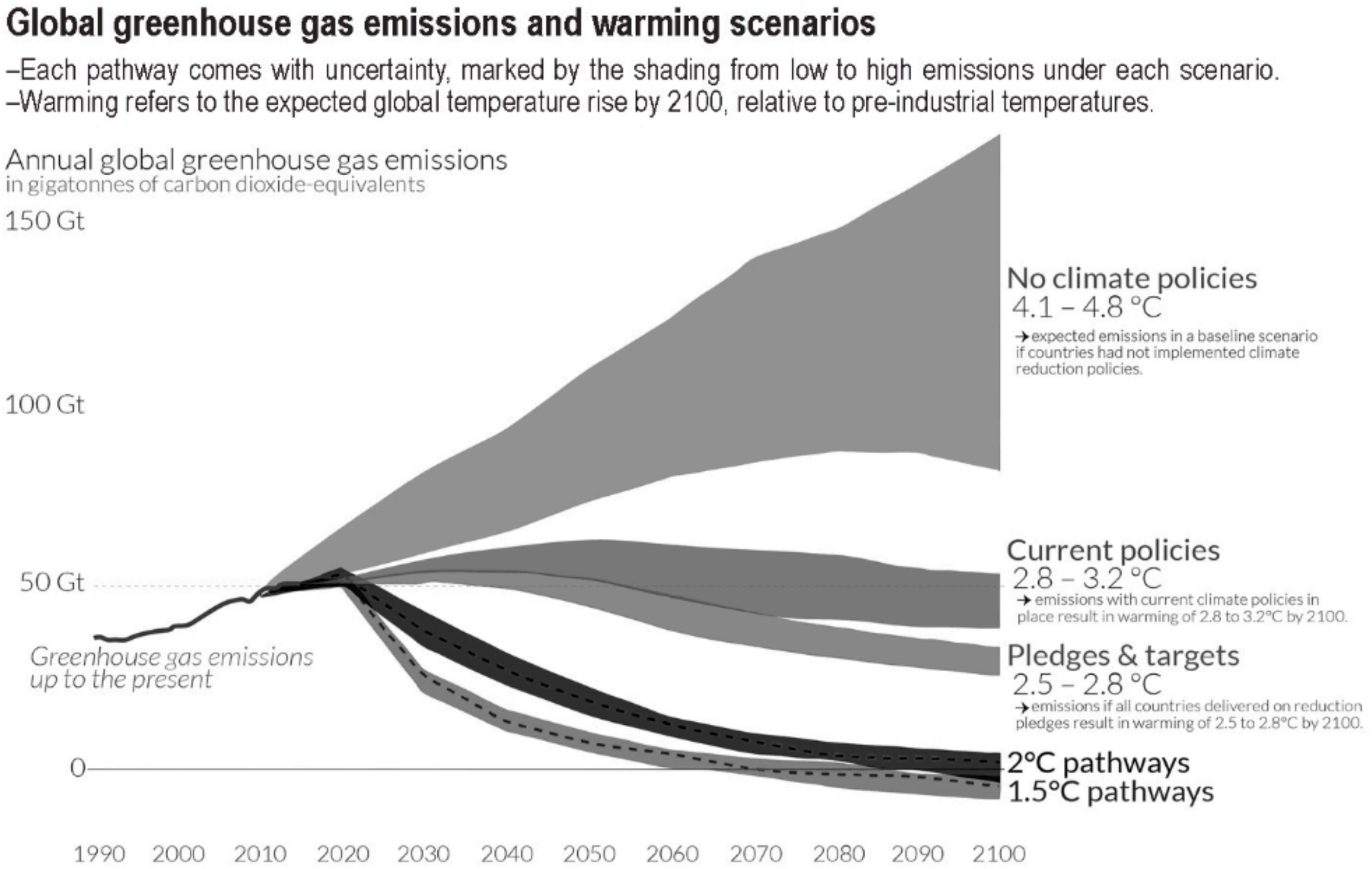
You can see that even if emissions don't keep growing, but also don't start falling, then by 2030 or so we would essentially have to turn off the spigot, which would be virtually impossible.

One can compare these hypothetical pathways with the actual range of likely trajectories given current global policies.

One of these futures will occur.

We might get to choose which one—if we actually think about it enough to make the choice (Figure 8.10).

Figure 8.10



Finally, we can extend the emission curve out to yet longer times.

Figure 8.11 shows what the range of cumulative human emissions looks like, from business as usual to the current and perhaps unrealizable global pledge of 2.6– 3.2°C warming.

Again, remember the pre-industrial CO_2 abundance in the atmosphere was 600 Gt.

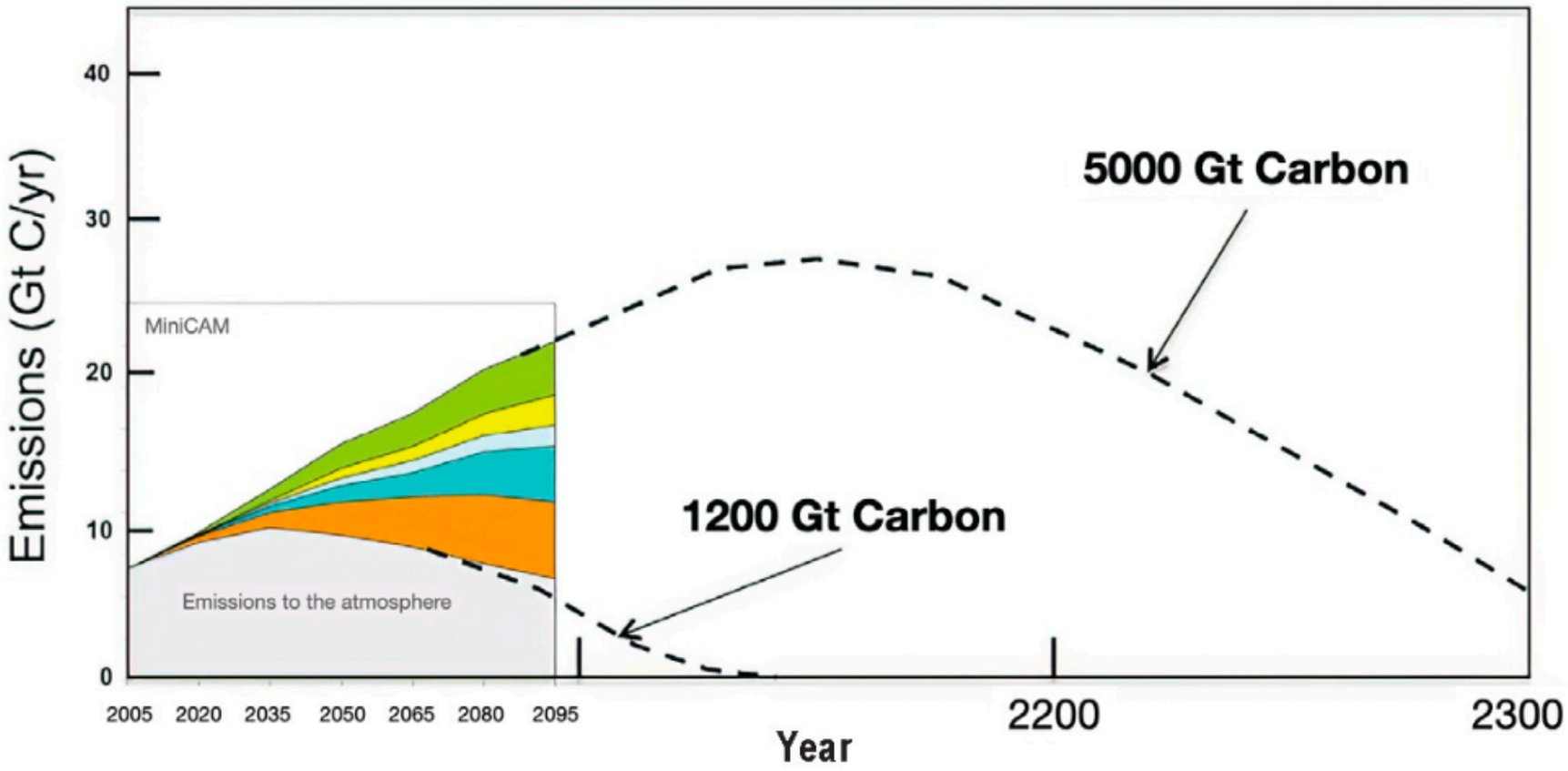


Figure 8.11

Returning to a theme from earlier in this class, the numbers demonstrate convincingly that human emissions are *not* insignificant on a global geological scale.

In two hundred years humans will have been able to emit at the very least twice as much CO_2 as existed in the Earth's atmosphere over much of the geological record.

These present a minimalist view of the future as it will be, one way or another, depending on our actions.

I have studiously avoided discussing how the increases I have described here will impact on humanity, and also about possible feedback or amplifying mechanisms that might exacerbate these impacts.

Next in our discussion, I will focus on one particularly disturbing aspect of the future as it *might be*.

In so doing, I will display the first graph I saw that convinced me that globally the risks are such that we would be wise to address climate change now.

9 - WHAT WILL BE, WILL BE?

There is a song “QUE SERA, SERA (WHATEVER Will Be, Will Be)” that seems to imply the future was beyond our ken.

I don’t buy it.

One of the purposes of science is to predict the future.

Many physicists are generally focused, with some exceptions, either on making predictions about experimental systems that are so well controlled that the implications of fundamental theory are directly discernible from observations, or they make predictions about the universe on its largest observable scales.

On these scales, fortunately, the universe can be modeled very simply.

It is far harder to model the full complexity of daily life on Earth, much less to predict what will happen tomorrow.

But when lives depend on it, we can rise to the task.

After centuries of effort, with help from modern satellites, supercomputers, and weather stations across the globe, meteorologists can now accurately predict the weather, at least on a statistical basis, up to ten days in advance.

Storms and hurricanes can be anticipated with a great savings in human life.

Weather is different than climate, however.

Weather takes place in a single place, at a single time.

Climate is longer term, and regional or global rather than local.

On global scales, the local variability from place to place can become unimportant.

Using general principles of work, heat, and energy flow, the global heat balance of the planet has been understood and measured.

And the predictive and experimental capabilities of physics and chemistry have been refined to test and confirm the fundamental principles governing the evolution of climate on our planet on its broadest scales.

Absorption of infrared radiation by greenhouse gases has been verified, and the energy balance governing the Earth/atmosphere/sun are understood.

The most direct prediction involves energy transfer, so we discovered radiative forcing.

From there, atmospheric models allowed a prediction of surface temperature changes that could be tested.

With temperature changes come a whole new set of phenomena.

Sea level rise due to thermal expansion is the most immediate consequence, but we have seen that the impact of the ocean is also significant because it is by far the largest heat sink governing the long-term evolution of climate in response to radiative forcing.

The differential heating of the planet as a function of latitude is the direct consequence of the asymmetric distribution of land mass and ocean as a function of latitude.

The enhanced immediate temperature rise of the Arctic is one result, and the long-term heating of the Antarctic will be another.

The extent to which this picture is accurate is of vital importance for the future of the planet in the coming millennium because the two greatest reservoirs of fresh water in the form of ice are located at or near the Earth's two poles: in Greenland and Antarctica.

And a significant temperature rise in one or both locations could have indirect consequences that may be more dramatic than that due to the direct increase in mean surface temperature alone.

The thermal expansion of water has dominated the measured sea level rise to date, and a continued expansion is basically guaranteed due to the additional heat already stored deep in the oceans.

But as has been mentioned, there are other contributions, primarily due to ice sheet disruption in Greenland and Antarctica.

While these have been subdominant over recent decades, there is every indication that this will not remain the case.

As the National Academy of Sciences put it in a study published in 2011:

> *The dynamic response of ice sheets to global warming is the largest unknown in the projections of sea level rise over the next century.*

Thus, even if it involves considerations beyond the direct consequences of radiative forcing, it is worth exploring an overview of the data, mechanisms, and predictions for significant disruption of glacial ice and its impact on expected sea level rise.

Glacial melting is accelerating, especially in Greenland.

The extent to which this melting will continue to accelerate will depend on a host of factors, and as the National Academy stressed, not all of these are easily estimated direct consequences of radiative forcing.

This is, after all, the future *as it might be*.

As far as I am concerned, the most chilling images, if you forgive the pun, come not from the predictive models of the future, but from the data about the past.

I remember when I first saw this data over a decade ago—before I had thought about the underlying physics in any depth—I was shocked out of my own complacency and decided there were realistically some serious risks worth worrying about.

Let's start with Greenland. By every standard one can devise, the melting of Greenland's ice sheet is accelerating.

Every year seems to top the report from the previous year.

First, consider ice mass loss.

Already a decade ago studies were demonstrating increasing loss, measured in gigatons/year.

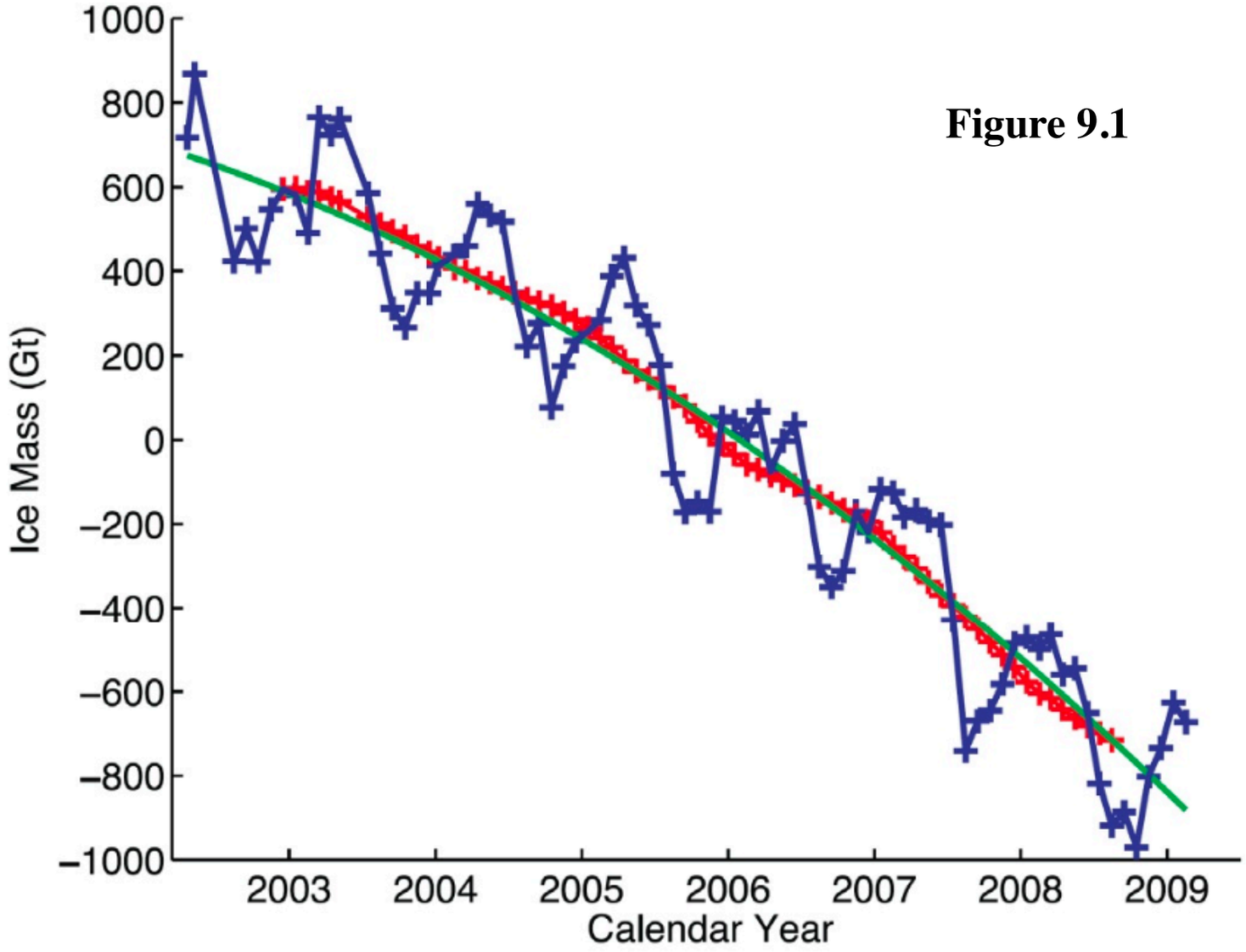
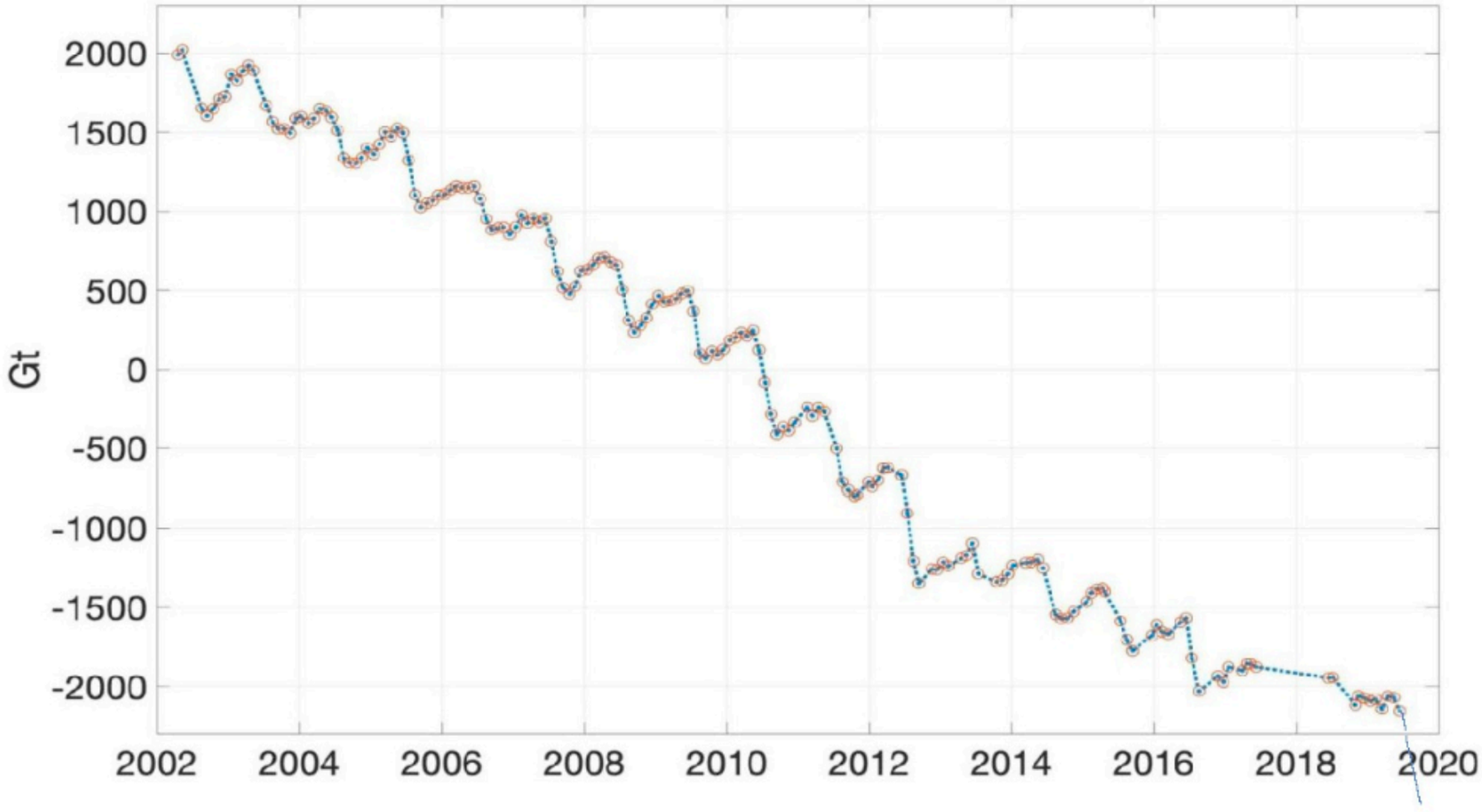


Figure 9.1

The Scripps Institution of Oceanography provided this Figure 9.1 in 2009, where the Greenland ice mass quantities are shown relative to the ice mass value in 2006.

The updated Figure 9.2 comes from the 2019 Arctic Report Card, where the timescale is set so that it matches the previous figure on the dates of overlap, and ice mass figures shown are now relative to 2010 and not 2006.

Figure 9.2



I added one more point to this graphing(see large drop at end), coming from two reports in April 2020 from labs that include Columbia University, University of California, and the Jet Propulsion Lab.

They reported that over the Arctic summer of 2019, Greenland lost approximately 600 Gt of ice, more than twice the average rate of ice loss, about 270 Gt/yr., between 2012 and 2017.

Between 2003 and 2007, by contrast, the average annual mass loss was 180 Gt, and between 2000 and 2003 the rate was 150 Gt/yr.

Between 1997 and 2003 the average loss was about 74 Gt/yr., and between 1993 and 1999, before the data shown on these curves, the average mass loss was recorded to be only about 54 Gt/yr.

The trend is clear.

Figure 9.3 comes from a satellite that measures the microwave brightness temperature of the surface. Shown below are the annual number of cumulative melt days on the ice.

On the left is the annual average from 1979– 2007, and the middle is 2012, the most severe melting year until 2019.

The image from 2019 includes only the period April–December.

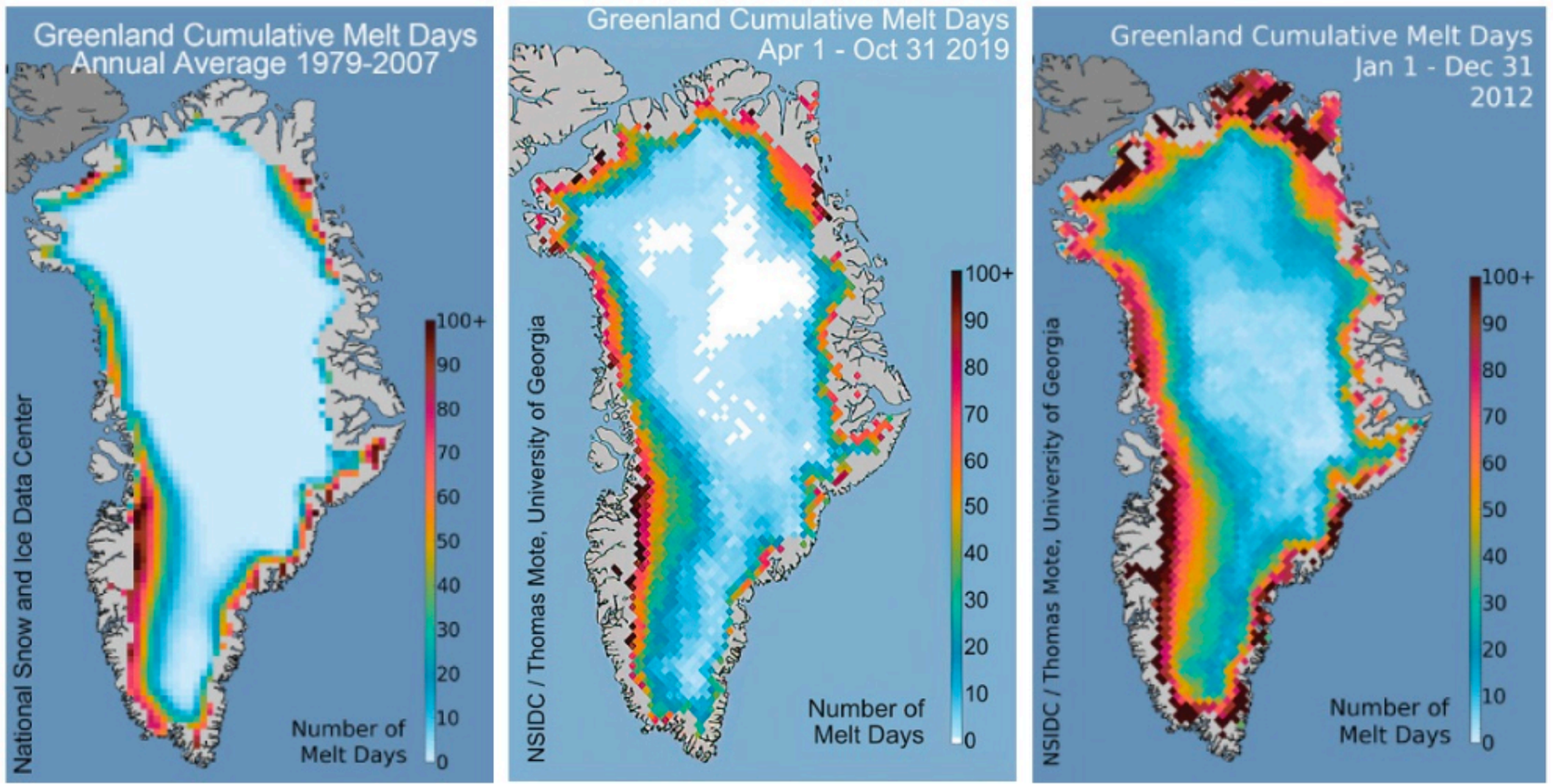


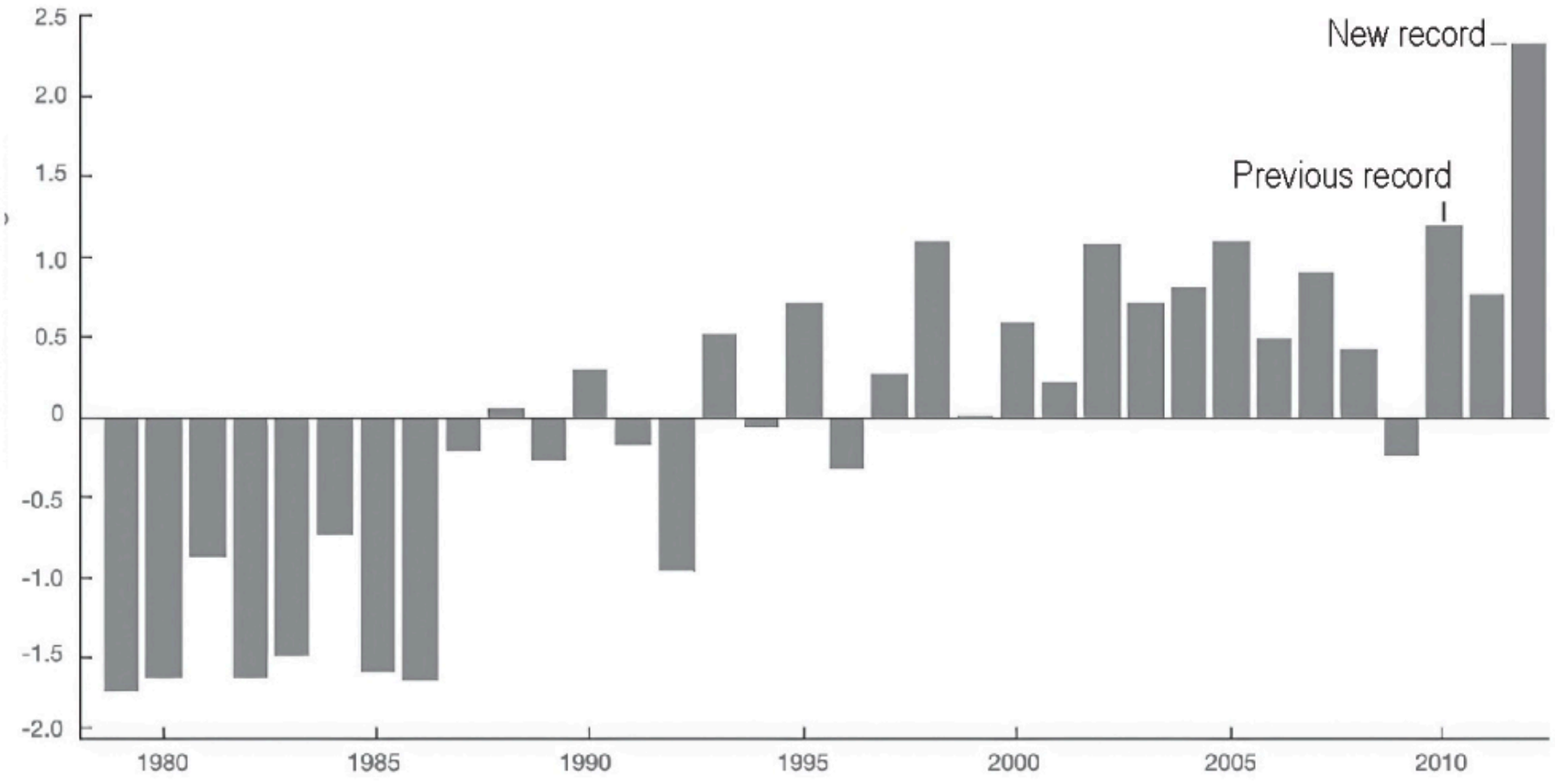
Figure 9.3

One can get another graphical representation of the data by multiplying the total number of cumulative melt days by the area that the melt was detected in.

This has been labeled the *standardized melt index*, when the yearly data is compared to the average of the full measurement period from 1979 onward.

Figure 9.4 is the index up to 2012.

Figure 9.4



This data make it unambiguously clear that the Greenland ice sheet is now melting at an accelerating rate, so Greenland is now contributing an ever-increasing amount to measured sea level rise.

Over the entire decade from 1993 to 2003, Greenland ice melt contributed about 0.21 mm to the rising sea level, whereas the 2019 melt in a single year raised sea levels by an order of magnitude more than this.

What can we expect in the long term?

Before we focus on that question, some comments about the interpretation of the current data:

First, an acceleration of Greenland's ice melt is to be expected based on the analysis of asymmetric global warming in response to radiative forcing I have already described.

We have observed and we expect more initial warming in the far north compared to that in the Southern Hemisphere.

Second, it is worth noting that the total ice mass loss in Greenland is not simply due to melting but an almost equal amount in the early 2000s was due to ice discharge, like the calving of glaciers into the sea, and also the interaction of warming ocean water with glaciers in contact with the ocean.

These factors are related to yet more complex climate effects.

That said, one should note that the ice discharge from Greenland over recent years has not increased noticeably, while melting has.

Thus, estimating the total ice mass loss in the future by considering melting alone might not lead to too significant an underestimate of resulting sea level rise.

While the data presented on the previous pages is consistent with what one would expect on the basis of radiative forcing and global temperature increases, a natural skeptic might ask, when presented with this data, about its historical context.

Is this anomalous?

Is there other evidence of Greenland ice mass as temperatures have varied in the past, and if so, by how much?

Fortunately, there has been a significant amount of data obtained that sheds light on the paleoclimate in Greenland.

The results are sobering.

Detailed measurements of marine deposits, coastal terraces, coral deposits, isotope abundances, and ice cores from Greenland make it quite clear that Greenland has been ice-free at times as recently as four hundred thousand years ago, and that the total Greenland ice mass correlates more closely to global temperature than anything else.

But, while rapid retreat of glacial ice near the edges at the coasts has occurred in the past over the course of decades, Richard Alley and colleagues concluded in a 2010 *Quaternary Science Review*: "major changes of central regions of the ice sheet are thought to require centuries to millennia."

A number of factors that might have been expected to significantly contribute to ice loss or gain and thus, which might add uncertainty to model estimates, have been shown to be largely absent.

Ice loss in response to warming has been relatively insensitive to sea level change, for example.

And, while precipitation is predicted to increase with increasing temperature, and this has been documented in increased snowfall in Greenland during periods of warming, that increased snowfall could not compensate for the increased melting and ice discharge, so the net ice mass decreased.

As summarized in the 2010 review (*Quaternary Science Reviews*, Vol 29) by Alley and colleagues:

Paleoclimatic data show that the Greenland Ice Sheet has changed greatly with time.

From physical understanding, many environmental factors can force changes in the size of an ice sheet.

Comparison of the histories of important forcings and of ice-sheet size implicates cooling as causing ice-sheet growth, warming as causing shrinkage, and sufficiently large warming as causing loss.

The evidence for temperature control is clearest for temperatures similar to or warmer than recent values (the last few millennia).

Snow accumulation rate is inversely related to ice-sheet volume (less ice when snowfall is higher), and so is not the leading control on ice-sheet change...sea-level change is not the dominant forcing at least for temperatures similar to or above those of the last few millennia...the limited paleoclimatic data consistently show that short-term and long-term responses to temperature change are in the same direction.

The best estimate from paleoclimatic data is thus that warming will shrink the Greenland Ice Sheet, and that warming of a few degrees is sufficient to cause ice-sheet loss.

The loss of the complete Greenland ice sheet would raise sea levels by about 7 meters, and significant changes in the ice sheet have accompanied periods of warming during the six major interglacial periods between ice ages on the planet over the past four hundred thousand years.

There is evidence that during at least the warmest period, about four hundred thousand years ago, when local temperatures were at least 2–4°C higher than the present time, that the ice sheet did disappear, with a consequent sea level rise of 7 meters, and during a more recent period, about a hundred thousand years ago when the local temperature was at least briefly in a similar range of 2–4°C above the present time, that sufficient melting occurred to raise sea levels by about 3.5 meters.

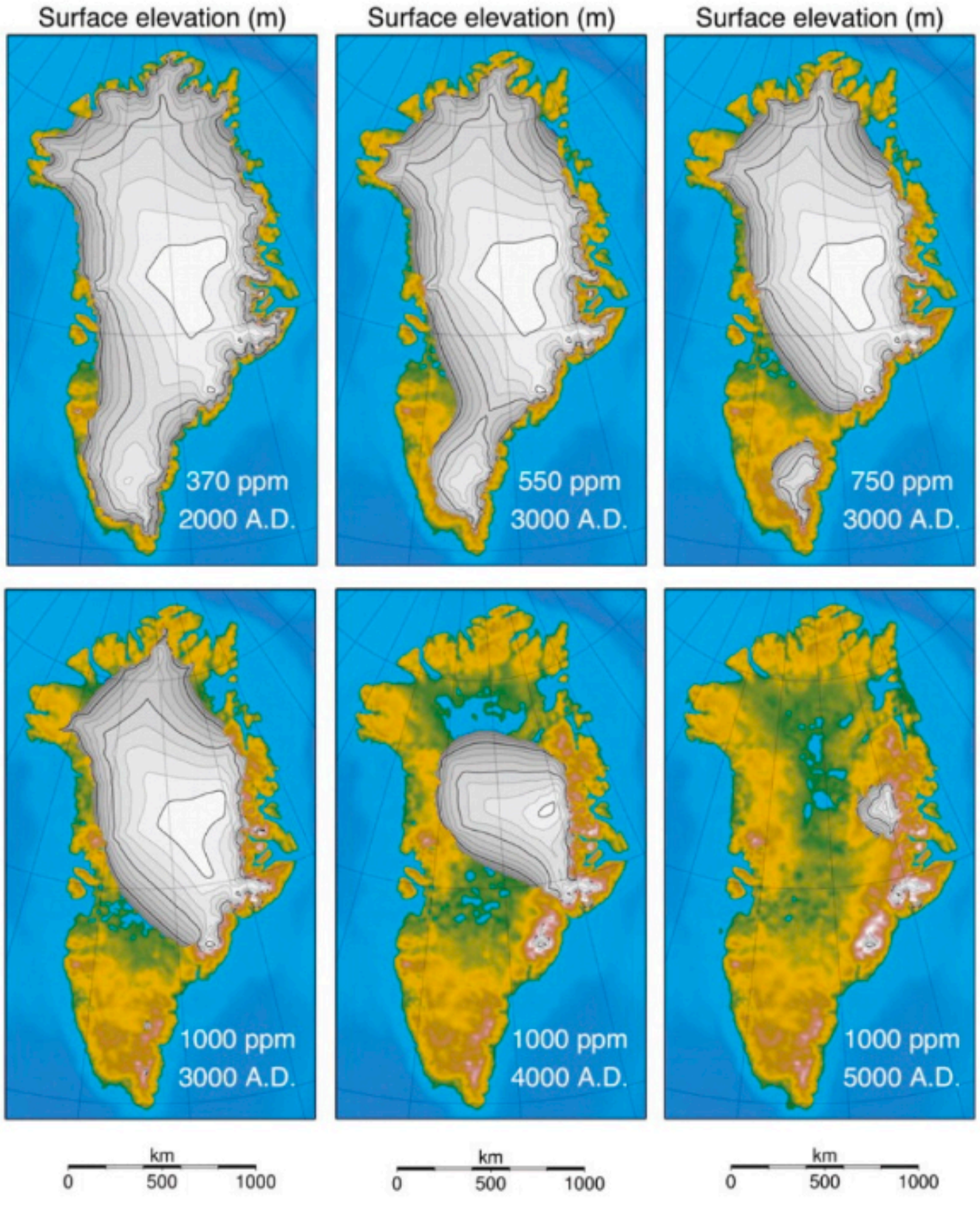
The uncertainties remain great, however, which should not necessarily be a source of solace.

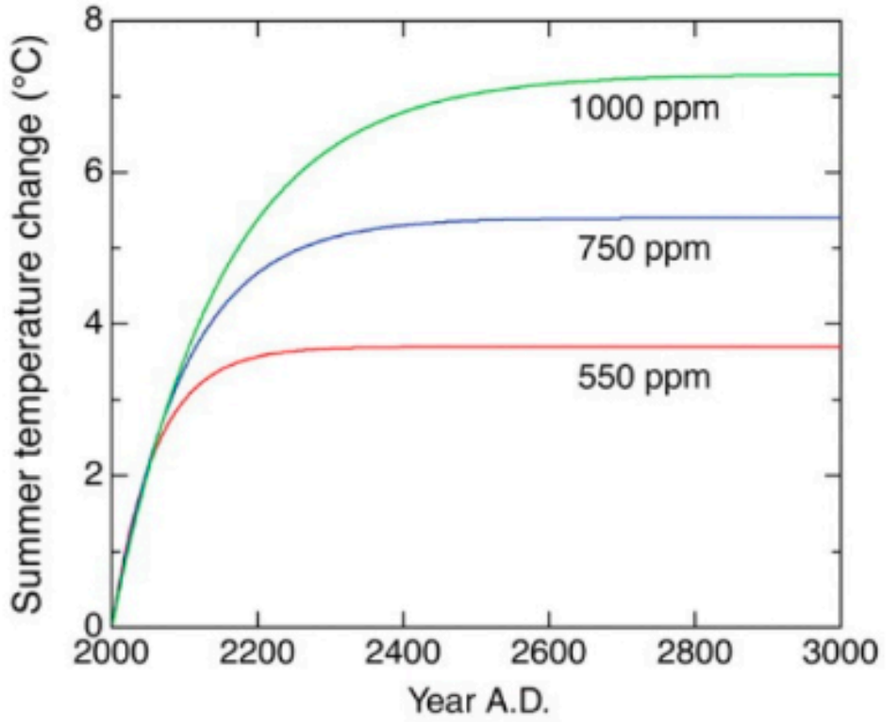
As stated in the abstract of the 2010 study:

> *The evidence suggests nearly total ice-sheet loss may result from warming of more than a few degrees above mean 20th century values, but this threshold is poorly defined (perhaps as little as 2°C or more than 7°C).*

Guided by this paleoclimate data, Richard Alley and colleagues have modeled the likely longer-term dependence of the Greenland ice sheet on temperature and, using radiative forcing estimates, on the CO_2 abundance.

Figure 9.5 shows that various values of the peak CO_2 abundance are assumed, depending on when global emissions might substantially cease.





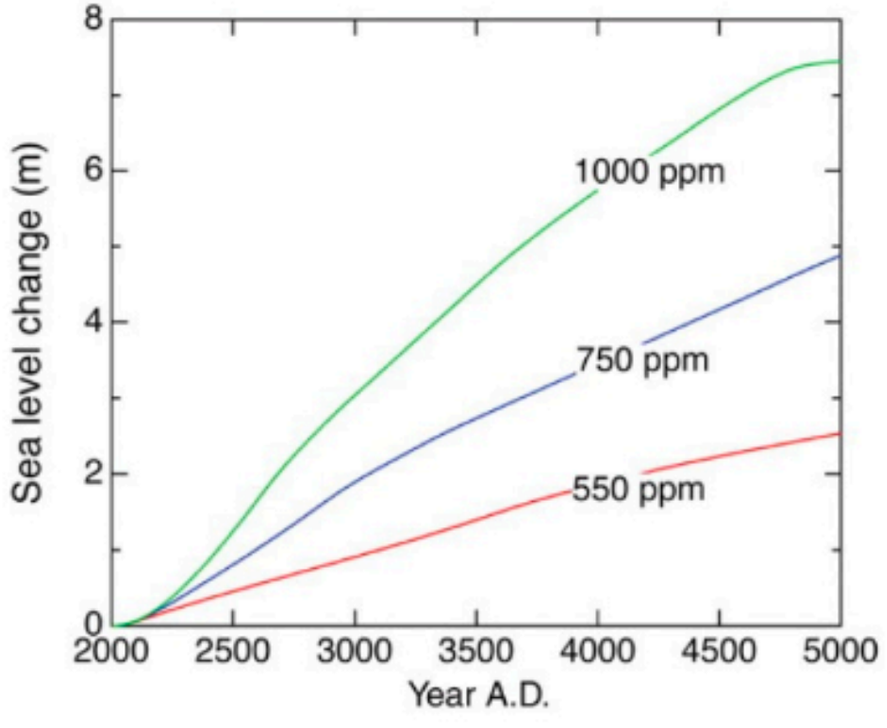


Figure 9.5

In the near term, the mean estimate in 2010 for the guaranteed sea level rise by 2100 due to Greenland ice sheet melting, given a more-or-less certain rise of 2°C before the end of the century, was about 0.05 meters.

However, since that time ice mass loss has doubled, reaching the maximum estimated melting rate predicted in models, so a newer estimate closer to 0.1 meters now seems more appropriate.

The National Academy quotes an upper bound of 0.16 meters, assuming continued increase in melting combined with a possible increase in ice discharge at the coasts, which currently contributes an equal amount to melting in the total ice mass loss.

As increased precipitation due to increased temperature is now known to not be a significant factor in reducing mass loss, other factors not included in models would tend to increase the actual mass loss rate and would bring them into closer alignment with recent observations.

Such factors include recently observed lubrication at the base of glaciers due to large surface melting, which can increase the rate of glacier movement into the sea.

And most recently in the 2019 Greenland data, it was observed that much of the extreme melting that year was not directly due to higher temperatures, but rather to high-pressure weather systems in the region that may have been an indirect result of temperature increase.

These blocked the formation of clouds, which allowed unfiltered sunlight to melt the snow.

In addition, with less snowfall, darkened dirt and soot-covered ice was exposed, which altered the albedo, absorbing more heat than would otherwise be the case.

As a result, a Greenland contribution of 0.1–0.16 meters to sea level rise seems a reasonably conservative range to expect this century.

Moving across the globe to Antarctica, the situation is quite different.

There, snowfall increases in East Antarctica have on average balanced ice loss in West Antarctica.

Again, this is not unexpected on the basis of our earlier predictions, as near-term warming is predicted to be less severe in the high-latitude South, and increased precipitation is predicted there as well.

Nevertheless, over the past two decades the West Antarctic ice sheet has undergone noticeable changes, which I became acutely aware of when a friend visited and sent me an image.

The immensity of these icebergs in that region is hard to fathom until you see them up close.

When one passes close by your boat, it stands taller than a skyscraper and stretches as far as you can see.

My friend tried to capture one from the bow of his vessel.



But these icebergs are small potatoes.

A few months before my friend's arrival, in the dark of the Antarctic winter, a giant iceberg, 225 square miles, about ten times the size of Manhattan, had broken off from the Pine Island Glacier on the Amundsen Sea.

It was not the first such iceberg, the largest, nor would it be the last.

Just this year it shed another smaller clump 120 square miles in area.

Since 2012 this single glacier has been shedding fifty-eight gigatons of ice a year, making the biggest contribution to sea level rise, of about 0.1 mm/yr., of any single ice floe on Earth.

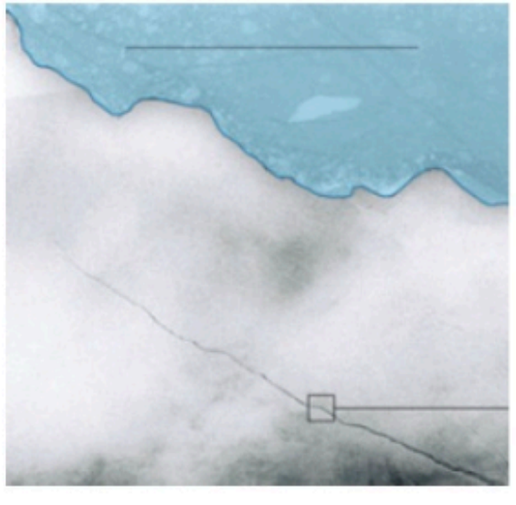
These are not the biggest recent icebergs to form off the West Antarctic coastline.

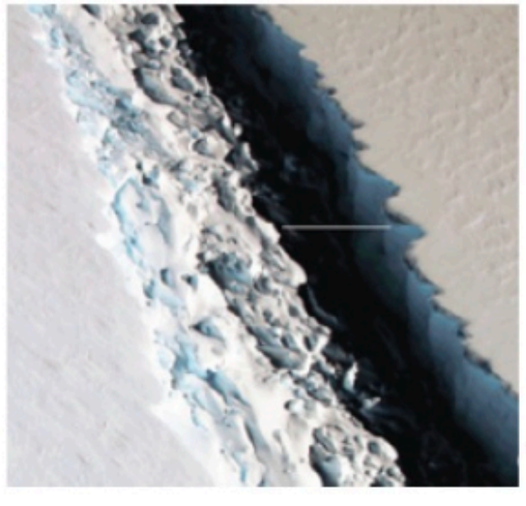
A massive ice shelf, called the Larsen Ice Shelf, on the eastern side of the West Antarctic peninsula, opposite the location of the Pine Glacier, has broken off three spectacular state-sized pieces since 2002, with the last and largest, called Larsen C, breaking off in 2017.

It weighed one trillion tons and measured about 2,200 square miles, the size of Delaware.

The photographs of the rift as it broke off were remarkable.







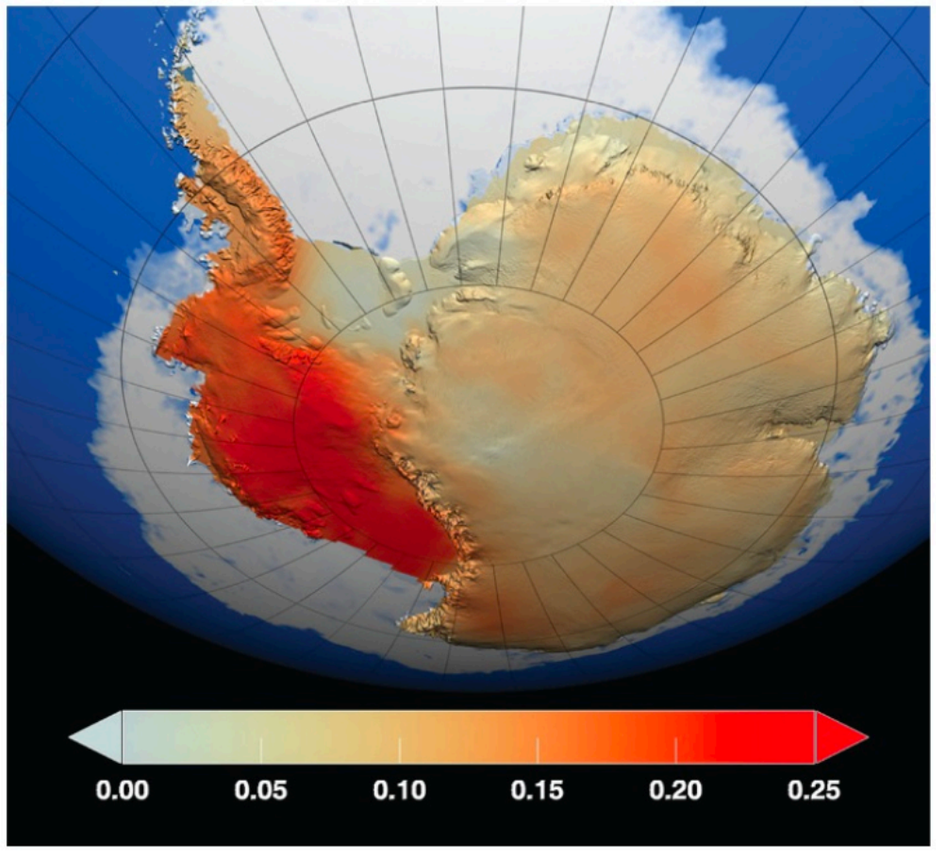
As large as these are, the Larsen shelf calvings do not have any impact on global sea levels because the original ice shelves were already floating on the water.

But nevertheless, what they may portend is somewhat more ominous.

While Antarctica has not warmed as extensively as the Arctic over the past decades, it has warmed.

Changing wind patterns and related warm deep-ocean currents being pushed southward, as predicted, have warmed West Antarctica most, as seen here with the average temperature change from 1957–2006 displayed (Figure 9.6).

Figure 9.6



This warming trend has accelerated ice discharge from West Antarctica.

In this case melting is not the primary cause of loss of ice mass.

A more extreme impact comes from the effects of warming from warm seas on ice shelves like Larsen.

As can be seen in Figure 9.7, the ice shelf, which is floating buoyantly on the water, can be undercut as warmer deep water melts the ice underneath it.

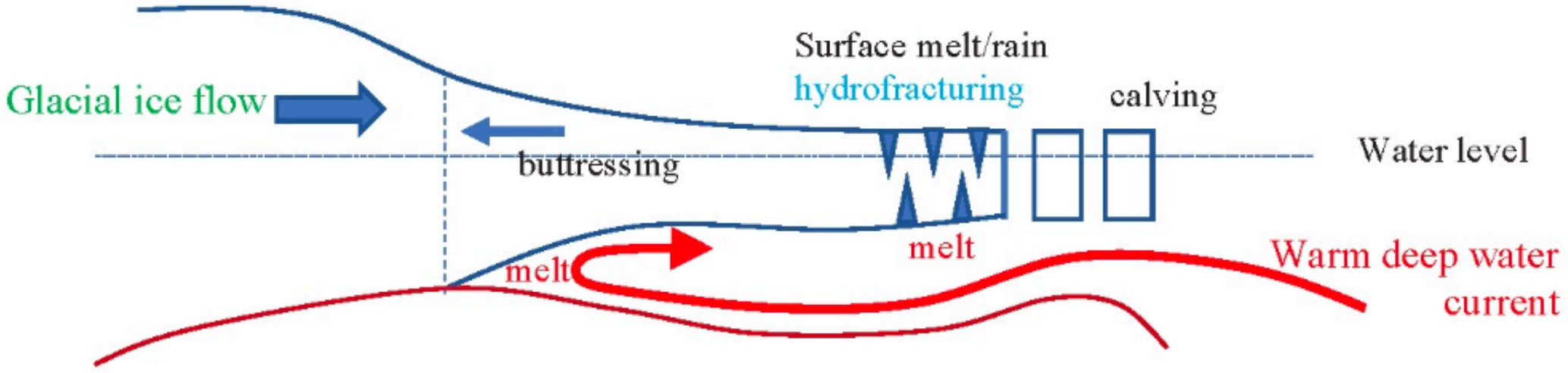


Figure 9.7

When combined with surface effects, the sheet becomes unstable and breaks apart in the process of calving.

Eventually the entire ice sheet disappears (as Larson A, B, and C have), as shown in Figure 9.8.

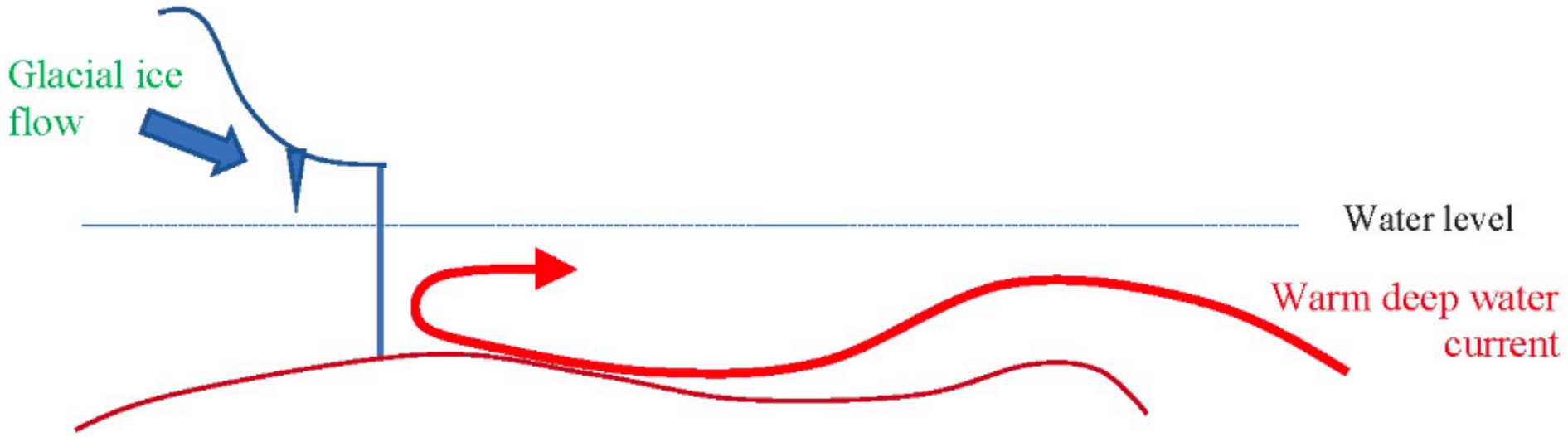


Figure 9.8

At this point, not only can the ocean current continue to undercut the remaining ice wall, but the buttressing provided by the floating ice shelf that has been holding the glacier back is removed.

This allows the glacier to flow unfettered into the sea.

After the Larson B ice sheet disappeared, the glaciers feeding it sped up by up to a factor of eight, contributing an additional .07 mm/yr. to sea-level rise.

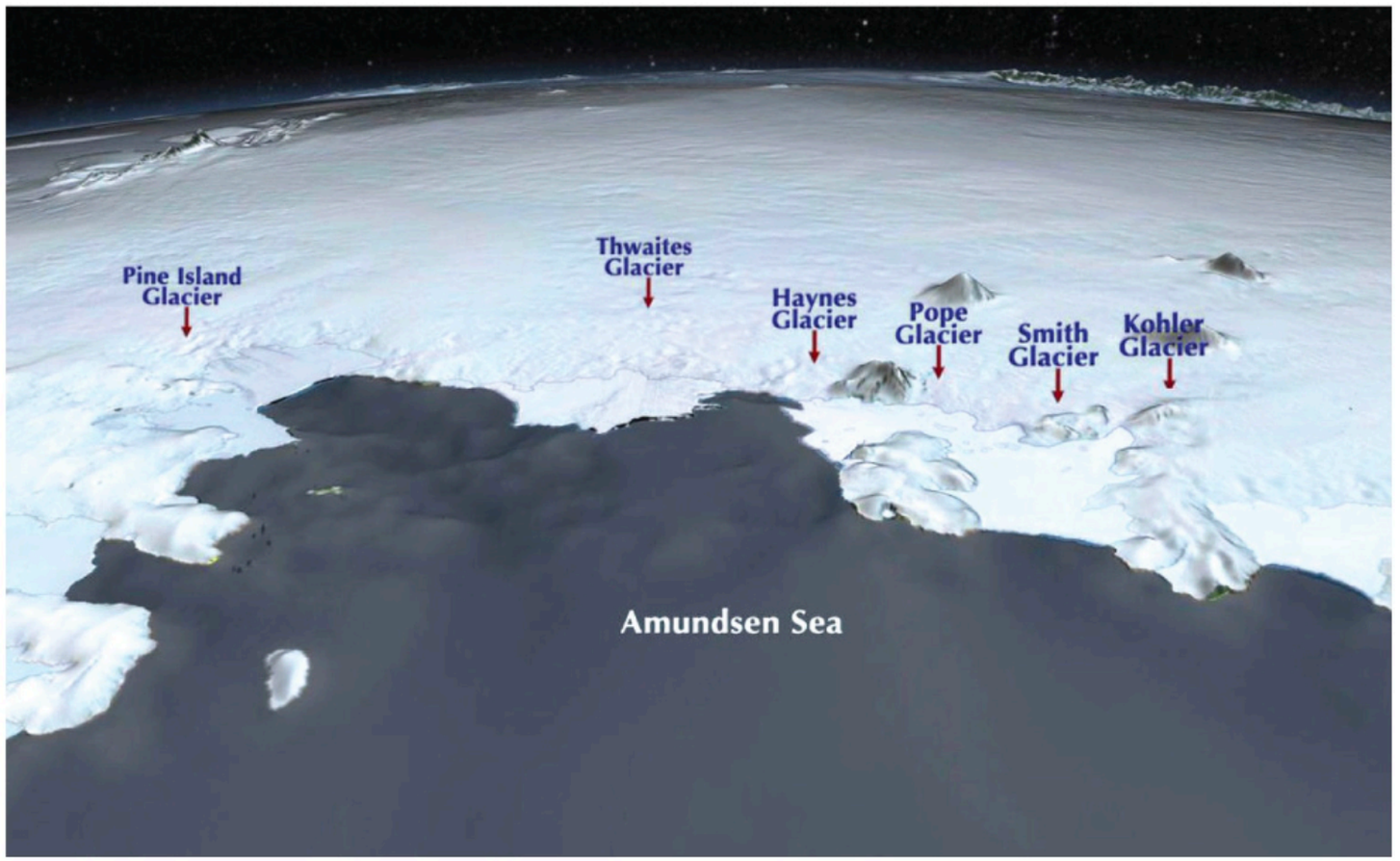
A similar effect was observed when Larsen A ice sheet was lost.

Of greatest concern, however, is the Amundsen Coast Basin that contains the Pine Island, Thwaites, and Smith Glaciers, among others.

These are not currently buttressed by large ice shelves and are accelerating and free-flowing at more than three times the speed of the average Antarctic glacier.

Even though the glaciers that would feed ice shelves on the coast have been accelerating, the ice shelf hasn't been growing, indicating that they are probably melting from below due to warmer water.

As this point in time, a *Proceedings of the National Academy of Sciences* study suggested even more rapid breakdown of both Thwaites and Pine Island shelves than previously thought.



Currently these are contributing an ice-mass loss in excess of accumulation that accounts for between 0.14 mm/yr. and 0.24 mm/yr.

Suppose just these three glaciers continue to accelerate at their current rate and double their current velocity by the end of this century, as models suggest, while the rest of the ice on the Peninsula continues to lose mass at a rate consistent with the current surface ice mass balance.

Then by 2100, mass loss in the Peninsula will contribute at least 0.12 meters to global sea level rise.

That is significant, but it could be much worse—albeit depending on the accuracy of detailed model estimates of the glacier dynamics—and on a quicker time frame than many centuries or millennia.

Two studies in 2014 suggested that these systems have already destabilized, and their disappearance within a century or two is inevitable.

What makes this worrisome is that if released, the ice currently stored in these basin glaciers would produce 1.5 meters of sea level rise.

The situation could get even worse.

A study in the *Proceedings of the National Academy* in 2015 modeled the glacier systems and fit them to their current rate of motion and melting.

It concluded that even if the temperature is turned back down to its value in the 1970s and 1980s by some as-of-yet- unknown effect, not only does the destabilization of theses glaciers still continue, but the entire West Antarctic ice sheet becomes destabilized.

Over centuries to millennia this would add an additional three meters to global sea levels.

We are now fully in the realm of disastrous “might be” Ghost of Christmas Future scenarios, which, while plausible, are not yet firmly established.

So instead let’s step back and tally up what we have ascertained is more or less guaranteed in the way of sea level rise by 2100.

If one extrapolates from current heating and current changes in ice mass loss rates in Greenland and Antarctica, and adds that to the already built-in thermal expansion of the ocean, as well as the impact of the melting of other glaciers and ice caps around the world, one arrives at:

Global Sea Level Rise Built-In by 2100:

Thermal expansion will contribute	≈0.23 ±.09 meters
Other glaciers and ice caps will contribute	≈0.37 ±.02 meters
With current acceleration of melting, Greenland will contribute	≈0.16 meters
With current acceleration of discharge, Antarctica will contribute	≈0.12 meters
Total	≈0.88 ±.12 meters

You get a sense of the extrapolation required to produce the above estimates for Greenland and Antarctic ice mass loss from Figure 9.9, which shows the relative measured contribution of each to ice mass loss and sea level rise for the period 1989–2018.

The growing significance of Greenland ice melt is clear, and the downward curvature gives a sense of the current acceleration in the rate of mass loss.

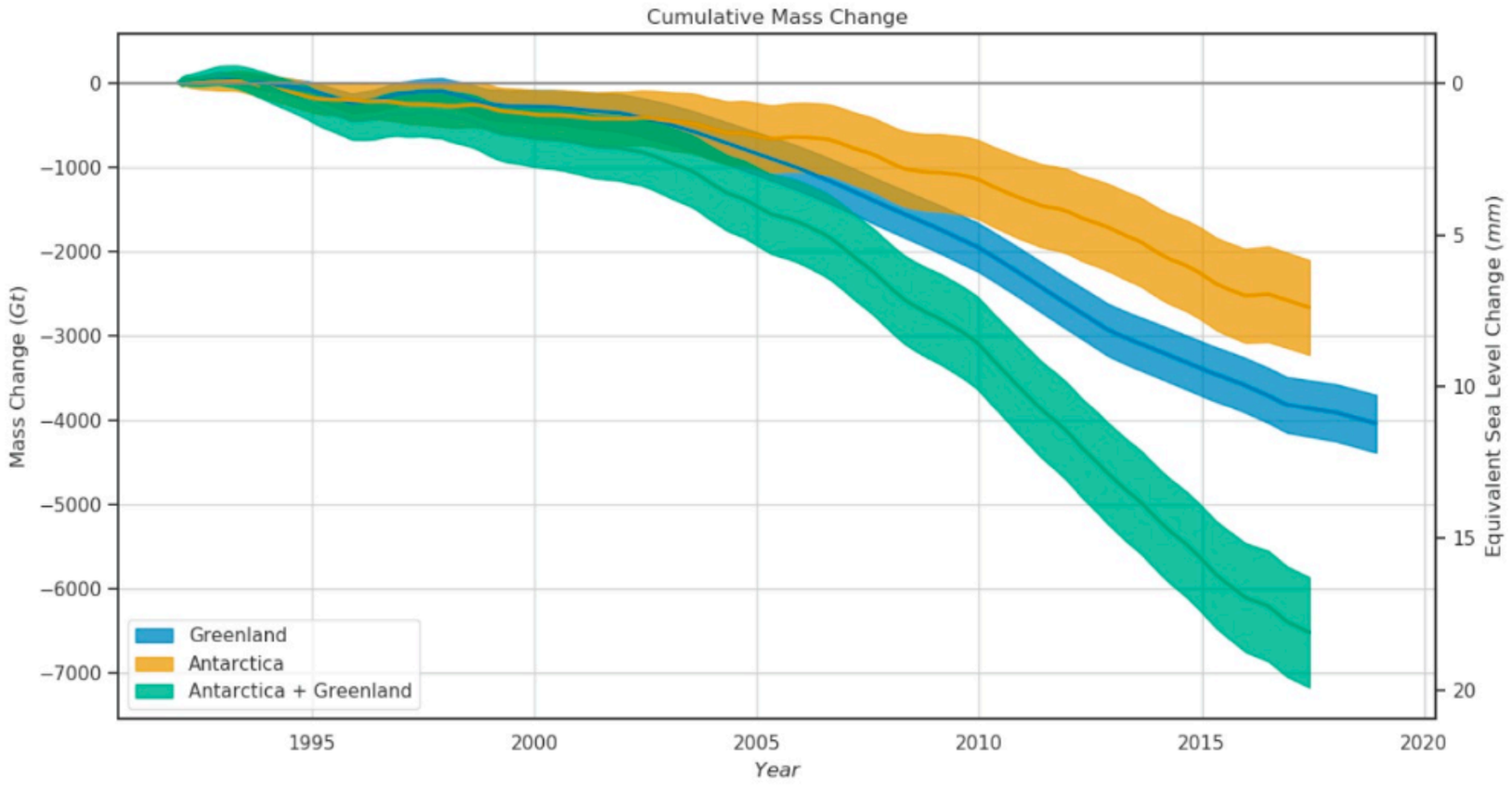


Figure 9.9

If instead the current accelerated ice loss in Greenland levels off, then it will contribute only 0.05 meters to global sea level rise by 2100.

If at the same time, the West Antarctic Peninsula stabilizes so that the East Antarctic ice growth balances the West Antarctic ice loss, and Antarctica as a whole contributes nothing to net sea level rise this century, then the *minimum expected sea level rise*, already built in to the planet on the basis of the radiative forcing thus far, will be 0.65 ±0.12 m by 2100.

In other words, the minimum sea level rise we can expect this century is 0.5 meters, while the maximum we can expect, assuming no catastrophic collapse of glaciers in Antarctica or increased acceleration of melting in Greenland due to other climate feedback effects, is 1.0 meter.

These two values bracket the expected range for the best-case scenario of the future *as it will be*.

Faced with measured sea level rise of a mere 3–4 cm since the turn of this century, it is easy for many people to view concerns about future sea level rise in the coming fifty to eighty years as premature.

And for some, it may seem that even a 0.5–1.0-meter sea level rise is not such a big deal.

But as I shall later elaborate, it is.

I have saved what may be the worst for last.

We have seen that increased CO_2, leading to increased global temperatures, will result in sea level rise due to the well-understood physics of thermal expansion and also due to more subtle effects from differential changes in climate and ocean currents across the globe.

In spite of the fact that the extrapolations to 2100 and beyond that I have described arise from a vast amount of data that has been accumulated over the past thirty to forty years, some skepticism about depending solely on climate models might still seem justified.

Significantly larger sea-level changes in this century or millennium may still seem like the stuff of science fiction.

There is, however, one more set of compelling data that I have not yet shown.

That involves paleoclimate measurements of the type I discussed for Greenland, but in this case ones that go back well beyond the direct CO_2 and temperature record and that also trace global sea levels over this time.

Using a host of markers, from coral reefs to isotopic abundances to coastal traces in stable geological structures, a clear correlation between CO_2 abundance, ice volume, and global sea levels emerges.

If one goes back more than sixty-five million years ago to the age of dinosaurs, when atmospheric CO_2 was as high as 1250 ppm, sea levels were as much as 73 meters higher than they are now.

By thirty-two million years ago, atmospheric CO_2 decreased to about 500 ppm—*comparable to what they are likely to be by the mid-twenty-first century*—and sea levels had subsided to merely 28 meters higher than at present.

On the other hand, during the last glacial maximum just twenty-one thousand years ago, the sea level was as much as 130 meters lower than it is now.

Again, it may be tempting to dismiss extrapolations from millions of years back when direct measures of CO_2 and global temperatures are not available.

However, the Red Sea is particularly sensitive to sea-level changes because of the narrow and shallow connection with the open ocean, so that lowering of the sea level decreases transport through the channel and increases the residence time of water in the Sea.

Careful modeling of flow in this channel, combined with oxygen isotope measurements in sediment cores, then give estimates of sea level as a function of time.

Estimates of sea-level low points during periods of maximum glaciation have been independently corroborated by other means, and the estimation method has been extended to allow continuous records of sea levels with careful attention to achieving time resolution on the order of a century where possible.

For the past twenty thousand years Red Sea estimates of global sea levels can be compared to coral reef data taken from around the world and good agreement is obtained.

Figure 9.10 is a recent compilation.

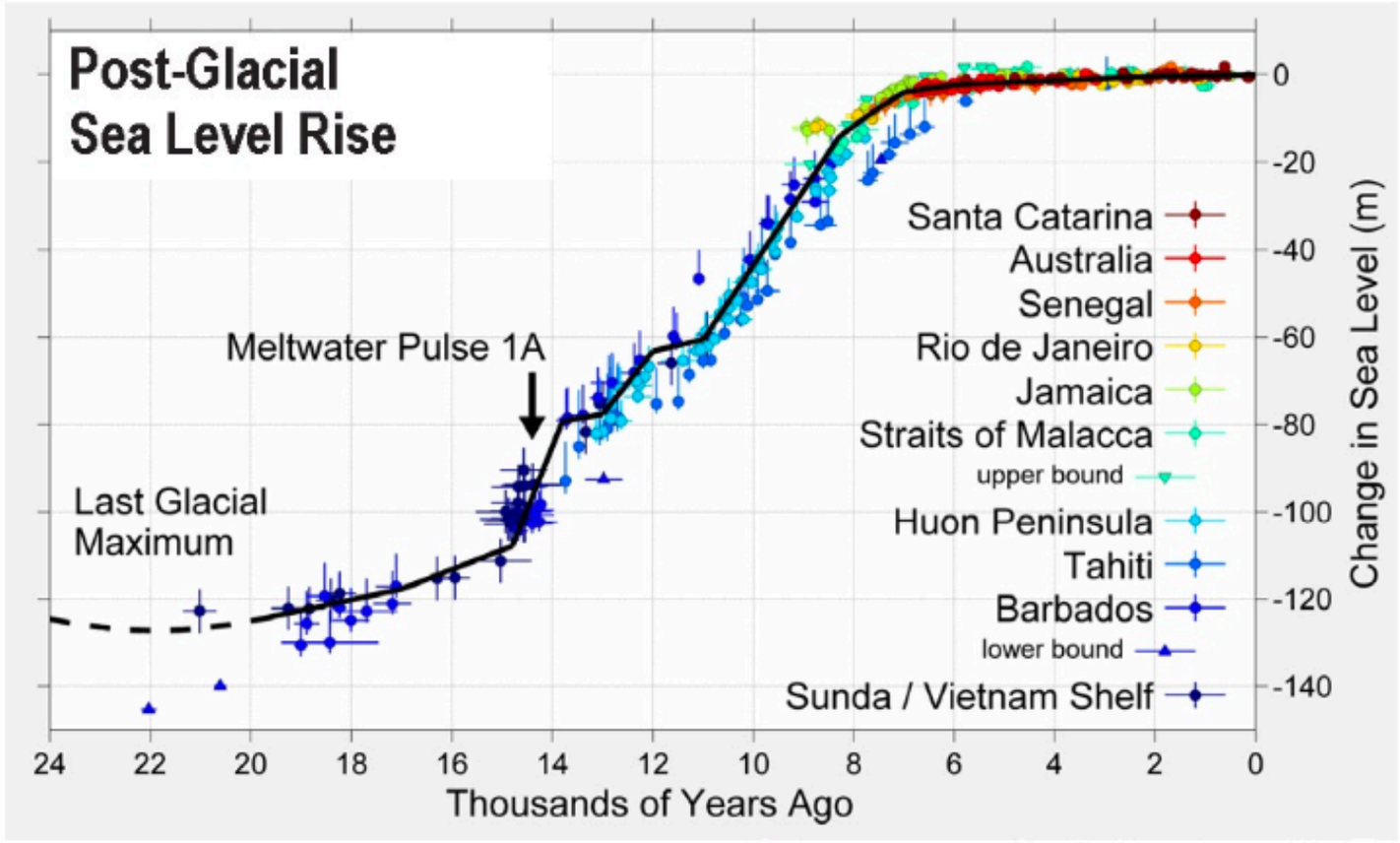


Figure 9.10

A reduction in sea level on the order of 120 meters during the period of maximum glaciation, mentioned earlier, can be clearly seen.

Equally important is the fact that the sea level has varied by more than 10 meters within a few centuries, as can be seen during the period between fourteen and fifteen thousand years ago—compatible for example with a total loss of sea ice in Greenland within a millennium.

For those who doubt significant sea-level change is possible, or who doubted the predicted relatively direct connection between CO_2 levels, global temperatures, and sea-level rise,

I now finally turn to the figure that first shook me out of my own complacency in this regard.

Using Red Sea data, one can produce a curve of inferred sea level going back 440,000 years, which allows a comparison with ice core data.

Figure 9.11 is from Vostok ice cores I showed earlier comparing measured CO_2 abundance with temperature.

Figure 9.11

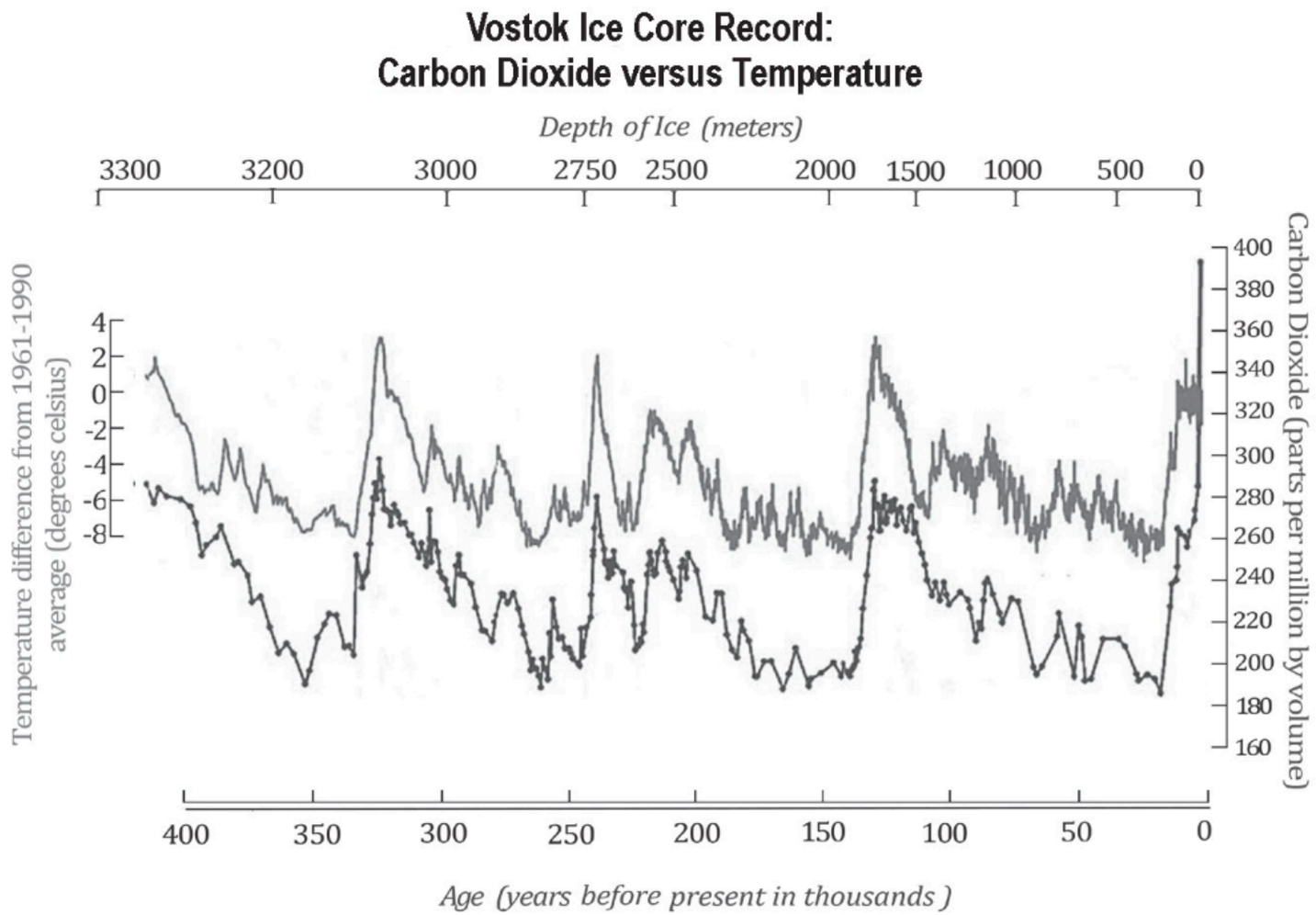
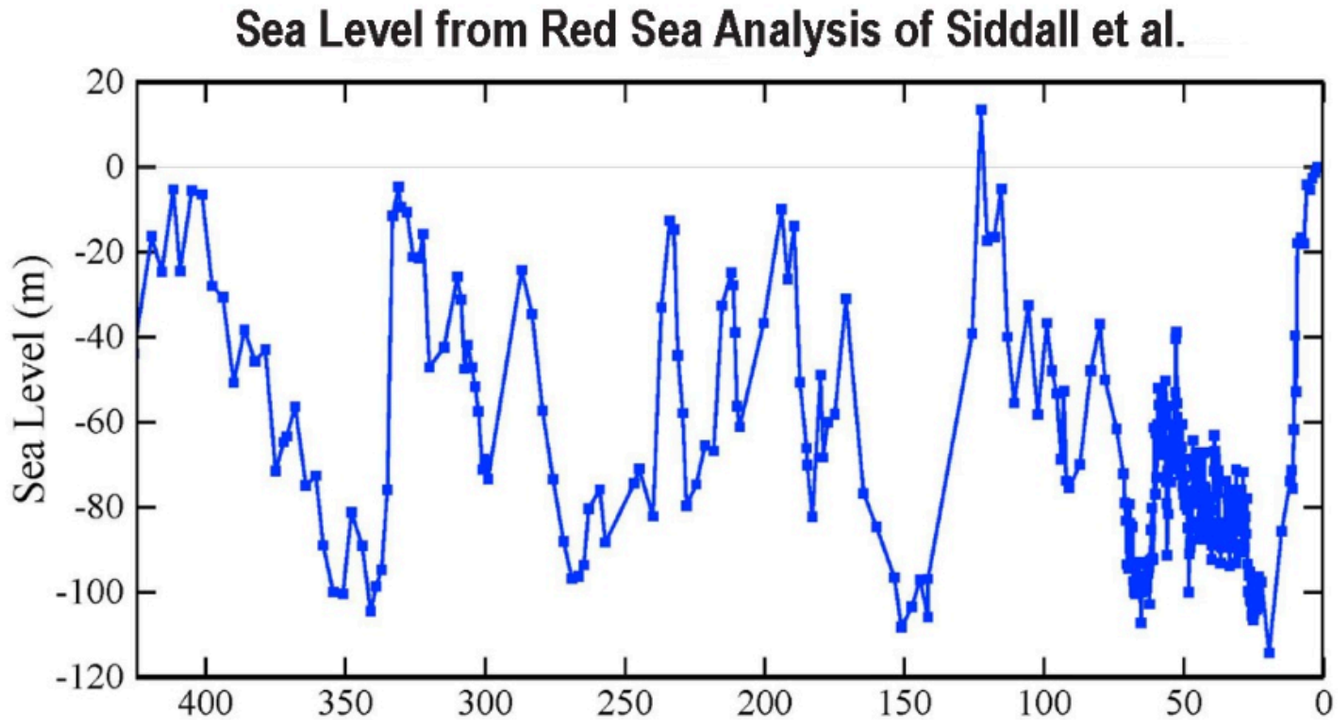


Figure 9.12, by Jim Hansen, is of the inferred global sea level, based on Red Sea data.

Figure 9.12



Global sea level extracted, via a hydraulic model, from an oxygen isotope record for the Red Sea over the past 470 kyr (concatenates Siddall's MD921017, Byrd, & Glacial Recovery data sets; AMS radiocarbon dating).
Source: Siddall et al., *Nature*, **423**, 853-858, 2003.

It is true that correlation is not causation, but if one has a physical model, as we do, that predicts such a correlation, then observing one as clear as the one shown here between CO_2 abundance and sea levels is highly suggestive, to say the least.

Finally, I remind you that the current CO_2 atmospheric abundance of 415 ppm is already off the upper chart and the predicted CO_2 in 2100 will be even higher, almost a factor of two larger than any datapoint shown over the past four hundred thousand years, during which sea levels varied by over 130 meters.

While the temporal resolution of these plots is not sufficient to reliably predict what is likely to happen over the course of a single century, we can, on the basis of the historical record, assess that a likely sea-level change on the order of meters in a few centuries is not implausible, and thus decide with more confidence whether planning or acting now in the face of such a possibility is worth the risk.

10 - CLIMATE CHANGE TODAY

Earlier I said that there is a difference between “weather” and “climate” because this distinction is relevant to how we interpret our daily lives.

One anomalously cold day in Washington, DC, doesn’t imply no global warming, and one ultra-hot week in the summer can’t necessarily be blamed on it either.

It is worth considering, however, whether any sort of weather measurements might ever yield a statistically significant probe of climate change.

If you measured the weather conditions everywhere around the world on a given day, or measured the weather in one location for a full season, could you discern definitive evidence that radiative forcing is causing climate change?

Recently a Swiss team of researchers claim to have demonstrated the first possibility.

Using advanced algorithms and data for both temperature and humidity for many locations around the world each day since 2000, they claim that they can pinpoint data for any day that provides statistically significant evidence of global climate change.

Their specific claim is that by using temperature and humidity readings from around the world on a single day, they could work backward to extract a “fingerprint” of radiative forcing and hence warming.

The fingerprint is determined by machine learning algorithms that explore the space of model predictions and compare these to the data obtained from around the globe.

This is a bold claim and the statistical techniques used by this group are far from intuitive.

Their analysis, to my knowledge, hasn’t yet been repeated by other groups.

Nevertheless, their result is perhaps not as unexpected as it sounds on first hearing.

In a single location, weather can vary by tens of degrees from the mean temperature (as measured over some sufficiently long time span) on any day from year to year.

This is why a snap freeze somewhere in April or a hot day in Washington in December doesn’t provide evidence for or against global climate change.

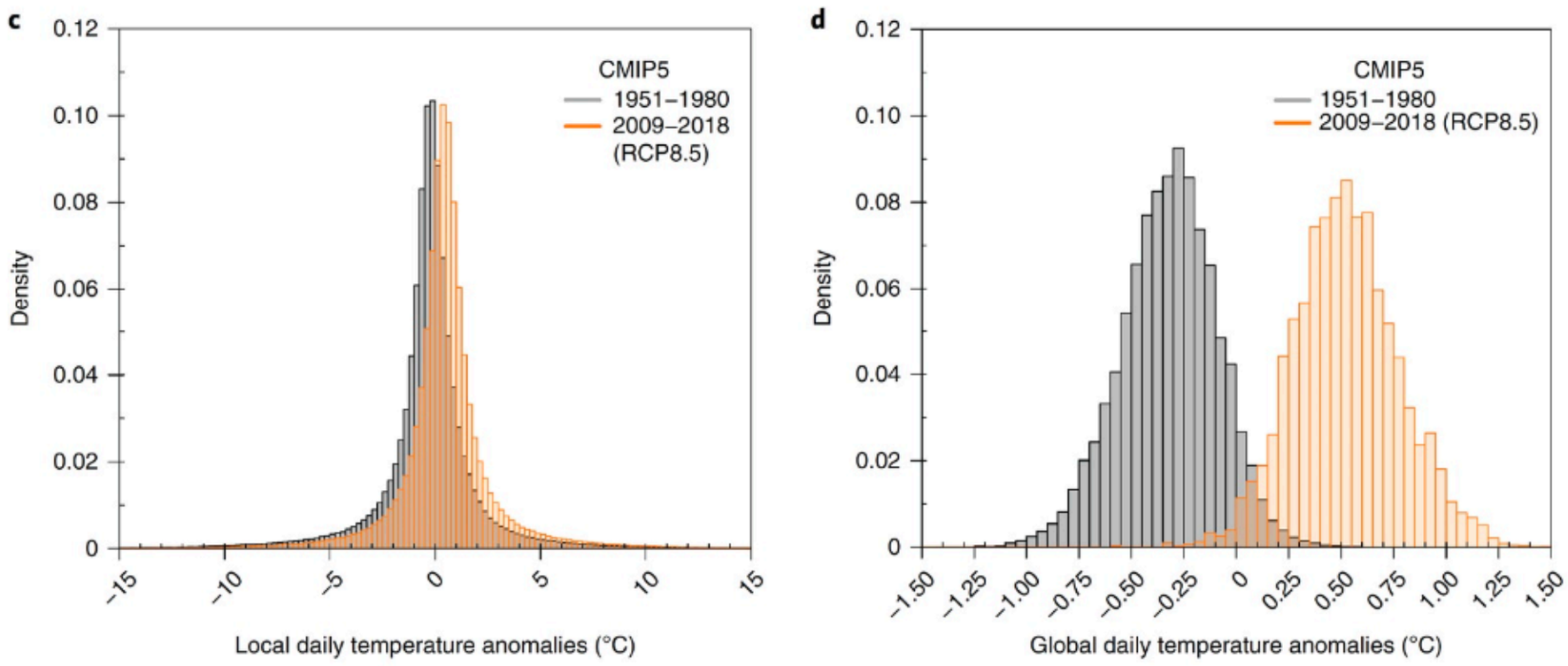
But if one compares temperatures from many different locations around the world on any day to mean temperature at each location that day, and then averages over the variations, one finds that large random daily variations in various single locations tend to cancel out from the full data set.

Basically, it is an example of the statistics of large numbers.

Just as individual humans often behave inexplicably when considered individually, when one averages over all of human population one may isolate statistically significant trends that yield insight into the human condition.

Shown in Figure 10.1 are their results from a simpler analysis than the full fingerprinting project that demonstrates the power of using a global analysis to explore even daily data.

Figure 10.1



CMIP5 in the legend stands for "Coupled Model Intercomparison Project Phase 5," which is a data archive used to analyze climate models and from which these data were extracted.

The figure on the left is built up by recording the variation in the local daily temperature and comparing it to the mean local daily temperature for that day in that location.

The distribution of such variations is then compiled for many different days.

Then this distribution is combined with similar data sets from other locations around the globe to get good statistics for the average distribution of daily local temperature variations.

The second method first averages the temperature deviation on a single day over many different locations around the globe.

Some will be positive and some negative, and the average variation will thus be smaller.

One then plots the distribution of these averages for many different days

Two things jump out.

First, the statistical distributions behave as expected for large data sets.

They are roughly symmetric about the mean and distributed with more or less Gaussian shapes.

Next, as expected, the distribution of average temperature variations on a single day for a single location has a much bigger spread than the distribution of the variations on that day when first averaged around the globe.

Because of the smaller spread of global daily temperature variations, the measured distributions for different eras can be more effectively compared.

Here the global average daily temperature distributions during two different periods, 1951–1980 and 2009–2018, are plotted versus the mean daily temperature from 1979–2005.

The fact that most of the range for the first period is negative implies that an average day was colder during the period 1951–1980 than it was from 1979–2005.

The opposite is true for the average day during 2009–2018.

The difference of about 0.75°C between the peaks of the two distributions from these different eras suggests that even daily estimates, normally taken as evidence of weather and not climate, can, if suitably globally averaged, provide a signal of climate change.

If global daily temperatures can provide clear evidence that global warming has recently occurred for periods separated by decades of time, can one find some other *local* evidence that a global warming signal can be clearly separated from natural variability from one year to the next?

The answer is yes, at least for some locations, provided one considers not daily temperatures but seasonal temperatures.

As the question suggests, the issue comes down to separating signal from noise.

Shown in Figure 10.2 is a distinct oscillating signal.

Below that, the same signal is displayed with significant noise overlaid on it.

If you had just the second piece of data would you be able to infer the original pattern of oscillation?

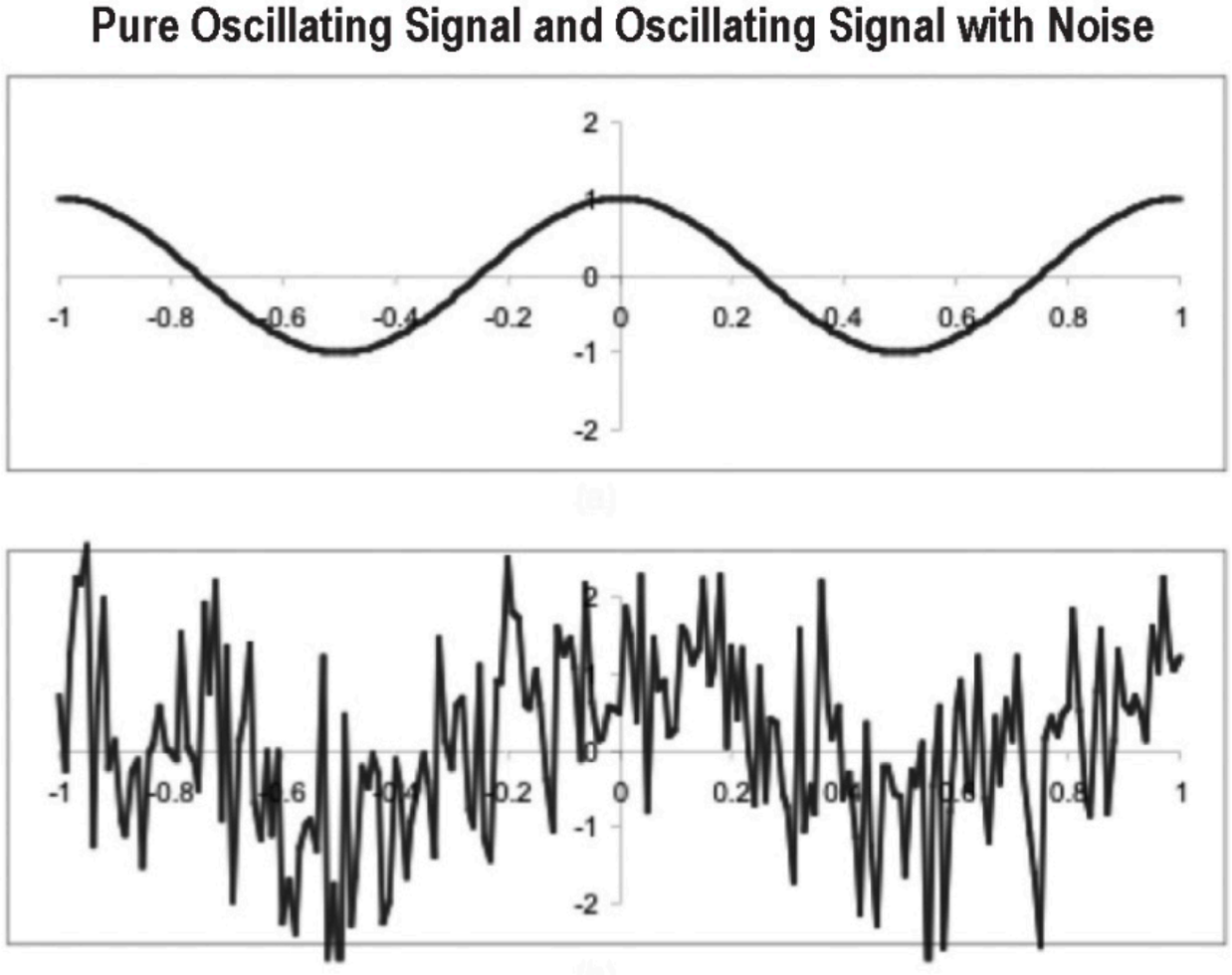


Figure 10.2

The climate scientist Susan Solomon and colleagues have explored separating signal from noise by minimizing noise.

They examined regions and seasons where natural year-to-year variability in temperature has been shown to be small and found that the warm season in equatorial regions seems to fit the bill.

Figure 10.3 shows results displayed with different shades representing the difference in average temperatures for 1990–1999 vs. 1900– 1999, for model predictions (left) vs. observations (right).

In all cases, observations are noisier than the models, as one should expect, since the models don't contain all the messiness of the real world.

But during the warm season in the tropics, circled in the left of the lower two sets, year-to-year variation is small in both the models and data.

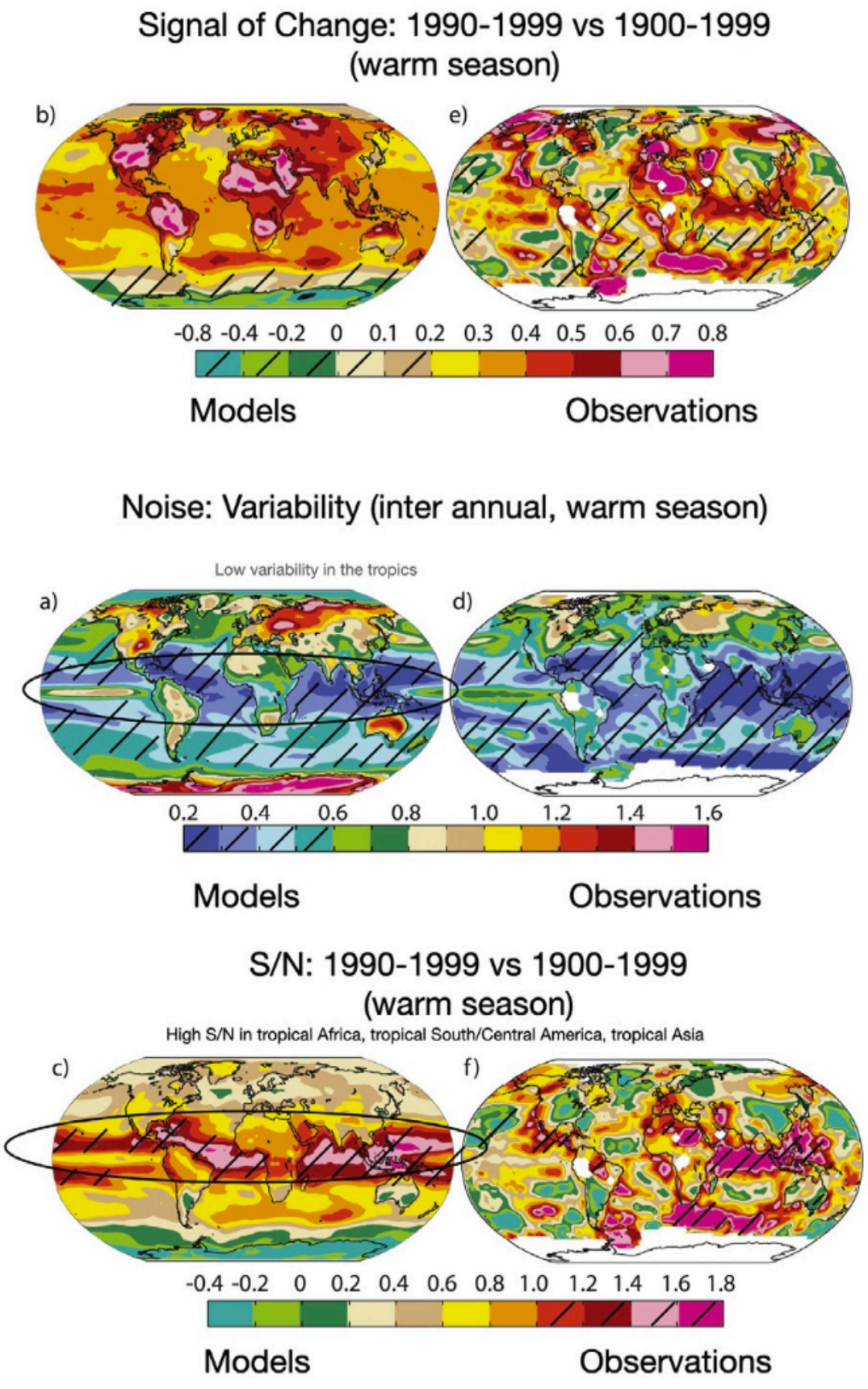


Figure 10.3

This is also not surprising.

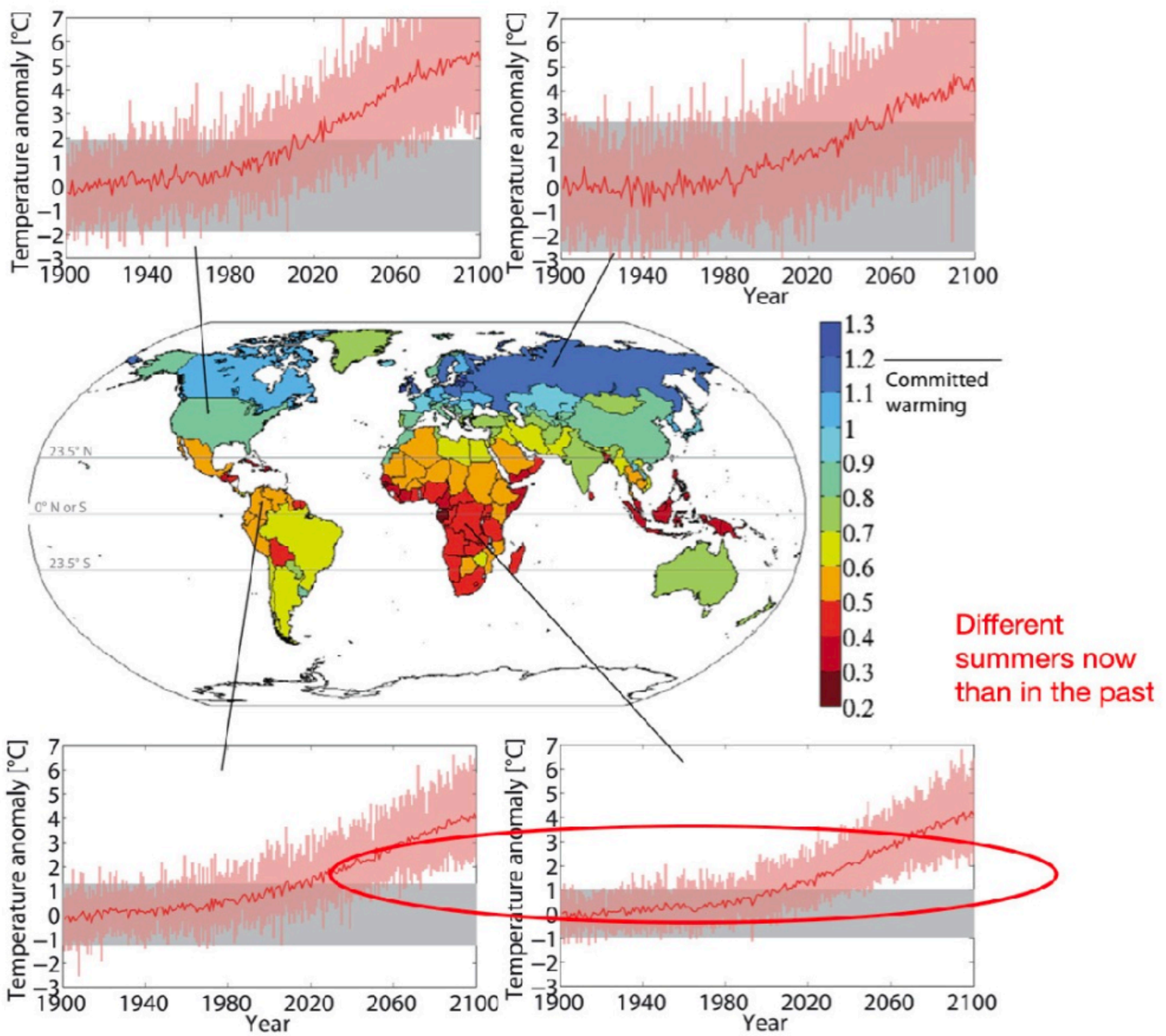
For anyone who has travelled in these regions, the weather report is almost always the same, something like eighty-eight degrees at 3:00 p.m. with a chance of rain, for example.

By comparing signal to noise in the model predictions, the significance of the signal in the tropics is clear but is visually less obvious for the actual data.

However, if one plots the observed warm season temperature range each year versus the more-or-less-constant variability from year to year for a variety of locations around the world —from the 2.5°C average variability from summer to summer in Russia and ±2°C in the US to the ±1°C average variability in Central America and equatorial Africa—a clear signal emerges in the equatorial region.

By 2020 the summer there is measurably different than it has been in the past (Figure 10.4).

Figure 10.4



The first regions that therefore are already experiencing clear effects of global warming on a year-to-year basis are the equatorial and tropical regions of the world.

What makes this particularly unfortunate is that these include the countries that have been, on the whole, least involved in contributing to the cause (in the black box on the graph), where CO_2 emissions in tons per year per person in 2006 are shown in Figure 10.5.

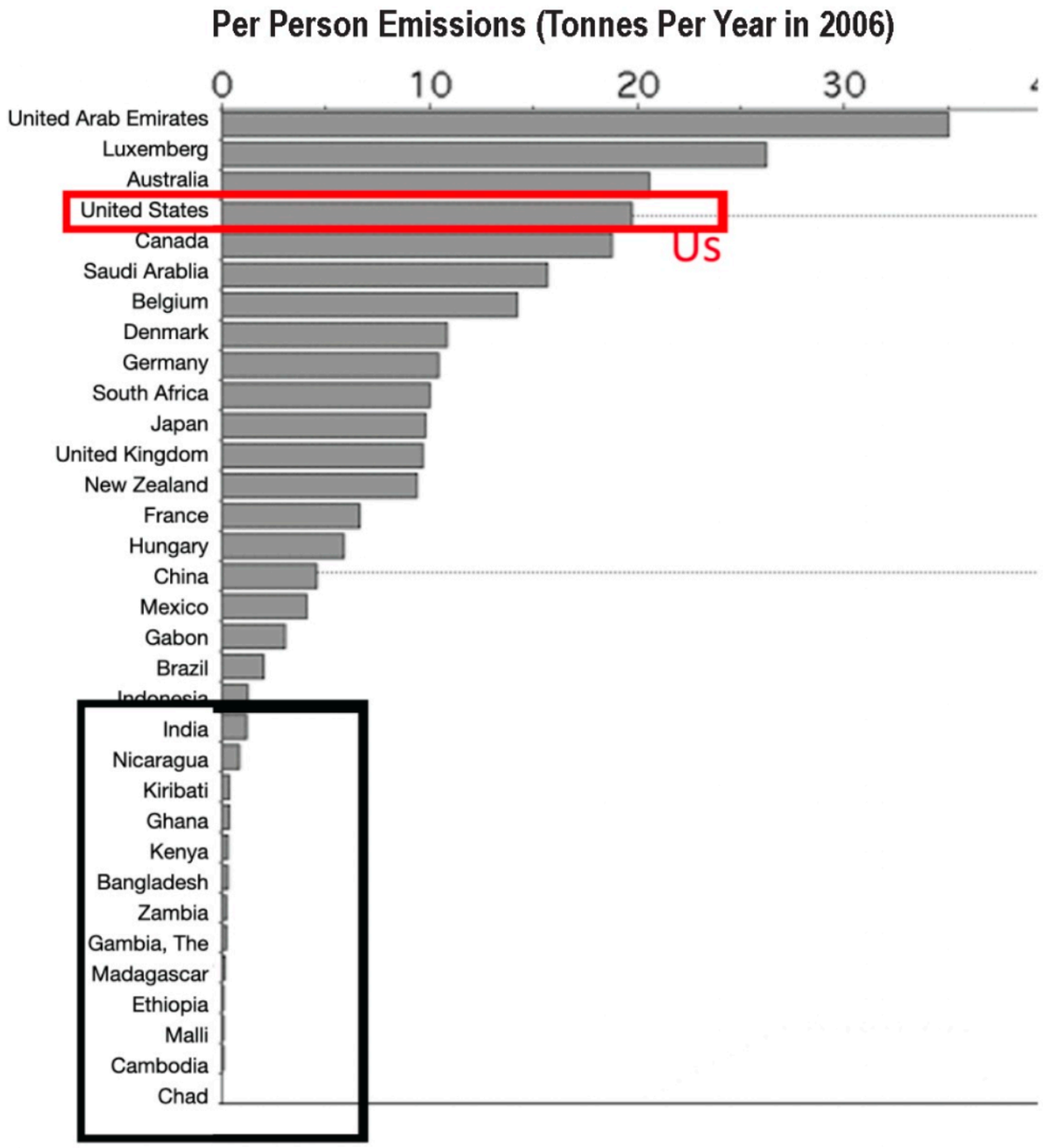
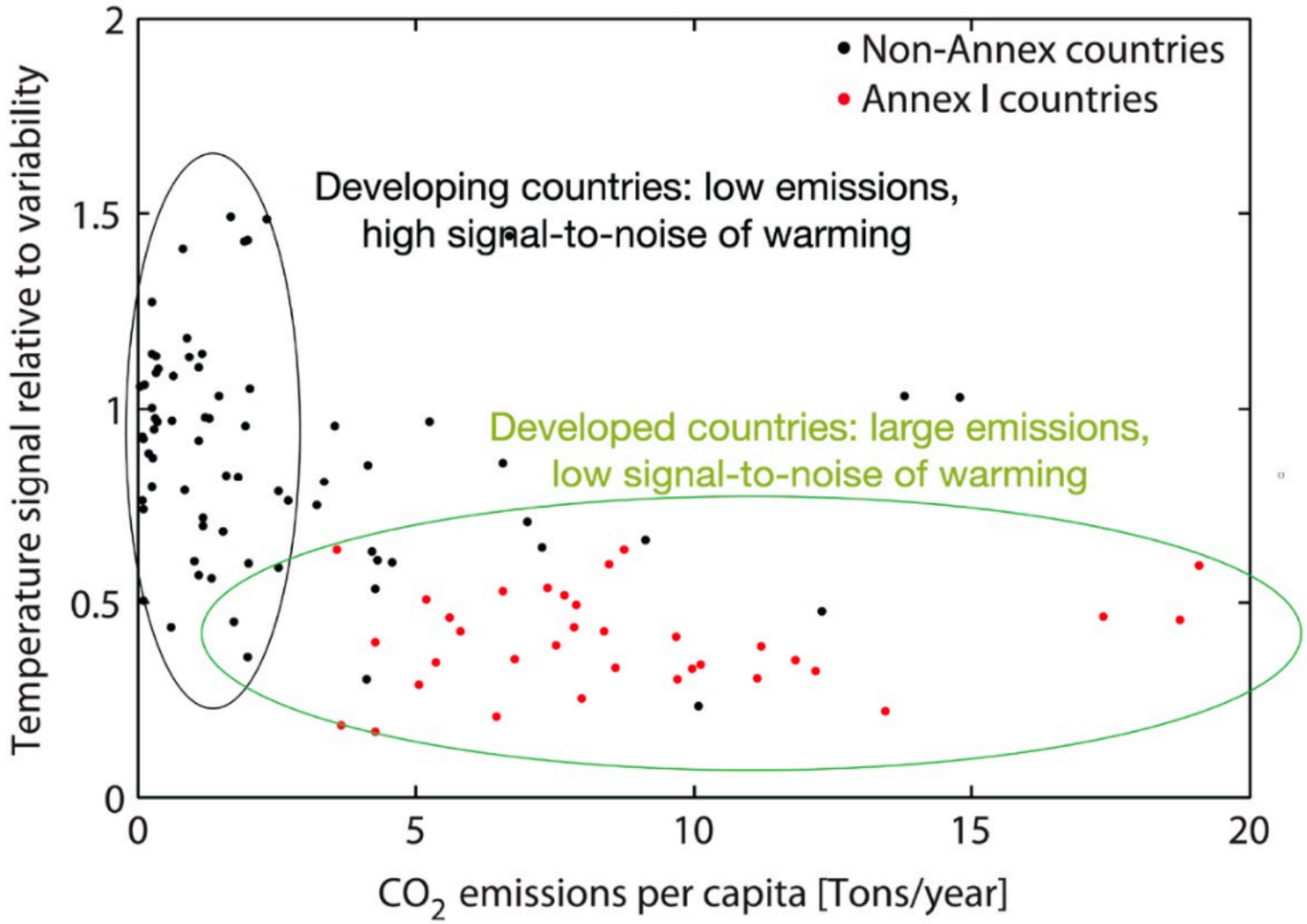


Figure 10.5

These low-emission countries are also probably the least capable of mediating the effects of global warming.

Unfortunately there seems to be a clear inverse relationship between the level of development in a country and signal-to-noise of global warming as Figure 10.6 conveys.

Figure 10.6



The situation is likely to only get worse for those regions experiencing the immediately detectable effects of global warming today.

By the end of the century, while it will be hotter everywhere on average, in the tropics and subtropics *the seasonal average summer temperature will be warmer than the warmest summer thus far almost 100 percent of the time*.

It is in these regions that the greatest percentage of currently malnourished individuals live and where the population depends most heavily on agriculture, which will be most strongly affected by global warming.

If the climate is broken in any way, it will have been the developed world that has broken it, but it is the developing world that may suffer first.

Ownership is clear, but who accepts ownership remains to be seen

11 - WORST CASE: FROM FEEDBACK TO TIPPING POINTS

Perhaps nothing strikes more terror into the heart of a numerical modeler than the words “nonlinear feedback.”

The reason numerical simulations are so useful is that they can handle many small timesteps, and during small time intervals generally all changes are linear.

Therefore even complicated behavior over the long run can be successfully simulated with relatively simple mathematical operations at each step.

The problem with nonlinear feedback mechanisms is that an initial small change can produce an effect that feeds back on the changing variable, producing yet a larger effect.

Pretty soon untreatably large effects can grow exponentially fast.

When this happens, predictive capabilities are rapidly lost.

So goes the famous example of chaos, where a butterfly flaps its wings in Kansas, ultimately causing a tornado to rip through Nebraska.

But nature doesn't care about the happiness of modelers or the utility of their modeling techniques.

There are undoubtedly feedback mechanisms that affect climate that can be difficult to model because of the sudden changes they may bring about.

Here is one possible example.

Heating causes ice to melt in Arctic waters.

But the melting ice exposes the darker water underneath, which reflects less sunlight and absorbs more, causing more heating that melts more ice, and so on.

Here is another: ice melts in Greenland and reduces the height of the central ice sheet, but the temperature at lower altitudes is higher, so as the top of the ice sheet lowers in elevation, it heats up yet more and melts more quickly.

Here is one ice possibility I previously described: warm water causes floating ice sheets to calve and eventually break apart, but these buttress large glaciers that now are unrestrained, and quickly flow into the sea, destabilizing the entire ice sheet behind them.

I hope that this class has demonstrated how an explanation of fundamental physics arguments can give interested laypeople the tools to:

(a) address questions they may have about the world and their place within it,

(b) assess claims about the world they may read about, and

(c) make decisions about their own actions and about public policy questions.

With this in mind, I have concentrated on the general physical principles and tools that give some basic perspective on how a predictive understanding of the causes and various effects of radiative forcing on climate can be obtained and how reliable it might be.

There are some important potential effects that are, however, difficult to connect directly to underlying physics or to include directly in models.

This does not mean we should ignore them.

One of the most powerful and important aspects of science is its acknowledgment of uncertainty.

The fact that all physical statements have uncertainties associated with them is not a failing but a virtue.

It doesn't mean that we cannot predict results reliably nor that we don't understand physical systems.

It means we must work to quantitatively understand and explicitly state just how reliable a prediction is likely to be.

Thinking deeply about the quantitative uncertainties in estimates or measurements is a central part of making the estimates or measurements.

That the basic predictions of climate science come from well-understood physical principles should make it clear that much of climate science is not some invisible voodoo practice or something that requires supercomputers to assess.

Because of this, we all should be able to appreciate and assess the causes, effects, and risks that arise from these predictions and that we face as a result of human industrial activity.

I would be remiss if I didn't follow up on the preceding discussions of these predictions by at least briefly pointing out some of the "known unknowns"—those less predictable consequences that are based on the interplay between numerous climate factors that feed back on each other.

Recent work highlighted describes a number of these.

We may not be able to state with authority when or if any of them will become significant, but in any risk analysis we would be negligent not to at least factor them in at some level into our decision-making.

The effects of greatest concern for climate modelers, and the rest of us, are what have become known as "**tipping points**"—those drastic changes, over various time frames, that once set in motion are likely to be irrevocable.

I have already discussed one such tipping point, the West Antarctic Ice Sheet.

Once destabilization of the glaciers in the Pine Island Bay of the Amundsen Sea begins in earnest, then at least one study suggests that even if temperatures return to pre-2000 levels, the entire West Antarctic ice sheet may be destabilized.

This would add perhaps three meters to sea level rise over the course of a millennium or two.

Adding some credibility to this possibility, there is paleoclimate evidence that such widespread collapse of the ice sheet has occurred more than once in the past.

Once floating ice shelves that buttress glaciers are lost, there is nothing to stop the irresistible pull of gravity bringing glaciers down into the sea.

Other feedback effects I have mentioned that happen to be related to ice could lead to a complete collapse of the Greenland Ice Sheet on a timescale shorter than that predicted due to temperature rise and ice mass melting.

That prediction suggests a temperature rise of 6–7°C would be necessary for irrevocable collapse to occur, and even then over a timescale of multiple millennia.

However, there is sufficient uncertainty so that a much smaller temperature rise could lead to collapse.

Recall that one unanticipated factor that played a role in the 2019 severe melting summer was the large high-pressure zone that centered on Greenland, probably due to ocean current changes, leading to more sunshine and melting and less precipitation.

Another effect mentioned earlier is the nonlinear feedback that occurs as ice sheets melt: their elevation decreases, putting the surface in contact with warmer air, which causes the sheet to melt faster.

Effects such as these, combined with the increased relative warming in the Arctic water due to changing albedo and increased absorption of light as sea ice melts, have led some researchers to suggest that the Greenland ice sheet threshold, beyond which it is doomed to melt, could be as low as 1.5–2°C.

This number is ominous because we have already basically committed the Earth to such a temperature rise before mid-century, given the CO_2 already in the atmosphere.

Of course even if Greenland ice-sheet melting is assured, with its consequent seven-meter sea level rise, the time period over which this could happen may depend on the future temperature profile of Earth.

Thus actions taken in this century could perhaps delay the inevitable beyond this millennium.

Another area where irreversible sudden shifts may be possible involves large-scale ecosystems, an area I have not focused on in this class, not because it isn't important, but because the issues here are more biological than physical.

One particular biological impact that has received some international attention is the mass bleaching of large coral reefs, with the Australian Great Barrier Reef being the most famous.

The severe bleaching of much of this reef has already been documented with photographs that have appeared in newspapers around the world.

Marine biologists project that if the temperature rises by a further 2°C, 99 percent of the world's tropical coral reefs could be lost.

Multiple factors could come in to play here beyond simple warming, although warming dominates.

One in particular is worth reviewing here because, like sea level rise, it involves straightforward physics and chemistry directly associated with increased CO_2 concentration in the atmosphere.

I have described how an equilibrium is built up between CO_2 in the atmosphere and the formation of carbonic acid in the oceans as CO_2 dissolves.

One of the important observations that can be used to argue against those who claim that increasing CO_2 is having little global effects (and by consequence, that global warming is related to something else, like rapid changes in solar luminosity, which, by the way, we can measure and which do not exist) is that the world's oceans are already responding to a different kind of forcing from increased CO_2 in the atmosphere.

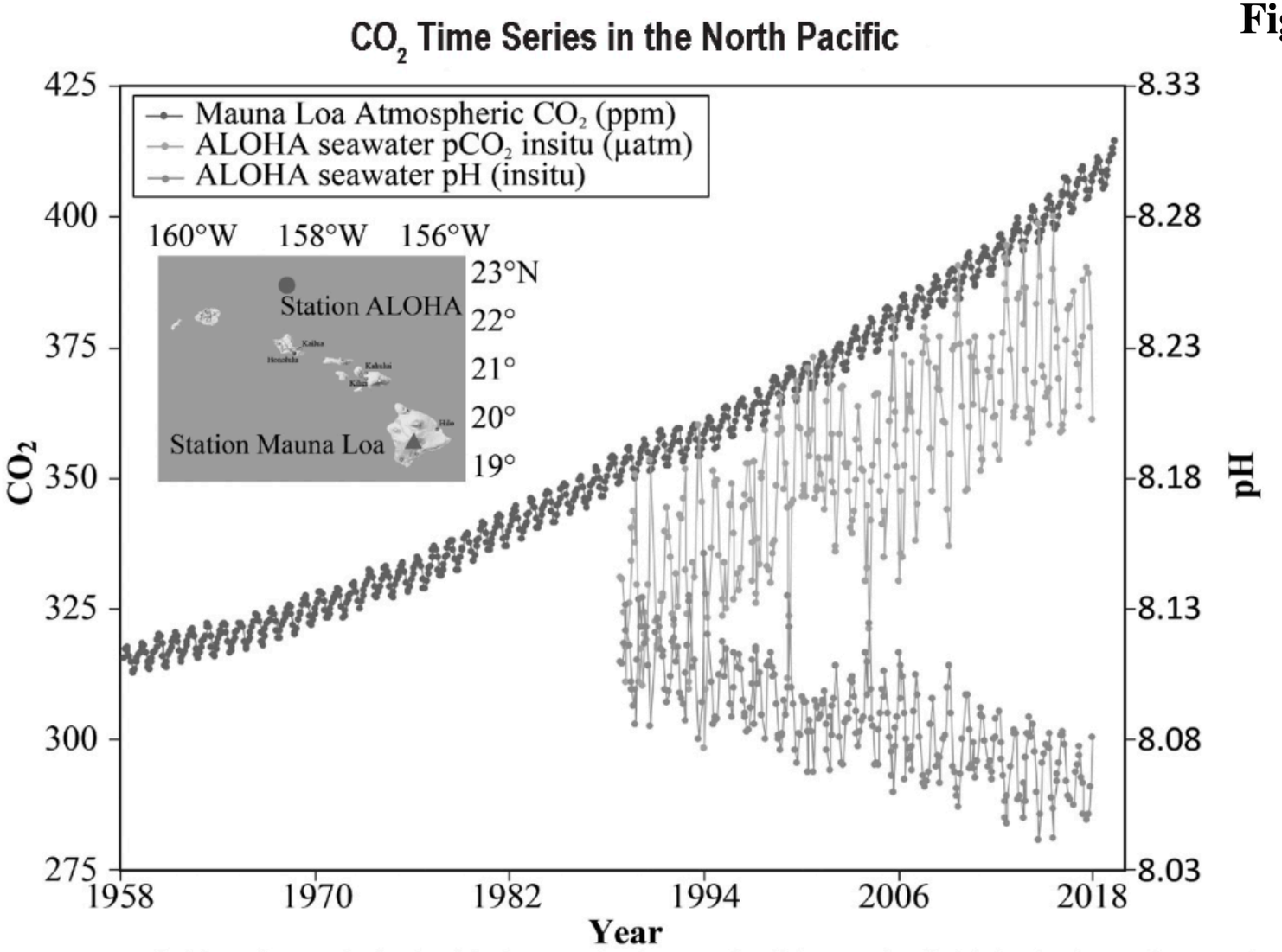
We are apparently entering a new phase, some call it Oceans 2.0, due to both ocean warming and ocean acidification.

As early as 2002, ocean measurements made it clear that the mean pH level of the world's oceans had been decreasing at a regular rate for over a decade.

Measurements have continued since then, and the chart in Figure 11.1 was prepared by the US National Oceanographic and Atmospheric Administration (NOAA) in 2018.

Shown in the figure is the CO_2 abundance from the Keeling curve in Mauna Loa we have already seen, and alongside it are measurements in the ocean off Hawaii of the amount of CO_2 dissolved in seawater and the seawater pH.

Figure 11.1



Data: Mauna Loa (ftp://aftp.cmdl.noaa.gov/products/trends/co2/co2_mm_mlo.txt) ALOHA (http://hahana.soest.hawaii.edu/hot/products/HOT_surface_CO2.txt)
Ref: J.E. Dore et al, 2009. Physical and biogeochemical modulation of ocean acidification in the central North Pacific. *Proc Natl Acad Sci USA* **106**:12235-12240.

As would be predicted in any high school chemistry text, as the CO_2 atmospheric abundance increases, the fraction of CO_2 dissolved in water in contact with the atmosphere will increase.

As that happens, the amount of carbonic acid, H_2CO_3, will increase, which in turn will lower the pH.

Decreasing pH in the oceans may have many complex biological effects, but one that involves simple chemistry is this: as oceans become more acidic, the increase of positive hydrogen ions will allow their increased binding with pre-existing negatively charged carbonate (CO_3) ions.

But the carbonate ions are what both coral and shellfish use to build their calcium carbonate for their exoskeletal structures.

With fewer carbonate ions free for the taking, corals will suffer as will shellfish.

Indeed, a study of coastal waters off the west coast of the US in 2011 surprisingly found that 53 percent of tiny pteropods, very small marine snails, already had severely dissolved shells.

NOAA has done experiments on these small snails by placing them in water prepared with the ocean acidity predicted in 2100 for forty-five days.

Figure 11.2 is a photograph of the time series.

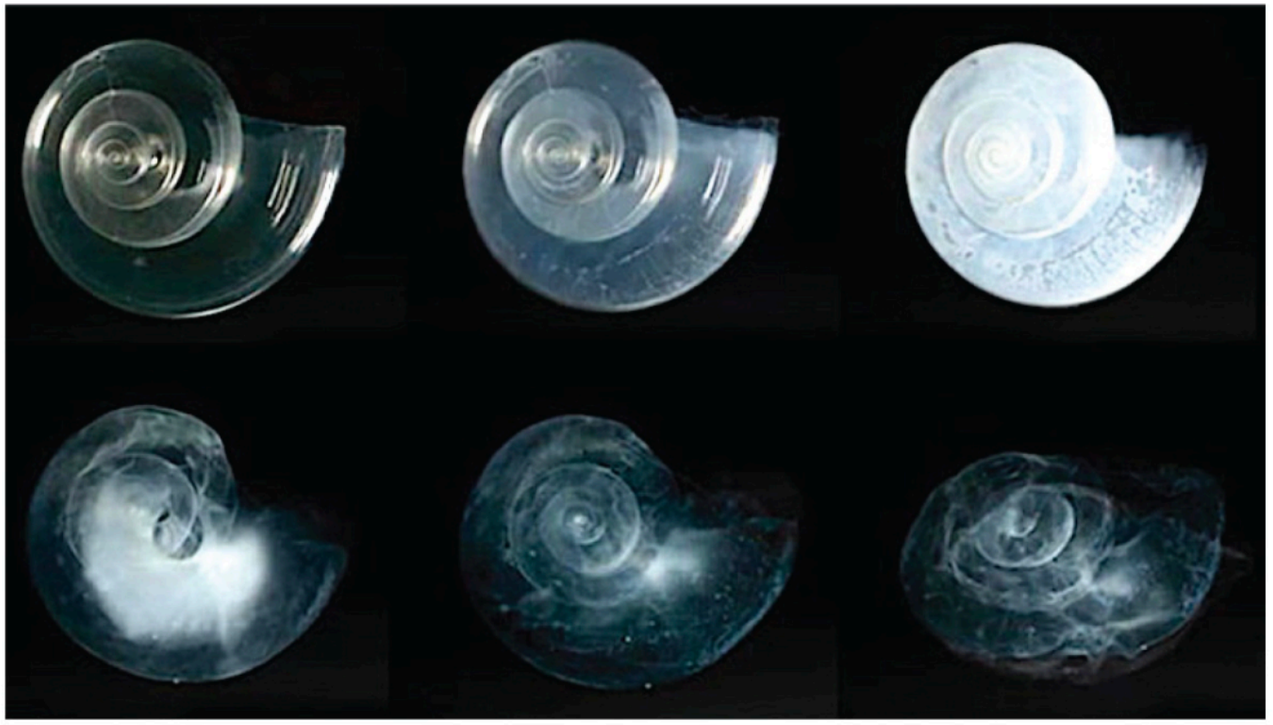


Figure 11.2

As dismal as these photographs are, to be fair one must be careful before inferring increased ocean acidity will immediately bring disaster.

The consequences of a reduction in oceanic biodiversity involving a possibly reduced fraction of pteropods and a closely associated shelled zooplankton called foraminifera could be severe, as these small animals are an early step in the food supply chain for almost all fish in the ocean.

But if they disappear, will that simply create an evolutionary niche for other non-shelled zooplankton to adapt to take up the same role?

Also, when these shelled animals die, they sink to the seafloor and carry their calcium carbonate with them.

This gets deposited as rock, sequestered, and eventually subducted into the mantle, or as with the White Cliffs of Dover, raised above sea level by plate tectonic movements.

How might their possible disappearance impact the amount of CO_2 taken up by the oceans?

Naively, one might think it would reduce this intake, which would lead to increased CO_2 in the atmosphere—another feedback loop.

But the carbonate chemistry of ocean water involves more than just calcium carbonate, so inferring a direct feedback on atmospheric CO_2 in this way is too simplistic.

A variety of competing effects will determine the ultimate impact, if any.

On larger scales, to what extent various large fish populations, a staple for feeding much of humanity, will survive a combination of increased temperature and increased acidity is currently not known.

But clearly a collapse of any major fish species would reverberate throughout the planet.

Feedback on land may also produce climate change tipping points, including the impact of warming on North American boreal forests, where large-scale insect infestations and increases in forest fires have caused dieback.

This turns forests from carbon sinks to carbon sources, not just directly through fires but also through decay of dead trees and plants.

A potentially more severe feedback loop may occur in the Arctic, where warming can melt the permafrost, which contains huge stores of CO_2 and methane—perhaps as much carbon could be released in this way from the permafrost of Siberia, full of organic material, as currently exists in the atmosphere. And methane is an even more potent greenhouse gas than CO_2.

Permafrost is already being observed to thaw, and at what temperature the thawing may become irreversible remains to be seen.

Increased greenhouse forcing is not the only impact of human industrial activity that may produce a nonlinear climate or biological tipping point.

Deforestation of the Amazon has been significant, with about a 17 percent loss of Amazon rainforest to date.

But the humidity in the region is strongly impacted by the existence of the rainforest itself.

At some tipping point, the forest stops increasing the humidity and rainfall so it is no longer self- sustaining, and it will degrade into a drier savannah.

This would release huge amounts of CO_2 currently stored in the rainforest trees and vegetation into the atmosphere.

One estimate, by Thomas Lovejoy, is that the Amazon tipping point could happen when 20–25 percent of the rainforest has been cut down.

Given the current 17 percent level of deforestation, this could happen before midcentury.

According to at least one economist, the combination of deforestation, large-scale forest fires, and global warming could accelerate the rate of destruction, meaning the tipping point could be reached in this decade.

As worrisome as each of these possible tipping points—some leading to short-term disasters and others, like the potentially irreversible melting of Greenland, taking centuries or millennia—may separately be, of even greater potential concern is a possible feedback between them.

There is increasing evidence that such feedback may be common.

A 2018 study of thirty possible systems at risk found that in 45 percent of cases, a tipping point in one area had a domino effect, increasing the risk of exceeding a tipping point in another.

As an example, here is one scenario that can be argued is already currently underway:

> *"Arctic sea-ice loss is amplifying regional warming, and Arctic warming and Greenland melting are driving an influx of fresh water into the North Atlantic. This could have contributed to a 15% slowdown since the mid-twentieth century of the Atlantic Meridional Overturning Circulation (AMOC), a key part of global heat and salt transport by the ocean.*
>
> *Rapid melting of the Greenland ice sheet and further slowdown of the AMOC could destabilize the West African monsoon, triggering drought in Africa's Sahel region.*
>
> *A slowdown in the AMOC could also dry the Amazon, disrupt the East Asian monsoon and cause heat to build up in the Southern Ocean, which could accelerate Antarctic ice loss."*

Lest any or all of these seeming doomsday scenarios make you want to crawl under a desk (or leave this class, except for your worries about that contribution to the greenhouse effect), I want to stress that all of these examples are of a future that *might be*.

None can be as simply or directly modeled as the impacts I have focused on in the rest of this class, and none of them are guaranteed.

But it is worthwhile recognizing that any one of them could have a short-term impact that could be more devastating than the slow and steady rise in global temperature might otherwise suggest.

It will be up to decision-makers to determine if we are willing to empirically test these ideas in the coming decades.

Or whether the risks, which could be small, when multiplied by the possible devastation, which in some cases could be immense, combine together to convince politicians and the public that we need to act globally now.

That is, I hasten to add, if it is not already too late.

12 - BACK TO THE MEKONG

I began this part of the class with a description of that mighty intertwined system of river and humanity, along with the paradox of its existence.

It is worth repeating here:

This diurnal mother-of-a-tide ought, of course, to spell disaster to the Delta.

A salty inundation, albeit only once a day, would soon sour the world's most productive ricebowl and turn the green dazzle of paddy into maudlin thickets of mangroves like those along the Donnai below Saigon.

What prevents such a disaster is the power of the mighty Mekong.

The inrushing tide meets the outrushing river, and in the best traditions of ecological equilibrium they compromise.

The river rises, its progress barred by the tide.

The backing-up of the river by a big 'diurnal' is measurable as far upstream as Phnom Penh and beyond.

But there and throughout the three to four hundred kilometers down to the sea, salination is barely detectable…

The river thus defends the Delta from its deadliest foe since the rising waters are overwhelming its own, not the China Sea's.

Sadly, this is a battle the Mekong is destined to lose.

Its demise will be swift, brought on by forces that originate from far away.

The story I will now relay involves two completely independent scientific studies, each done with tools as different as one can imagine.

They were reported within weeks of each other in 2019, both leading to startling conclusions with implications not just for the Mekong but for the rest of the planet.

The first study involves my favorite kind of science, at least when I pursue it myself: feet on the ground and low tech.

Dutch geographer Philip Minderhoud, a postdoctoral researcher based at Utrecht University, noticed something strange while visiting Vietnam a number of years ago to investigate land subsidence—the lowering of ground level due to natural and human-induced processes.

Comparing what he saw on the ground to satellite maps, he realized something was wrong:

"I saw, for example, roads that were only slightly higher than the water that was directly connected with the sea, whereas according to the maps, they should have been located higher."

He also saw strange patterns of contour lines on the maps, linear stripes that clearly weren't real traversing riverbeds, for example.

Clearly the satellite maps themselves were suspect.

A technique called LIDAR, which stands for Light Detection and Ranging, uses the reflection of pulsed lasers from the air to measure ground elevation, and it is accurate to thirty centimeters or better.

While there are some LIDAR systems in space which have been used to measure elevations of the Earth's polar regions, generally LIDAR has been used by planes and is expensive.

Traditionally, therefore, large-scale global land elevation maps are done using other kinds of satellites which create Digital Elevation Models (DEMs), most often using radar techniques.

The resolution of these DEMs typically range between one and thirty meters.

For example, in February of 2000 the space shuttle *Endeavour* engaged in a mission called the Shuttle Radar Topography Mission (SRTM), a global survey sponsored by the Department of Defense.

Its data is publicly available and has been widely used by various organizations, including the World Bank, to decide where to allocate flood control resources.

However, there are problems using this data to model low-lying deltas.

In particular, there is a difference between absolute elevation and relative elevation.

To turn one into the other, one must have calibration points— predetermined zero points of elevation, which themselves can vary around the Earth due to local effects.

Earth isn't a perfect sphere, and the gravity of land masses will impact on local sea levels as well.

What one needs is a good local calibration.

Without this, satellite elevation readings can be inaccurate, especially for low-lying areas.

And things become even less accurate if calibration points are based on land masses that may themselves have been subsiding.

Because the Mekong Delta is built up from soft river sediments, when it compacts it can subside.

In addition, upstream dams can block the flow of new sediments to replenish those lost by erosion.

These effects can be exacerbated when large urban areas pump up groundwater from below and where urban infrastructures interfere with the natural seepage of rainwater into the Earth.

In Mexico City, one of the world's largest cities, built on what was originally a large underground aquifer now being drained to provide fresh water for over ten million people, settling is occurring at a rate of up to twenty centimeters per year.

It is quite noticeable in the sagging of building foundations as you walk down various streets.

In the Mekong, subsidence in some areas has been measured at rates approaching five centimeters per year.

It was in this context that Minderhoud began his studies.

Satellite and space shuttle data had put the elevation of the delta at 2.6–3.3 meters, uncomfortable but tolerable given sea level rise estimates for this century.

Apparently, the Vietnamese government had already known that the elevation maps used by the West did not conform to their own elevation readings, but the detailed elevation data was a military secret.

It was here that Minderhoud exploited the human side of science.

He built trust with the Vietnamese authorities, pointed out how his research might contribute to their own goals, and offered to combine his research program with their efforts.

In the end, he was given the data for twenty thousand separate elevation points throughout the Delta.

In return, he significantly improved the calibration of these points.

The Vietnamese had used a point near Hanoi, two thousand kilometers to the north.

Instead, Minderhoud's group used a local point for calibration at an island location, Hon Dau.

The result of all of this on-the-ground effort was a new topographical Digital Elevation Model tailored to the region, and the change was staggering.

It turns out the new mean elevation of the delta is about 0.8 meters above sea level, about 2 meters below the older satellite estimate.

Figure 12.1, from Minderhoud's research, gives a sense of the difference between the old satellite-based estimates of elevation (labeled STRM), a more recent model based on that data (MERIT), and the new more accurate topological model (Topo).

Shown by black lines are two cross-sections of elevations across the country.

The earlier DEMs overestimate the elevation at every single location.

Minderhoud demonstrated that reasonable agreement between MERIT and Topo could be obtained by systematically lowering the MERIT elevation at every point by 2.5 meters.

Figure 12.1

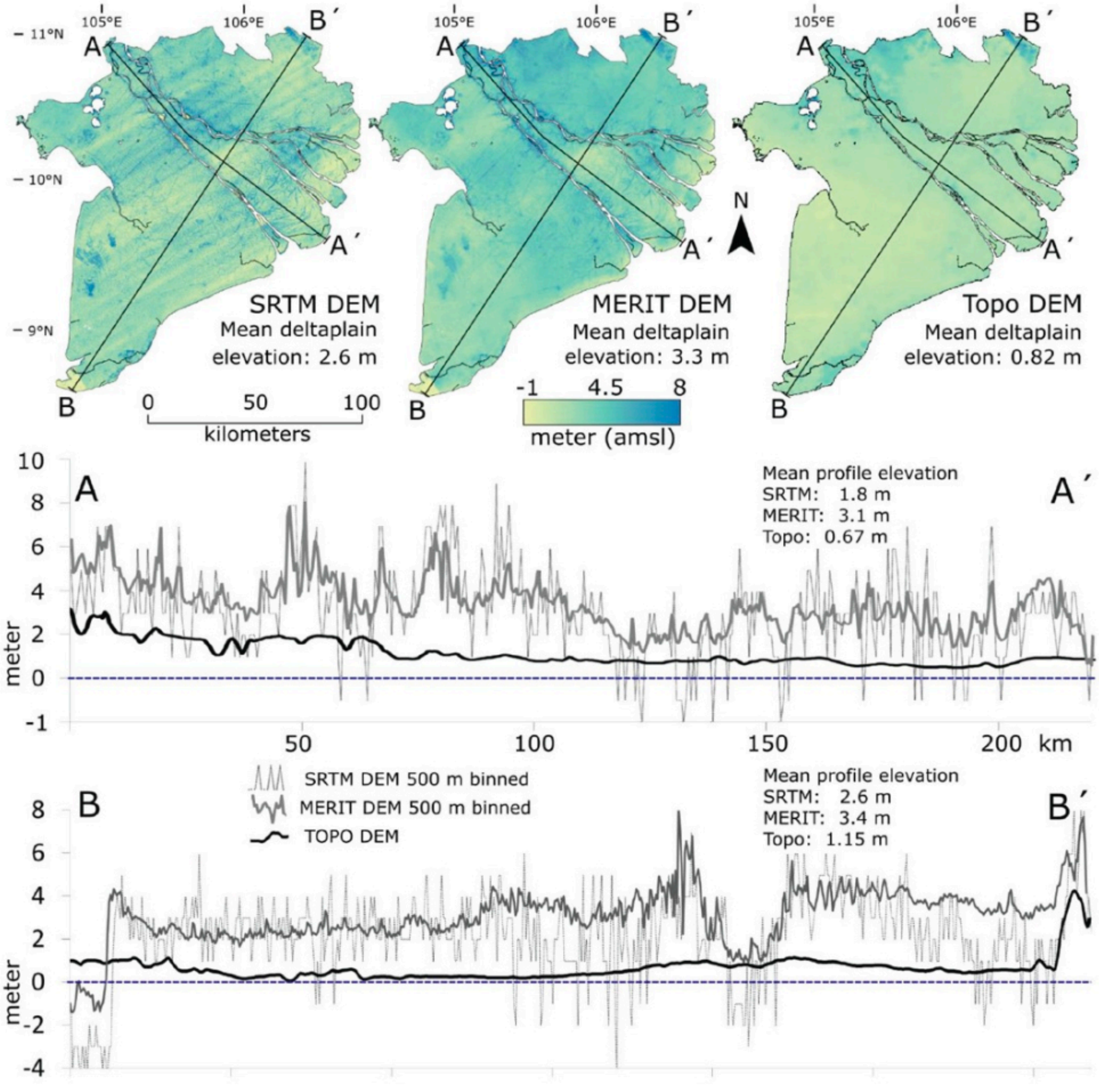
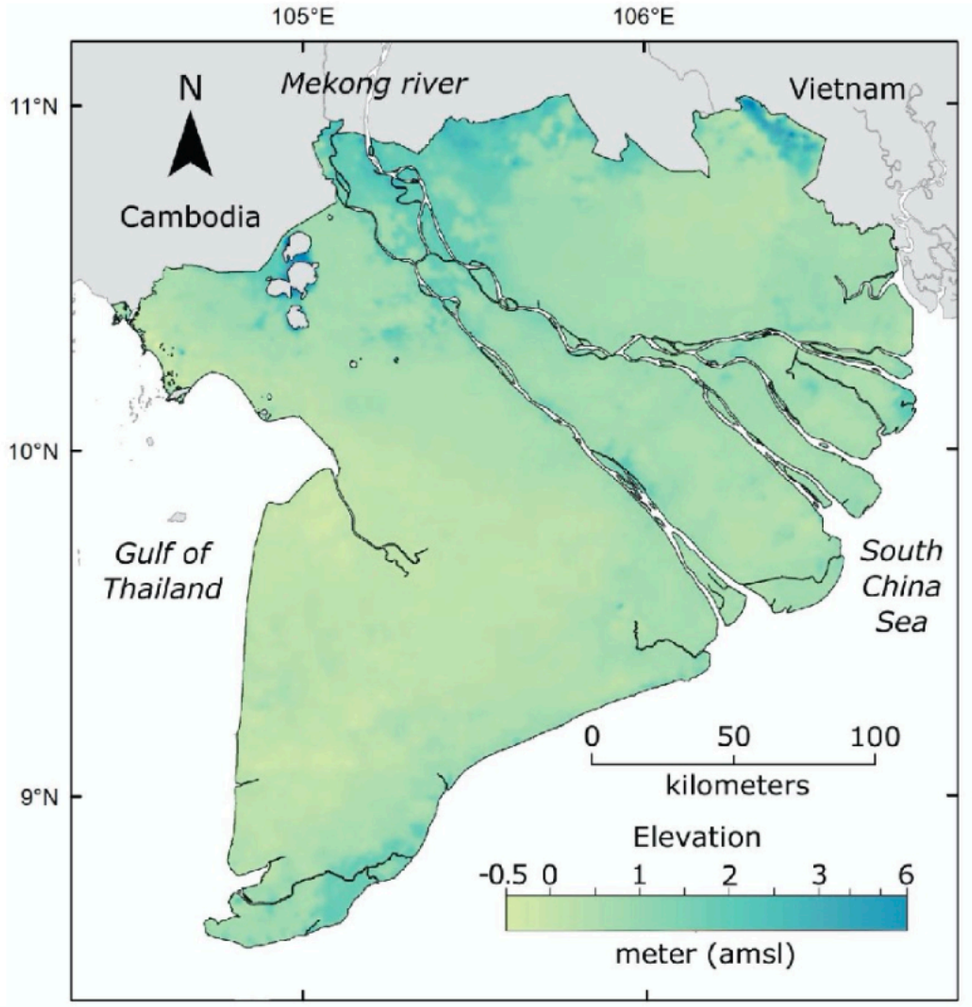

105°E
106°E
11°N
10°N
9°N
A
A´
B
B´
SRTM DEM
Mean deltaplain
elevation: 2.6 m
MERIT DEM
Mean deltaplain
elevation: 3.3 m
Topo DEM
Mean deltaplain
elevation: 0.82 m
N
0
50
100
kilometers
-1
4.5
8
meter (amsl)
meter
km
SRTM DEM 500 m binned
MERIT DEM 500 m binned
TOPO DEM
Mean profile elevation
SRTM: 1.8 m
MERIT: 3.1 m
Topo: 0.67 m
Mean profile elevation
SRTM: 2.6 m
MERIT: 3.4 m
Topo: 1.15 m

Figure 12.2 is an aerial map obtained using Minderhoud's full data set, with different shadings displaying land elevation across the entire delta region.

Figure 12.2



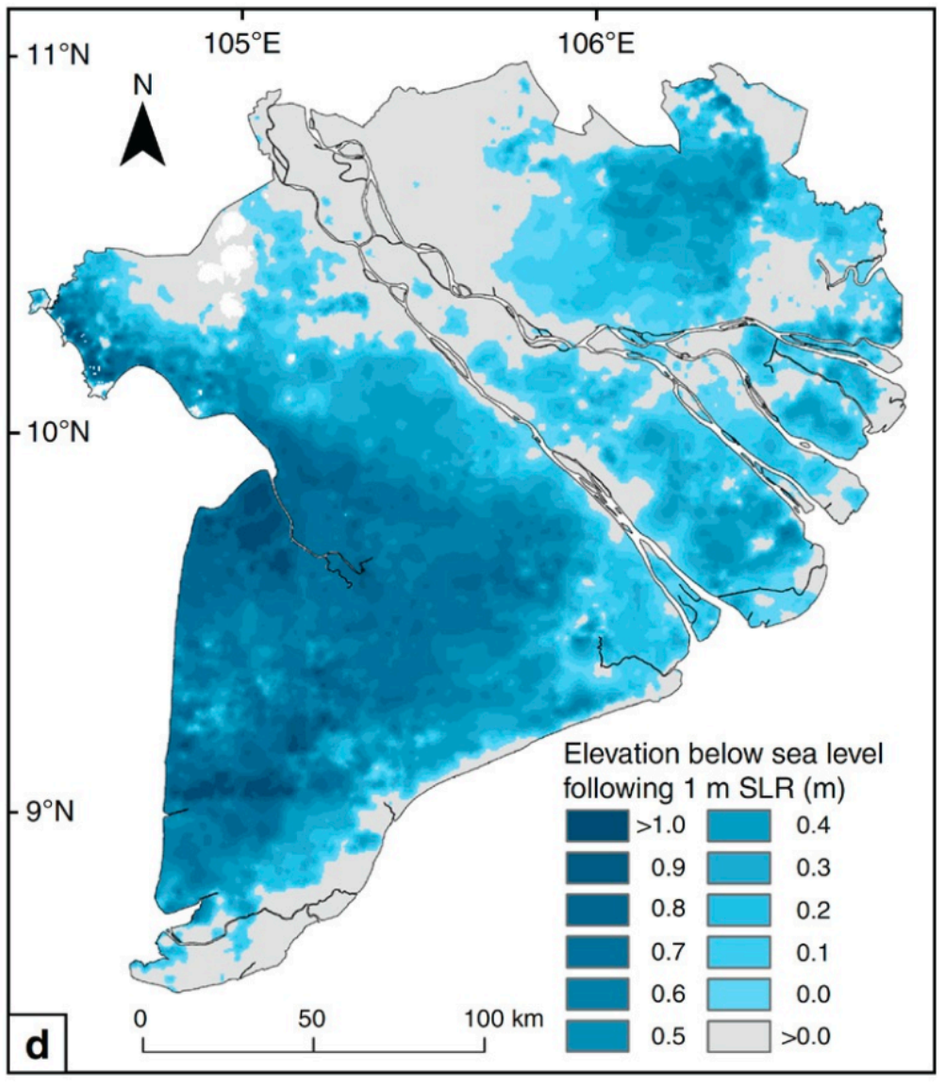
The implications of this for the region are staggering.

As we have seen, sea level rise is guaranteed to be at least 0.5 meters by 2100.

But worse still, including the average current subsidence rate of the land of ≈1.0 cm per year, the relative sea level rise compared to the land elevation in this region will be at least 0.8–1.0 meters by century's end.

Figure 12.3 is a plot showing how much of South Vietnam will then be below sea level.

Figure 12.3



The relevant statistics, assuming 1.0 m relative sea level rise are as follows: 75 percent of the total area of the Mekong Delta plain, comprising most of South Vietnam, will be below sea level.

The number of people who currently live in this area is about 12.3 million people, about 70 percent of the total population of the region.

The situation is actually even worse.

The Delta currently provides a food store for a far larger population, through freshwater fishing and rice production.

Both are vulnerable to collapse if the area is inundated with seawater.

And finally, to exacerbate the situation, there is the incessant removal of sand from the Mekong itself.

As I noted earlier, this has reduced the Mekong riverbed elevation by 1.4 meters since 2008.

Climate change, combined with other impacts of human activity in the region, does represent a Perfect Storm closing in on the Mekong Delta.

While the daily battle between the flow of one of the mightiest rivers in the world and the incoming daily tide of the China Sea goes on each day, the war has already been lost.

The richest rice-producing area in the world could be submerged under salty water by the end of this century.

A river with more freshwater fish than any other river in the world will likely become brackish far from its current mouth.

And an agricultural population that depends on both will lose their food source, and much of the land on which they currently reside.

It may seem premature to draw these damning conclusions on the basis of a single study, however impressive.

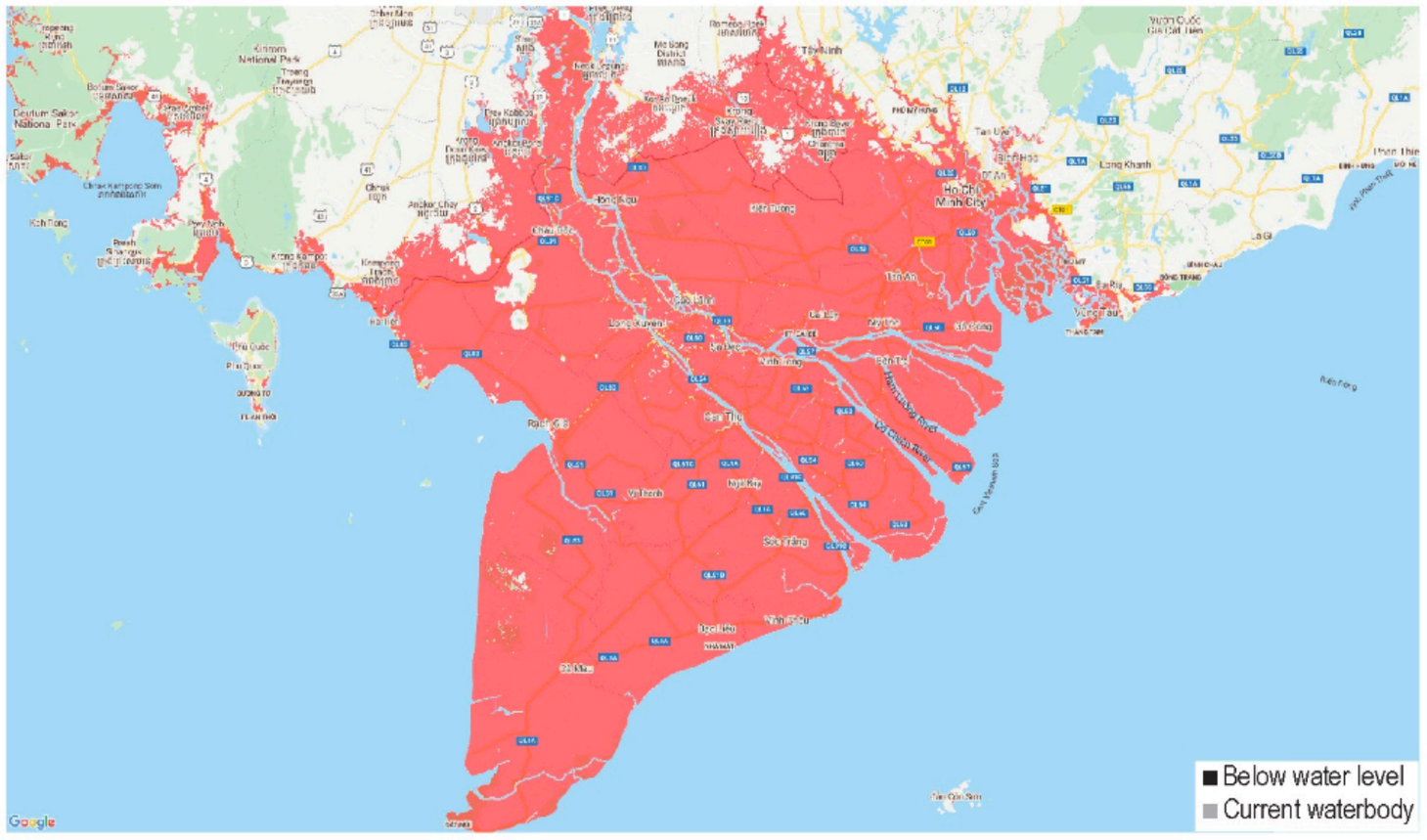
But within weeks of the publication of this work, a completely independent and broader research program published a paper describing the use of new neural networks to recalibrate and globally reduce errors in the Shuttle Radar Topography Mission DEM, producing a new model called CoastalDEM.

In the region in which this analysis overlaps with the Dutch team analysis, the agreement is remarkable.

CoastalDEM concludes a population between nineteen and thirty million in Vietnam currently reside on land that will be below the future high tide line as early as 2050.

Figure 12.4 is a pictorial representation based on their new data of the land (shaded) that will be below high tide and annual floods in 2050 in South Vietnam—essentially the whole country below and including the currently bustling megalopolis of Ho Chi Minh City (assuming no new man-made coastal flood defenses implemented).

Figure 12.4



While many factors present in the Mekong Delta make it particularly sensitive to the impact of rising sea level, the Mekong represents, in microcosm, the threat facing much of the world's population from sea level rise.

The CoastalDEM study explored the global impact of their new elevation data.

Here is their summary abstract (bolded text is mine):

> “Here we show – employing CoastalDEM—that **190M people (150–250M, 90% CI) currently occupy global land below projected high tide lines for 2100 under low carbon emissions**, up from 110 M today, for a median increase of 80 M. These figures triple SRTM-based values. **Under high emissions, CoastalDEM indicates up to 630M people live on land below projected annual flood levels for 2100**, and up to 340M for mid-century, versus roughly 250M at present. We estimate one billion people now occupy land less than 10m above current high tide lines, including 230M below 1m.”

These are not the numbers of some remote future.

They describe a future we are now more or less committed to in this century.

It is hard to know how to adequately express the true human toll associated with staggering figures like this.

I have focused here on the Mekong Delta, not only because it will be hit so severely, but also because these people had suffered in the past century from the very worst atrocities of war and genocide, and yet emerged from this maelstrom without bitterness or a sense of victimization.

Rather, they retain an optimism that the future can be better than the past, and they willingly shared with me the pleasure of our common humanity.

Cambodia rebuilt after a disaster largely of its own making in which perhaps a third of its citizens were massacred.

Vietnam suffered a loss of millions of people in a war that was prompted from abroad to emerge as a thriving economic powerhouse today.

Both countries now face a threat far greater than they have ever faced, and one largely not of their own making.

The question of how these countries and their cultures might survive through to the other side of this century will reflect, in microcosm, how humanity in general will deal with a global phenomenon that is unprecedented in modern human history and about which we have had fair warning.

Context is everything.

One meter does not seem like a huge elevation, unless perhaps someone is holding a bowling ball at this height above your feet.

Or unless it is all that stands between you and the loss of your home, your livelihood, and in some cases much of your country by the middle or end of this century.

Vietnam is not alone.

Without intervention, much of Bangladesh will disappear at high tide, as will large swaths of land in India, Indonesia, Malaysia, Papua New Guinea, and China in Asia; the Netherlands, Northern Germany, France, Spain, England, and Northern Italy in Europe; southern Iraq and northern Egypt in the Middle East; Mozambique in Africa; Argentina and Brazil in South America; Mexico in Central America; and Florida, Louisiana, Massachusetts, and North Carolina in the USA.

Beyond this, almost every major coastal metropolitan area in the world will be partially submerged at some time during the year by 2100, including London, Tokyo, Miami, New York, Boston, and Shanghai, to name just a few.

The policies we adopt in the face of the information I have presented here —from the fundamental and well-tested predictions of radiative forcing to the more speculative possibilities of tipping points, and from the long-term global risks we face to the more immediate devastation facing places like the Mekong—are not for me to address here.

They will be determined by the citizens of the world and the governments they elect or under which struggle.

I hope for the sake of Cambodians and the Vietnamese, and also for hundreds of millions of Bangladeshis, Indians, Japanese, Chinese, Africans, Polynesians, Indonesians, Middle Easterners, North and South Americans, Europeans, and all the rest, that technological ingenuity combined with rational action, tempered by simple grace and empathy, might supersede our longstanding human traditions of xenophobia, greed, and violence in the face of the national and international challenges we now face.

But hope and expectation are two different things.

13 - FORTUNE FAVORS THE PREPARED MIND

This class isn't primarily a call to action, but neither is it a requiem.

As Richard Dawkins often stresses, humans have a capacity for planning the future that is probably unique among all lifeforms on Earth.

Evolution, while often anthropomorphized, has no plan by which it proceeds.

Nature is sublimely indifferent to whether we survive as a species, or even whether our planet continues to exist.

But we humans, with our potentially unique capacity for self-awareness and rational thinking, can plan and strategize for the future.

Moreover, we can develop scientific tools that allow us to predict the outcome of our actions as well as the outcomes of things we have no control over.

And we can develop technologies that change the present and the future, that give us unprecedented control over our environment.

These remarkable gifts have brought us to the present moment, which some might view as a precipice and others view as a mountaintop, from which we can see farther than we have ever seen before.

In the process, they have created a global civilization and a global economy we are still learning how to deal with as social beings.

I am offering this class after the first global pandemic in which the interconnectedness of humanity has been manifest, and the impact of acting quickly in the face of evidence has never been clearer.

I can't say that the global report card in this regard has yet been very impressive.

But I also know that most of us feel certain that this, too, shall pass.

We had different levels of confidence on our political leaders, but had confidence that our scientists and doctors can recommend courses of action that make us safer, and they ultimately developed treatments and vaccines that ameliorated the levels of illness and death that the virus might have otherwise caused.

The discomfort of our regimen of enforced isolation and social distancing were real, but as we expected they were temporary.

Life after the pandemic will undoubtedly be changed, but will it change more than superficially?

Imagine instead a global change far more pervasive and potentially far more long-lasting.

Imagine not just your immediate livelihood under threat, but your home and the homes of your family and friends.

Imagine the long- term livelihood of everyone you know potentially destroyed, with no safety net to fall back on, with no place to retreat.

This is the near-term future that seems likely for some significant fraction of the world's population, independent of the vagaries and vicissitudes of the longer-term impacts of climate change.

We need to face that future with open eyes if we are to plan for how to overcome it, or at the very least, ameliorate it.

The only way to do that is to understand where we have come from and where we are heading, as well as what the impacts of our action, or inaction, might be.

I have many friends who are climate scientists, geophysicists, and oceanographers, and I am struck by their different emotional responses to the same reality, ranging across the gamut from stoicism to public protestation.

But they share a fundamental optimism that rational action based on empirical evidence and the well-tested scientific models is the best hope we have to address the challenges we face in the coming century.

Among these individuals, perhaps the most optimistic, is the view is that technology provides tremendous opportunities for economic and social well-being in the face of climate change.

These include transitions to more sustainably distributed energy resources, agricultural transitions that can feed more with less, and ultimately, if necessary, proposals to mediate radiative forcing to slow the temperature rise or even cool the planet—including geo-engineering options that involve adding things like aerosols to the atmosphere or, less realistically perhaps, global capture of carbon from the atmosphere followed by carbon sequestration.

I am not sure I am as optimistic, but neither do I see us traveling down an inexorable road to oblivion.

The future is charging at us like a freight train, but it is doing so on tracks we have built.

We may have time to divert the train, or perhaps build a bridge so it safely bypasses us.

We will never know unless we try.

I wish I had some silver bullet, but all I can offer is what I have tried to do here, to provide some perspective on the realities we face and how we can understand them.

Scientists like me, not elected by the public, should not be making public policy.

But we do have an obligation to use the results of our work to help inform public policy and, where possible, create technologies to aid in its enactment.

Technology could, in principle, ameliorate some of the inevitable impacts of climate change.

But the greatest challenges we face are likely to be political and economic rather than technical.

Developed nations may be able to protect areas that may lie below future high tide levels—as the Dutch did in the 20th century—at an increasingly arduous cost.

But what about less developed countries?

Will assistance be provided to avoid mass migration?

Improved agricultural technology could help keep family farms afloat in areas that will be impacted by climate change, but the hardest hit regions may also be the poorest, with least access to new technologies.

Understanding the science of climate change and its likely impact is the first step.

Informed and willing policymakers will then need to weigh the safety, sociopolitical, and economic risks associated with future amelioration challenges versus the challenge of enacting potential preventative economic and technical measures now.

I have had cause numerous times over the past months to think back to the example of Louis Pasteur, the French biologist and chemist whose groundbreaking work contributed to the principles of vaccination, as well as the technique of pasteurization, named after him.

His breakthroughs didn’t end disease, and they didn’t force government officials to listen to reason.

The current nonsensical anti-vaxxers who are rabidly fighting against reality and the social contract, testify to the limitations of rationality among the human species.

Yet Louis Pasteur anticipated that when he said fortune *favors* the prepared mind.

It makes no guarantees.

It just offers better odds.

I’ll take that any day.

CLIMATE CHANGE

Pass #3 - Different Approach and Many More details

In this class, I will used well-known(and studied) data up to the middle/end of the last decade; occasionally I will add data(less studied) from later up to the present when needed.

This works because the data is fully complete and the same stuff has been happening since the data was collected, which usually says things are even worse now!!!!

Climate Change - 2nd Pass

Part 1 - The Basics — GLOBAL WARMING IN A NUTSHELL

1 - Climate Change: A Primer

KEY QUESTIONS AND ANSWERS

Before exploring the various aspects of climate change in depth, let's quickly answer some of the most frequently asked questions about the issue.

The following slides will bring you up to speed with the current situation and the future outlook.

THE BIG PICTURE ——— Is the planet really warming up?

In a word, yes!!.

Independent teams of scientists have laboriously combed through more than a century's worth of temperature records (in the case of England, closer to 300 years' worth).

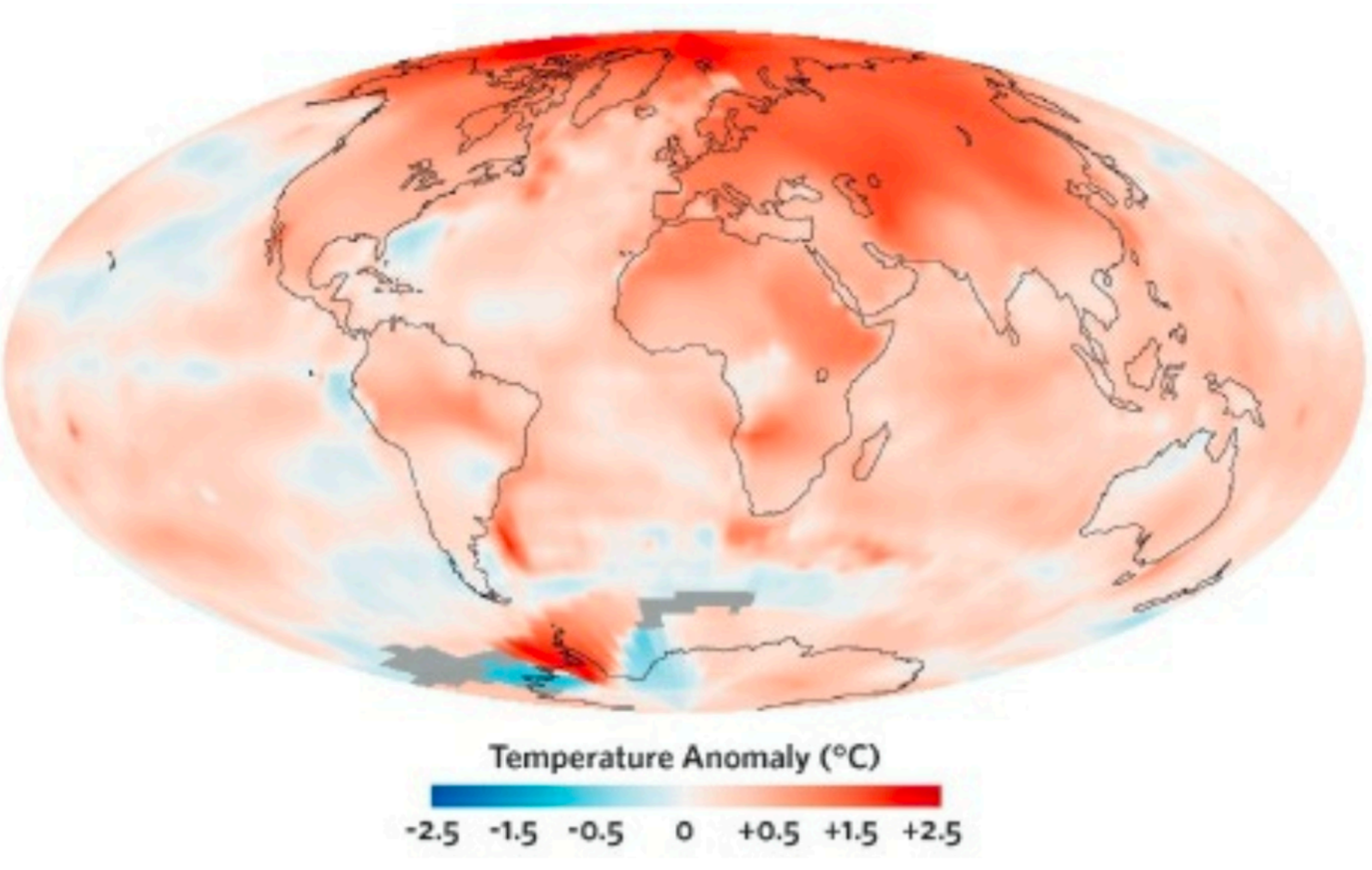
These analyses all point to a rise of around 0.8°C (1.4°F) in the average surface air temperature of Earth when comparing the period 2003–2012 to 1850–1900.

The estimated linear trend from 1880 to 2012 is a bit higher, roughly 0.85°C (1.53°F).

When the record warmth of the mid-2010s is included, the linear trend since 1880 is closer to 1°C (1.8°F).

Later we will show how this average is calculated.

The map on the slide below shows how warming since the 1970s has played out regionally.



The change in surface temperature relative to 1951-1980 (gray indicates no measurements)

In recent decades, global surface air temperatures have spiked dramatically, reaching a new highs every year since 1998.

The second decade of the twenty-first century was the hottest on record—warmer than any others in the past millennium—and the 2020s are on track to be warmer still.

Apart from what temperatures tell us, there's also a wealth of physical and ecological evidence to bolster the case that Earth as a whole is warming up.

Ice on land and at sea is melting dramatically in many areas outside of interior Antarctica and Greenland.

The major glaciers remaining in Montana's Glacier National Park could be reduced to fragments before the mid-twenty-first century.

Arctic sea ice has lost nearly three-quarters of its average summer volume since 1980, and the next several decades may see a point in summer when open water covers nearly all of the Arctic Ocean, perhaps for the first time in tens of thousands of years.

The growing season has lengthened across much of the Northern Hemisphere, with a boost from urban heat islands in some areas.

The average blooming date of Japan's famed sakura (cherry blossoms) is now earlier than at any time in more than a thousand years of record-keeping, pushing the traditional festivals from April to March.

According to the various Climate Impacts Program, spring is arriving two to six days earlier, and autumn two days later, each decade.

Mosquitoes, birds and other creatures are being pushed into new territories, driven to higher altitudes and latitudes by increasing warmth.

The range of twelve bird species in Britain shifted north in the 1980s and 1990s by an average of 19km (12 miles).

And Inuits in the Canadian Arctic report the arrival over the last few years of barn swallows, robins, black flies and other previously unseen species.

Many forms of marine life are moving poleward even more dramatically.

A 2017 review found that marine-based species are heading toward the poles at an average rate of 72 km (45 mi) per decade, or about four times faster than land-based species.

But don't many experts claim that the science is uncertain?

There is plenty of uncertainty about details in the global-warming picture: exactly how much it will warm, the locations where rainfall will increase or decrease, and so forth.

Some of this uncertainty is due to the complexity of the processes involved, and some of it is simply because we don't know how individuals, corporations and governments will change their greenhouse emissions over time.

But there's near-unanimous agreement that global climate is already changing and that fossil fuels are at least partly to blame (see later discussion).

The uncertainty that does exist has been played both ways in the political realm.

Skeptics use it to argue for postponing action.

Is a small temperature rise such a big deal?

A degree or so of warming may not sound like such a big deal, given that temperatures can vary by much more in a single day.

However, little increments make a big difference when they're in place day after day.

The average annual temperature in Boston is only about 4°C (7°F) below that of Baltimore.

Recent overall warming has been larger in certain locations, including the Arctic, where small changes can become amplified into bigger ones.

The planet as a whole could warm by 5°C (9°F) or more by the **end of this century**—hardly a small rise.

That's more than half of the difference between current temperatures and the depths of the last ice age.

Most of human history has unfolded within a narrow climatological window well short of such levels.

Any warming also serves as a base from which heat waves become that much worse, especially in big cities, where the **heat-island effect** comes into play.

Like a thermodynamic echo chamber, the concrete canyons and oceans of pavement in a large urban area heat up more readily than a field or forest, and they keep cities warmer at night.

During the most intense hot spells of summer, cities can be downright deadly, as evidenced by the hundreds who perished in Chicago in 1995 and the thousands who died in Paris in 2003 and Moscow in 2010 (see later discussion).

How could humans change the whole world's climate?

Humans have transformed Earth's atmosphere by adding enormous quantities of carbon dioxide (CO_2) and other **greenhouse gases** to it over the last 150 years.

As their name implies, these gases warm the atmosphere, though not literally in the same way a greenhouse does.

The gases absorb heat that is radiated by Earth, but they release only part of that heat to space, which results in a warmer atmosphere (see later).

The amount of greenhouse gas we add to the air each year is staggering—in carbon dioxide alone, the total is more than 30 billion metric tons, or more than four metric tons per person.

And that gas goes into an atmosphere that's remarkably shallow.

If you picture Earth as a soccer ball, the bulk of the atmosphere would be no thicker than a sheet of paper wrapped around that ball.

Even with these facts in mind, there's something inherently astounding about the idea that a few gases in the air could wreak havoc around the world.

However, consider this: the eruption of a single major volcano—such as Krakatoa in 1883—can throw enough material into the atmosphere to cool the global climate by more than 1°C (1.8°F) for over a year.

From that perspective, it's not so hard to understand how the millions of engines and furnaces spewing out greenhouse gases each day across the planet, year after year, could have a significant effect on climate.

If automobiles spat out chunks of charcoal every few blocks in proportion to the invisible carbon dioxide they emit, the impact would be more obvious.

Despite this, many people respond to the threat of global warming with an intuitive, instinctive denial.

When did we discover the issue?

Early in the twentieth century, the prevailing notion was that people could alter climates locally (for instance, by cutting down forests and plowing virgin fields) but not globally.

Of course, the ice ages and other wrenching climate shifts of the past were topics of research.

But few considered them an immediate threat, and hardly anyone thought humans could trigger worldwide climate change.

A few pioneering thinkers saw the potential global impact of fossil fuel use (see later discussion), but their views were typically dismissed by colleagues.

Starting in 1958, precise measurements of carbon dioxide confirmed its steady increase in the atmosphere.

The first computer models of global climate in the 1960s, and more complex ones thereafter, supported the idea floated by mavericks earlier in the century: that the addition of greenhouse gases would indeed warm the climate.

Finally, global temperature itself began to rise sharply in the 1980s, which helped raise the issue's profile among the media and the public as well as among scientists.

Couldn't the changes have natural causes?

The dramatic changes in climate we've seen in the past 120 years are not proof in themselves that humans are involved.

As contrarians are fond of pointing out, Earth's atmosphere has gone through countless temperature swings in its 4.54 billion years.

These are the results of everything from cataclysmic volcanic eruptions to changes in solar output and cyclic variations in Earth's orbit (see later discussion).

The existence of climate upheavals in the past raises a legitimate question: how can we be sure that the current warming isn't "natural"—that is, caused by something other than burning fossil fuels?

That query has been tackled directly over the last couple of decades by an increasing body of research, much of it compiled and assessed by the **Intergovernmental Panel on Climate Change (IPCC)**, a unique team that has drawn on the work of more than 1000 scientists over more than 25 years.

We'll refer often throughout this class to the IPCC's work.

As shown below, the IPCC's second, third, fourth, fifth and sixth assessments each issued progressively stronger statements on the likelihood of human involvement on the climate change of recent decades.

1995: “The balance of evidence suggests a discernible human influence on global climate.”

2001: “There is new and stronger evidence that most of the warming observed over the last 50 years is attributable to human activities.”

2007: “Human-induced warming of the climate system is widespread.”

2013: “It is extremely likely that human influence has been the dominant cause of the observed warming since the mid-twentieth century.”

2023: “Climate is already changing in every region.

Many of these changes have not been seen in thousands of years.

Many of them such as sea-level rise are irreversible over hundreds of thousands of years.

Strong reductions in greenhouse gas emissions would limit climate change.

But it could take 20-30 years for the climate to stabilize Climate change due to human activities that is already affecting the lives of billions of people and disrupting nature.

The world faces unavoidable hazards over the next two decades even with global warming of 1.5°C.

It will be impossible to limit warming to 1.5°C without immediate and deep cuts in greenhouse gas emissions.

It is still possible to halve emissions by 2050”



The densest part of the Earth’s atmosphere extends only a few kilometers above the surface, as shown in this photo taken from the international Space Station.

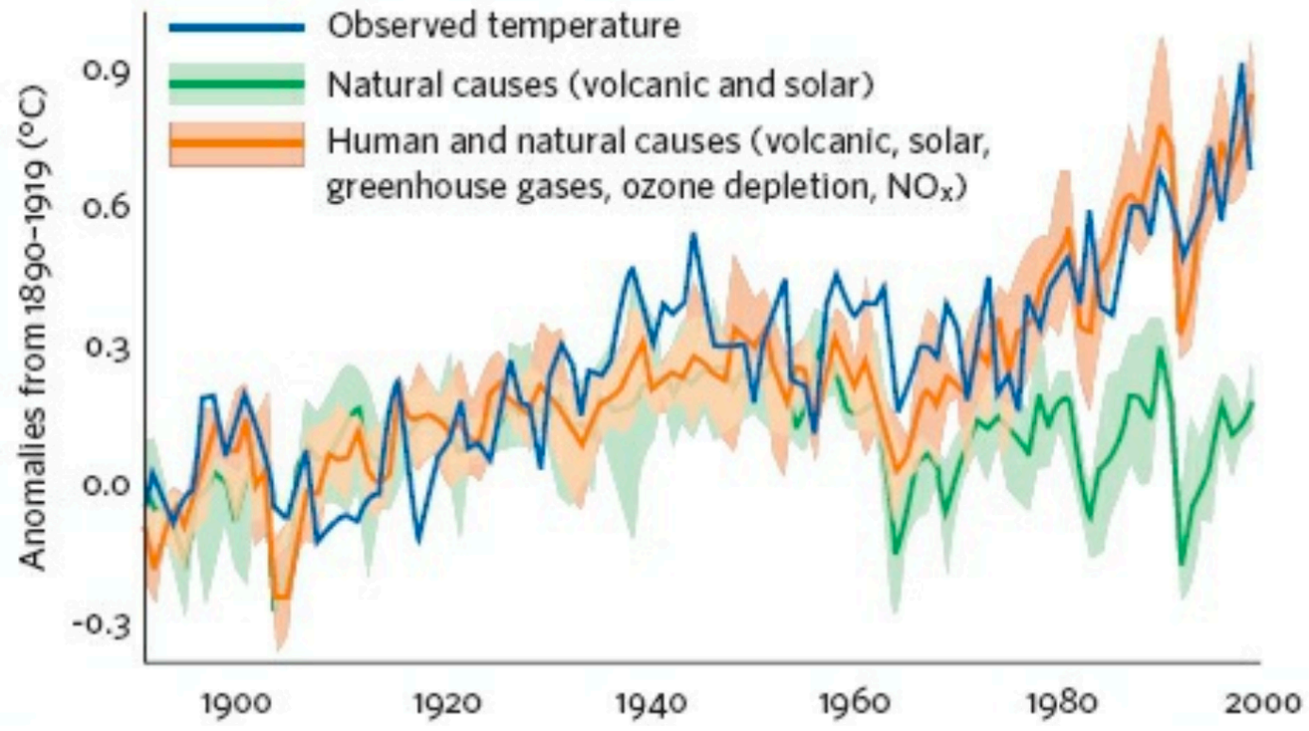
The orange sky near the bottom represents the troposphere, or “weather layer”, where temperatures have been warming .

Above this layer is the stratosphere, seen here as a light gray region, where temperatures have been cooling and ozone depletion has been a concern over recent decades.

To support claims like these, scientists call on results from two critical types of work: **detection** and **attribution** studies.

Detection research is meant to establish only that an unusual change in climate has occurred.

Attribution studies try to find the likelihood that particular factors, including human activities, are involved.



The orange and green shaded areas show the departure of 20th century global temperature from the 1890-1919 average, as produced by four computer simulations.

The orange and green lines show averages of the four models .

The green runs factored in natural agents of climate change only; the orange runs include both human and natural factors.

The blue line shows the temperature measured in the real world — —> WE ARE THE CAUSE!!

One way to attribute climate change to greenhouse gases is by looking at the signature of that change and comparing it to what you'd expect from non-greenhouse causes.

For example, over the past several decades, Earth's surface air temperature has warmed most strongly near the poles and at night.

That pattern is consistent with the projections of computer models that incorporate rises in greenhouse gases.

However, it agrees less well with the warming that might be produced by other causes, including natural variations in Earth's temperature and solar activity.

Clues like these strengthen the case that the changes we're seeing can be chalked up to greenhouse gases.

As computer models have grown more complex, they've been able to incorporate more components of climate.

This allows scientists to tease out the ways in which individual processes helped shape the course of the last century's warm-up.

One such study, conducted at the National Center for Atmospheric Research (NCAR), examined five different factors: volcanoes, sulfate aerosol pollution, solar activity, greenhouse gases, and ozone depletion.

Each factor had a distinct influence.

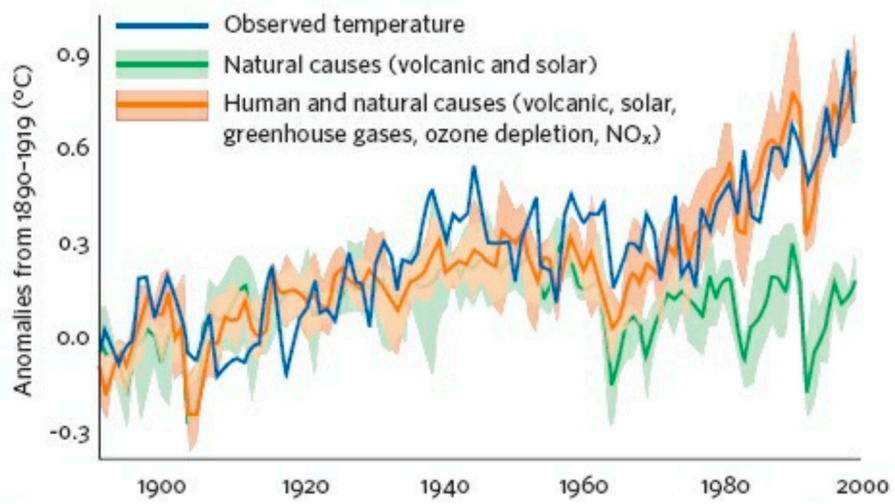
The eruption of Mount Pinatubo in 1991 helped cool global climate for several years.

Sulfate pollution peaked in the middle of the twentieth century, between World War II and the advent of environmentalism, and it may have helped produce the slight drop in global temperatures observed from the 1940s through the 1970s.

Small ups and downs in solar output probably shaped the early-century warming and perhaps the midcentury cooling as well.

However, the sun can’t account for the pronounced warming evident since the 1970s.

The bottom line (see graphic) is that the model couldn’t reproduce the most recent warming trend unless it included greenhouse gases.



Couldn’t some undiscovered phenomenon be to blame?

Although many people would love to find a “natural” phenomenon to blame for our warming planet, such as the relationship between clouds and cosmic rays, the chance that a suitable candidate will emerge has gotten extremely remote.

Even if it did, it would beg a tough question: if that newly discovered factor—something we’re not aware of now—accounts for the climate change we’ve observed, then why aren’t carbon dioxide and the other greenhouse gases producing the warming that basic physics and chemistry tell us they should be?

And there’s another catch.

Any mysterious, yet-unknown process hidden in the weeds could just as easily be a cooling as a warming agent.

If it were to suddenly wane, it could leave us with even greater warming than we imagined possible.

Trusting in the unknown, then, is a double- edged sword.

Most scientists in the trenches look instead to Occam’s razor, the durable rule credited to the medieval English logician and friar William of Occam:

One should not increase, beyond what is necessary,
the number of entities required to explain anything.

How do rainforests fit into the picture?

As a whole, Earth's land-based ecosystems absorb more CO_2 than they release, but when deforestation is considered on its own, it's a major source of carbon dioxide.

Though deforestation has eased significantly over the last few years, it still accounts for roughly 10%–15% of recent human-produced CO_2 emissions.

When tropical forests are razed or burned to clear land, the trees, soils, and undergrowth release CO_2.

Even if the land is eventually abandoned, allowing the forest to regrow, it would take decades for nature to reconcile the balance sheet through the growth of replacement trees that pull carbon dioxide out of the air.

Rainforests also cool the climate on a more local level: their canopy helps trap moisture in the atmosphere, providing a natural air-conditioning effect.

When the rainforest has been slashed and burned over large areas, hotter and dryer conditions often set in.

By contrast, in midlatitude and polar regions, it appears that forests actually tend to warm the climate, as they absorb more sunlight than bright, reflective snowfields do.

Didn't we have "global cooling" a while back?

The planet did cool slightly from the 1940s to the 1970s, mainly in the Northern Hemisphere and most likely related to the postwar boom in industrial aerosol pollutants that bounce sunlight away from Earth.

Despite a flurry of 1970s media reports on an imminent ice age, far more studies at the time were concerned about the eventual impact of greenhouse gases—and greenhouse warming has long since eclipsed the midcentury cool spell.

After temperatures reached a new global high in 1998, the following decade remained warm, but with smaller ups and downs that continued into the early 2010s.

The absence of another show-stopping record led many who deny climate change to claim that the supposed global warming "stopped" or went on "hiatus" in 1998.

To the extent atmospheric warming had indeed slackened (a topic itself debated widely), it was largely because the oceans—which take up more than 90% of the energy trapped by human-produced greenhouse gases—were stowing away even more heat than usual.

Some of that heat escaped into the air in the mid-2010s, which stoked a new round of global record highs.

In truth, nobody expects the global average temperature to rise smoothly from one year to the next.

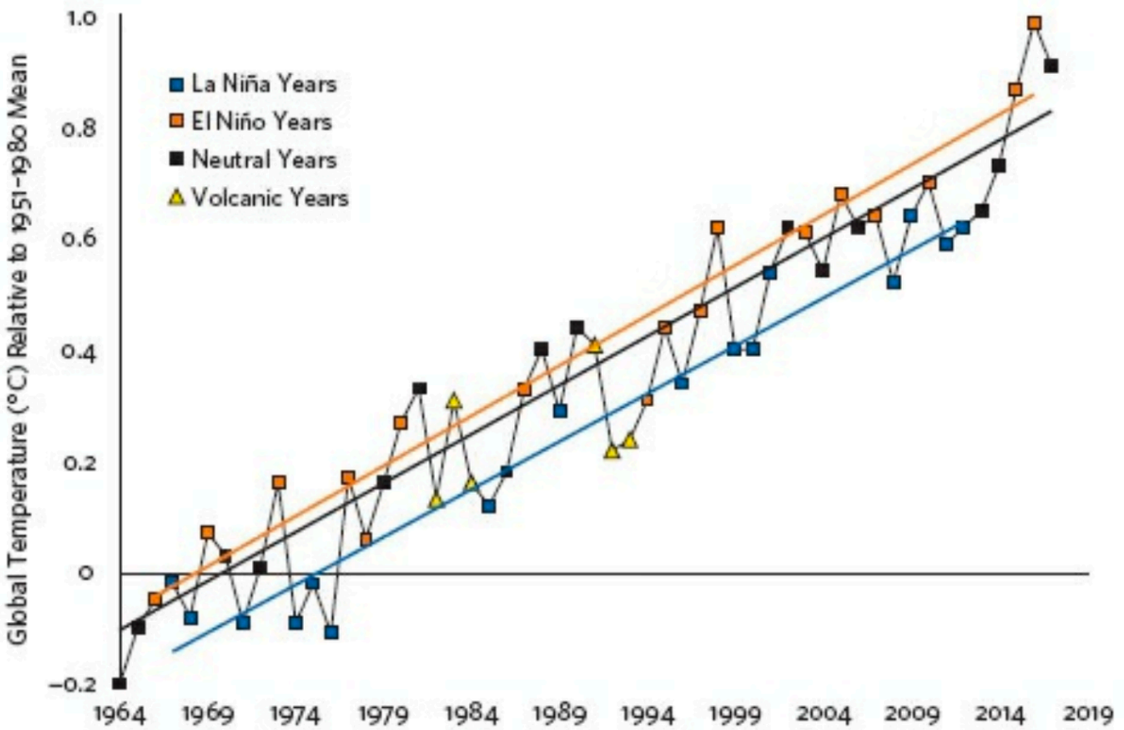
Just as any spring in London, New York, or Beijing will see a few cold snaps, we can expect long-term warming to be punctuated by periods with little rise, or even a slight drop, in global temperature.

There can be many factors driving such variability, but one big one appears to be the natural ebb and flow of heat between the oceans and atmosphere.

During El Niño, warmth held in the tropical Pacific Ocean flows into the atmosphere, and global temperatures typically rise by several tenths of a degree Celsius.

The opposite occurs during La Niña, when the Pacific draws more heat than usual from the global atmosphere.

The twenty-first century so far has seen an outsized number of La Niña years (see graph), which helps to explain the apparent temporary slowdown in atmospheric warming.



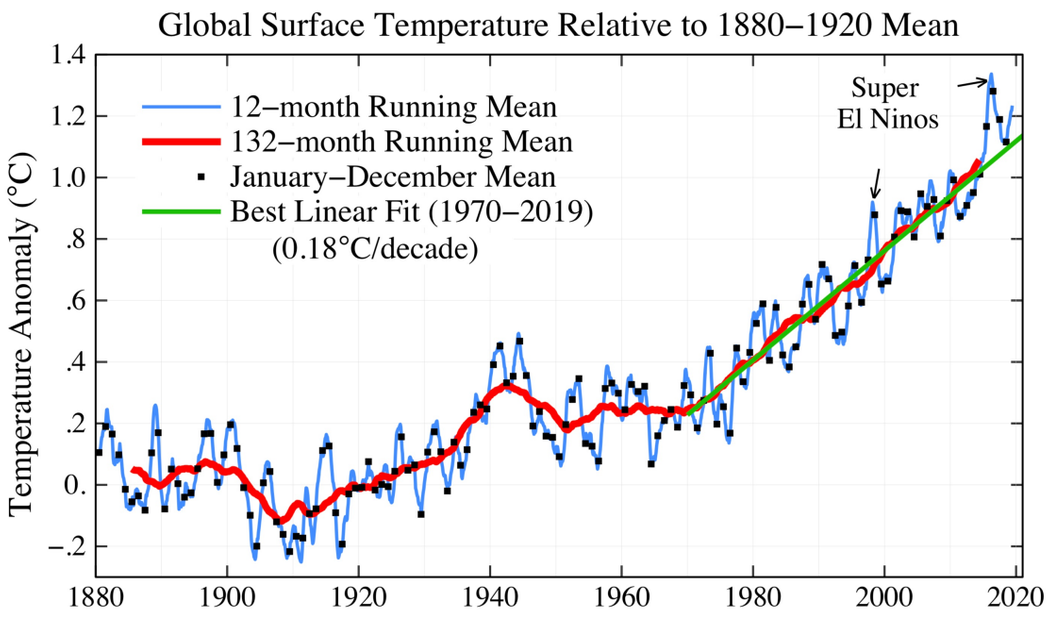
Global temperatures have trended above the 20th century averages since the late 1970s.

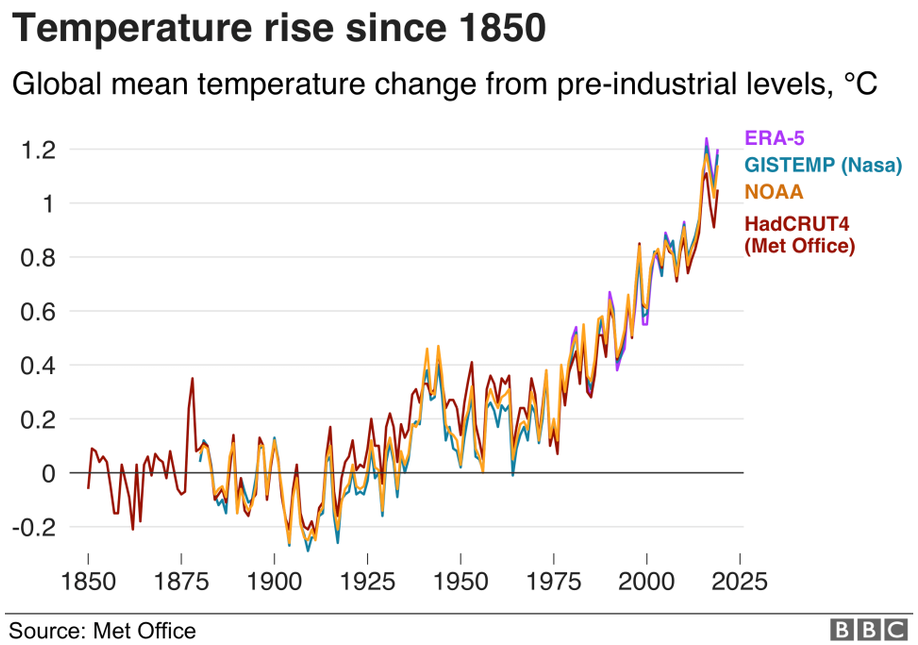
The blue squares show years when the influence of La Nina predominated.

Normally cooler than El Nino years (orange squares), La Nina years have been occurring more often this century, which appears to have helped temper atmospheric warming.

However, La Nina years are now warmer than they used to be. The mid-20210s saw the warmest years to date in all three categories. El Nino, La Nina, and neutral,

Additional graphs make it clear the trend is continuing.





None of this changes the longer-term outlook, and the extra heat tucked into the ocean during the first few years of the twenty-first century could influence global temperature far down the line.

As for individual record-warm years, these will continue to be most likely during El Niño events.

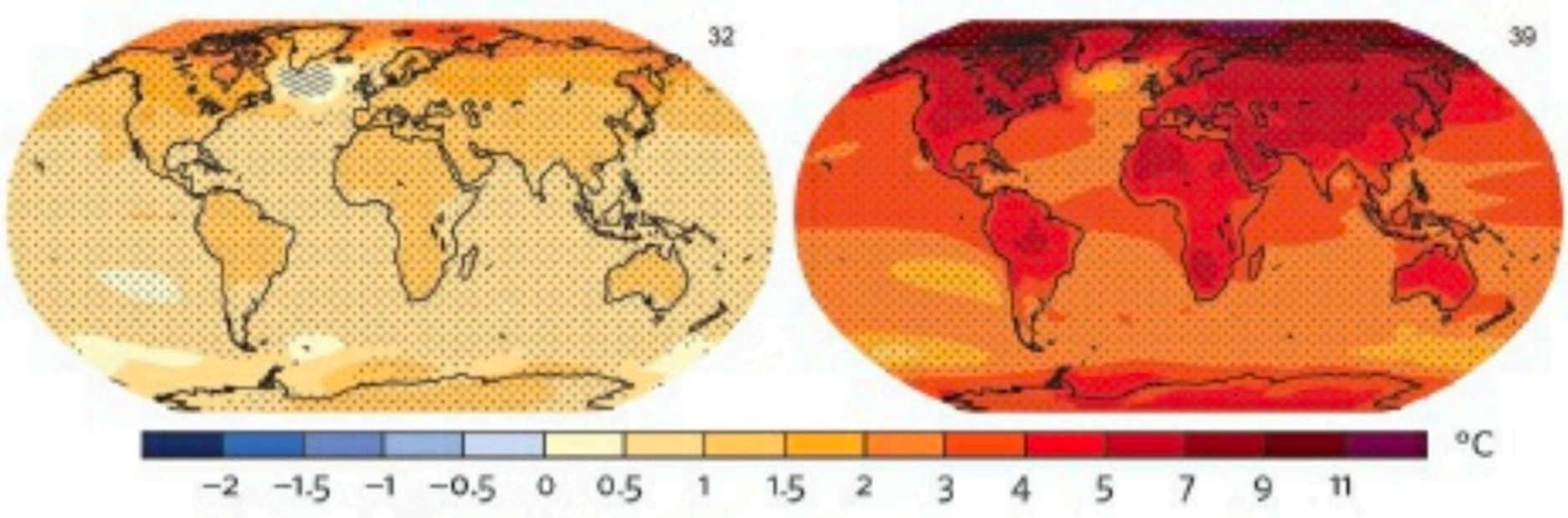
THE OUTLOOK

How hot will it get?

According to the 2013 IPCC Working Group I report, there's a wide range of possibilities for how far global average temperature is likely to climb above its 1986–2005 values.

The spectrum extends anywhere from 0.4° to 2.6°C (0.7°–4.7°F) by 2046–2065 and from 0.3° to 4.8°C (0.5°– 8.6°F) by 2081–2100.

Importantly, this would be on top of the warming of about 0.6°C (1.1°F) that occurred from the 1850–1900 period to the 1986–2005 comparison period.



Changes in annual mean surface temperature from the period 1986-2005 to 2081-2100 for two scenarios(low-end and high-end).

These represent the averages from more than 30 climate models.

In the area covered by dots, the indicated changes are large relative to natural variability and cross-hatched areas indicate where the change is small relative to natural variability.

Is global warming necessarily a bad thing?

Whether you view climate change as bad, good, or neutral depends on your perspective.

Some regions and economic sectors—and some species—are likely to benefit, at least temporarily.

Many others will suffer intense problems and upheavals, though, especially as they move further from anything in past experience.

And some of the potential impacts, such as major sea level rise, increased flood and drought risk, intensified hurricanes, and many species being consigned to extinction, are bad news from almost any perspective.

So while it may be a bit of a reach to think in terms of “saving the planet” from global warming, it’s perfectly valid to think about preserving a climate that’s sustaining to as many of Earth’s occupants as possible.

Perhaps the more pertinent question is whether the people and institutions responsible for producing greenhouse gases will bear the impacts of their choices, or whether others will— including those who had no say in the matter.

Indeed, people in the poorest parts of the world, such as Africa, will generally be least equipped to deal with climate change, even if the changes there are no worse than elsewhere.

Yet such regions are releasing only a small fraction of the gases causing the changes.

Will agriculture suffer?

The effects on agriculture will vary based on where, when, and how farming and ranching is done.

Global agricultural productivity is predicted to go *up* in some areas, at least in the near term, thanks both to the fertilizing effect of the extra CO_2 in the atmosphere and to now-barren regions becoming warm enough to bear crops.

However, the rich world looks set to reap most of the benefits, as crop yields in much of the tropics, home to hundreds of millions of subsistence farmers, are likely to experience negative impacts.

Unwanted plants could thrive in some areas.

Periodic shortages of some staple foods, such as corn, may become likely, as huge swaths of cropland are used to grow biofuels and as heat and drought become more intense.

Once global temperatures rise beyond a certain point—perhaps as soon as the 2030s, the combined regional effects of climate change are expected to become a net drain on global crop production, one that only worsens over time.

How will the changes impact wildlife?

Because climate is expected to change quite rapidly from an evolutionary point of view, we can expect major shocks to some ecosystems—especially in the Arctic—and possibly a wholesale loss of species, especially given the other stresses on biodiversity now being exerted by increasing global population and development.

To keep up with twenty-first-century warming, a wide array of species would need to evolve at rates up to 10,000 times more quickly than they did during past evolutionary shifts triggered by climate change.

Climate change between now and 2050 may commit as many as 37% of all species to eventual extinction—a greater impact than that from global habitat loss due to human land use.

Recent studies suggest a wide range of possibilities as warming proceeds.

One concluded that "a large fraction of both terrestrial and freshwater species faces increased extinction risk under projected climate change during and beyond the 21st century."

Will rising seas really put cities such as New York and London under water?

Not right away, but it may be only a matter of time.

Depending on how much greenhouse gas is emitted this century, the IPCC's 2013 assessment concluded that sea levels by 2081–2100 could be anywhere from 0.26 to 0.81 m (10–32 in.) higher than they were in the period 1986–2005.

Especially toward the high end, that's enough to cause major complications for millions of people who live and work near coastlines.

More recent work points to a risk of substantially bigger rises this century—on the order of 1 to 2 m (39–79 in.)—should the huge ice shelves and cliffs that hold back Antarctica's ice sheets become destabilized.

Changes in ocean circulation are expected to push sea level even higher than the global average in some areas, including the northeast U.S. coast.

Another complication for many towns and cities is that a rise of just 0.5 m will bring everyday water levels close to the range of some current high tides—which means that by later this century, vulnerable roads and buildings might see "nuisance" flooding every few days instead of once or twice a year.

Unfortunately, sea level won't stop rising in 2100, which means coastal areas will have to contend with ever-worsening problems.

If emissions continue to rise unabated through this century, the Greenland and/or West Antarctic ice sheets could eventually be thrown into an unstoppable melting cycle that would raise sea level by more than 7 m (23 ft) each.

This process would take some time to unfold—possibly a few centuries, though nobody can pin it down at this time, and it appears that some of the ice cliffs ringing West Antarctica and holding back glaciers are more prone to sudden collapse than earlier believed.

The upshot is that many of the world's most beloved and historic coastal cities may be hard-pressed to survive without drastic measures to hold back the sea.

Will the Gulf Stream stop, freezing the United Kingdom and northern Europe?

The Gulf Stream and North Atlantic Current bring warm water (and with it warm air) from the tropical Atlantic to northern Europe.

This helps keep the United Kingdom several degrees warmer than it would otherwise be.

Although this system is unlikely to collapse entirely, there is a possibility that it could be diminished by climate change.

One reason is that increasing rainfall and snowmelt across the Arctic and nearby land areas could send more freshwater into the North Atlantic, possibly pinching off part of the warm current.

The Atlantic meridional overturning circulation—an index often used to gauge the northward flow of warm water—is already at its weakest in more than a thousand years, and it might weaken further this century by anywhere from 12% to 54%, on the basis of the highest-emission scenario considered in the 2013 IPCC assessment.

That's probably not enough to offset global warming completely for the United Kingdom or northwest Europe, although it could certainly put a dent in it.

Will we reach a "tipping point"?

Some aspects of climate change are expected to unfold in a roughly linear fashion.

Computer models, and the changes we've seen to date, suggest that many of the regional and seasonal fingerprints of our future climate will be roughly twice as strong for 2°C of global warming as for 1°C.

On top of these incremental changes, there's a real risk of even bigger trouble because of positive-feedback processes.

Although “positive feedback” may sound like a good thing, in the physical world this process tends to amplify change, pushing conditions well beyond what one would otherwise expect.

The challenge is to identify the points at which the most dangerous positive feedbacks will kick in.

For instance, scientists consider it likely that the Greenland ice sheet will begin melting uncontrollably if global temperatures climb much more than 2°C (3.6°F) beyond preindustrial readings.

Because of the implications for coastal areas, as noted above, this is a particularly worrisome threshold.

As each positive feedback has its own triggering mechanism, some would no doubt occur sooner than others.

Thus, there is no single temperature agreed upon as a tipping point for Earth as a whole.

However, scientists, governments, and activists have worked to identify useful targets that might focus attention and action.

The longtime goal of the European Union and many environmental groups—limiting global temperature rise to no more than 2°C (3.6°F) over preindustrial levels, as specified in the Paris Agreement—looks increasingly unrealistic.

We're already more than halfway there, and only the most optimistic of the latest IPCC projections keep us below the 2°C threshold by century's end.

Another approach is to set a stabilization level of greenhouse gases—a maximum concentration to be allowed in the atmosphere, such as 500 parts per million (ppm) of CO_2 as compared to 270–280 ppm in preindustrial times and around 410 ppm today.

Still another concept is to limit the total amount of carbon added to the air over time.

This perspective is especially useful in demonstrating that we likely have far more oil, coal, and gas available than we can afford to burn without hitting the 2°C point.

WHAT CAN WE DO ABOUT IT? What would it take to stop climate change?

It isn't possible to simply "stop" climate change.

The century-plus lifespan of atmospheric CO_2 means that the planet is already committed to a substantial amount of greenhouse warming.

Even if we turned off every fuel-burning machine on Earth tomorrow, climate modelers tell us that the world would warm at least another 0.5°C (0.9°F) as the climate adjusts to greenhouse gases we've already emitted.

The best the world can do, therefore, is adapt to whatever warming can't be avoided while trying to limit the potential amount of additional warming.

The latter will only be possible to the extent changes in technology and lifestyle enable us to pull back far beyond our current emission levels, or unless we find some safe method to remove enormous amounts of carbon from the atmosphere, or both.

That's a tall order, but if we're determined to reduce the risk of a wide range of climate impacts, we have no choice but to fulfill it.

Given the global nature of the climate problem, slashing emissions on the scale needed will almost certainly only be possible as part of an ambitious global accord, such as the Paris Agreement together with any potential successors.

Otherwise, there's a big risk: if one region or nation cuts back emissions on its own, that could actually bring down the price of fossil fuels on the open market and thus spur consumption elsewhere.

Which countries are emitting the most greenhouse gases?

For many years the United States was in first place, with 30% of all of the human-produced greenhouse emissions to date and about 20% of yearly totals, despite having only a 5% share of the global population.

However, China has now taken the lead.

Its emissions are considerably lower per capita, but because of its growing population and affluence, China leapfrogged the United States as the world's leading greenhouse emitter around 2006.

By 2014, China accounted for more than 28% of the planet's annual energy-related carbon dioxide emissions, according to the Netherlands Environmental Assessment Agency, while the U.S. share had dropped to around 14%.

Together, the United States and China account for an ever-increasing slice of the globe's greenhouse emissions—more than 40%, as of 2014.

Won't nature take care of global warming in the long run?

Only in the *very* long run - **if humans disappeared, the Earth will fix itself!!!!**

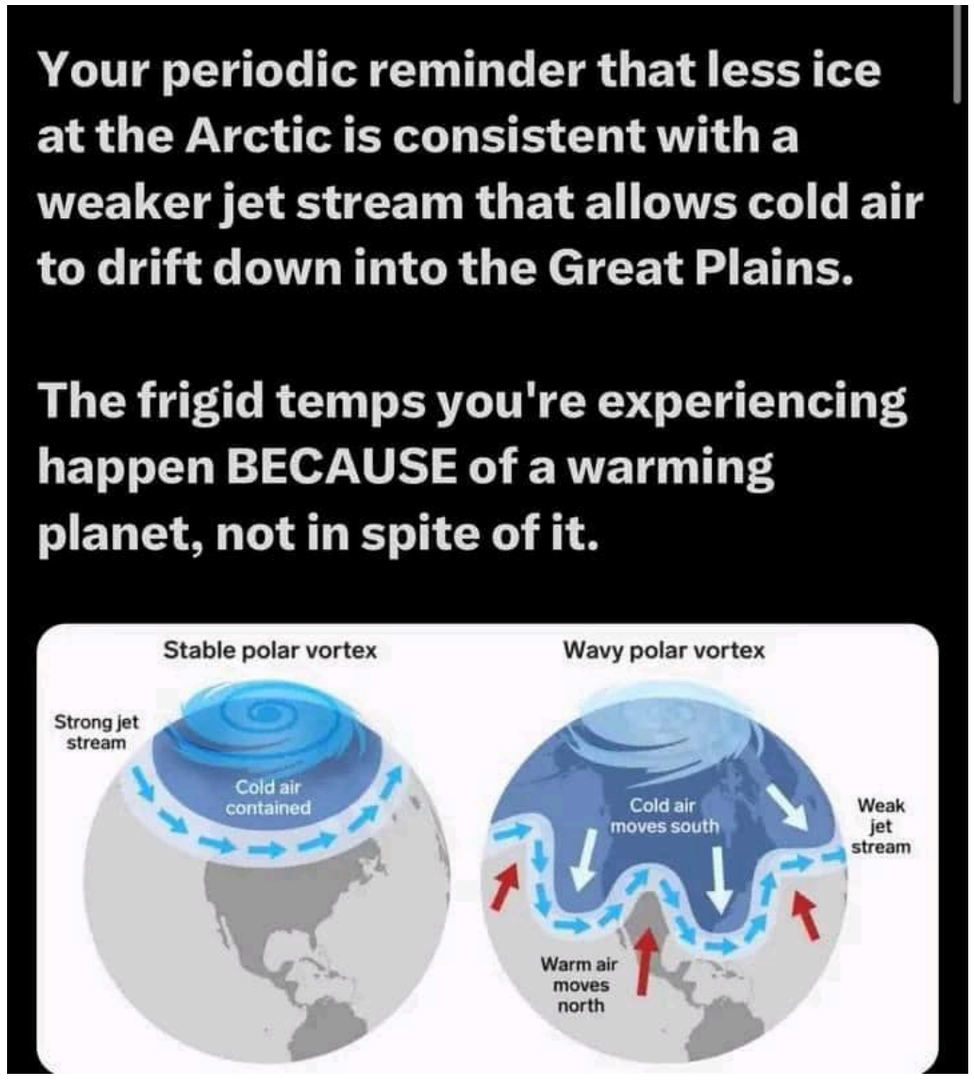
Assuming that it takes a century or more for humanity to burn through whatever fossil fuels it's destined to emit, it would take hundreds of years before most of the CO_2 emissions were absorbed by Earth's oceans and ecosystems, and some of the remaining human-produced CO_2 would remain in the air for many thousands of years.

There are few analogs in the geological past for such a drastic change in global climate over such a short period, so it's impossible to know what will happen after the human-induced greenhouse effect wanes.

All else being equal, cyclical changes in Earth's orbit around the sun can be expected to trigger an ice age sometime within the next 50,000 years, and other warmings and coolings are sure to follow.

In the meantime, we will have our hands full dealing with the decades ahead of us and the serious climate changes that our way of life may help bring about.

An Aside: When you have a cold winter and your friend says "what happened to global warming?"



Imagine our planet suddenly stripped of its atmosphere—a barren hunk of rock floating in space.

If this were the case, then Earth's near-ground temperature would soar by day but plummet by night.

The average would be something close to a bone-chilling –18°C (0°F).

In reality, though, Earth's surface temperature now averages a much more pleasant 14.5°C (58.1°F).

Clearly, there's something in the air that keeps things tolerably warm for humans and other living things.

But what?

One of the first people to contemplate Earth's energy balance was the French mathematician and physicist Joseph Fourier.

His calculations in the 1820s were the first to show the stark temperature contrast between an airless Earth and the one we actually enjoy.

Fourier knew that the energy reaching Earth as sunlight must be balanced by energy returning to space, some of it in a different form.

And though he couldn't pin down the exact process, Fourier suspected that some of this outgoing energy is continually intercepted by the atmosphere, keeping us warmer than we would otherwise be.

Harking back to experiments by others on how a glass box traps heat, Fourier likened the atmosphere to a hothouse (or greenhouse), and voilà—the concept of the **greenhouse effect** was born.

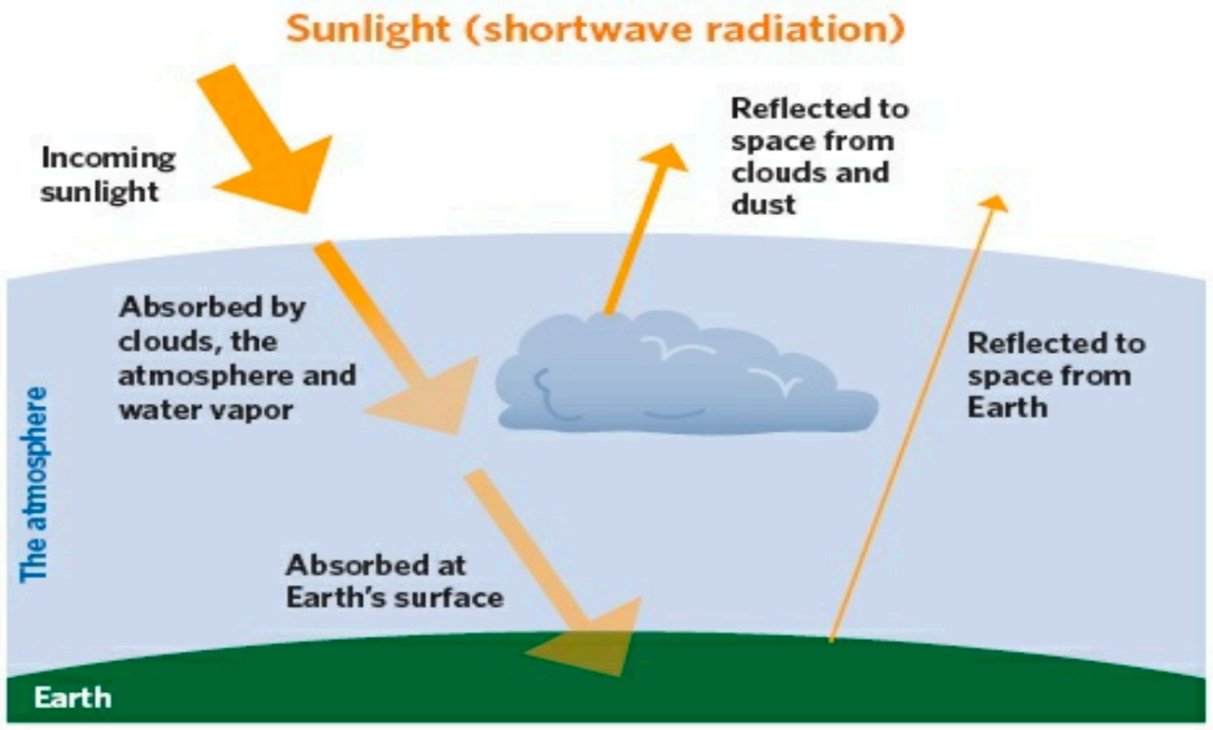
It's been with us ever since, even though it's a flawed analogy.

The atmosphere doesn't imprison the air the way a glass box does.

Instead, it absorbs **infrared radiation** rising from Earth's sun-warmed surface.

All else being equal, the more greenhouse gas there is, the less radiation can escape from Earth to space, and the warmer we get (but there are a few twists along the way, as we'll see).

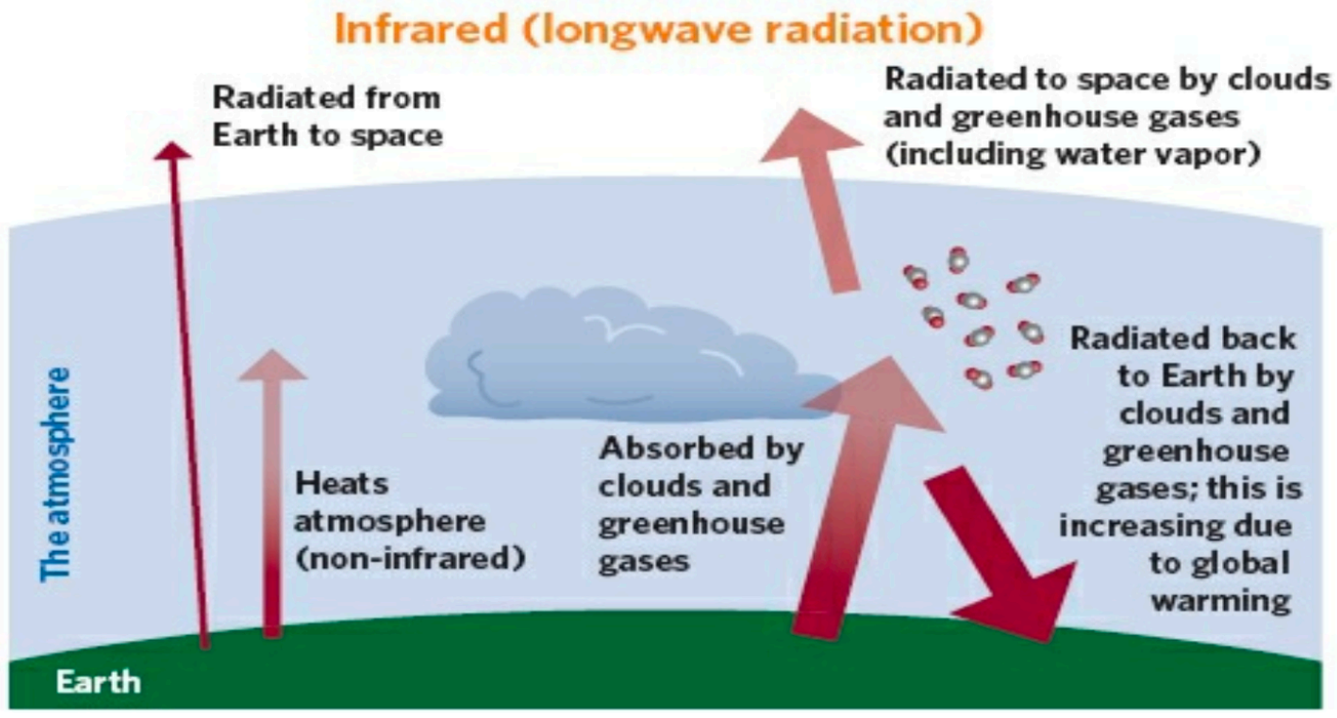
The diagrams show what happens, on average, to the sunlight that reaches our planet:



Incoming and outgoing radiation: the diagram left shows sunlight entering the atmosphere.

The light absorbed (as opposed to reflected) is given out as infrared radiation as shown in the diagram below.

The widths of the arrows in both diagrams reflect the warming power of each process, relative to the total sunlight coming in.



About 30% gets reflected or scattered back to space by clouds, dust, or Earth's surface. (Light- colored features, such as snow and ice, reflect far more than they absorb, while dark features such as forests absorb more than they reflect.)

More than 20% is absorbed in the atmosphere, mainly by clouds and water vapor.

Almost 50% gets absorbed by Earth's surface—land, forests, pavement, oceans, and the rest.

The incoming radiation from the intensely hot sun is mostly in the visible part of the spectrum, which is why you shouldn't stare at the sun.

Being much cooler than the sun, Earth emits far less energy, most of it at infrared wavelengths we can't see.

Some of Earth's outgoing radiation escapes through the atmosphere directly to space.

Most of it, though, is absorbed along the way by clouds and greenhouse gases (including water vapor), which in turn radiate some back to the surface and some out to space.

Thus, Earth's energy budget is maintained in a happy balance between incoming radiation from the sun and a blend of outgoing radiation from a warm surface and a cooler atmosphere (an important temperature distinction, as we'll see shortly).

The air's two main components, nitrogen (78%) and oxygen (20%), are both ill-suited for absorbing radiation from Earth, in part because of their linear, two-atom (diatomic) structure.

But some other gases have three or more atoms, and these branched molecules capture energy far out of proportion to their scant presence.

These are the **greenhouse gases**, the ones that keep Earth inhabitable but appear to be making it hotter.

Most greenhouse gases are well mixed throughout the **troposphere,** the lowest 8–16 km (5–10 mi) of the atmosphere.

Water vapor is the big exception; it's much more concentrated near ground level.

As mountain climbers well know, the troposphere gets colder as you go up, and so these greenhouse gases are cooler than Earth's surface is.

Thus, they radiate less energy to space than Earth itself would.

That keeps more heat in the atmosphere and thus helps keep our planet livable.

That's all well and good, but the more greenhouse gas we add, the more our planet warms.

As carbon dioxide and other greenhouse gases accumulate, the average elevation from which they radiate energy to space ascends to even higher, and colder, heights.

This means less energy leaving the system, which in turn causes the atmosphere to heat up further.

Once the added gases get the ball rolling through this process, a series of atmospheric readjustments follows.

On the whole, these tend to be **positive feedbacks**—ones that amplify the warming instead of keeping it in check.

More water evaporates from oceans and lakes as temperatures warm, for instance, and this roughly doubles the impact of a given increase in carbon dioxide.

Melting sea ice reduces the amount of sunlight reflected to space.

Some feedbacks are less certain: we don't know whether cloud patterns will change to enhance or diminish the overall warming.

And, of course, we're not talking about a one-time shock to the system.

The whole planet is constantly readjusting to the greenhouse gases we're adding.

A rogue's gallery of gases

In the 1850s, an amateur U.S. scientist named **Eunice Foote** discovered that the air inside sunlit glass jars warmed to different temperatures based on whether the jars contained pure hydrogen, carbon dioxide, or ambient air.

Although Foote's measurements weren't enough to isolate what we've come to know as the greenhouse effect, her 1856 paper "Affecting the Heat of the Sun's Rays" was apparently the first to hypothesize that Earth's temperature might be warmer if the atmosphere contained more carbon dioxide.

Several years later in England, **John Tyndall** became the first to explore and document the remarkable power of greenhouse gases in the laboratory.

Tyndall put a number of gases to the test in his lab, throwing different wavelengths of light at each one to see what it absorbed.

Almost as an afterthought, he tried coal gas and found it was a virtual sponge for infrared energy.

Tyndall went on to explore carbon dioxide and water vapor, both of which are highly absorbent in certain parts of the infrared spectrum.

The broader the absorption profile of a gas—that is, the more wavelengths it can absorb—the more powerful it is from a greenhouse perspective.

Here are the main greenhouse gases of concern.

Carbon dioxide (CO_2), the chief offender, accounted for roughly 410 of every one million molecules in the air, or 410 **parts per million (ppm)**, by the year 2018, as measured on top of Hawaii's Mauna Loa.

That number varies by about 8 ppm with the seasonal cycle: it first touched the 400-ppm threshold midway through 2013, then sank below it later in the year.

The yearly average has been climbing by 1–3 ppm, or about 0.25%–0.75%, per year (see later).

The worldwide **emissions** of CO_2 from fossil fuel use increased by several percent a year up through the early 2010s.

The rate of emission increase dropped to near zero in the mid-2010s, most likely because of global shifts away from coal and toward natural gas and renewable energy, before it again topped the 1% mark in 2017.



Both a pollutant and a natural part of the atmosphere, carbon dioxide is produced when fossil fuels are burned, as well as when people and animals breathe and when plants decompose.

It's the energy-packed quality of fossil fuels (containing millions of years of stored carbon) that allows them to add so much CO_2 to the air so quickly when we burn them.

Plants and the ocean can soak up huge amounts of carbon dioxide, which is helping to keep CO_2 levels from increasing even more rapidly (see later).

Because the give-and-take among these processes is small compared to the atmospheric reservoir of CO_2, it takes a long time for an increment of carbon dioxide added to the air to leave it.

More than half of the pulse will still be there a century later, and some of the rest will remain in the atmosphere for thousands of years more.

Atmospheric levels of CO_2 held fairly steady for centuries—at around 270–280 ppm—until the Industrial Revolution took off.

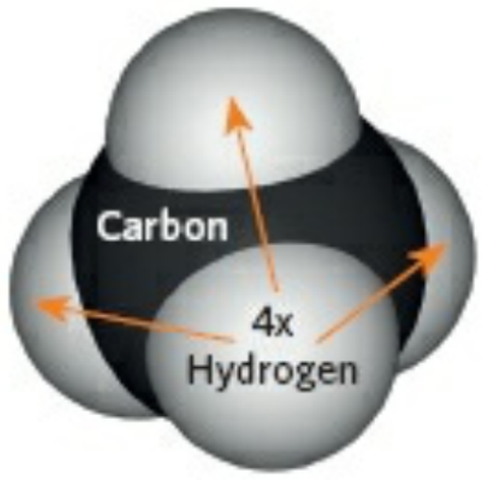
In past geological eras, the amount of CO_2 has risen and fallen in sync with major climate changes.

Although factors other than carbon dioxide can kick off some of these transitions, CO_2 is a critical factor in making them more powerful.

Methane emerges from a wide array of sources—including rice paddies, peat bogs, belching cows, and insect guts, as well as vehicles, homes, factories, wastewater, and landfills—making it far more challenging to analyze globally than CO_2.

Methane is a greenhouse powerhouse: though it stays in the air for less than a decade on average, its effect on global warming is roughly 30 times that of carbon dioxide per molecule when both are considered over a century-long span.

Methane's total annual contribution to the greenhouse effect is thought to be roughly a third as big as carbon dioxide's, even though it makes up only around 2 ppm of the atmosphere.



After soaring in the last few decades, the amount of atmospheric methane increased more slowly in the 1990s; it then held almost constant through much of the 2000s before resuming a strong upward trend into the 2010s.

Because methane is shorter-lived than carbon dioxide and doesn't mix as thoroughly in the atmosphere, it's been a challenge to nail down these trends and their causes.

Some of the decade-plus slowdown in net emissions may have been related to more frequent dry conditions in the tropics, along with sagging economies across Eastern Europe and Russia in the 1990s.

As for the renewed rise in emissions, a satellite-based analysis published in 2018 attributed the jump mainly to expanded natural gas operations in North America related to hydraulic fracturing **(fracking)**, together with increased emissions from moistened rice paddies and wetlands.

Biomass burning, another source of methane, appears to be on the decrease globally.

Studies thus far have produced wildly varying results on the percentage of methane lost to the atmosphere by fracking, largely through pipeline leaks.

Estimates vary from as low as 0.3% to as high as 9%, with the actual leakage rates likely depending on local regulation and other factors.

Methane releases from a warming Arctic are not expected to become a major issue for some time, although the longer-term risks are sobering indeed.

Ozone, like a versatile actor, takes on multiple roles, but in the greenhouse drama, it's only a supporting player.

Instead of being emitted when fossil fuels are burned, ozone forms when sunlight hits other pollutants and triggers ozone-making reactions.

Its presence can spike to unhealthy levels when stagnant air takes hold for days near ground level, where people, animals, and plants live and breathe.

Fortunately, ozone survives only a few days in the troposphere.



This makes it hard to assess its global concentration.

Ozone varies greatly by location, altitude, weather conditions, and time of day.

Concentrations can range from less than 10 parts per billion (ppb) in clean ground-level air to more than 100 ppb in polluted regions.

There's apparently been little overall change in tropospheric ozone amounts since the 1980s, perhaps because of cleaner-burning cars and factories, but models hint at a global increase of about 30% since the Industrial Revolution.

Higher up, a protective natural layer of ozone warms the stratosphere while helping to shield Earth from ultraviolet light.

The human-induced hole that appears each year in this layer of “good” ozone has led to record-low temperatures in the lower stratosphere.

Water vapor isn’t a very strong greenhouse gas, but it makes up for that weakness in sheer abundance.

If you’re on a warm tropical island with 100% relative humidity, that means the balmy breezes are wafting as much water vapor as they can—perhaps as much as 5 out of every 100 molecules in the air as a whole.

When averaged globally, the concentration of water vapor is much less, and it varies greatly by location, altitude, and time of year, with colder air able to carry less of it.



You can’t see water vapor itself, only the haze, clouds, rain, and snow it produces.

Water vapor works to propagate itself and boost global warming through an interesting twist.

As temperatures rise, the atmosphere can accommodate more water vapor—up to 7% more for every degree Celsius of warming.

In response to this, more water vapor tends to evaporate from oceans, lakes, and other moist surfaces—and the added water vapor enhances the warming cycle via the greenhouse effect.

This is one of several positive feedbacks critical to the unfolding of future climate change.

Changes in satellite-based sensors have complicated attempts to analyze water vapor trends across the globe, but the bulk of evidence points toward an overall increase since the 1990s, especially across the oceans.

A few other gases, which are extremely scant but extremely powerful, make up the rest of the greenhouse players.

Chlorofluorocarbons (CFCs) and related compounds increased rapidly until they were identified as a key player in stratospheric ozone depletion.

Under the Montreal Protocol, they've begun to level off.

Along with helping to destroy "good" ozone, they are also powerful, long-lived greenhouse gases—another good reason we're phasing them out, though some replacements are just as risky for climate.

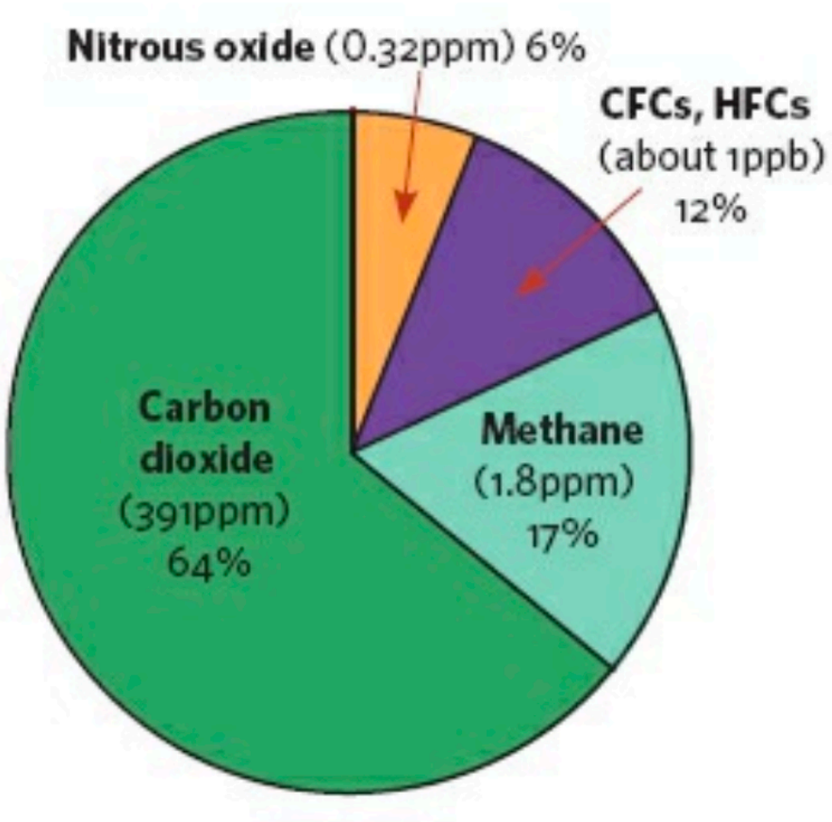
These include **hydrofluorocarbons** (HFCs), which are themselves being phased out through the UN's **Kigali Amendment** starting in 2019.

Nitrous oxide is another industrial and agricultural by-product, commonly referred to as laughing gas.

It shows up at only about 330 ppb, but with about 300 times the molecule-for- molecule effect of CO_2 over its century-long lifespan in the atmosphere.

Though its immediate effect on radiative balance is less than that of CFCs, its long lifetime makes it far more important than CFCs when the effects are measured over a 100-year time frame, and its atmospheric concentration in the 2010s was increasing at roughly 0.3% per year.

The diagram shows the prevalence of each gas and how much of an impact it's had in enhancing the overall greenhouse effect.



The prevalence of human-produced greenhouse gases and the best estimate of their relative importance in affecting Earth's radiative balance in the year 2011.

The percentages of radiative forcing depends on each gas's prevalence and its molecular ability to trap energy.

The numbers total less than 100% because of rounding and other trace gases.

The percentages do not account for the longevity of each gas: when considered over its longer lifetime, carbon dioxide's total radiative impact per molecule is larger than methane's, for example.

Not shown are tropospheric ozone (produced indirectly) and water vapor.

Radiative effects from airborne particles (aerosols) and clouds are omitted.

Greenhouse pioneers

Virtually no one at the peak of the Victorian age reckoned that burning coal or oil would tamper with our climate, but it was becoming clear that great changes in climate had occurred before.

Chief among those were the ice ages, which coated North America and Eurasia with kilometer-thick sheets of ice as far south as modern-day Germany and Illinois.

Fossils proved that this ice cover had persisted well into the era of early human civilization, some 12,000 years ago.

What made the climate plummet into such frigidity and then recover?

And could it happen again?

In the mid-1890s, the Swedish chemist **Svante Arrhenius** took a look at how carbon dioxide might be implicated in ice ages.


Svante Arrhenius

His unexpected results provided our first real glimpse of a greenhouse future.

Scientists already knew that volcanoes could spew vast amounts of greenhouse gas, so Arrhenius wondered if a long period of volcanic quiet might allow carbon dioxide levels to draw down and perhaps help plunge Earth into an ice age.

He set out to calculate how much cooling might result from a halving of CO_2.

When he included the role of water vapor feedback, Arrhenius came up with a global temperature drop of around 5°C (9°F).

Soon enough, a colleague inspired him to turn the question around: what if industrial emissions grew enough to someday double the amount of carbon dioxide in the air?

Remote as that possibility seemed, Arrhenius crunched the numbers again and came up with a similar amount, this time with a plus sign: a warming of about 5°C (9°F).

Amazingly, this century-old calculation isn't too far off the mark.

Scientists now use a **doubling of CO_2** as a reference point for comparing computer models and a benchmark for where our atmosphere might stand by around 2050 or a little afterward, if present trends continue.

Most models have pegged the **equilibrium warming** from a doubling of CO_2 (the point, many years after the doubling, when temperatures have stabilized) at somewhere between 1.5° and 4.5°C (2.7°–8.1°F).

The 2013 IPCC report reinforced this range as the most likely.

All in all, Arrhenius's initial forecast was quite impressive for the pre-computer era.

In his own time, though, Arrhenius was a lone voice.

He himself far underestimated the pace of global development, figuring it would take until the year 4000 or so for his projected doubling to occur.

Even then, he figured, a little global warming might not be such a bad thing.

Europe was just emerging from the hardships of the Little Ice Age, which had much of the continent shivering and crops withering from about 1300 to 1850.

Arrhenius figured a little extra CO_2 might help prevent another such cold spell.

Another greenhouse pioneer came along in the 1930s, by which time the globe was noticeably heating up. **Guy Stewart Callendar**, a British engineer who dabbled in climate science, was the first to point to human-produced greenhouse gases as a possible cause of already observed warming.

Callendar estimated that a doubling of CO_2 could produce roughly 2°C (3.6°F) of warming.

Though he continued to study the problem into the 1960s, Callendar's work gained little notice at the time, perhaps in part because global temperatures were cooling slightly at that point.

In any case, many scientists discounted the message of Arrhenius and Callendar because of the fact that CO_2 and water vapor absorbed energy at overlapping wavelengths.

Laboratory tests seemed to show that the two components of the atmosphere were already doing all the absorbing of infrared energy that they could: enlarging the atmospheric sponge, as it were, could have only a minuscule effect.

Only after World War II did it become clear that the old lab tests were grievously flawed because they were carried out at sea level.

In fact, carbon dioxide behaved differently in the cold, thin air miles above Earth, where it could absorb much more infrared radiation than previously thought.

And CO_2's long lifetime meant that it could easily reach these altitudes.

Thus, the earthbound absorption tests proved fatefully wrong—one of many dead ends that kept us from seeing the power of the greenhouse effect until industrialization was running at full tilt.

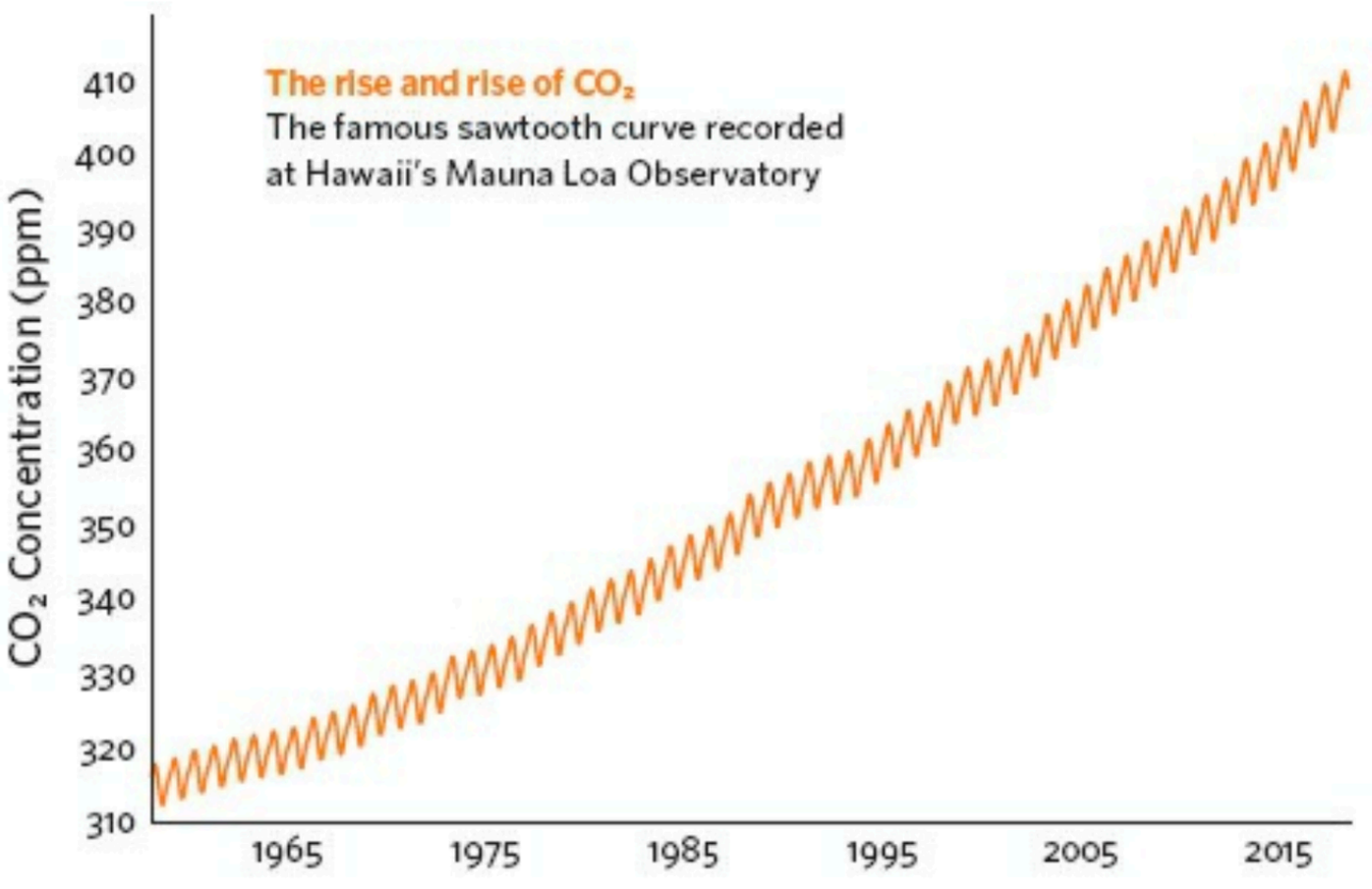
The tale told by a curve

If there's one set of data that bears out the inklings of Arrhenius and Callendar, it's the record of CO_2 collected on top of Hawaii's Mauna Loa Observatory since 1958 (see graph below).

Charles Keeling convinced the Scripps Institution of Oceanography to fund the observing site as part of the International Geophysical Year.

Because of CO_2's stability and longevity, Keeling knew that the gas should be well mixed throughout Earth's atmosphere, and thus data taken from the pristine air in the center of the Pacific could serve as an index of CO_2 valid for the entire globe.

After only a few years, Keeling's saw-toothed curve began to speak for itself (see graph).

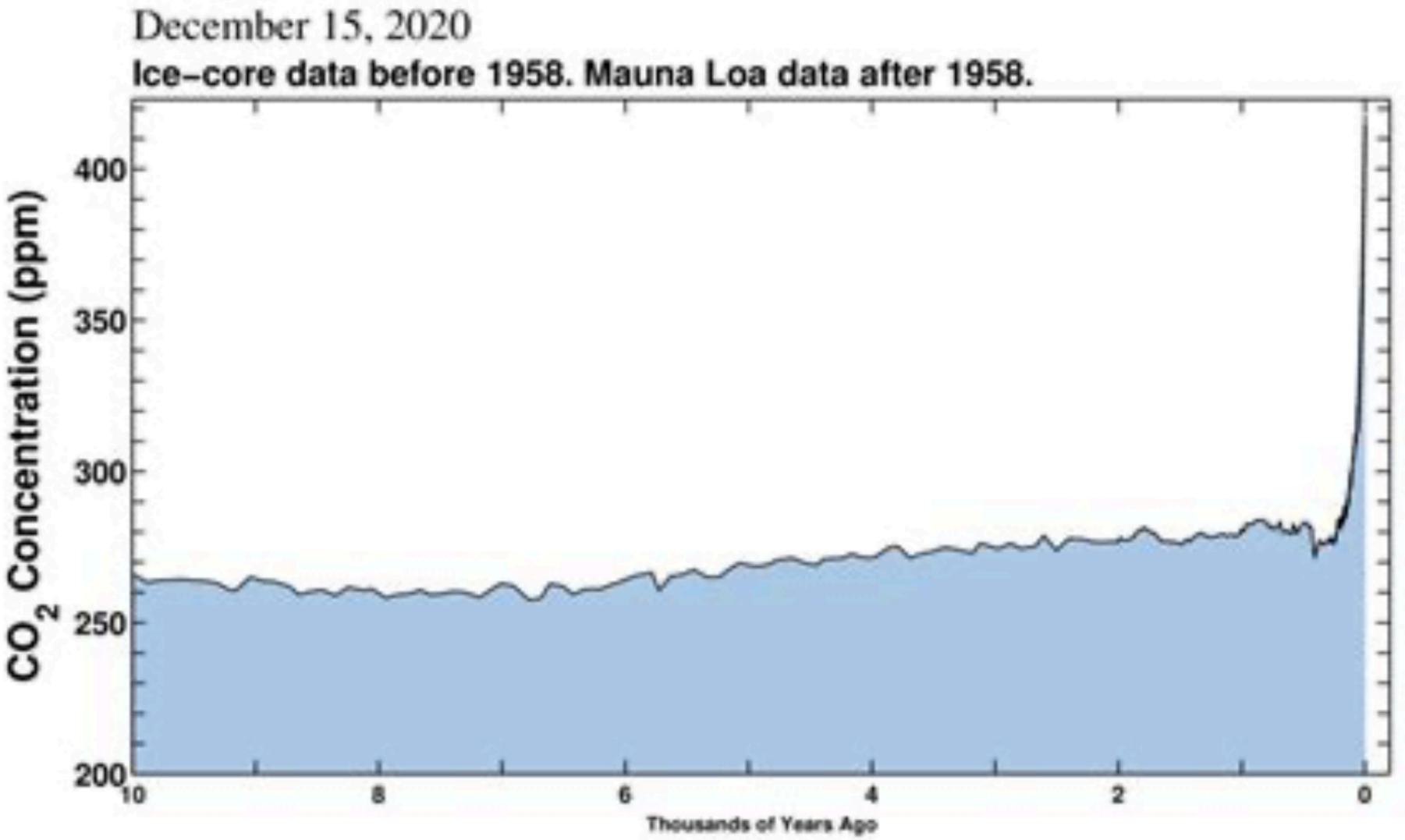


Carbon dioxide measurements taken on top of Hawaii's Mauna Loa since 1958 have produced this famous saw-toothed curve.

The annual ups and downs reflect the cycle of plant growth and decay in the Northern Hemisphere, where land areas are more extensive.

It showed a steady long-term rise in CO_2, along with a sharp rise and fall produced each year by the wintering and greening of the Northern Hemisphere (whose land area and plant mass far outstrip those of the Southern Hemisphere).

As a prelude to future class discussions consider this graph: look carefully at the end!



With Keeling's curve as a backdrop, a growing number of scientists began to wonder whether human-induced warming might take shape much sooner than Arrhenius predicted, as population grew and industrialization proceeded.

In the 1960s, climate scientists devised their first primitive renditions of Earth's atmosphere in **computer models.**

Like the first crude motion pictures, even these early models had a compelling quality.

One model created by Syukuro Manabe and Richard Wetherald at the U.S. Geophysical Fluid Dynamics Laboratory confirmed in 1967 that a doubling of CO_2 could produce warming of 2.0°C (3.6°F).

The increasingly complex models developed since the 1960s have refined the picture, but they haven't changed it all that much (see later discussion).

The most widely accepted estimate of warming for a doubling of CO_2 is in the ballpark of 3.0°C (5.4°F).

The Mauna Loa data continue to show carbon dioxide concentrations rising inexorably, and despite ratification of the Kyoto Protocol and the Paris Agreement, the world remains perilously far from even beginning to stabilize CO_2 levels in the atmosphere.

Although Arrhenius foresaw a century ago that humans could accentuate the natural greenhouse effect, he would doubtless be amazed at how quickly the process is unfolding.

3 — Who's Responsible?

WHICH COUNTRIES, REGIONS, AND ACTIVITIES ARE WARMING THE WORLD?

It took an awful lot of coal, gas, and oil passing through engines and stoves to boost the atmosphere's level of carbon dioxide by roughly 40% over the last century and a half.

There's been able assistance in warming the planet from other heat-trapping greenhouse gases, such as methane, as well as from deforestation and other changes in land use.

Still, most of the human-produced global warming to date appears to be due to CO_2.

Where did it all come from, and what are the main sources today?

Asking these questions isn't merely an exercise in finger pointing.

In order to get a handle on where global emissions are headed, it's critical to know what parts of the world and what economic sectors are contributing the most to the current situation.

When it comes to the greenhouse effect, one nation's emissions are everyone's problem.

Ordinary pollution disperses in a few days and tends to focus its worst effects in the region where it was generated, plus some areas downwind.

Carbon dioxide is much longer lived: once added to our current atmosphere, a pulse of carbon dioxide will take more than a century just to dwindle by half.

That gives the natural mixing processes of the atmosphere time to shake and stir CO_2 evenly around the globe.

Some of the countries producing minimal amounts of greenhouse gases, such as tiny island nations, are among the most vulnerable to the climatic fallout from emissions produced largely by a few big countries.

Scientists can't yet draw a straight line from emissions to a specific bout of extreme weather, but for more than a decade they've been able to gauge the extent to which greenhouse warming hiked the odds of a particular event, such as the 2003 European heat wave (see later) or China's extreme rainfall of 2016.

It's not too much of a stretch to imagine courts in the future using such findings, along with the tally of who emits what and where, in order to assign partial responsibility for major climate shocks.

How much greenhouse gas is in the air right now?

The global atmosphere currently carries about **3200 metric gigatons** of carbon dioxide, which includes about 870 metric gigatons of carbon.

That's by far the largest presence of any human-produced greenhouse gas.

The atmosphere also holds about four metric gigatons of carbon in the form of methane, which is a much stronger but shorter-lived greenhouse gas.

A metric gigaton is a billion metric tons, or about 1.1 billion short tons, the measure used in the United States.

How much are we adding each year?

As of 2016, humans were directly putting just over 35 metric gigatons of carbon dioxide into the atmosphere per year—a startling increase from around 15 metric gigatons in 1970 and about 26 metric gigatons as recently as 2002.

This is mainly from burning fossil fuels, though about 3% of the total comes from **cement production**.

Almost half of this total yearly output remains in the atmosphere, while the rest is absorbed (more on this later).

Deforestation adds still more carbon dioxide to the air— anywhere from 6% to 20% beyond the totals above, depending on the acreage and density of the forests felled in a given year.

The global **emission rate** of CO_2 doesn't go up every year.

During economic slowdowns, the rate can actually decline slightly, as it did in 1992 and again in 2009. CO_2 emissions held virtually steady from 2014 to 2016 despite an economy growing at more than 3% per year—an encouraging sign that prosperity is no longer wedded to fossil fuel use.

However, even if the emissions rate were to hold constant for years on end, we'd still be adding lots of carbon dioxide to the atmosphere.

The **total amount** of CO_2 in the air—as seen in the earlier graph has risen every year since measurements began in the late 1950s.

To put these figures in perspective, consider **per-person emissions.**

On the basis of the figures above, each human on Earth is responsible for putting an average of 4.7 metric tons of carbon dioxide into the air each year.

That's 4700 kg, or about 10,400 lb.

In other words, someone who weighs 68 kg (150 lb) puts the equivalent of their weight in carbon dioxide into the atmosphere in less than a week.

Of course, the real figures vary enormously on the basis of where and how you live.

For most people in Europe and North America, the average is much higher, as shown in the chart later.

What happens to the greenhouse gas we put into the air?

It's easy to assess how much CO_2 is burned each year, at least in principle.

And because CO_2 is mixed throughout Earth's atmosphere over its long lifetime, it's also fairly straightforward to measure the global amount of CO_2 in the air by taking measurements at a few well-chosen observation points.

From these two factors, we know that in a typical year, about 45% of the carbon added to the atmosphere each year from human activity stays there, boosting airborne CO_2 for a century or more.

Individual molecules cycle through the system more quickly than that, however.

Imagine water in a park fountain that's continually going down the drain and returning through the fountain pump.

If you increase the inflowing water without changing the outgoing drainage, the water level in the fountain's pool will slowly rise, even though large amounts of water are coming and going all the time.

The other 55% of human-produced emissions is either absorbed by the ocean or taken up by land-based ecosystems—trees, crops, soils, and the like.

Plants take up carbon dioxide when they photosynthesize and return it to the soil and the atmosphere when they die and decompose.

Plants and soil-based microbes also add some CO_2 to the atmosphere through respiration, just as people do when they breathe.

Of the carbon that a typical tree takes up, about half is quickly returned to the air by respiration; the rest goes into wood or leaves that store the carbon until they decompose.

This simplified picture gets a bit more complex when you consider that Earth's blanket of vegetation goes through changes from year to year and decade to decade.

El Niño and other atmospheric cycles can spawn drought over huge areas.

The warmer, drier conditions make for struggling plants that respire more CO_2 and absorb less of it.

These conditions also allow for more deforestation and large-scale fires that add CO_2 to the atmosphere.

Recent work hints that the absorption of CO_2 by plants and oceans could decrease in the long term as the atmospheric load increases.

It's very hard to measure and separate out all these effects; for one thing, what we add is folded into a much larger exchange of carbon that takes place naturally.

Over the last few years, the balance has played out as follows, as calculated by the Global Carbon Project:

Land-based ecosystems took up around 30% of the CO_2 emitted by fossil fuel use over the period 2007–2016.

However, there are large variations from year to year due to climate cycles and changes in land use, and scientists aren't yet sure which regions account for most of the absorption (see below).

Also, this doesn't factor in the CO_2 returned to the air by deforestation and other changes in land use, an amount equal to about 10% of fossil fuel emissions.

Oceans absorbed roughly 24% of CO_2 emitted by fossil fuels.

The atmosphere retained the rest of the emitted CO_2.

On average, that's about 46%, though in a given year the percentage can range anywhere from around 25% to 100% depending on how much CO_2 is taken up by plants and soils that year.

Because most of Earth's plant life is north of the equator, and because land-based uptake of CO_2 is more seasonally driven than oceanic uptake, the northern spring produces a "breathing in" effect (a dip in CO_2 levels), with a "breathing out" (a rise in CO_2) each northern autumn.

These ups and downs show up as the sawtooths atop the steady multiyear rise in CO_2 visible on the earlier graph.

Comparing Greenhouse gases

Because greenhouse gases vary so widely in the power of the their climatic effects, one often relies on a unit called **global warming potential**.

The global warming potential of a gas is a measure of its contribution per unit mass to greenhouse warming in the atmosphere over a given time span as compared to that of carbon dioxide.

Methane, for example, is shorter-lived than CO_2 but much more powerful in its ability to trap heat in the atmosphere.

Thus, over a century's time, methane's global warming potential is estimated to be somewhere between 28 and 35, compared to a value of 1 for carbon dioxide.

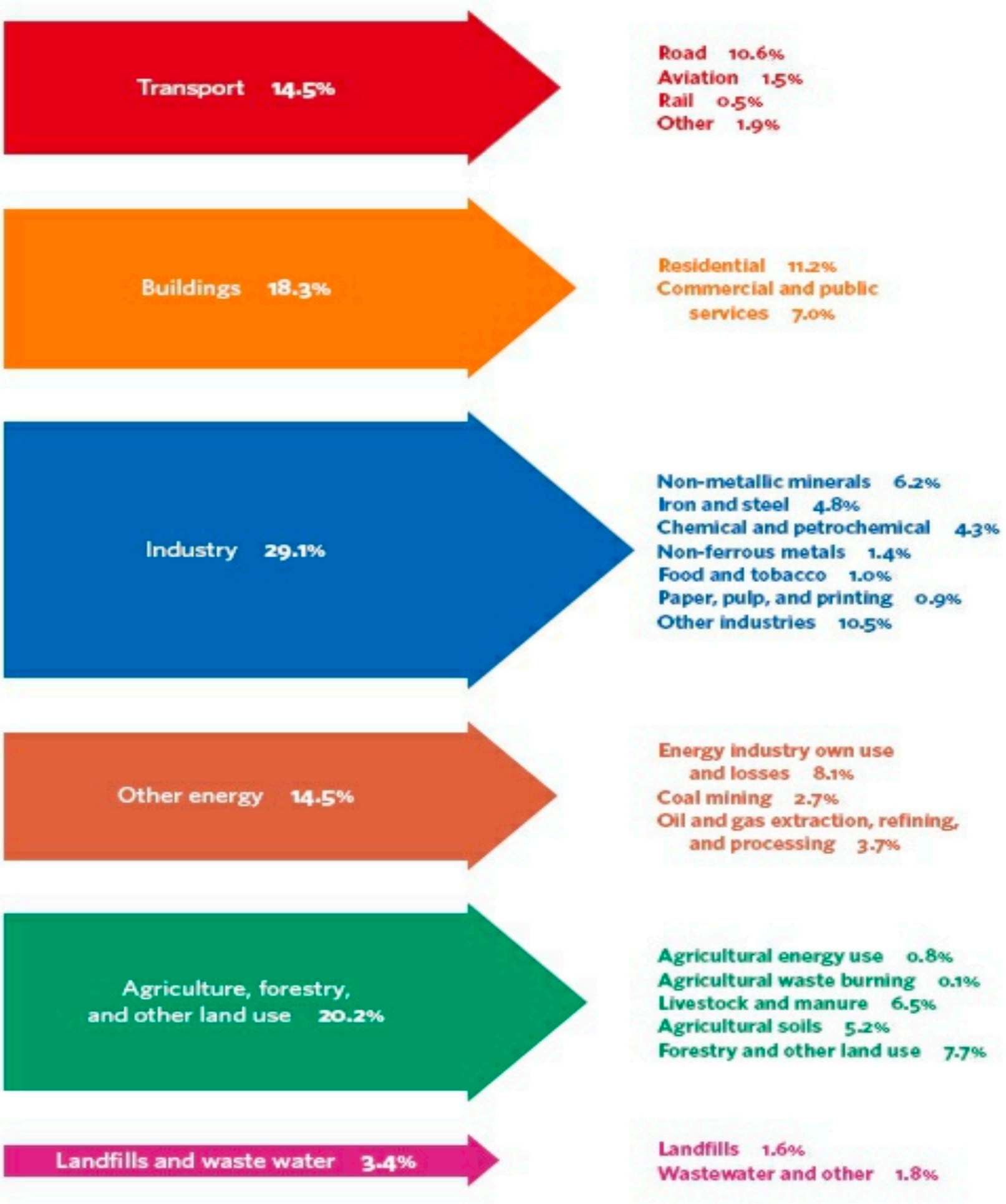
These figures can be multiplied by the prevalence of each gas to produce a **carbon equivalent** that enables all emissions to be considered as a group.

By using carbon equivalents, one can get a better sense of the impact of the atmosphere's total greenhouse burden.

How much do different activities contribute?

The chart shows two different ways to analyze the sources of human-produced greenhouse gas: where the emissions come from (left) and the activities that generate them (right).

As our emissions climb at anywhere from 0.5% to 2% per year, the balance among various sources is slowly shifting.



Breakdown of total global greenhouse gas emissions (CO_2 equivalent) by sector (left) and end use (right) as of 2012. Because of rounding, the numbers in the right-hand column do not total 100%.

By and large, the world's most technologically advanced nations have become more energy efficient in recent years, with **industrial emissions** actually declining a few percent in some developed countries since 1990.

Counterbalancing this is the explosion of industry in the world's up-and-coming economies.

Another area of progress is deforestation, which accounted for more than 25% of emissions in the 1990s but closer to 10%–15% in recent years.

The emissions produced by heating, cooling, and powering **homes** have been increasing by more than 1% per year.

Even though energy use per square meter has been dropping, today's homes have gotten bigger, and they're stuffed with more energy-hungry devices.

Consumer electronics, including computers, now make up as much as 15% of global household electricity use, according to the International Energy Agency, although smarter devices are starting to chip away at that total.

Meanwhile, **commercial buildings** are tending to become more efficient more quickly than homes, with companies motivated by the potential for long-term savings.

Emissions in the **transportation** sector are also climbing at more than 1% per year, stoked by the global growth in motor vehicle use and the ever-longer distances many people are driving and flying.

Transport-related emissions are growing most quickly in the developing world, as millions of people in newly prospering nations take to the roads for the first time ever in their own cars.

Which countries are most responsible?

Establishing which countries are most responsible for climate change is in some ways an arbitrary exercise, since the atmosphere isn't particularly concerned about national boundaries.

In any case, it's more complicated than simply adding up the amount of fossil fuels that each nation burns each year.

First, there's the fact that a populous country's total greenhouse gas emissions may be relatively large even if its per capita emissions are relatively small, as is the case with China.

Consider this: if China were suddenly divided into four nations, each with roughly the population of the United States, then U.S. emissions would far outstrip those from each of the four new countries.

Then there's the issue of greenhouse gases emitted in the past (some of them before global warming was a known problem), many of which still remain in the atmosphere.

The question of how countries are using—or abusing—their landscapes also needs to be considered.

Countries undergoing major deforestation, such as Brazil and Indonesia, would rank significantly higher in the list shown later of nations with the highest total emissions if the greenhouse-boosting effects of forest destruction were taken into account.

To add another level of complexity, there's the claim that emissions figures are currently rigged in favor of nations that tend to import, rather than export, goods.

When it comes to overall greenhouse gas output, however, two countries stand above the rest.

The United States is responsible for around 25% of the **cumulative** CO_2 emissions to date (that is, all emissions since the industrial era began) and continues to generate almost 15% of annual greenhouse gas emissions—more than three times its share, considering that the United States represents less than 5% of the world's population.

Some of the nation's outsized emission rate is clearly due to a lack of emphasis on energy efficiency and a focus on economic growth as opposed to environmental virtue.

There are also historical factors in the mix that are difficult to change: a car-loving and car-dependent culture, an economy built on vast reserves of fossil fuels, and the simple fact that the United States is large in both population and land area.

The country now has more than 60 years' worth of suburban development that virtually forces millions of Americans to drive to work, school, and just about anywhere else.

Despite all these factors, and to the surprise of many, U.S. emissions actually dropped by roughly 14% from their 2005 peak up to 2016.

Among the factors at play: the major recession from 2008 onward, increased use of mass transit, higher auto efficiency standards, a boom in solar and wind energy, and—perhaps most influential—a major shift from coal to natural gas, which produced about 30% of U.S. electricity in 2017 compared to 19% in 2005.

Quickly outpacing the United States in terms of current emissions is China.

With its billion-plus citizens, the country's emissions per person are still comparatively low.

But as the nation began to industrialize in earnest, its total emissions climbed at a staggering pace - close to 10% annually for the decade 2002–2011.

At the turn of this century, it was thought that China might surpass the United States in total emissions by the year 2025.

That point was actually reached in 2006, according to the Netherlands Environmental Assessment Agency.

A burst of growth in renewables, together with a shift away from coal, helped China to rein in its emissions growth in the mid-2010s: the year 2016 actually saw a net drop in China's emissions over 2015 (albeit just 0.7%).

Percentage of global CO_2 emissions (fossil fuels, cement, and gas flaring only)		Emissions per capita (metric tons of CO_2 from fossil fuel use)	
China	28.5	Qatar	49.6
United States	14.5	Curaçao	37.8
India	6.2	Trinidad and Tobago	34.2
Russia	4.7	Kuwait	25.4
Japan	3.4	United Arab Emirates	23.2
Germany	2.0	Bahrain	23.0
Iran	1.8	Brunei	21.8
Saudi Arabia	1.7	Sint Maarten (Dutch)	19.5
South Korea	1.6	Saudi Arabia	19.5
Canada	1.5	Falkland Islands	18.9
Brazil	1.5	Luxembourg	17.3
South Africa	1.4	New Caledonia	16.5
Mexico	1.3	Gibraltar	16.5
Indonesia	1.3	United States	16.2
United Kingdom	1.2	Australia	15.3
Australia	1.0	Canada	15.1
Turkey	0.9	Estonia	14.8
Italy	0.9	Oman	14.5
Thailand	0.9	Kazakhstan	14.3
France	0.8	Bonaire/St. Eustatius/Saba	13.3

The **United Kingdom** has produced about 5% of the world's cumulative emissions, thanks to its head start as the birthplace of the Industrial Revolution.

In the course of a major industrial transformation, the United Kingdom has cleaned up its act significantly.

It now generates less than 2% of the yearly global total of greenhouse gases.

Still, that's about twice its rightful share, considering that the United Kingdom houses less than 1% of Earth's population.

In the list above of the world's top 20 greenhouse emitters, measured in two different ways: percentage of all global emissions and emissions per capita.

For total emissions, the list is a mixed bag, with the contributions slightly weighed toward developing versus developed nations.

The **emissions per capita** column tells a different story, with the list topped by the tiny oil- producing nations of **Qatar, Curaçao, Trinidad and Tobago, Kuwait**, and the **United Arab Emirates.**

These countries have so few residents that their contributions to the global greenhouse effect remain small, but because they are heavy producers and consumers of oil and gas, they have a high per capita emissions rate.

Otherwise, industrialized nations lead the way, with the sprawling, car-dependent trio of the **United States, Australia**, and **Canada** not far behind the oil producers.

Part 2 - The Symptoms

WHAT'S HAPPENING NOW, AND WHAT MIGHT HAPPEN IN THE FUTURE

4 - Extreme Heat
TOO HOT TO HANDLE

To those who think of Russia as a perpetual icebox, the summer of 2010 came as a sizzling shock.

The nation roasted in unprecedented heat from July into early August.

Moscow, whose summer afternoons are typically close to room temperature, soared above 38°C (100°F for the first time on record and hit 30°C (86°F every day for more than four weeks.

Smoke from peat and forest fires blocked out skyscrapers and clogged lungs.

Even before the heat wave subsided, it was clear that thousands of lives had been lost, and Russian leaders—long skeptical on global warming—openly expressed concern about the nation's climatic future.

The horror in Russia triggered memories of the super-hot European summer of 2003—a relentless barrage of heat that produced an even greater slow-motion catastrophe, killing more than 50,000 people.

In at least one study, a group of climate scientists confirmed the suspicions held by many: the odds of getting the great heat wave of 2003 were boosted severalfold by the helping hand of fossil fuels.

Heat waves themselves are nothing new, of course.

As of 2018, many central U.S. states had yet to top the all-time records they set during the 1930s Dust Bowl, including a reading of 49.5°C (121°F) as far north as North Dakota.

In this context, an all-time record is the warmest reading observed at a given location on any day of any year.)

Other measures of increasing heat have been pushing the envelope, though.

In the contiguous United States, the summer of 2018 brought minimum daily temperatures (i.e., overnight lows) that averaged higher than in any other summer since records began in 1895.

In 2016, many cities, including Atlanta and Nashville, stayed above 70°F (21°C) for their longest periods on record.

Climate-modeling studies and our understanding of the greenhouse effect both indicate that the next few decades could bring hot spells that topple many long-standing records across midlatitude locations.

In 2016 alone, 22 nations—an unprecedented number, most of them in Asia and Africa—set all-time temperature peaks (their highest single readings on record).

The roll call included 49°C (120.2°F) in Niger, 51°C (123.8°F) in India, and a blistering 53.9°C (129.0°F) in Iraq.

More nations set all-time highs in 2016 than set all-time lows in the entire preceding decade.

This has gotten even worse during the period up to 2024!

Striking as these statistics are, a heat wave doesn't have to bring the warmest temperatures ever observed to have catastrophic effects.

All you need is a long string of hot days combined with unusually warm nights.

Stir in a complacent government, cities that were built for cooler times, the lung-clogging effects of air pollution, and a population that can't or won't respond to the urgency, and you have a recipe for the kind of disaster that 2003 brought to Europe and 2010 delivered to Russia.

Off the charts

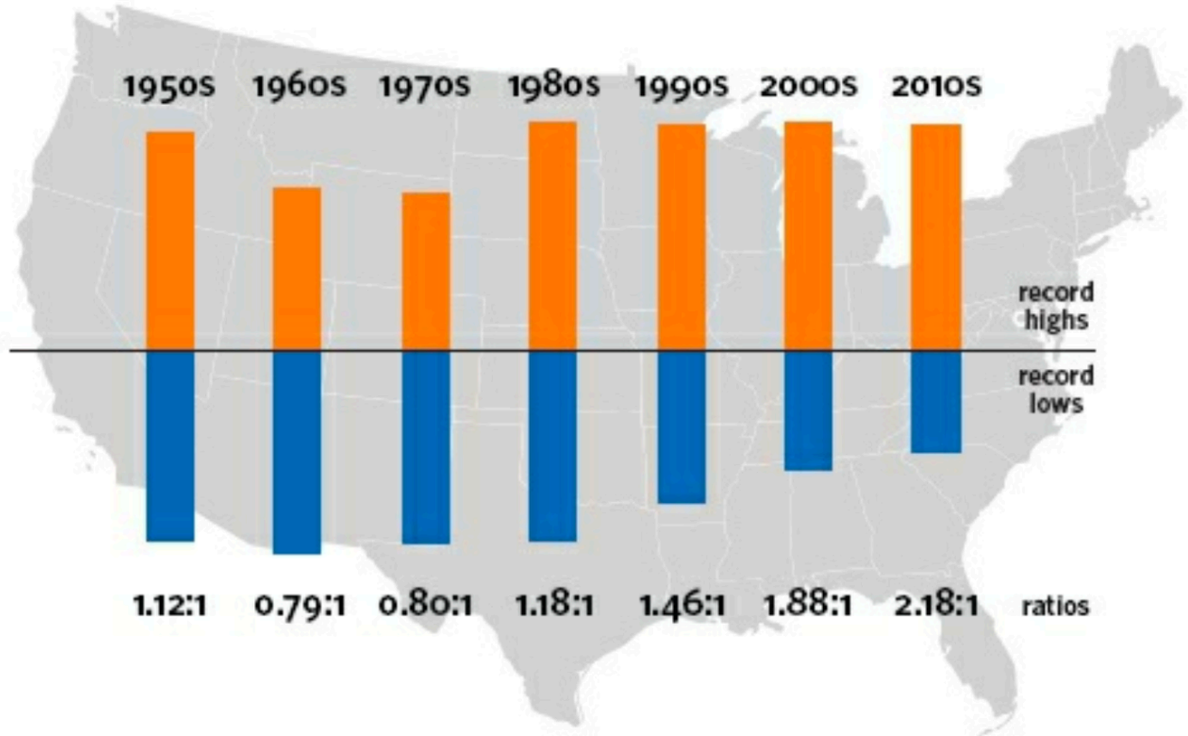
Extremes in summertime heat are one of the long-anticipated outgrowths of a warming planet.

Part of this is due to simple mathematics.

If the average temperature goes up, then the most intense spikes in a hot summer ought to climb as well.

Computer models and recent data bear out this intuitive concept.

In the contiguous United States, daily record highs have been outnumbering daily record lows by an increasing margin in each decade from the 1980s onward (see graphic).



The ratios are even more skewed for monthly and all-time records.

Looking ahead on the global scale, the 2013 IPCC report considered it virtually certain that most of Earth's land areas will see more frequent warm extremes and less frequent cold ones, on both a daily and seasonal basis.

Of course, it'll still be possible to have a record-setting blast of cold every now and then, just as a baseball player on steroids will occasionally strike out.

Over time, though, warm records should increasingly outpace the cold ones.

Could heat spikes themselves become more intense, relative to a rising baseline?

That's a different and more complex question.

Any rise in average summer temperature—the base on which any heat wave builds—is itself enough to cause plenty of trouble.

But if heat waves also gained amplitude, thus straying even further from a rising average, then the resulting double whammy would have huge implications for people and ecosystems.

Only in the last few years have scientists looked more closely at temperature variability and whether it might change as the planet warms.

Many of the model-based projections of greater heat extremes can be explained simply by a rising baseline temperature.

But heat waves could intensify more quickly than average temperatures in places where the landscape sometimes dries out, a process that allows more heat to go into warming the atmosphere instead of evaporating water.

In the long run, hot is hot, whether it manifests because of intensified spikes, a rising baseline, or both.

While we may want to sort out the ‘variability issue’ for scientific reasons, it does not alter the fact that the future will be bleak with regard to heat extremes.

However high it might get, the peak temperature on any given day isn’t the ideal measure of a heat wave’s ferocity.

When it comes to the toll on people, animals, and plants, the **duration** of a heat wave and the warmth of the **nighttime lows** are the real killers.

A weeklong stretch of severe heat can be far more deadly than just two or three days of it.

If a heat wave is defined at a certain location by a certain number of days with readings above, say, 35°C (95°F), then a warming climate would be expected to bump more days above such a threshold over time.

There’s a different connection between global warming and steamy nights.

As will be shown in later discussions, higher temperatures have boosted **atmospheric water vapor** on a global scale.

When there's more water vapor in the air, nights tend to stay warmer.

Moreover, cool nights act as a critical safety valve, giving people a physiological and psychological break from the intense daytime stress of a heat wave.

If the nights warm up too much, that safety valve goes away.

When a fever hits

Europe's 2003 heat onslaught is a textbook example of how the variables above can add up to disaster.

During the first two weeks of August, large swaths of the continent topped 35°C (95°F) and found themselves hard-pressed to cool below 20°C (68°F).

Such temperatures might not sound like a big deal to an American from the Sunbelt, but most of Europe's urban areas lack the ubiquitous air conditioning of the hottest U.S. cities, and certainly most Europeans aren't accustomed to dealing with weather like this.

Among the history-making statistics from 2003:

(1) August 10 was the hottest day on record in **London**, with highs of 37.9°C (100.2°F) at **Heathrow Airport** and 38.1°C (100.6°F) at the Royal Botanic Gardens.

Not to be outdone, Faversham, Kent, climbed to 38.5°C (101.3°F) that same afternoon.

It was the first day in nearly 300 years of record keeping that any place in Britain had topped the 100°F mark.

(2) All-time national records were also broken with 40.4°C (104.7°F) in **Roth, Germany;** 41.5°C (106.7°F) in **Grono, Switzerland;** and a scorching 47.3°C (117.1°F) in **Amareleja, Portugal.**

(3) Germany notched its warmest night on record, with the weather station at **Weinbiet**, a mountain near the Rhine, recording nothing lower than 27.6°C (81.7°F) on August 13.

Numbers can't do full justice to the impact of this atmospheric broadside on society.

In **England**, railway tracks buckled, tube stops resembled ovens, and the London Eye and its glassed-in Ferris wheel pods had to shut down.

Schools and offices closed for “heat days” in parts of Germany.

Perhaps no place suffered so acutely as **Paris.**

With little air conditioning and block after block of tall, closely packed stone buildings, the French capital became a heat-trapping machine.

The city ended up falling a whisker short of its all-time high of 40.4°C (104.7°F), which was set in 1947, but it did endure nine consecutive days that topped 35°C (95°F).

Even worse for many were the nights of surreal heat. On August 12, the temperature never dropped below 25.5°C (77.9°F), making it Paris’s warmest night on record.

The early-August heat was so sustained that it kept the temperature within the intensive care unit of one suburban Paris hospital, which lacked air conditioning, at or above a stifling 29°C (84°F) for more than a week.

Russia’s heat wave of 2010 brought similar misery.

A vast dome of high pressure encased Moscow and surrounding areas from mid-July to early August, setting the stage for a vicious brew of heat, smog, and smoke from nearby fires.

During the worst bouts of pollution, visibility dropped to a few city blocks.

On these days, Muscovites faced the no-win choice of opening windows for a smoky breeze or keeping their homes sealed and sweltering.

Wind shifts pushed the smoke away from Moscow at times, but the heat was virtually constant, with high temperatures running more than 10°C (18°F) above average for weeks on end.

Alexander Frolov, the head of Russia's weather service, dubbed the heat wave the worst in the nation's 1000-year history.

Relief wasn't easy to come by, as smoke and heat also plagued the thousands of summer homes (dachas) where Muscovites traditionally flee from urban swelter.

Many people took to the beach after drinking, which—combined with the disorientation that extreme heat can produce— helped lead to more than 1200 drownings in June alone.

The heat also decimated wheat crops across a huge swath of Russia's breadbasket, which sent world grain prices soaring.

All told, the months of July and August brought 56,000 more fatalities across central Russia in 2010 than in the year before, with most of those deaths presumably related to the heat and pollution.



Many Muscovites wore masks in August 2010 when a heat wave—accompanied by smog and smoke from nearby wildfires—gripped the Russian capital. The city's mortality rate doubled as the concentration of atmospheric carbon monoxide soared to more than double acceptable safety norms.

The disaster galvanized Russian leaders, who'd long viewed global warming science with suspicion.

Early in 2010, President Dmitry Medvedev had dubbed climate change "some kind of tricky campaign made up by some commercial structures to promote their business projects."

Only a few months later, the heat had produced a remarkable about-face. In late July, Medvedev said, "What's happening with the planet's climate right now needs to be a wake-up call to all of us."

In 2012, it was North America's turn to experience unprecedented heat on a grand scale.

The first salvo was pleasant rather than painful: an astoundingly warm March that sent temperatures rocketing to summerlike levels—in many cases, soaring above 27°C (81°F) for days on end— across much of the midwestern and northeastern United States and southeastern Canada.

At least three U.S. locations experienced evening lows that broke daily record *highs*.

In Canada, the city of Saint John, New Brunswick, saw temperatures as high as 27.2°C (80.1°F), balmier than anything it had previously observed in March *or* April.

Perhaps because it was so tempting to simply enjoy the warmth, the event got little serious scrutiny at the time, but it was actually one of the most extreme temperature departures in North American history.

Millions of dollars in Midwestern fruit crops were lost when premature blossoms froze in April.

Things got considerably worse that summer, as drought and heat punished much of the central and eastern United States.

July 2012 ended up as the hottest month ever observed in cities from Denver to St. Louis to Indianapolis, as well as for the United States as a whole.

That entire year was also the nation's warmest, by a full 0.6°C (1.0°F)—a huge margin for such a record—and the summer of 2012 was the nation's second hottest, behind only the Dust Bowl summer of 1936.

The massive 2012 drought became one of the nation's costliest natural disasters on record, with crop losses totaling tens of billions of dollars.

Huge wildfires also devastated large parts of the U.S. West.

Surprisingly, and fortunately, less than 100 deaths were initially confirmed from the summer heat of 2012.

Though air conditioning is far more common in the States than in Europe, the low death toll also suggests that heat wave safety practices are gaining ground.

Pollution: Heat's hidden partner in crime

Many victims of heat waves die not because the air is so warm, but because it's so dirty.

The sunny, stagnant conditions prevalent during heat waves makes an ideal platform for sun-driven processes that create ozone, which is a blessing in the stratosphere as we discussed earlier, but a dangerous pollutant at ground level.

Ozone irritates the lungs and makes them more vulnerable to other chemicals.

Moreover, the relative calm of a heat wave allows tiny bits of heavy metals, sulfates, nitrates, dust and other substances to accumulate in the air.

The smallest of these particulates can sneak past the body's natural respiratory filters, causing a variety of lung problems and raising he risk of heart attacks.

The human cost of heat

It can take a surprisingly long time to gauge how many people die in a heat wave, but the broad strokes of the disaster often show up quickly.

Moscow's death rate in July 2010 doubled that of the preceding July, and the number of funerals soared.

At the height of the heat, one Moscow crematorium operated around the clock for days.

In a gruesome but telling bit of irony, its furnaces overheated.

Likewise, it was clear in early August 2003 that people were dying in large numbers across Europe, but it took several weeks to start getting a handle on the heat's human cost.

Simply adding the various national counts would imply that more than 50,000 people died as a result of the heat wave.

Later estimates brought the toll as high as 70,000, though the exact number is difficult to discern (see below).

No heat wave in global history has produced so many documented deaths.

How could heat victims go untallied for a year or more?

The answer lies partly in the way mortality statistics are compiled and partly in the way heat waves kill.

Many people die indirectly in heat waves—from **pollution**, for instance, or from preexisting conditions that are exacerbated by the heat—so it's not always apparent at first that weather is the culprit.

And unlike a tornado or hurricane, a hot spell doesn't leave a trail of photogenic carnage in its wake.

People tend to die alone, often in urban areas behind locked doors and closed windows.

The piecemeal nature of heat deaths over days or weeks makes it hard to grasp the scope of an unfolding disaster until the morgue suddenly fills up.

Furthermore, it often takes months or even years for countries to collect and finalize data from the most far-flung rural areas.

On top of this, heat is a selective killer.

It targets the very young and especially the very old, people whose metabolisms can't adjust readily to temperature extremes.

In Rome, more than half of the 1000-plus deaths attributed to 2003's heat were among people older than 85.

Because the elderly die so disproportionately in heat waves, it's tempting to assume that weather is simply claiming those fated to die shortly anyhow.

If that were the case, however, then you'd expect mortality to dip below average in the months following a heat wave.

Careful studies of major heat disasters have, in fact, shown that such dips typically account for only about 20%–30% of the spikes above average, or what epidemiologists call the **excess deaths,** that were observed in the preceding heat.

Thus, it seems that most victims of heat waves, even the elderly, die well before their time.

In France, the situation was exacerbated by local culture.

The worst of the heat struck during the August vacations, when much of professional Paris heads out of town.

Left behind to fend for themselves were many thousands of elderly Parisians, the ones most susceptible to heat-related health problems and least able to seek refuge in a cool location.

As many as half of France's fatalities occurred at rest homes, few of which had air conditioning.

Improved **warning systems** for heat have been saving lives in the United States, where many areas can easily top 35°C (95°F) in a typical summer.

After years of perfunctory heat advisories, issued by the U.S. National Weather Service (NWS) but seldom heeded, the country was shocked into awareness by a 1995 heat wave in Chicago that killed more than 700 people.

That same year, Laurence Kalkstein of the University of Delaware teamed up with the NWS to launch a new watch–warning system for extreme heat.

The scheme also included intensive **neighborhood interventions**, such as block captains assigned to check on neighbors.

It's been estimated that Philadelphia's heat plan saved more than 100 lives in four years.

A 2014 study in *Environmental Health Perspectives* found that the number of heat-related U.S. deaths for every 10°F (6°C) increase in temperature dropped by more than 60% from 1987 to 2005.

U.S. heat can still be deadly, though, especially when it strikes where people are not accustomed to it.

Hundreds of Californians died in an intense 2006 heat wave, more than half of them in homes where temperatures reached insufferable levels.

Attributing the heat

How much of today's heat can be pinned on greenhouse gases?

The European disaster of 2003 brought that question to the fore.

The ever-growing heat islands of modern cities no doubt played a role, and these are a form of climate change in themselves.

Yet, the continent-wide, rural–urban scope of the heat pointed to something more going on.

Many activists trumpeted the heat wave as a classic example of global warming at work, but it was a year later before scientists had marshaled a statistical case for this accusation.

The resulting study was a landmark—one of the first to lay odds on whether a specific weather or climate event was rendered more likely to occur by global warming.

Using a climate model from the Met Office Hadley Centre, the research team simulated European summers since 1900 a number of times, sometimes including the century's ramp-up in greenhouse gases and sometimes omitting it.

They found that human-induced greenhouse emissions made it roughly four times more likely that a summer like 2003 might occur.

Such efforts—often dubbed attribution studies—are a crucial way that scientists and policymakers get perspective on the effect of climate change on extreme weather events.

Temperature projections based on different scenarios, are called Representative Concentration Pathways (RCPs).

However, the results can be open to varying interpretations.

For example, National Oceanic and Atmospheric Administration (NOAA) researchers asked whether the Russian heat wave of 2010 could have been foreseen on the basis of twentieth-century temperature trends.

They concluded that there was enough variability inherent in the region's July climate to allow for such an intense event to occur regardless of climate change.

However, using a different technique, researchers at the Potsdam Institute for Climate Impact Research found that the odds of setting a July heat record in Moscow have been boosted fivefold by longer-term warming.

Such studies may not be as contradictory as they seem.

Even where a record-smashing heat wave can occur without a boost from climate change, long-term warming may still make a particular record value far more likely to be achieved.

Out of 16 heat-related published attribution studies in 2016 and 2017, all but one found that the heat event in question was made more likely through the presence of human-produced greenhouse gases.

In one case, a team led by researchers at Japan Meteorological Agency went even further: they concluded that Asia's extreme warmth of 2016 "would not have been possible without climate change."

The future of summer sizzle

On a broader scale, the European disaster of 2003 and Russia's 2010 meltdown were merely blips—albeit spectacular ones—in a long-term warming trend that spans most of the world, particularly large parts of North America, Eurasia, and Australia.

While the planet as a whole has warmed around 1.0°C (1.8°F) since 1900, the rise has generally been greater over the Northern Hemisphere continents (especially at higher latitudes) than over the tropics.

That simplified picture doesn't hold quite as well, though, when you go beyond the average temperature and look at various types of heat extremes, some of which have increased in the tropics as well as midlatitudes.

It’s taken a long while for scientists to verify how temperature extremes have changed globally over the last few decades, not to mention how they might evolve in the future.

Why is this so?

After all, the global average has been tracked for decades, and individual stations keep close tabs on daily weather.

Between the local and the global, though, it’s surprisingly difficult to assess regional trends in temperature and precipitation extremes.

Nations have traditionally shared more general data, such as the average high and low temperatures for each month at major reporting sites.

But knowing that the average August high in a given city rose, say, 1°C in 50 years doesn’t tell you whether stretches of the most intense heat are becoming more frequent, intense, or longer lived.

To find that out, you need a consistent way to measure heat extremes, and you need day-by-day data extending over decades.

That’s been hard for researchers to obtain.

Some nations balk at releasing such data, considering it valuable intellectual property.

Others aren't set up to provide the data easily.

Climate scientists made big headway on this problem in the 2000s through a series of regional workshops.

These meetings, accompanied by the creation of an Expert Team on Climate Change Detection and Indices, provided a way in which countries could freely share some parts of their daily-scale data for research without jeopardizing the revenue they earn by selling those data.

The expert team also forged a set of 27 core indices of heat, cold, and precipitation that were used in preparing the 2007 and 2013 IPCC reports.

Emerging from this and subsequent work is a more detailed picture of how temperature-related extremes are evolving across the world and what we can expect going forward.

Drawing on these core indices, a 2013 analysis of climate trends from 1901 to 2010, published in the *Journal of Geophysical Research: Atmospheres* and led by Markus Donat, found the following:

- Stations around the globe are seeing an average of roughly 16 more “warm days” per year than they did in 1900. (Warm days are defined as those when the high temperature falls in the top 10% of the spectrum for a given calendar date; you’d expect 36 such days in a typical year.) Cool days are down by about 14. Changes after dark are more dramatic: warm nights have surged by about 30 per year, and cool nights are down by about 26.
- Averaged across the planet, the hottest day recorded at each spot each year rose on the order of 1°C (1.8°F) since 1900, while the coldest night jumped by a full 6°C (10.8°F). Russia, in particular, is seeing nighttime winter temperatures often holding well above where they used to bottom out.

Not all geographic areas have seen temperatures climb.

One region where the twentieth century’s overall warming barely made a dent is the southeastern United States.

Donat and colleagues found a small but significant decline in the number of warm days across the U.S. Southeast since 1950, although warm nights have increased there.

Parts of this “warming hole” saw average temperatures drop by as much as 1°C (1.8°F) between 1930 and 1990.

Why the hole?

It's still a topic of active research.

Some scientists have found possible links to sea surface temperatures in the Atlantic and Pacific, while a study led by Harvard University points to sunlight-blocking particulate pollution across the region, which grew during the postwar years but is now on the wane.

The climate modeling featured in the latest IPCC assessment shows many of the major observed twentieth-century trends increasing over the next few decades.

How much they accelerate depends largely on how much and how soon greenhouse emissions can be controlled. The Canadian Centre for Climate Modelling and Analysis led a 2013 paper in the *Journal of Geophysical Research: Atmospheres* detailing the globally averaged picture painted by these models for the period 2081–2100, as compared to the period 1981–2000.

Some results include the following:

- The total percentage of nights that fall in the "warm nights" category, as defined above, ranges from 30% if greenhouse gas concentrations begin to drop by midcentury (the pathway dubbed RCP2.6) to almost 70% if emissions rise unchecked (RCP8.5). Warm days will range from around 25% to 60%, depending on RCP. In contrast, only about 1%–4% of all days would qualify as cool days. Cool nights would virtually disappear under RCP8.5 and would drop to less than 4% even for RCP2.6.

- The hottest day of each year will warm by about 1.5°C (2.7°F) for RCP2.6 and by more than 5°C (9°F) for RCP8.5. The coldest night will be about 2°C (3.8°F) warmer for RCP2.6 and more than 6°C (10.8°F) warmer for RCP8.5.

Again, as one might expect, there is liable to be a good deal of variety in how and where these temperature changes play out.

Researchers found that large parts of Brazil and Sub-Saharan Africa could see as many as 100 or more additional “tropical nights” per year, when temperatures fail to drop below 20°C (68°F).

Major boosts in the number of such balmy nights may also occur as far poleward as the southeastern United States and southern Europe.

The number of nights that dip below freezing should see a drop worldwide, but the decrease could be as much as 60 nights or more per year in a coastal belt stretching from the Pacific Northwest states to Alaska.

As for heat waves themselves, the 2013 IPCC report concluded that by the 2080s and 2090s, it’s very likely that heat waves will occur with a higher frequency and duration, with somewhat more uncertainty on how much their intensity might be boosted.

Although there wasn't a specific outlook provided for earlier in the century, the summary did note that "models project near-term increases in the duration, intensity, and spatial extent of heat waves and warm spells."

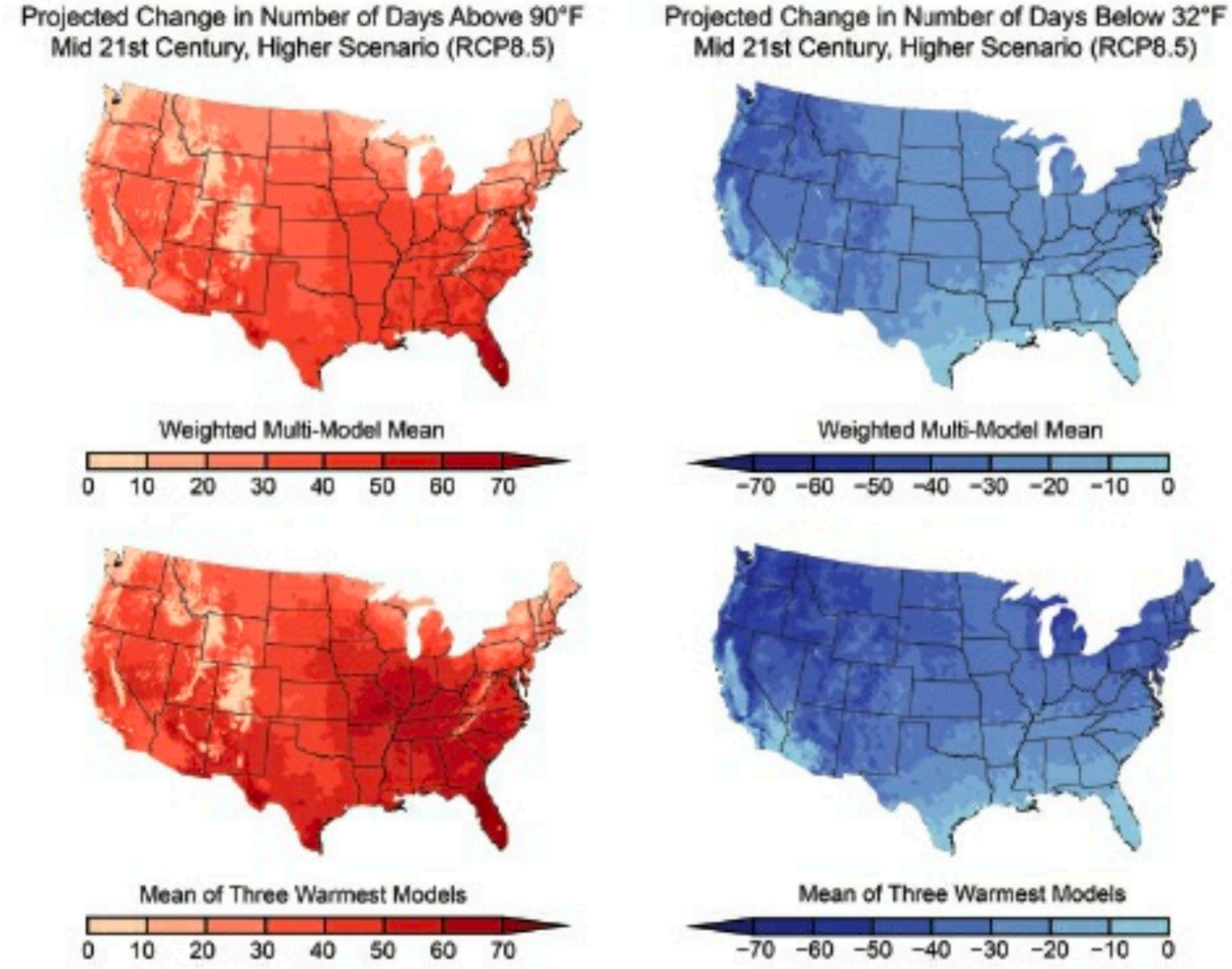
More recently, scientists have zeroed in on how heat extremes might intensify even faster than average temperature.

One big factor is soil moisture.

Dry soils allow more incoming sunlight to go into warming the soil and the atmosphere above it, which can reinforce local and regional heat waves.

When they sifted through the 2013 IPCC modeling for a 2017 paper they found that the single hottest day of each year could warm faster than the annual average over large sections of Europe, North and South America, and southeast China.

On a monthly scale, it's the tropics of South America and Africa— including many countries ill equipped to adapt to weather and climate shocks—that will see the biggest increase in heat variability, according to a 2018 study from Wageningen University.



Maps from the 2017 Climate Science Special Report showing projected changes in the number of days that will soar above 90°F (left) or drop below 32°F (right) by the middle of this century (2036–2065, compared to 1976–2005), based on the average result from a variety of climate models using scenario RCP8.5. (U.S. National Climate Assessment/ NOAA National Climatic Data Center/Cooperative Institute for Climate and Satellites, North Carolina)

A wild card in the Atlantic

There's one character waiting in the wings that could change this picture, at least in theory: a slowdown or shutdown of the **Atlantic thermohaline circulation**, which helps keep Europe unusually warm for its latitude.

This is the scenario made famous in the 2004 film *The Day After Tomorrow*.

Although evidence shows that the thermohaline circulation has ground to a halt more than once in climate history, it's believed that this process takes at least a few years to play out, and sometimes many decades, rather than the few days portrayed in the movie.

In any case, a complete shutdown of the thermohaline circulation isn't expected any time soon.

However, the 2013 IPCC report deems a slowdown "very likely" this century.

According to the IPCC, that weakening could be 20%–30% if emissions begin dropping by midcentury (the RCP4.5 pathway) or 36%–44% if they continue rising through the century (RCP8.5).

A slowdown of this magnitude could largely erase long-term warming across the United Kingdom and diminish it across many other parts of Europe.

However, summers could still be warmer and more drought-prone across the United Kingdom and Europe than they are now.

Rather than fully counteracting greenhouse warming, then, a thermohaline slowdown would only make the picture even more complex.

We will have a full discussion of the thermohaline shutdown hypothesis later.

Handling the heat

In the long run, even if the world manages to cut greenhouse emissions over the next few years, Europe will clearly need to adapt to the risk of heat waves like the one it endured in 2003.

Better warning systems will help; more air conditioning, and the associated cost, seems inevitable.

The latter may put a dent in Europe's goal of reducing greenhouse emissions, though it's possible that some of the energy spent to cool the continent will be counterbalanced by a drop in the need for wintertime heating fuel.

However, nobody knows how many of the poorest and most vulnerable of Europeans will simply be left to suffer through future hot spells in un-air-conditioned misery.

Poverty is certainly a major co-factor in heat deaths across the **developing world.**

A blistering two-month-long heat wave in 2013 brought major suffering to hundreds of millions of people across eastern China, including rural residents as well as the urbanites of Shanghai.

That city beat its all-time temperature record, set in 1942, at least five times during the summer, finally topping out at 40.8°C (105.4°F) on August 7.

In South Asia, people are well accustomed to spells of intense heat during the late spring, just before the monsoon arrives, when temperatures can soar well above 40°C (104°F) across wide areas.

As is the case elsewhere, it's the extremes on top of that already scorching norm that cause the most suffering.

On May 26, 2010, Mohenju Daro, Pakistan, reported an astounding 53.5°C (128.3°F)—the highest temperature ever reliably observed in Asia up to that point.

Several premonsoonal heat waves in recent years have each killed more than 1000 people across India, many of them landless workers forced by circumstance to toil in the elements.

At least 2500 Indians perished as a result of the scorching premonsoonal heat wave of May 2015.

As for the United States, much of its Sunbelt, including the vast majority of homes and businesses in places like Atlanta and Dallas, is already equipped with air conditioning.

Even as temperatures soar further, these regions may prove fairly resilient—as long as the power stays on.

Older cities in the Midwest and Northeast appear to be more at risk of occasional heat crises, even though their average summer readings fall short of those in the South, where air conditioning is the accepted standard.

Indeed, U.S. fatalities appear to be more common in places where intense heat is only an occasional visitor.

Another factor is diabetes, a global epidemic hitting the United States especially hard.

Because diabetes inhibits sweating and destabilizes blood sugar levels, it could put millions of Americans at an increased risk of heat-related illness, according to a 2010 study by the Mayo Clinic and NOAA.

High temperatures may also be related to a mysterious new strain of chronic kidney disease striking young, healthy workers who toil in the heat of Central America and South Asia.

Lack of access to water and/or reliance on sugary drinks are some of the other possible risk factors being considered.

Such detective work foreshadows the kind of research into heat stress that a warming world will increasingly demand of us.

5 - Floods and Droughts - TWO SIDES OF A CATASTROPHIC COIN

Although it's natural to think of temperature first when we think of global warming, the impact of climate change on precipitation may be even more important in the long run for many places and many people.

Too much rain at once can cause disastrous floods, while too little can make an area unproductive or even uninhabitable.

Anyone who was caught in the epic 24-hour rainfall of 944 mm (37.2 in.) in Mumbai, India, in July 2005, or who experienced Britain's destructive summer rains in 2007, might wonder if the planet's waterworks are starting to go into overdrive.

That thought also must have crossed the minds of people in Pakistan during the summer of 2010, as torrential monsoon rains inundated an area the size of England.

At least four million people were left homeless—one of the worst such displacements in modern times.

And in 2017, Hurricane Harvey slowed to a crawl as it weakened near Houston, producing the heaviest rains ever to swamp a U.S. city. Totals of 750 mm (29.5 in.) covered the metro area, and a mind-boggling peak of 1539 mm (60.58 in.) was reported northeast of the city.

NOAA pegged Harvey's damage at $125 billion, mainly as a result of massive flooding.

Full-globe observations of rainfall are difficult to obtain, in no small part because 70% of the planet is covered by oceans.

For the world's land areas as a whole, the available data do not point to any clear global trend in total precipitation over the last 50–100 years.

However, some regional patterns have emerged, as noted in the 2013 IPCC assessment.

There's generally more rain and snow falling across middle and higher latitudes of the Northern Hemisphere.

In the United States, average precipitation has increased by about 5% over the last century, according to the National Oceanic and Atmospheric Administration.

Meanwhile, the distinction between the wet tropics and dry subtropics has intensified over the last 30–60 years, with the wet areas tending to moisten and the dry areas trending drier.

Over the next century, researchers expect a further increase in rain and snow, especially in winter, poleward of about 50°N (including most of Canada, northern Europe, and Russia), with rainfall also increasing in tropical areas that benefit from monsoons.

Meanwhile, precipitation may decrease—particularly in the summer—across the subtropics and in drier parts of the midlatitudes, including much of Australia, Southern Africa, the Mediterranean and Caribbean, and the southwestern United States and Mexico.

While confidence in this general picture has become stronger, there's still important uncertainty about just where the transition zones between increasing and decreasing precipitation will lie.

When it comes to floods, it's the character of the precipitation that counts the most.

How often is rain concentrated in short, intense bursts that produce flash floods, or in multi-day torrents that can cause near-biblical floods across entire regions?

And how often is it falling in more gentle, nondestructive ways?

Here, the global-change signal is worrisome.

Data show a clear ramp-up in **precipitation intensity** for the United States, Europe, and several other areas over the last century, especially since the 1970s.

When it rains or snows in these places, it now tends to rain or snow harder, over periods ranging from a few hours to several days.

Projections from computer models suggest that this trend will intensify in the decades to come, although the details will vary from place to place and not every location will follow the trend.

Modeling carried out in support of the 2013 IPCC assessment indicates that the heaviest five-day precipitation events one might expect per year at a given spot could increase by anywhere from 5% to 20% this century, depending on the pace of greenhouse emissions.

There’s a cruel twist to this story.

You might expect droughts to diminish on a global basis as rainfall intensity goes up, but higher temperatures not only allow more rain-producing moisture to enter the atmosphere, they also suck more water out of the parched terrain where it hasn’t been raining.

Atmospheric circulations associated with a warming planet also tend to reduce the number of lighter precipitation events.

Thus, in addition to triggering more intense rainfall, global warming could also increase the intensity of drought impacts, a seeming paradox that already appears to be taking shape in some areas, including California (see sidebar).

Where and how the worst droughts manifest hinges on a number of factors, including a tangled web of influence through which the world’s tropical oceans can help trigger multiyear drought in places as far flung as the southwestern United States and southern Africa.

Even in a single location, both drought and downpours can become more common if the timing of rainfall shifts in not-so-fortunate ways.

A wetter world (but not everywhere)

The stakes are high when it comes to assessing the future of floods and droughts, because both are notoriously deadly and destructive.

A 2015 United Nations report estimated that the global economic toll from flooding averages more than US$30 billion per year.

China's Yangtze River flows over its banks so reliably during monsoon season (its floods and related crop failures have claimed more than a million lives since 1900) that it prompted the world's largest flood-control project, the colossal and controversial Three Gorges Dam.

When it comes to landfalling hurricanes and typhoons, high water is typically far more deadly and destructive than high wind.

Many other flood victims die on a smaller scale, unlucky enough to be caught solo or in a small group along a rampaging creek during a lightning-quick flash flood.

Though drought is a less dramatic killer than flooding, its toll can also be staggering.

The World Meteorological Organization estimates that more than 650,000 people died in Africa during brutal famines that peaked in the wake of devastating droughts of 1972–1975 and 1984–1985.

Southern and eastern Africa have also been prone to recurring drought over the last several decades.

Looking further back, the history of civilization is checkered with cultures believed to have met their downfall partly or mainly because of drought (see later discussion).

The roots of both flooding and drought lie in the physical process known as evaporation.

As global warming heats the world's oceans, the water molecules near the sea surface become more energetic and tend to evaporate into the atmosphere more readily.

Thus, the air gains water vapor —or, as it's often put, the air "holds" more water vapor.

For every degree Celsius of warming, the water vapor in the atmosphere is able to increase by up to 7%.

Rain and snow tend to develop where air is converging and rising.

If the air is warmer and has a bit more water vapor, it ought to rain or snow more intensely, all else being equal.

This appears to be exactly what's happening on a global average, as noted above.

However, rainfall and snowfall often vary greatly over small distances.

Thus, a modest increase in global precipitation intensity could mask regional and local trends that are more dramatic, downward as well as upward.

To complicate the picture further, rain or snow totals at some locations can rise or fall sharply for a year or more because of the climate cycles such as El Niño discussed below.

To help quantify the links between downpours and climate change, the scientists who came up with a set of criteria for measuring temperature extremes (see earlier) did the same for precipitation.

Those indices are a key part of research that fed into the IPCC's 2007 and 2013 assessments.

In 2002, Povl Frich (Met Office Hadley Centre) and colleagues confirmed that precipitation extremes seem to be increasing in general, especially across the planet's higher latitudes.

Highlights from Frich's study, which looked at the second half of the twentieth century, included the following:

- **The number of days that see at least 10 mm (0.4 in.) of rain or melted snow** rose by 5%–15% across large stretches of the United States, Europe, Russia, South Africa, and Australia. These locations also tended to see upward trends in the peak five-day totals of rain or snow in a given year.
- **Much of North America and Europe** showed jumps in the fraction of rainfall and snowfall that fell on the soggiest 5% of all days with precipitation. In other words, the wettest days became wetter still.

Subsequent research, including modeling in support of the 2013 IPCC assessment, backs up the general picture of global trends toward more intense rain and snow events.

A 2013 overview led by the University of New South Wales found a particularly strong jump since the 1980s in the ratio of total annual rainfall to the number of wet days—in other words, a jump in the average amount of rain per wet day.

However, there's still a lot of smaller-scale variability in the global patchwork of wet and dry.

Even a region where precipitation is generally on the increase can have some locations that buck the overall trend.

Are floods increasing?

Among the lineup of atmospheric villains often associated with climate change, river and stream flooding is one of the toughest characters to nail down.

There's no doubt that rising seas will enhance the flood threat from hurricanes and coastal storms along low-lying shorelines, but inland flooding is another matter.

Although there's a good deal of confidence that the most intense bouts of rain and snow are getting heavier, it's surprisingly hard to compile a global picture of whether river floods are becoming more frequent or intense.

In part, that's because the chain of events that leads from an unusually heavy rain to a flood involves many factors other than immediate weather: for example, how wet the region's soils already are, how high rivers and reservoirs are running, what kind of flood-control devices are in place and—perhaps most critically—how much the landscape has been altered by development.

Putting aside any potential contribution from climate change, development and population trends alone will boost the world's financial exposure to flood risk severalfold from 2010 to 2050, according to a study led by the University of Amsterdam that considered both river and coastal flooding.

One of the first attempts at a global flood census was published in 2002.

Led by the U.S. Geological Survey, the study examined 29 of the world's largest river basins.

It found that 16 of their 21 biggest floods of the last century occurred in the century's second half (after 1953).

However, as noted above, a warming climate makes both downpour and drought more intense, so an increase in floods doesn't necessarily mean rivers are running consistently higher around the globe.

A 2009 study from the National Center for Atmospheric Research estimated that about a quarter of the world's largest rivers saw a reduction in streamflow between 1948 and 2004, while only about 10% of the rivers showed an increase.

Land-use changes play a huge role in flooding potential.

Deforestation tends to exacerbate the risk of flooding and landslides, as the water that falls is less likely to enter the soil and more likely to move unimpeded toward the nearest stream or low area.

Water flows particularly easily across the acres of pavement that are laid down as cities expand into the countryside.

Faster-flowing water is especially likely to feed into small-scale flash floods, which are even harder to monitor and analyze than larger river floods.

There's even wildfire to consider: paradoxically, an intense wild-land fire tends to raise the odds for subsequent flooding.

The fire's heat can effectively seal the soil, making it more water-repellent, and the fire clears the landscape of vegetation that would otherwise impede runoff.

Timing is another important factor when looking at how global precipitation and flood risk have evolved.

For instance, are the heaviest one-hour deluges getting heavier, which would presumably help make flash flooding more likely?

This is virtually impossible to calculate for many parts of the globe, because many national meteorological centers do not compile such fine-scale data or won't release them without collecting a hefty fee.

Also, the heaviest of heavy downpours tend to be so localized that they can be measured only with the help of the dense observing networks found mainly in highly industrialized countries.

This leaves us in the dark as to changes in flash floods across poorer nations.

It will still be a challenge for scientists to untangle any greenhouse-related trends in river flooding from the myriad other factors involved.

In its special 2012 report on managing risks from extreme events, the IPCC noted that "[m]any river systems are not in their natural state anymore, making it difficult to separate changes in the streamflow data that are caused by the changes in climate from those caused by human regulation of the river systems."

The report expressed low confidence in analyses of recent change in the global magnitude and frequency of floods, as well as in projections of future change, albeit with a few regional exceptions.

Some progress has been made since then, especially with new chains of models that link meteorology, hydrology, river flow, and population.

A comprehensive 2016 study led by the European Commission—focusing on large river systems, and holding population and development constant—found that even a 1.5°C rise in global temperature over preindustrial levels could more than double the planet's risk of flood damage as compared to 1976–2005.

The largest anticipated impacts are in perennially flood-vulnerable China, but some of the largest ramp-ups in projected risk were found in Bangladesh and India.

Defining drought

Anybody can see drought at work: wilting crops, bleached grass, trickling streams, a receding lakeshore.

Yet scholars have debated for many years how best to define this strange condition— the result of the absence, rather than the presence, of something.

Meteorological drought is the most obvious type: the rains stop, or at least diminish, for a given length of time.

In places where it normally rains throughout the year, such as London or Fiji, you might see regular rain and still find yourself in a meteorological drought if the amounts don't measure up.

One way to gauge a drought in such places is to add up the cumulative rainfall deficit over some period of time (for example 30, 60, or 90 days).

Another would be to calculate how long it's been since a single day produced a given amount of rain.

In semiarid or desert locations, such as Melbourne or Phoenix, you could measure how many days it's been since it's rained at all.

No matter where you are, it's the departure from average that makes the difference, since flora and fauna adapt to the long-term norms at a given locale.

A year with 500 mm (20 in.) of rain would be disastrously dry for Tokyo but a deluge in Cairo.

Since drought affects us by drying up water supplies and damaging crops, experts now classify drought not only by rainfall deficits but also in terms of **hydrological** and **agricultural drought.**

How low is the water level in the reservoir that your city relies on?

Are the lands upstream getting enough rain or snow to help recharge the system?

These factors, and others, help determine if a hydrological drought is under way.

Especially when a dry spell first takes hold, it may take a while for rivers and lakes to show the effects.

Likewise, long-parched ground may soak up enough of a drought-breaking rain to keep the hydrologic system from responding right away.

Thus, hydrological drought often lags behind meteorological drought, a process that may unfold over a few weeks in a fairly moist climate or across months or even years in a semiarid regime.

Crops can be affected in even more complex ways by drought, making a declaration of agricultural drought highly dependent on what's normally planted in a given location.

If a dry spell is just starting, the topsoil may be parched but the deeper subsoil may still be moist enough to keep an established crop going.

Conversely, a long-standing drought may break just in time to help germinating plants get started, even if the subsoil remains dry.

As it draws moisture out of vegetation, drought is an ideal booster of **wildfire** risk, especially when paired with unusually high temperatures.

In California, "hot droughts" have become far more common than cooler ones over the last century, according to a 2015 paper in the *Proceedings of the National Academy of Sciences of the United States* led by Stanford University.

A record-hot summer and a parched autumn teamed up in 2017 to produce California's third deadliest wildfire on record—the Tubbs Fire, which killed 22 people in Napa and Sonoma Counties in October—and its largest, the Thomas Fire, which rampaged across Santa Barbara and Ventura Counties in December.

On the other side of the United States, in and near the Great Smoky Mountains National Park, unprecedented autumn heat and intense drought fed a catastrophic fire in November 2016 that killed 14 and injured more than 100.

What do oceans have to do with drought?

Oceans might seem to be the last thing that would control the onset or departure of a drought, but the more scientists learn about the 70% of Earth's surface that is covered by water, the more they learn about how oceans can dictate the amount of water falling on the other 30%.

And as greenhouse gases heat them up, the oceans may start influencing drought in hard-to-predict ways.

Some parts of the world are dry by virtue of their location.

Most of the planet's major deserts are clustered around 30°N and 30°S, where sinking air predominates at many longitudes, while rising air keeps most of the deep tropics between 15°N and 15°S quite rainy on a yearly average.

The fuel for those showers and thunderstorms is the warm tropical water below.

Should a large patch of ocean warm or cool substantially—because of a transfer of heat by ocean currents, for example—it can take months or years for that region to return to more typical conditions.

While they last, though, the unusual water temperatures can foster atmospheric circulations that bring persistent wetness or dryness to various parts of the globe.

The supreme examples of such an ocean-driven shift are El Niño and La Niña, which bring warmer- or cooler-than-normal surface waters, respectively, to the eastern tropical Pacific.

This region of warming or cooling can be as large as the continental United States, and the resulting shifts in rainfall patterns can affect much of the globe.

Typically, El Niño hikes the odds of drought across Indonesia, Australia, India, southeast Africa, and northern South America, and it can cause winter-long dryness across parts of Canada and the northern United States.

Most El Niño droughts only last a year or so, but that's enough to cause major suffering where societies can't easily adapt (by changing crops or shifting their planting schedules, for instance).

Some of the worst North American droughts have occurred during multiyear periods when La Niña predominated.

Although El Niño and La Niña are the undisputed champions in drought-making, other oceanic cycles also have an influence.

Two of the most important are the Atlantic multidecadal oscillation (AMO) and the Pacific decadal oscillation (PDO; also called the interdecadal Pacific oscillation when both the North and South Pacific are analyzed).

Despite the contrast in their names, both play out over more than a decade: the PDO tends to switch phases about every 20–30 years (though with a good deal of irregularity in the mix) while the AMO takes closer to 30–40 years.

There's some evidence that the warm phase of the AMO and the cool phase of the PDO are linked to enhanced odds of drought in the U.S. Southwest.

Both of these phases predominated from around 2000 through the early 2010s, with the PDO making a sharp switch to its positive phase in the mid-2010s.

On top of these natural waxings and wanings, there's human-induced climate change.

Warmer temperatures help promote more evaporation and thus create a tendency for droughts to have a higher impact where they do occur.

Global-scale warming also fosters a poleward shift of arid subtropical zones.

These two factors are expected to hike the drought risk substantially along a belt from the southwestern United States to the northern Mediterranean.

Already, years of on-and-off drought have left Arizona's Lake Powell and Nevada's Lake Mead—both critical for water supply across the populous U.S. Southwest—struggling to remain even half full.

Scientists from Lamont–Doherty Earth Observatory and Cornell University have voiced concern that the southwestern United States and northwestern Mexico may enter a state of semipermanent drought later this century.

Factors such as El Niño will still lead to periodic wet years, and substantial water may still flow to the region via the Colorado River from the high-elevation Rocky Mountains, where precipitation appears less likely to decrease over time.

Still, the long-term trends suggest major trouble for agriculture and water use in the U.S. Southwest.

Similar predicaments are liable to strike along and near the Southern Hemisphere's subtropics, including Australia as well as South Africa.

After three years of major drought, Cape Town tiptoed to the brink of shutting down its municipal water supply in 2018, as reservoirs dipped close to their minimum operating levels.

With the help of emergency-level water conservation by residents, business, and government, the city managed to postpone the feared “Day Zero” for at least a few months, which provided crucial time for Cape Town to bolster its portfolio of water sources.

How will broad shifts like subtropical drying intersect with oceanic cycles and changes?

That picture remains fuzzy, because it’s exceedingly difficult for climate models to replicate ocean cycles and how they might evolve and interact in a warmer climate.

This, in turn, makes it challenging to know when and where oceanic influences on precipitation may shift.

Some researchers have argued that the Pacific might move toward a long-term state that resembles El Niño, while others see a greater risk of an ongoing La Niña–like state.

All in all, it seems there is much yet to learn about how ocean-modulated droughts will unfold as the world warms, with the stakes running high for many millions of people.

The plow and its followers: Farming and rainfall

Perhaps no other part of weather and climate humbles human ambition as does rainfall and the lack of it.

For centuries we've flattered ourselves by thinking we can control the atmospheric tap, whether it be through rainmaking rituals or cloud seeding.

Indeed, one of the first ways in which Western thinkers conceptualized human-induced climate change was the notion that "rains follow the plow"—an interesting spin-off from the "manifest destiny" mindset that fueled U.S. and Australian expansion in the 1800s.

As Europeans moved into progressively drier parts of North America and Australia, they believed that cultivating the prairies and bushland—watering the crops and thus moistening the air above them—would help bring rainfall.

Occasional setbacks, like Australia's vicious drought of 1888, didn't quash the belief at first.

It wasn't until the 1930s, when severe multiyear drought created the Dust Bowl across the U.S. heartland, that farmers and land managers began to rethink their relationship to climate.



A dust storm approaches the Texas Panhandle town of Stratford in April 1935. (NOAA/ George E. Marsh Album)

The flip side of this faith that humans could produce a rainy regime via agriculture was the belief that improper use could dry out the land permanently.

The notion of **desertification** got its start in Africa's Sahel, where it was postulated by colonial explorers as far back as the 1920s and long supported by actual events.

During moist years, government policy in some African nations had encouraged farmers to cultivate northward toward the desert's edge. e time, it was accepted by many that the disaster was being produced, or at least perpetuated, by its victims.

In turn, that pushed nomadic farmers and their herds into even more marginal territory farther north.

When the drought hit, both grazing and farming were too intensive for their land's now-fragile state.

Wind and erosion took hold, and the zone of infertile land grew southward.

At the time, it was accepted by many that the disaster was being produced, or at least perpetuated, by its victims.

Now it seems that researchers have come around to the idea that neither drought nor rain may follow the plow in a predictable way.

It stands to reason that both dry and wet conditions have a certain self-sustaining quality.

All else being equal, when the ground is soaked, the air above it is more likely to produce further rain than when the ground is dusty.

But the big question is, how powerful is this process against the vast energy reserves of the oceans and the natural variations of weather?

Thus far, there's no consensus on exactly how much of a difference land use can make.

One possibility is that human-produced changes to the landscape can sometimes modulate a drought that's driven by broader forces.

For example, some computer models fed with ocean conditions from the 1930s produce multiyear drying across the southern U.S. Great Plains.

However, the actual Dust Bowl was larger and farther north than these model-generated versions.

One reason, according to Lamont–Doherty's Richard Seager, might be the influence of dust from overplowed, desiccated farms.

As the dust blew across a vast area, it likely fostered dry, subsident conditions that suppressed rainfall over a larger swath than usual.

No matter what might cause their rains to stop, drought-prone areas can take simple, useful steps to reduce their vulnerability.

Some of the techniques that proved helpful in the U.S. Great Plains, such as wind-breaking belts of trees, were successfully adapted to the Sahel.

Prudent land management can also help people get the most benefit out of the rain that does fall.

At the same time, a bit of modesty is in order: oceanic shifts and other natural processes appear to be able to produce drought in some cases no matter how much we do to keep it away.

6 - The Big Melt - CLIMATE CHANGE IN OVERDRIVE

If there's any place where global warming leaps from the abstract to the instantly tangible, it's in the frigid vastness of the Arctic.

Here, the ice is melting, the ground is heaving, plants and animals are moving, and people are finding themselves bewildered by the changes unfolding year by year.

Because of a set of mutually reinforcing processes, climate change appears to be progressing in the Arctic more quickly than in any other region on Earth.

Not that some Arctic residents haven't had to adapt to climate change before.

Signs of Stone Age life dating back nearly 40,000 years—well before the last ice age ended—have been found near the Arctic Circle in northwest Russia.

What appears to be different this time is the pace of the change and the sheer warmth now manifesting.

Long-time residents of western Canada and Alaska have seen wintertime temperatures climb as much as 5°C (9°F) since the 1950s.

That's several times beyond the average global pace.

Much of the permafrost that undergirds the Alaskan and Canadian Arctic warmed to unprecedented levels in the 2010s.

Sea levels are slowly rising across Arctic coastlines, as they are elsewhere around the globe, and longer ice-free stretches are increasing the risk of damaging, coast-pounding waves.

Looking ahead, experts believe the next few decades will likely bring a summer when the Arctic Ocean is virtually free of ice, perhaps for the first time in more than 100,000 years.

As we will see, these changes threaten to reorder Arctic ecologies in dangerous and unpredictable ways, with potentially vast long-term consequences for other parts of the planet.

The big melt isn't limited to the poles.

Many glaciers in the highest mountains of the tropics and midlatitudes are receding at a startling pace.

It's tempting to chalk this up to the basic effects of a warmer atmosphere, but drying may in fact be a bigger culprit than warming in the erosion of glaciers on some tropical peaks, including the one best known to many: Africa's Kilimanjaro.

Meanwhile, the biggest ice repositories of all—the vast sheets that cover most of Antarctica and Greenland—are sending frozen and melted water to the ocean at accelerating rates, an ominous development that will touch every corner of the world's oceans and coastlines.

On thin ice

The North Pole sits at the heart of an ocean ringed by the north fringes of eight countries: **Canada, Greenland, Finland, Iceland, Norway, Russia, Sweden**, and the **United States.**

All this adjoining land provides plenty of toeholds for sea ice, which overspreads the entire Arctic Ocean and portions of nearby waters each winter.

This ice takes on many roles: a home base for polar bears, a roadbed for people, and a shield for coastal towns and underwater creatures.

In the summer, as warm continental air sweeps poleward from North America and Eurasia, the ice releases its grip on land and contracts into the Arctic Ocean, with only the oldest, most resistant sections clinging to Greenland and the islands of the Canadian Arctic.

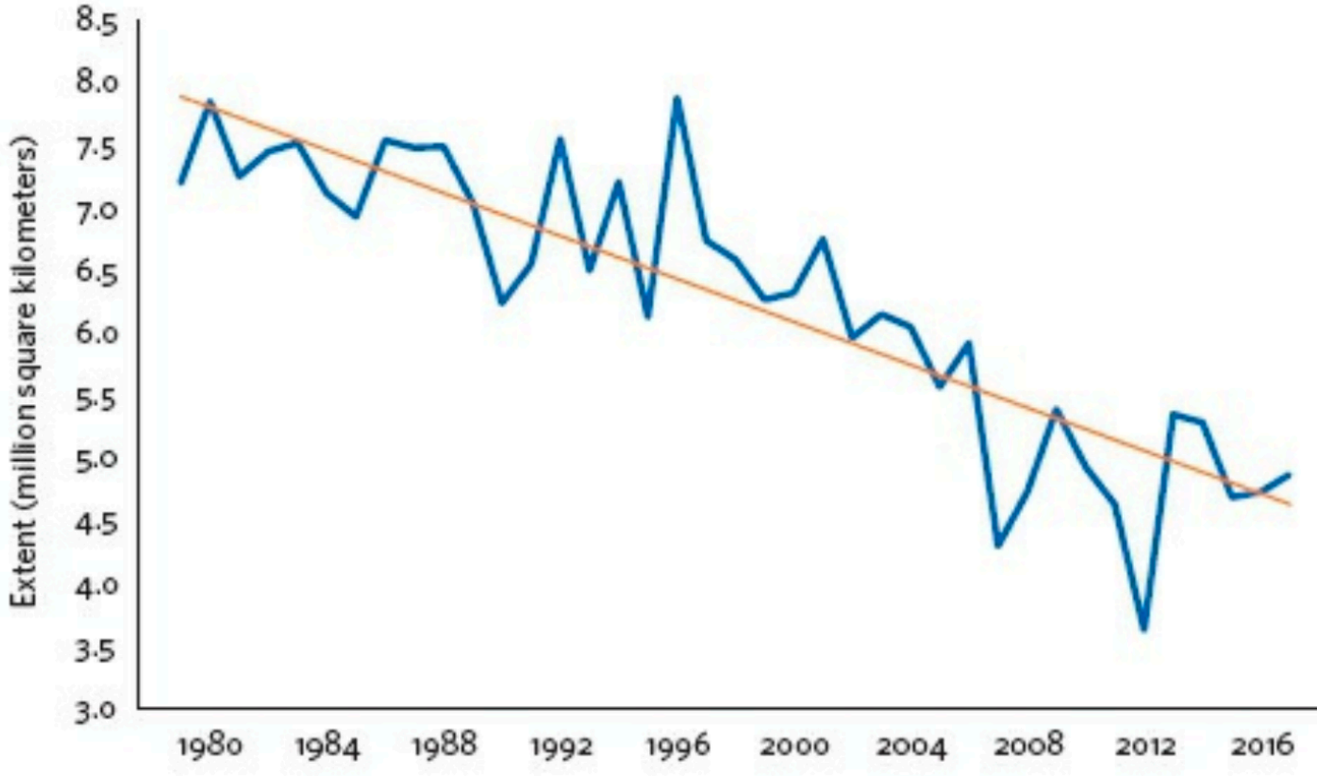
Much of the rest disintegrates into a messy patchwork that allows icebreaker cruise ships to penetrate all the way to the North Pole.

While some patches of ice erode in place, winds and waves push other fragments south to meet their doom, primarily through the Fram Strait between Greenland and Spitsbergen, Norway.

Eventually, winter returns, and by February the Arctic Ocean is once again encased in ice.

Nobody expects the Arctic to completely lose its wintertime ice cover anytime soon, but summer is another matter.

As of 2017, the average coverage in September, when the ice usually reaches its minimum, was only about **65%** of what it was in the late 1970s, according to the U.S. National Snow and Ice Data Center.

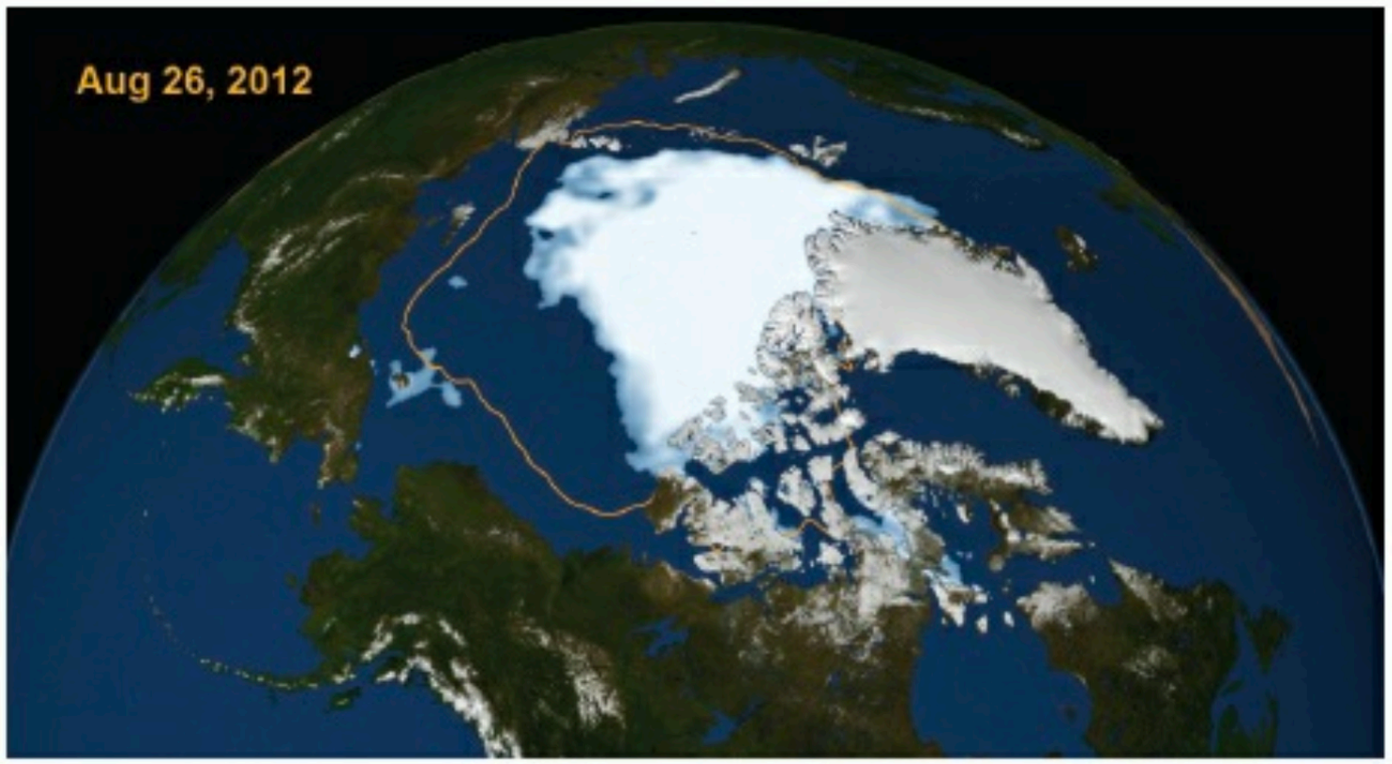


The typical extent of Arctic sea ice—averaged across September of each year, when the ice normally bottoms out—has plummeted during the last several decades, though not every year has shown a drop. (U.S. National Snow and Ice Data Center)

Even the wintertime ice extent is shrinking by 2%–3% per decade. Satellite images paint a stark portrait of these trends, which have reached truly dramatic proportions.

The summer of 2007 smashed all prior records, with even the most experienced ice watchers surprised by the pace of melt.

By that September, the ice extent was barely over 4,000,000 km^2, more than 1,000,000 km^2 less than its previous low, set in 2005.



The yellow contour shows the average minimum extent for Arctic sea ice achieved each year from 1979 to 2010. Though the sea ice normally doesn't reach its minimum until September, it had already shrunk to the lowest value on record by the time this satellite image was taken on August 26, 2012. (Scientific Visualization Studio)

Only two-thirds of the usual late-summer pack remained, and ice-free waters stretched from Alaska and Siberia nearly to the North Pole.

Scientists later connected the melt of 2007 to unusually sunny skies in early summer, along with an ice-threatening pattern of atmospheric pressure called the Arctic dipole.

Summer winds normally tend to encircle the Arctic, which helps preserve ice locked at high latitudes.

But when the dipole sets up, warm air shuttles across the Arctic from North America to Eurasia, and ocean currents push ice toward the North Atlantic, where it's more likely to melt.

A team from the University of Alaska Fairbanks targeted the dipole pattern as evidence of a drastic change in Arctic climate.

The sea ice didn't retreat quite so dramatically the next summer, a point that was trumpeted to bolster climate change denial. However, it's now clear that 2007 marked a turning point for the Arctic.

In every September since then (up through at least 2017), the minimum ice extent has dipped below anything observed prior to 2007.

Another new record was set in 2012—followed once again by a modest recovery, in what's become a two-steps-downward, one-step-upward pattern.

Ship traffic is one real-world marker of the huge changes afoot in the Arctic.

In almost every year since 2007, the fabled Northwest Passage north of Canada—long considered a no-go zone due to near-permanent ice cover—has lost enough sea ice to make the passage's two routes navigable.

Meanwhile, the Northern Sea Route north of Siberia, once limited to icebreaker-guided traffic, has seen dozens of ship transits since the late 2000s, including a commercial tanker that made the trek in 2017 without any icebreaker assistance.

In 2010, two yachts —*Northern Passage* and *Peter 1*—became the first to circumnavigate the Arctic in a single summer.

The Arctic's ice isn't just contracting horizontally; it's also getting shallower.

Widespread summer melt and shifting ocean currents are eliminating much of the ocean's thick multiyear ice, which can extend 2 m (6.6 ft) or more below the surface.

The thinner coat that forms in place of multiyear ice is easier to erode when summer returns.

Data on ice thickness collected by U.S. nuclear submarines from the late 1950s to 2000 show that sea ice thinned by up to 40% over the period.

The submarines traversed only a few select routes below the ice, though.

Estimating ice thickness over a broader area from space isn't easy, but NASA did so from 2003 to 2010 using a laser-based sensor on board a satellite called *ICESat*, and similar data have been collected by a European satellite launched in 2010, *Cryosat-2*.

Computer models are stitching together the limited thickness observations with depictions of ocean, ice, and atmospheric behavior to produce a comprehensive portrait of ice decline in three dimensions.

At the University of Washington, a model called the Pan-Arctic Ice Ocean Modeling and Assimilation System (PIOMAS) indicated that in 2017 the April maximum in ice volume was down almost 25% from its 1979–2012 average, and the September minimum had plunged by roughly 65%—far more than the surface data in the graph below might lead one to believe.

Even the stubborn remnants of multiyear ice hanging on for dear life aren't what they used to be.

A team from University of Manitoba trekked deep into the Arctic Ocean on board an icebreaker in September 2009.

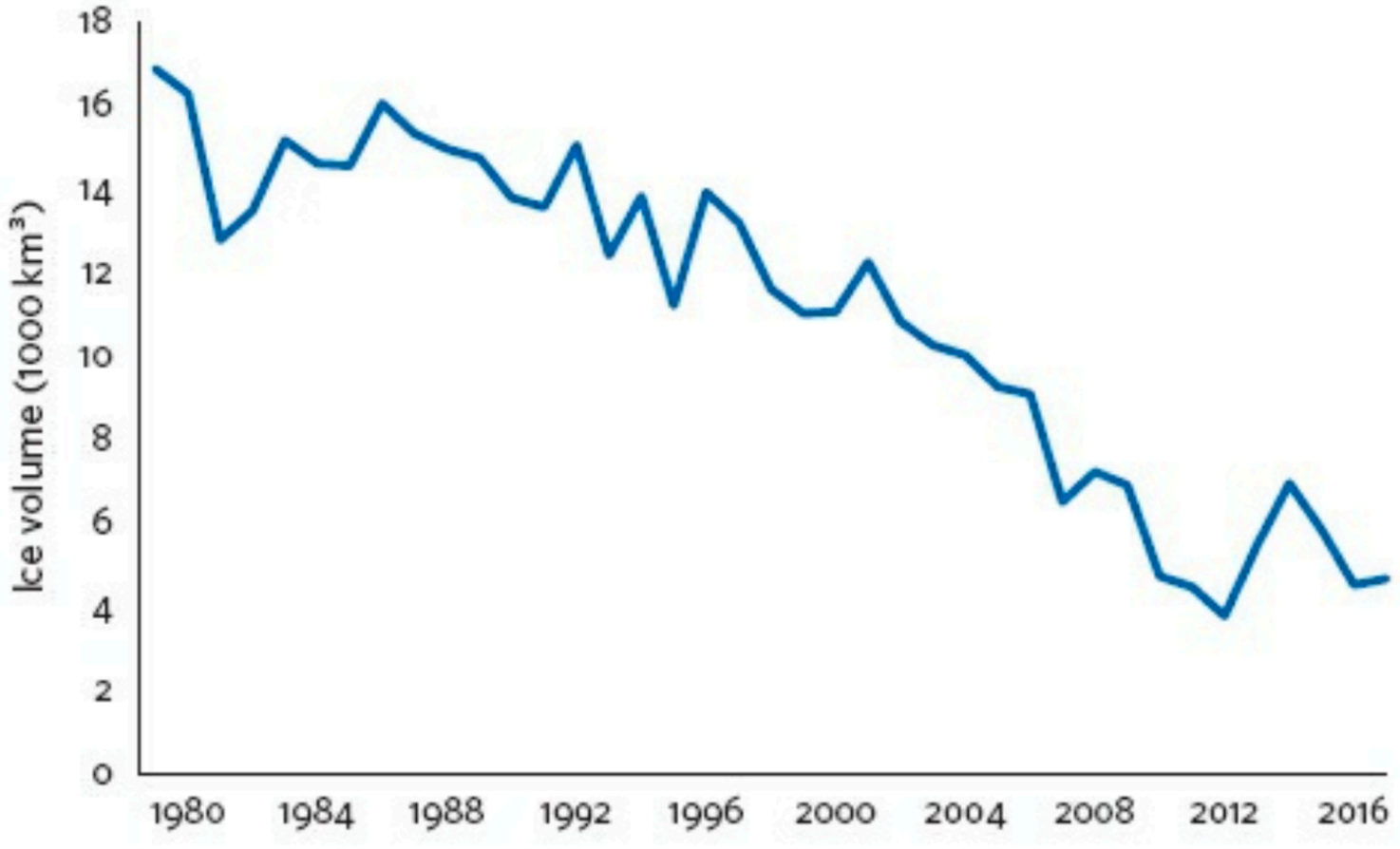
They expected to grind their way through multiyear or thick first-year ice along most of their route. Instead, they encountered chunks of older, thicker ice beneath a coating of fresh ice only about 5 cm (2 in.) thick. University of Washington).

When scanned by NASA's satellite-based sensor, this new icescape resembled a more typical, multiyear ice field—a masquerade hiding a far more serious decline than scientists had realized.

The tenuous state of today's Arctic ice was also highlighted by the Great Arctic Cyclone of August 2012.

This unusually large and intense storm delivered several days of strong wind that fragmented large swaths of ice across the central Arctic.

Before-and-after satellite images showed a huge impact on the ice pack, but it looks as if the storm mainly hastened the destruction of ice that was already doomed, according to a general analysis by the University of Washington.



Until 2004, the volume of ice remaining at the last day of each melt season, as computed by the University of Washington PIOMAS model, topped 10,000 km^3 each year. In 2012, that volume was a mere 3261 km^3. If this pace of decline had continued, the Arctic Ocean would have been virtually ice-free at some point before 2020. However, the rate of ice volume loss slowed measurably in the 2010s, providing at least a temporary reprieve. (PIOMAS/University of Washington)

How long will the Arctic Ocean hang onto its summertime ice?

For a long while, the best consensus among top climate modelers was that the Arctic would retain some ice throughout each summer until at least 2080, and perhaps into the next century.

But with more recent data and model results, that consensus has fragmented even more quickly than the ice itself.

The 2013 IPCC report asserted that the Arctic is likely to see a September virtually free of sea ice before midcentury, assuming that emissions follow a middle-of-the-road path.

That's based largely on global climate models that show improved skill in depicting Arctic ice.

However, even the best models still tend to deplete sea ice at a pace slower than that actually seen over the last few years.

What makes the Arctic so vulnerable?

Trouble seems to beget trouble when it comes to far-northern climate.

As a team of scientists noted in the American Geophysical Union's weekly newspaper, Eos, "The Arctic system balances on the freezing point of water. Each summer, the system swings toward the liquid phase; each winter, it returns to the solid phase."

While a tropical city might warm by a degree or two without obvious effects, the same amount of warming could transform parts of the Arctic if it were to bring the system from just below to just above the freezing point.

If the tropics are the lumbering long-haul trucks of the climate world, responding slowly but powerfully to greenhouse gases, then the Arctic is a finely tuned sports car, reacting sensitively to even small changes.

Much of this responsiveness is because of positive feedbacks: self-reinforcing processes that tend to amplify change.

In a definitive 2005 study, the Arctic Climate Impacts Assessment (ACIA) noted several positive feedbacks that allow the Arctic to warm more quickly than other parts of the globe ni response to a given increase in greenhouse gases.

These include the following:

A darkening surface. Several kinds of change in the Arctic al help to produce a darker surface that absorbs more sunlight and, in turn, warms the air above it more strongly.

Open ocean typically absorbs more than 90% of the solar energy reaching it, while snow and ice absorb as little as 10%, reflecting the rest of the sunlight of their bright white surface.

Thus, a patch of ocean that remains unfrozen can soak up as much as nine times more solar energy.

Likewise, when snow cover over land disappears for longer periods, the exposed tundra absorbs more of the sunlight than the snow did.

Shrubs and forests, in turn, absorb more solar energy than the lighter tundra does, so as these larger plants migrate northward, they act to further warm the region.

This is in contrast to the tropics, where trees tend to cool the climate as they catch rainfall and evaporate it back into the air.

Pollution wafting into the Arctic from more populated areas, including fast-industrializing Asia, helps darken some of the snow and ice, increasing its absorptivity.

A more direct route to warming. As the sun passes over the tropics, much of its energy goes into evaporating water, but the evaporation rate in the frigid Arctic is far less.

Thus, of the sunlight that reaches the Arctic, a higher percentage goes directly into warming the air.

A thinner atmosphere. The Arctic's troposphere, or "weather layer" typically extends about 8-10 km (5-6 mi) up, compared to 16-18 km (10-11 mi) in the tropics.

In the shallow troposphere of the poles, it takes less energy to produce a given amount of warming.

There's another wrinkle as well.

Especially during clear, calm, cold spells, Arctic landscapes often experience a thin, ground-hugging layer of air, called an inversion, that's much colder than the air just a few hundred meters higher up.

When the inversion breaks up, the surface can warm dramatically.

Thus, anything that punctures inversion or makes them form less often, such as a change in atmospheric circulation or a loss of snow cover, could produce major ground-level warming.

Potential changes in heat-trapping clouds are a huge uncertainty in polar regions and elsewhere across the globe.

Over the Arctic, summer cloudiness appears to be increasing.

This could help tamp down snow and ice loss, since low clouds and fog in summer screen out the midnight sun, whereas clear skies in winter allow heat to easily escape to space.

However, the variegated complex behavior of clouds can lead to unexpected results.

One analysis found that distinctive low-lying clouds over central Greenland in July 2012 allowed sunlight in while blocking outgoing heat, which helped lead to record-breaking ice melt.

With all of this in mind, we might hope to find a few negative-feedbacks - processes that tend to dampen warming over time, such as the apparent cloud cover changes as noted above.

One possibility is the vast flow of heat from the tropics towards the poles.

If the Arctic continues to warm at a faster pace than lower latitudes, as expected, then the pole-to-equator contrast would be weakened and the flow of heat ought to slacken.

However, these weakened winds will carry more moisture than before - and that's a warming influence, all else being equal.

All of these feedbacks, plus the tendency of glacial melting to accelerate, point toward a potential cascade of effects - a so-called "tipping point", as its been characterized by some.

In fact, there may be a number of tipping points, each with its own setting.

If so, each point we pass makes it that much harder to avoid the others.

Conversely, the sooner that greenhouse warming is stabilized, the less likely the Arctic will enter a period of runaway change.

As pointed out by in a 2013 review for *Geophysical Research Letters*, there are three common approaches to predicting how quickly ice loss will proceed.

- If you simply extrapolated from trends in sea ice volume, which is diminishing more rapidly than sea ice extent, one might have concluded in the early 2010s that a nearly ice-free Arctic would arrive by 2020, or even sooner (see graphic above).
- However, weather and climate don't usually follow straight lines or smooth curves. Natural variability in the Arctic can easily lead to plateaus in ice loss; in fact, sea ice volume stayed above its 2012 minimum through at least 2017. Factoring such plateaus in mind, you might expect it to take a while longer to reach the nearly ice-free point—perhaps around 2030.
- If you went strictly by climate models, you'd call for the ice-free benchmark to occur still later —perhaps around midcentury, with some models pegging a date as early as 2040 and others around 2060 or later.

Natural variations alone may make it impossible to nail down the first appearance of essentially ice-free Arctic waters with anything more precise than a 20-year window, according to a 2016 analysis led by the University of Colorado Boulder.

That window widens another few years because of uncertainty about how quickly we will or won't cut fossil fuel emissions.

Clearly, to predict when we'll see the first summer days with a wide-open Arctic is one of the great forecasting endeavors of our time, especially given the rollercoaster developments of the last decade-plus.

But this is much more than an intellectual challenge.

As stated by many, “Whether a nearly sea ice-free Arctic occurs in the first or second half of the twenty-first century is of great economic, social, and wildlife management interest.”

People, animals, and ice

So what’s wrong with a few extra weeks of open ocean in a place that’s usually locked tight in ice?

Indeed, the prospect of ice-free summers has many industries and entrepreneurs salivating, especially those hoping to cash in on trans-Arctic shipping.

Along the Arctic coast, however, the increasingly tenuous state of the ice during autumn and spring complicates life for indigenous residents.

Subsistence whalers and seal hunters use sea ice to venture beyond the coastline.

The timing of those seasons, and the quality of the ice, has now become less dependable.

Also, the nearshore ice normally helps keep wind-driven waves well offshore during the violent weather of autumn and spring, but storms now smash into coastal towns unimpeded by ice well into autumn.

The resulting erosion threatens the very survival of some Arctic settlements.

The loss of sea ice has especially grave implications for **polar bears,** which have become a familiar symbol of climate change for good reason.

In some areas, these white giants remain on the sea ice throughout the year, retreating to land only when forced there by melting ice.

Land areas are also used by pregnant polar bears, who then bring their hungry young onto the sea ice to find seals who are bearing and raising pups in the spring.

In areas where the ice melts completely in summer, the bears go without their usual food supplies while on land, largely relying on fat reserves built up during the spring.

Now that ice-free periods are lengthening, the reduced duration of sea ice duration (earlier breakup and later freeze-up) is a stressful affair for both mothers and nursing cubs.

At Hudson Bay, near the southern end of the polar bears' range, the summer fast is now a month longer than it was decades ago.

The body condition of bears arriving on land has declined and has resulted in both lower survival (especially for young and old bears) and lower reproductive rates.

Greater extremes in precipitation are also a concern.

Fires that rage through drought-stricken boreal forests destroy permafrost and the dens lodged within it, and more frequent spring rains have collapsed some of the bears' dens.

Rain in place of snow can also affect the availability of prey for polar bears, because ringed seals rely on snow lairs to give birth to and raise their pups.

When sea ice forms later than usual, there's less of it, and less time for winter snowpack to accumulate on what does form, thus posing additional risk for the seals' lairs.

Then there's the threat of interbreeding between polar bears and grizzlies now taking shape in Canada, as a warming climate shifts the grizzlies' range toward coastal lands where polar bears roam in between shorter sea-ice seasons.

One bear shot in 2006 in northern Canada was found through DNA testing to be a grizzly/polar hybrid—a "pizzly."

At least eight other pizzlies, representing at least two generations, have been confirmed since then.

This isn't the first time that grizzly and polar bears have mated. Genetic evidence has been found of several other periods of interbreeding over the hundreds of thousands of years since the two species first diverged.

However, a warming climate throws the pizzly phenomenon into an entirely new and uncertain context.

Despite all of these concerns, the polar bear is hardly disappearing from Earth just yet.

Because the bears' territory is so remote, it's tough to obtain a firm count, but the most common estimate is that there are now between 20,000 and 25,000 of them.

That's more than twice the number tallied in the 1960s, but the increase has been attributed to careful management of hunting that allowed depleted populations to recover.

Going forward, the bears face a very different set of challenges, and there is no guarantee their current resilience will continue.

If in fact polar bears become less of a presence in the Arctic over time, there's another predator waiting in the wings.

Killer whales, which normally avoid sea ice because of their long dorsal fin, are now expanding into polar bear habitat during the ice-free season.

Killer whales would likely become the top predator in the Arctic marine ecosystem if sea ice were to disappear.

Recent genetic studies indicate that polar bears as a species have some 600,000 years of history behind them, which implies that they've survived interglacial periods when temperatures were slightly warmer than today.

However, the pace of warming is much faster this time around (with still higher temperatures on the way), and there's added stress from human civilization that didn't exist back then.

With all this in mind, it's unclear how well the bears will be able to adapt.

Only a few months after the Arctic's dramatic melt of 2007, the U.S. government listed polar bears as a threatened species.

A 2016 study in *Biology Letters* led by the U.S. Fish and Wildlife Service found it likely that sea-ice loss will bring down global polar bear numbers by 30% over three generations, or roughly by midcentury.

There's a small but real chance (perhaps 5%–10%) that polar bear populations will be cut in half by that point.

The safest haven for the species may be in the waters adjoining northern Greenland and the archipelago of far northern Canada, where year-round sea ice is expected to linger the longest.

The World Wildlife Fund is working with nearby residents and the government of Canada on proposals for designating a potential "last ice area" that could serve as a refuge for the bears as well as for the ways of life practiced by indigenous residents.

Several species of **seal** are also at serious risk from ice loss.

Ringed seals, the Arctic's most common type, are a common food source for many Inuit communities as well as for polar bears.

As noted above, the seals nurse their pups in lairs on top of the snow-coated ice sheet, so the stability of springtime ice is critical for them.

Fortunately for the seals, Arctic sea ice hasn't diminished as rapidly in spring as it has in summer.

One survey of ringed seal populations near the northwest coast of Hudson Bay showed a seesawing pattern, with the numbers dropping in the 1990s but rising in the 2000s, when it appears seals were moving into the study area.

As the ice retreats, ringed seals may get a short-term boost from richer marine life, but those benefits will vary from place to place.

Pups may find themselves exposed to the cold ocean—and to predators—at earlier points in the spring.

In 2012 the U.S. government listed three subspecies of Arctic ringed seal and two populations of bearded seals as threatened, noting the same climatic pressures that drove the polar bear listing.

There's no evidence that ringed seals will take to land, which goes for several other seal species as well, so in the long run the loss of summertime ice may prove devastating.

In contrast, gray seals and other temperate species already accustomed to land have a better shot at migrating northward and/or dealing with reduced Arctic ice successfully.

Birds—made for mobility—appear to have a head start in adapting to changes across the Arctic.

The task won't be easy for all species, though.

Ivory gulls are scavengers that feed mainly on the carcasses of ringed seals left by polar bears, so a decline in those species will at least force the gulls to find other meal tickets.

Fish and other marine creatures may also find their diets shaken up, as schools shift with the warming oceans and as industrial fishing moves northward into formerly ice-encased areas.

One of the major challenges facing birds in the far north is mismatched ecosystem timing, which is especially critical here given the Arctic's short window of warmth each summer.

Some species, such as Arctic terns, travel all the way from the Antarctic to exploit the brief seasonal abundance of food in the Arctic.

If the timing of the productive season shifts, many species are expected to be out of sync with their food.

It's hard enough for some forms of Arctic wildlife to deal with climate change, but **pollution** may be making matters worse.

A witch's brew of contaminants, including mercury and other heavy metals, has drifted into the Arctic over the years from both developed and recently industrializing nations.

Many of these contaminants can work their way up the food web, and they've been found at dangerously high levels in the fat reserves of some creatures near the top of the chain, most notably polar bears.

Wildlife stressed by climate change may be less able to deal with the negative consequences of these imported toxins.

A softening landscape

As the Arctic Ocean slowly sheds its mantle of year-round ice, terrestrial life nearby is taking on new forms.

Perhaps the most emblematic is the "drunken forest"—larch and spruce trees that lean at crazy, random angles as the soil beneath them thaws out.

Permafrost—land that's been frozen for at least two years—holds the key to this and many other transitions across the Arctic landscape.

Permafrost covers about a quarter of the Northern Hemisphere's land area, including half of Canada, most of Alaska, and much of northern Russia, as well as a small part of Scandinavia and several mountainous regions closer to the equator.

On the fringes of the Arctic, the permafrost is **sporadic** or **discontinuous**, forming in a mosaic of favored locations.

In the coldest areas, **continuous permafrost** extends throughout the landscape.

Permafrost is topped by an **active layer** that typically thaws and refreezes each year.

The permafrost itself, which starts anywhere from a few centimeters to a few tens of meters below ground, remains below freezing year round.

It can extend hundreds of meters farther down, until the soil temperature again rises above freezing en route to Earth's molten center.

The thicknesses of the active layer and the permafrost depend on both recent and past climate, making them convenient diaries of global change.

As part of the International Polar Year (which ran, despite its name, from 2007 to 2009), scientists drilled more than 500 boreholes across the frozen north to gauge the health of its permafrost.

Most areas showed a continuation of the the soil- temperature rises documented over the last few decades, with the coldest permafrost warming the most quickly—a finding reinforced in other recent studies.

It's the deepening of the active layer, and the thawing of permafrost just below it, that's causing problems in certain parts of the Arctic.

Tucked within the permafrost is water, some of it in crystals bonded to the soil structure and some in much larger ice beds or wedges.

As thawing gets deeper, it can liquefy ice deep within the ground, leading to slumping of the surface and the formation of marshy hollows and small hummocks.

When such features come to dominate a landscape, it's referred to as **thermokarst.**

Such is the process that has put many trees and buildings in a slow-motion fall, leaning at ever-sharper angles over time, especially across the fast-warming lower Arctic between about 55° and 65°N.

Because of the variability of year-to- year warming and the random locations of ice pockets, it's almost impossible to predict which house or tree in a melt zone will start leaning next.

In **Yakutsk**, the major city of eastern Siberia, many buildings are on stilts, which can not only stabilize a structure but also prevent it from warming the ground further.

More than 300 structures and an airport runway have been compromised by unstable permafrost.

Less-than- optimal building practices are likely behind some of the structural failures.

Moreover, much of the observed thawing of permafrost around Yakutsk can be directly related to soil disturbances triggered by forest fires, construction, and agriculture.

Even so, future warming is bound to make things worse for Siberian cities that sit atop permafrost, as concluded in a major 2016 study by Russian and U.S. scientists published in *Geographical Review.*

Across large stretches of the Arctic, it's evident that rising air temperatures are indeed having an impact on permafrost.

Many Arctic locations have seen a warming of permafrost—measured below the active layer, at depths of 9–26 m (30–85 ft)—on the order of 1°–2°C (1.8°–3.6°F) or more since the 1970s.

As a rule, higher-latitude, colder permafrost has seen the greatest warming.

Atop these zones, the active layers are more mercurial, thinning and deepening mainly in response to the pace of ice melting within them, the warmth of a given summer, and the depth of the previous winter's snow.

However, even active layers have shown longer-term deepening in some regions, including northeast Siberia and interior Alaska, where **Fairbanks** is a poster child for permafrost trouble.

Fairbanks lies on top of a fairly thin permafrost layer—only a few meters deep in many locations.

Record warmth in the past 20 years has percolated downward, heating permafrost across Alaska by 1°–2°C (1.8°–3.6°F).

Around Fairbanks, some sections of permafrost have warmed by more than 3°C (5.4°F), bringing its temperature above –0.5°C (31°F) in some locations.

The results are startling: drunken spruce trees, heaving bike trails, and sinkholes pockmarking the landscape.

Thus far, the sinking has been patchy, focused where digging and construction have opened up layers of landscape to warming and melting.

More widespread problems could occur if and when the bulk of Fairbanks' permafrost gets to the dangerous side of the freezing point.

As one might expect, experts and climate models agree that the Arctic's permafrost will continue to degrade through this century and beyond.

The overall pace will depend largely on how quickly the atmosphere warms.

By 2100, the global area covered by permafrost within 3 m (10 ft) of the surface could shrink by anywhere from 37% to 81%, depending on greenhouse emissions, according to the 2013 IPCC report.



Hundreds of buildings in Yakutsk, Russia, have been compromised by permafrost instability, although construction practices are likely a factor as well. (Will Rose/ futureworldproject.org)

While this would be a major ecosystem change in its own right, the repercussions would be much larger if thawing permafrost were to release huge volumes of carbon dioxide and methane.

Some 1700 metric gigatons of carbon—about twice as much as the entire global atmosphere now holds—are estimated to be trapped in the top 3 m (10 ft) of northern permafrost.

Is it possible this storehouse will be released, triggering a massive positive feedback that makes climate change even worse?

This is a critical long-term question that's yet to be comprehensively answered.

The amount of carbon that might escape from permafrost depends greatly on the extent and evolution of features such as thermokarsts, which can generate new methane-producing wetlands, and waterways, which can carry organic matter out to sea before it has a chance to enter the atmosphere.

The main player in the mix, however, is most likely soil-based microbes that release carbon dioxide.

Right now, the icy, saturated conditions in and near permafrost inhibit the growth of these microbes, but as temperatures rise, the soil should tend to dry out (despite the increased precipitation projected for polar areas).

Warmer, less-soggy soil, in turn, could allow CO_2-spewing microbes to proliferate.

One of the first major analyses of permafrost–climate feedback, from scientists from NOAA and the University of Colorado Boulder, was published in 2011 in the journal *Tellus*.

The study estimated that enough carbon could be released through permafrost melt by 2200 to boost atmospheric CO_2 concentrations by some 60–120 ppm.

The high end of that range is on par with the full amount of CO_2 added to date by human activity.

Other studies have pointed to the potentially dire consequences if a large pulse of methane, on the order of 50 metric gigatons, were to escape to the atmosphere over the next several decades from permafrost or seabeds.

However, as yet there's no widely accepted mechanism by which rising temperatures would trigger such a vast and geologically quick release of methane.

Most global climate models don't yet simulate permafrost—much less such features as thermokarsts, or the potential effects of permafrost thaw on the carbon cycle—so the IPCC didn't include possible effects from permafrost–climate feedback in its 2013 projections.

The panel did note that such effects could be significant.



Using 29 years of data from Landsat satellites, researchers at NASA have found extensive greening in the vegetation across Alaska and Canada, including grassy tundra changing to shrubland and shrubs growing bigger and denser. From 1984 to 2012, extensive greening occurred in the tundra of Western Alaska, the northern coast of Canada, and the tundra of Quebec and Labrador. (NASA Goddard Space Flight Center/Cindy Starr)

Northern fire

Trees aren't just toppling across the Arctic; they're also burning.

A 2013 study led by the University of Illinois at Urbana–Champaign found that boreal wildfires in Alaska's interior are burning at rates unprecedented for at least 10,000 years.

In 2004 and 2005 alone, **fires** swept across more than 10% of the tundra and forest in Alaska's interior.

Across Siberia, which holds half of the world's evergreen forest, losses have reportedly risen more than tenfold over the last few decades.

In 2012, Russia experienced its worst fire season on record, with flames consuming an area larger than Colorado (300,000 km^2, or 115,000 sq mi).

Sharply warmer summers in recent years have helped dry out the landscape much sooner than usual, lengthening the fire season and making the fires that start that much more intense.

In July and August 2013, Fairbanks saw a record 23 consecutive days reaching 70°F (21°C), with 11 of those days hitting 80°F (27°C).

That same summer, temperatures topped 30°C (86°F) in far northern Russia, and wild-land fire reached well poleward of its usual territory—extending within 200 km (125 mi) of the Arctic Circle.

The wrath of high-latitude fire hit home for tens of thousands of Canadians on May 3, 2016, when a fierce early-season fire swept through Fort McMurray, Alberta, during an unprecedented spring heat wave.

Some 3000 structures were lost across the region.

Total damage of C$9.9 billion made the disaster the most expensive in the nation's history.



Flames and smoke confront motorists fleeing the Fort McMurray, Alberta, area during a catastrophic wildfire on May 3, 2016. (DarrenRD/Wikimedia Commons)

Insects are playing a part in the onslaught of northern fire.

Epic infestations of pine and spruce bark beetles have swarmed across western North America over the last couple of decades.

The bugs have killed off millions of trees and increased the forests' vulnerability to intense fires for several years afterward.

These infestations have been driven in part by warmer temperatures, especially in winter, which reduce the likelihood of insect die-off and help foster multiple breeding cycles in a single year.

To be sure, a milder Arctic paves the way for the northward migration of many flora and fauna.

The problem is that many parts of the ecosystem can't change with the fluidity of the climate itself.

For instance, the weather in 30 or 40 years may be conducive to boreal forests across some regions now covered with nutrient-poor tundra.

However, it might take decades longer for decaying grasses to prime the soil so it can support tree growth.

Shrubs might fill the gap more quickly: their coverage across northern Alaska has increased by more than 1% per decade since 1950, and satellites indicate that Siberia might also be seeing more shrub-covered areas.

Still, models suggest a northward migration of high-latitude forest later this century, especially across western Siberia and north-central Canada.

This would pull an increasing amount of carbon dioxide out of the atmosphere, though probably not enough to counteract the warming effect caused by the forests' propensity to absorb sunlight.

Increased shrubbery may also have its own mix of impacts, shading the ground and insulating snowfall but also absorbing sunlight—again, perhaps enough to provide a net warming where the shrubs thrive.

The Arctic's signature land beasts, **reindeer** (known as **caribou** in North America), face their own challenges.

Heavier spring runoff and rains across the North American Arctic could widen rivers enough to jeopardize the seasonal trek of the continent's vast herds of caribou, some of which hold more than 100,000 animals.

In some parts of the Arctic, autumn and early winter are already proving treacherous.

Caribou and reindeer normally browse below the surface of early- season snow to reach the tender lichen that spread at ground level.

In the past few years, however, layers of ice within the snow—fostered by increasingly prevalent warm spells and bouts of freezing rain—have encased the lichen at times, blocking the animals from feeding.

This process has been implicated in the decline of Peary caribou, whose numbers across far north Canada and Greenland have plummeted by more than 80% over the last 50 years, according to the Committee on the Status of Endangered Wildlife in Canada.

Reindeer in Lapland have experienced similar problems in accessing winter food, raising concerns among ecologists and local herders.

The Arctic's patchwork of **freshwater lakes** and **rivers** is also being rewoven as the climate warms. Across a Texas-sized patch of Siberia, researchers used satellite photos to identify more than a hundred small lakes that disappeared between the 1970s and 1990s.

These were victims of thermokarsts that allowed the waters to drain out while new lakes opened to the north.

How will Greenland's fate affect ours?

The biggest chunk of land in the far north, **Greenland** has languished in frozen obscurity for many centuries.

Two areas on its chilly southwest coast were populated by Erik the Red and a few hundred other Danes and Icelanders around AD 1000.

The Western Settlement disappeared around AD 1350 during a marked cooldown, while the Eastern Settlement hung on for some 200 years more before giving way.

Since then, the island's brutal cold has kept human occupation limited to a scattering of Inuit towns and, in the last few decades, a sprinkling of seasonal research camps.

Several of these are parked on top of the colossal **ice sheet** that covers about 85% of the island.

Greenland's vast coating of ice could end up shaping where millions of people live a few generations from now.

Actually, "sheet" may be too modest a label for this icy monolith.

It extends over 1,300,000 km^2 (500,000 sq mi)—the size of France and Spain combined—and the bulk of it extends more than 2 km (1.2 mi) high.

That makes the ice itself a topographic feature on a par with the Appalachian Mountains.

Although some of the sheet presumably melted in between each pair of ice ages, scientists believe it's been at least three million years since Greenland has been ice-free.

Under the sheer weight of all its ice, Greenland's center has sunk roughly 300 m (1000 ft) below sea level.

Until the 2000s, scientists were debating whether the amount of water locked up in Greenland ice had increased or decreased over the last several decades, but data since then show that the continent is indeed losing ice mass in most years (see below).

If and when the entire Greenland ice sheet was to melt into the sea, it would trigger a truly catastrophic sea level rise of close to 6 m (19 ft).

Such an inundation would swamp coastal cities around the world and, in some low- lying areas, close the book on millennia of human history.

Thankfully, most glaciologists agree it would take many centuries for the Greenland ice sheet to fully melt.

A major ice-core analysis released in 2012 bolstered that view.

During the Eemian, the period between the last ice age and the one before that (about 130,000 to 115,000 years ago), temperatures averaged as much as 8°C (14°F) warmer than today at the location in north Greenland where the ice core was drawn (a much greater rise than occurred globally).

But it appears that only about a quarter of the ice sheet's volume was lost during the interglacial period, adding perhaps 1–2 m (3–6 ft) to global sea level, and it took roughly 6000 years for that to happen.

Although this finding offers at least a bit of short-term reassurance, it's worth keeping in mind that human-produced climate change is amplified strongly in the Arctic, so any global rise in temperature could be much larger in Greenland.

Moreover, human-caused climate change probably won't follow the Eemian script to the letter.

Greenhouse gases are accumulating far more rapidly now than they did during the Eemian, and there are small but important differences in Earth's orbit around the sun now versus then.

The 2013 IPCC report notes that if global temperatures were sustained above a certain threshold—not yet identified with confidence, but perhaps 1°–4°C (1.8°–7.2°F) above preindustrial values—the Greenland ice sheet could almost completely melt over the course of a millennium or more.

Right now, with rapid cuts in global emissions looking unlikely, we're on a trajectory that could bring global temperature well within that temperature range this century.

For Greenland's ice, the point of no return wouldn't be obvious at first.

Once under way, though, the melting would become virtually unstoppable through a set of positive feedbacks.

For example, as meltwater runs off from the top of the ice sheet, the surface would gradually lose elevation and thus encounter progressively warmer air temperatures.

This would allow for still more melting, and so on, putting the ice in a cascade toward oblivion.

Even if that scenario isn't expected to culminate for at least a thousand years, there's still cause for concern in the shorter term, because Greenland's peripheral glaciers appear to be losing ice at a faster clip than either scientists or their computer models had reckoned.

Some of the ice processes now at work aren't represented well—or at all—in computer models because of their complicated nature.

Moreover, some dramatic thaws have been occurring across Greenland as a whole.

In the summer of 2007, melting occurred across fully half of the ice sheet's surface.

The thaw was even more jaw-dropping in 2012, when locations that normally see only a few days of melting experienced two months of it.

On July 12, satellites detected wet snow across 97% of the entire Greenland ice sheet (see graphic).

Ice-core analysis shows that such widespread melts have happened only eight times in the last 1500 years.

Apart from such dramatic short-term events, the overall rate of melting in recent years tops anything evident in ice cores going back more than 450 years, and perhaps much longer.

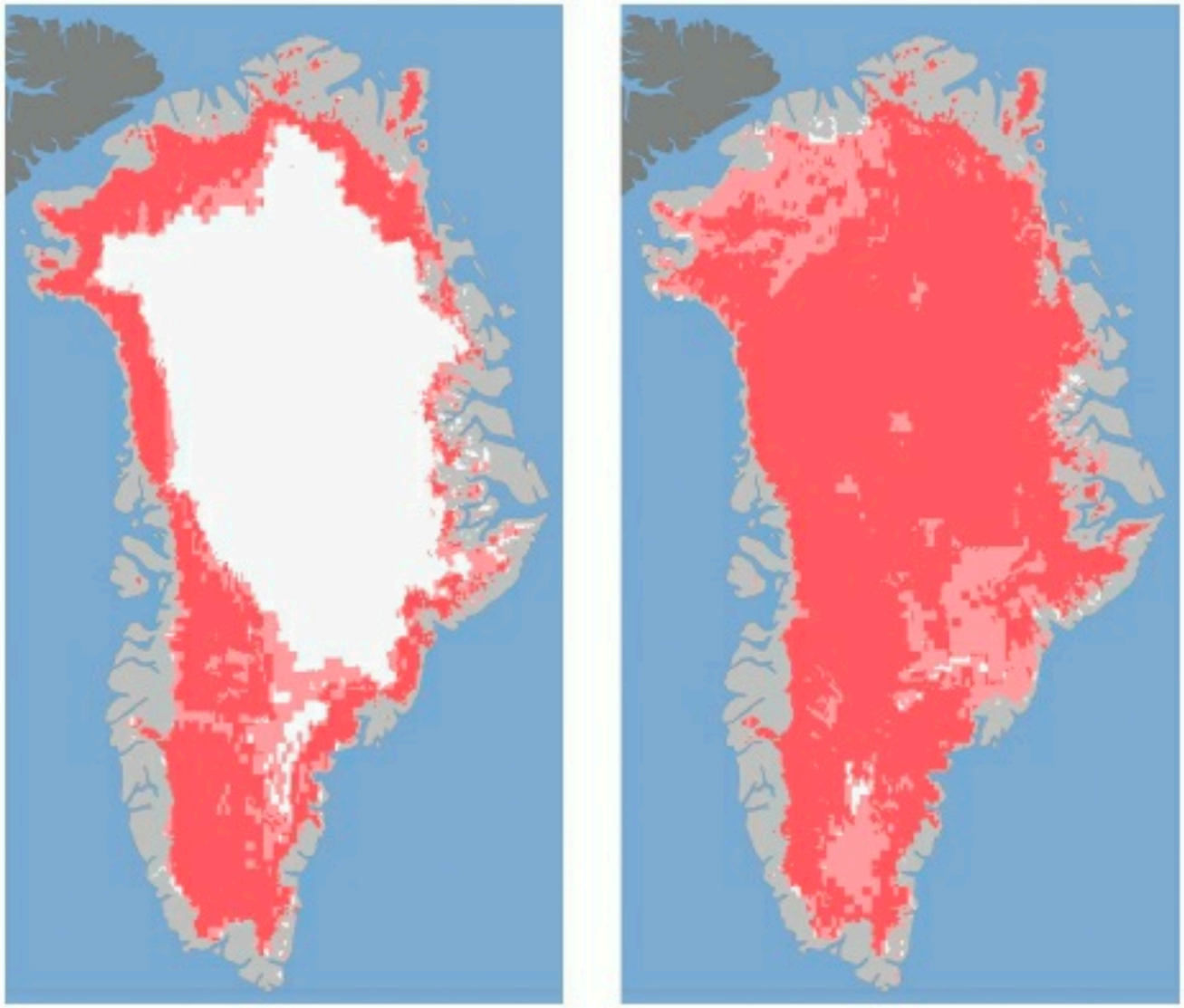
It's quite possible that the Greenland melt is already being hastened by soot from forest fires and fossil fuels landing on the ice sheet and boosting its solar absorption.

Even an increase in algae blooms atop meltwater-slathered ice may be having an impact.

Another tool for diagnosing the health of Greenland's ice sheet has been the Gravity Recovery and Climate Experiment (**GRACE**), twin ice-monitoring satellites launched by the United States with German instrumentation.

The original GRACE operated from 2002 to 2017, with a follow- on pair of satellites launched in 2018.

Analyses of GRACE data by Isabella Velicogna (University of California, Irvine) show that Greenland lost an eye-popping one trillion metric tons of frozen water from late 2006 to mid-2010.



In 2012, surface thawing of the Greenland ice sheet (pink colors) exploded in extent from about 40% of the sheet on July 8 (left) to roughly 97% on July 12 (right). Areas in light pink (probable melt) show where at least one satellite detected surface melting, while the dark pink shows areas where at least two satellites detected melt. (Nicolo DiGirolamo, SSAI/NASA GSFC, and Jesse Allen, NASA Earth Observatory)

That amount was nearly matched over the following two years alone.

According to an exhaustive 2012 analysis published in *Science* that drew on GRACE and other data, meltwater from Greenland accounted for 10%–15% of sea level rise from 1992 to 2011.

Data from 2012 suggest an even higher percentage.

The IPCC's 2013 report expressed high confidence that surface melt from Greenland will continue to increase at a faster pace than snowfall, keeping the continent's water budget in the red over the long term.

The years 2016–17 and 2017–18 threw in several curve balls, in the form of powerful Atlantic cyclones that dropped massive amounts of snow on southern and eastern Greenland.

Thanks in large part to the remnants of Hurricane Nicole in 2016—which brought record precipitation to parts of southeast Greenland—the winter of 2016–17 was the first since 2002 in which Greenland gained as much or more water in snow and freezing rain as it lost to the sea via melting and iceberg calving.

The main routes for Greenland water loss appear to be outlet glaciers, the frozen superhighways along which ice flows to the sea.

Several of Greenland's largest glaciers have been moving seaward at a snappier pace than expected over the last 15–20 years.

At the same time, so much melt is occurring at the forward edge of these glaciers that the net effect can be a retreat of that edge.

The bottom surface of the immersed ice tongue is especially vulnerable to melting because the added pressure of water lowers the melting-point temperature.

In some cases, a glacier's rush to the sea can be bolstered through a process called **dynamic thinning.**

The idea is that, as meltwater pours into a flowing glacier through moulins (deep, river-like channels), it can lubricate the base and hasten the glacier's seaward flow.

Because it has been difficult to quantify this process and incorporate it into ice-sheet models, it has remained something of a wild card.

While most research has emphasized the potential of dynamic thinning to push sea level rise well beyond standard projections, a dissenting view appeared in a modeling study in 2013 led by the University of Bristol.

It suggests that the changes in ice flow triggered by dynamic thinning, though widespread, might only add a small fraction to Greenland's total contribution to sea level rise by 2100.

This supports the idea that Greenland's contribution to sea level this century could end up being more than minor but less than catastrophic.

Because observations can't easily confirm everything going on at the base of ice sheets and glaciers, there will no doubt be much more research on this evolving topic.

The last few years have seen Greenland's largest, speediest glaciers flowing seaward more quickly, yet more erratically.

It's now apparent that these highly dynamic systems can move seaward in pulses that strengthen and weaken from year to year on top of longer-term depletion induced by climate change.

One of the world's fastest-moving glaciers is **Kangerdlugssuaq**, along Greenland's southeast coast.

Some 6% of the island's ice flows through the Kangerdlugssuaq Glacier toward the Atlantic.

This glacier moved seaward at about 6 km (3.7 mi) per year during the late 1990s and early 2000s, but more than twice as quickly—14 km (9 mi)—in 2005.

Also that year, the glacier's front edge pulled back a full 5 km (3 mi), by far its largest retreat on record. Kangerdlugssuaq then settled down over the next few years, still flowing more quickly than in the pre-2005 period, but with its forward edge actually advancing slightly.

On the west coast, the **Jakobshavn Glacier**, the largest of Greenland's ice-draining glaciers, more than doubled its seaward pace between 1996 and 2009, when it moved a total of nearly 30 km (19 mi).

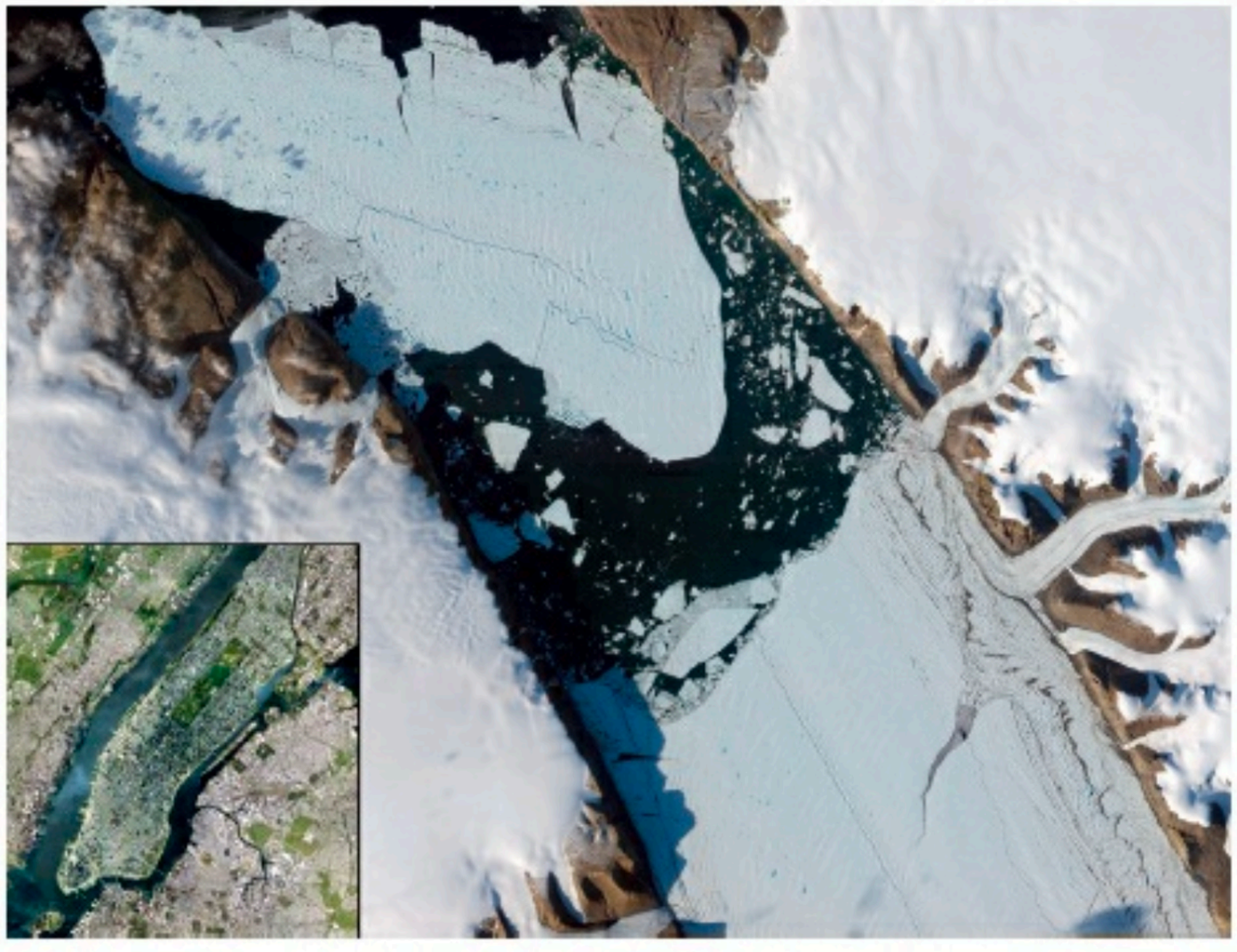
The glacier's motion slowed dramatically in 2011 but surged to a new record speed in 2013.

Calving from Jackobshavn hit a dramatic point in August 2015, when the glacier spat out enough ice in two days to cover 12.5 km^2 (4.8 sq mi) with ice as thick as a skyscraper.

There's been action in far northwest Greenland as well.

This region appears especially prone to multiyear spells of highly dynamic ice loss, with relatively quiet periods in between.

In the summer of 2010, Greenland's north coast came into the media spotlight when a gigantic "ice island" four times the size of Manhattan broke off from the **Petermann Glacier.**



The cleaving of the Petermann "ice island"—satellite image of Manhattan to scale inset. (NASA Earth Observatory/U.S. Geological Survey)

It was the Arctic's largest calving event since 1962, which also made it the largest to be portrayed through vivid satellite imagery.

Though calving is a natural process, the sheer scope of this one served as visual shorthand for the larger issue of long-term ice loss.

It's fair to say that the future of Greenland's ice sheet is one of the biggest unknowns looming over our planet's coastlines.

This uncertainty is boosted by the hundreds of less-studied Greenland glaciers for which there's little more than a decade of data on hand.

According to work from the University of Washington, the short record thus far "shows a complex time-dependent glacier response, from which one cannot deduce how the ice sheet will react in the long run to a major climatic warming, say over the next 50 or 100 years."

What about Antarctica?

Strange as it may seem, Antarctica has stayed mainly on the sidelines in the saga of global warming—at least until recently.

This vast, lonely land took center stage in the mid-1980s, when the infamous ozone hole was discovered a few kilometers up.

The Montreal Protocol and resulting actions by world governments are expected to eliminate the hole by later in this century.

Antarctica is surrounded by a vast ring of ice that waxes and wanes even more than the Arctic Ocean's ice, in part because of its overall lower latitude.

In contrast to the dramatic, high-profile decline in northern sea ice, Antarctic ice extent quietly grew by about 1% per decade from the 1970s into the 2010s.

In late 2012, it set a new high, topping 19,000,000 km^2 (even as Arctic sea ice extent fell to a new low), followed by even higher values in 2013 and 2014.

Some climate models didn't see the long-term increase coming, and many climate contrarians cited the growth as some type of natural compensation for the Arctic's ice loss.

However, the two basins are virtually independent—ice lost in one doesn't pop up in the other.

A more likely explanation is a strengthening of the winds ringing Antarctica due to both tropical variability and ozone loss; these winds tend to push sea ice outward and expand its range.

A spectacular shift occurred in 2017 and 2018, as those two years brought the two lowest minimum extents on record; 2017 also saw the second-lowest winter maximum.

Time will tell if this signals the beginning of the eventual long-term Antarctic sea ice decline anticipated by models.

A much greater concern is the state of the ice on Antarctica itself.

Almost two-thirds of the planet's fresh water is locked in Antarctica's ice sheet, poised to raise sea levels by spectacular amounts if the ice were to melt.

Fortunately, nobody expects the entire sheet to start melting anytime soon.

Temperatures in the interior of Antarctica are shockingly cold: Vostok Station, for example, has never recorded a temperature above freezing.

Glaciers flowing from the interior off the shores of East Antarctica are sensitive to multiyear temperature swing.

Some of these East Antarctic glaciers are advancing, while others retreat, but thus far there's little sign of a consistent retreat.

The main risk for now is at the continent's western end, where glaciers appear even more vulnerable than once thought and major collapses this century can't be ruled out.

It's been hard to determine exactly what's happening to temperatures across the great bulk of Antarctica, which has only a few reporting stations, especially in the interior of the continent.

But there are good reasons why parts of this frozen realm might run counter to Earth's long-term warming trend.

For one thing, Antarctica is surrounded by the vast, cold Southern Ocean, which absorbs some of the heat that would otherwise boost air temperatures across the region.

In addition, overall global warming has led to a strengthening of the Antarctic vortex, the ring of upper-level winds that encircles the continent.

This helps keep the coldest air focused over the continent and inhibits warmer air at midlatitudes from working its way southward.

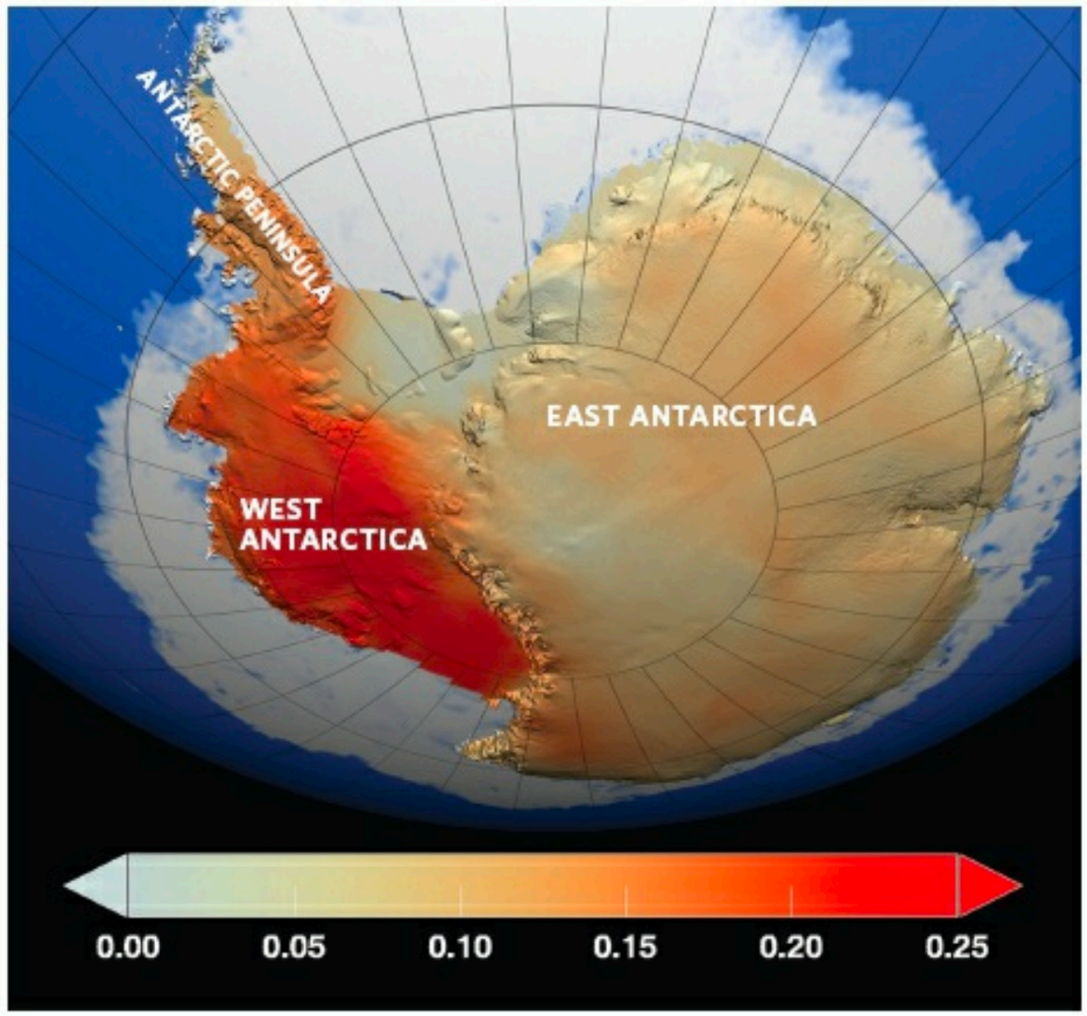
Moreover, ozone depletion in the Antarctic stratosphere over the last several decades has helped to tighten the Antarctic vortex, shifting the strongest winds poleward.

In East Antarctica, which spans the great bulk of the continent, satellite measurements of the ice sheet's temperature show cooling over some areas in recent decades, mainly during the summer and autumn.

However, at the South Pole itself, surface readings that had been cooling have recently begun to warm, leading to little net change there since the 1950s.

As opposed to its eastern counterpart, West Antarctica was solidly in the late-twentieth-century warming camp—so strongly, in fact, that it more than compensated for any cooling to the east.

The biggest changes in recent decades are taking place on the **Antarctic Peninsula**, which juts northward toward South America.



Red represents areas where temperatures increased the most from 1957 to 2006, particularly in West Antarctica, where some points warmed even more than shown here. Temperature changes are measured in degrees Celsius. (Trent Schindler, NASA Goddard Space Flight Center Scientific Visualization Studio)

Average temperatures at weather stations along the peninsula's west coast—one of the few parts of the continent that sees major thawing and refreezing each year—soared by as much as 2.8°C (5.0°F) from the 1950s through the 1990s, as rapid a warming as anywhere on Earth during that span.

Highlighting the complexity of climate variability, surface air temperatures across the peninsula then dropped by 0.5° to 1.0°C (0.9°– 1.8°F) from 1999 to 2014.

A team at the British Antarctic Survey chalked up the drop to short-term variability, perhaps goosed by the slowdown in ozone depletion noted above.

In a 2016 paper, the team reiterated that the long-term effects of climate change are expected to increasingly outweigh decadal-scale variations as the century unfolds.

During the second half of the 20th century, there was also marked warming *above* Antarctica, as revealed in a 2006 *Science* paper.

Looking at the Amundsen–Scott South Pole Station plus eight coastal stations, The researchers found that the air at a height of about 5.5 km (3.4 mi) above sea level had warmed by 1.5°–2.1°C (2.7°–3.8°F) since the 1970s, although it appears there was little warming in the decade following the 2006 paper.

It’s expected that warming will accelerate across most of Antarctica later this century, hastened by the globe-spanning effects of greenhouse gases as well as by a long-distance relationship with rising sea surface temperatures in the tropical Pacific.

The healing of the ozone hole later this century may also reduce the cooling effect of the polar vortex.

Across the interior, the warm-up may do little at first other than slowly build the ice sheet, in much the same way as Greenland’s ice sheet appears to be growing on top.

Slightly warmer air (that is, air that’s less bitterly cold) should bring more moisture with it, perhaps resulting in a boost to the tiny snowfall amounts across East Antarctica.

Overall, the net amount of ice and snow in the East Antarctic Ice Sheet appears to have grown modestly since the 1990s, though there’s been spirited debate among experts on exactly how much it’s grown (a difficult thing to measure via satellite over this data-sparse region).

Meanwhile, it’s far more evident that West Antarctica (including the Antarctic Peninsula) has been shedding ice and meltwater at an accelerating rate.

Because each method of assessing Antarctic ice has its pros and cons, scientists launched the Ice Sheet Mass Balance Inter-Comparison Exercise (IMBIE) in 2011 to obtain a firmer consensus.

IMBIE's work to reconcile dozens of studies has confirmed that the West Antarctic melt far exceeds the relatively modest gains of the East Antarctic Ice Sheet.

In a special 2018 issue of *Nature*, the IMBIE team estimated that East Antarctica gained mass from 1992 to 2017 at a roughly steady pace of 5 billion metric tons per year, a relative pittance across the vastness of that ice sheet.

At the same time, the continent as a whole was losing around 76 billion metric tons of ice per year—until the pace accelerated to an eye-popping 219 billion metric tons per year during 2012–2017.

The paper states:

"The continent is causing sea levels to rise faster today than at any time in the past 25 years".

"This has to be a concern for the governments we trust to protect our coastal cities and communities."

The Antarctic Peninsula's huge Larsen B ice shelf made headlines in early 2002 when a Delaware-sized chunk—perhaps 12,000 years old—broke off.

The adjacent Larsen C shelf made its own splash in mid-2017, releasing an iceberg almost twice as big as Larsen B's. Compelling as they are, such calving events (as with the 2010 Petermann calving in Greenland noted earlier) have only minor overall effects on sea level, since most of the ice is already in the water.

More concerning is whether such breakups might open the door to a faster seaward flow of ice from glaciers upstream.

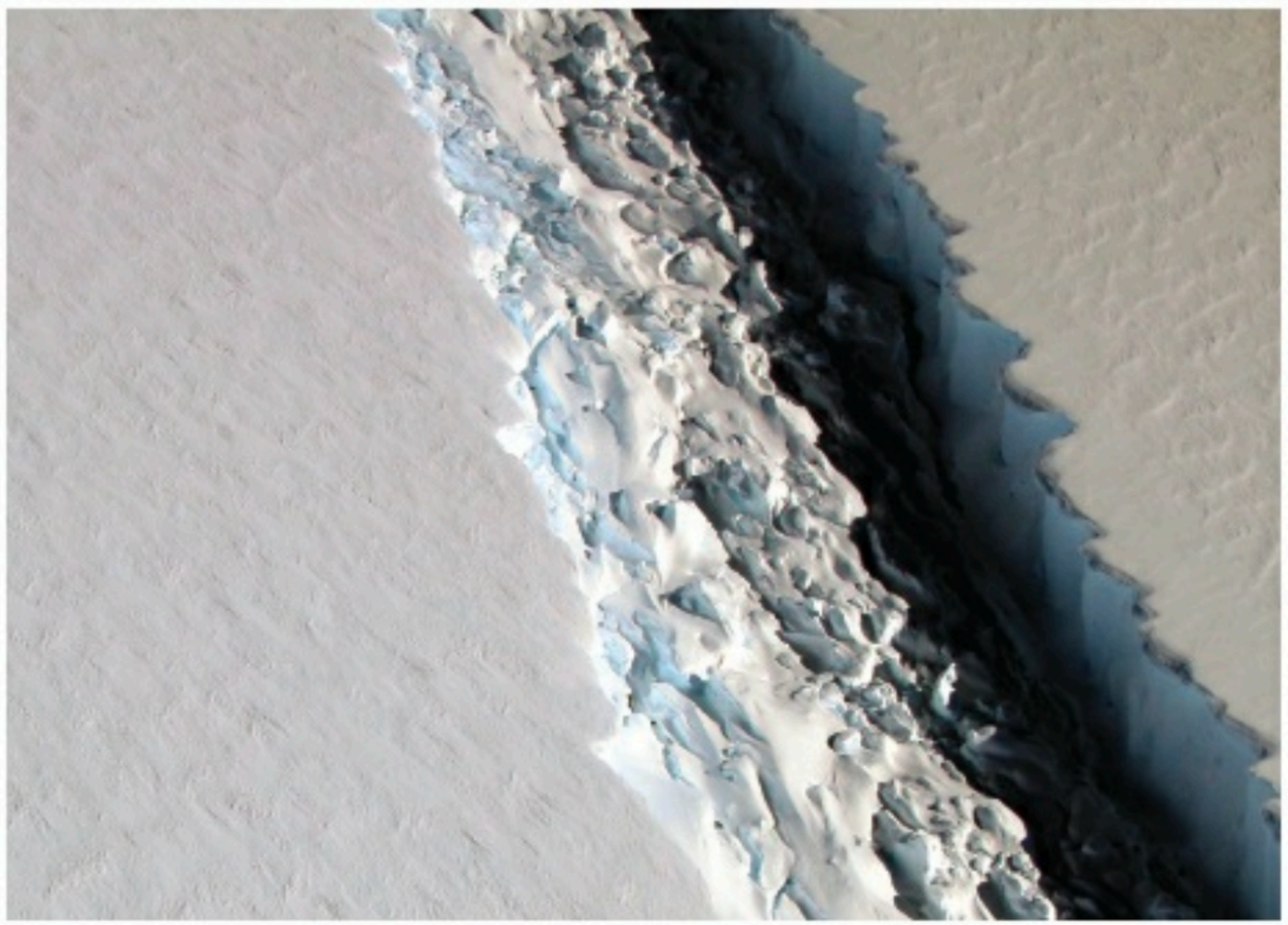
The hydrostatic force of the ocean pushes back on large ice shelves that are grounded on offshore bedrock; this process helps stanch the flow of ice from upstream.

When the shelves break up, the adjacent glacier flows more easily into the sea.

Six years after the Larsen B calving, the region's glaciers were moving roughly two to five times more quickly toward the sea.

Glaciers elsewhere in West Antarctica are also speeding up, largely because of changes in ocean circulation that surround the glaciers' offshore tongues with relatively warm water.

Interestingly, both oceanic and atmospheric shifts in West Antarctica are closely linked to the tropical Pacific, where recurrent clusters of thunderstorms can shape atmospheric circulation far afield in much the same way as El Niño and La Niña do.



The Larsen C ice shelf released a massive iceberg from the Antarctic Peninsula a few months after this rift—about 100 meters wide and 500 meters deep—was photographed from a NASA research aircraft on November 10, 2016. (NASA/John Sonntag)

As the tropics are naturally variable, their influence on West Antarctica can wax and wane for periods of years.

Thus, it's unclear how soon local changes might be enough to cause major ice shelf calving or collapse.

That said, the fact that so many West Antarctic glaciers extend into a warming sea is major cause for concern.

Glaciers on the Antarctic Peninsula are particularly vulnerable to future warming, given that the peninsula sits farther north and juts into the path of strengthening winds that bring warm air from lower latitudes.

It's a good thing that most climate models project the vast bulk of Antarctica's ice to stay onshore for at least the next century and probably much longer than that, since a complete melt would raise global sea levels by a cataclysmic 53 m (174 ft).

However, the IPCC's 2013 report notes that if a collapse of the West Antarctic Ice Sheet were to get under way later this century, it would raise the estimated upper limit of sea level rise at century's end by several tenths of a meter, causing significant impacts to coastal regions worldwide.

More recent studies brought in a worrisome new element: the risk that gigantic ice cliffs holding back West Antarctic glaciers might collapse.

This could potentially allow for several more feet of sea level rise by 2100 than captured in even the most pessimistic IPCC scenarios.

Tropics and midlatitudes: Goodbye glaciers?

Far above the lush forests and warm seas of the tropics, a few icy sentries keep watch on our planet's climate from a unique vantage point.

These are Earth's tropical glaciers, perched atop the Andes and other peaks that extend from 4600 m to more than 6900 m (15,100–22,600 ft) above sea level.

Farther north, in the midlatitudes, thousands of small glaciers and a few larger ice caps are sprinkled across parts of the Rockies and Sierra Nevada, the Himalayas and other Central Asia ranges, and the Alps.

These scattered outposts of ice aren't very good at hiding the strain they're under.

For several reasons, including their year-round exposure to warm air and their typically steep-sided structure, glaciers at lower and middle latitudes often respond especially quickly to climate change.

A melting trend that kicked in during the mid-1800s has accelerated in recent decades across many of the tropical and midlatitude glaciers monitored by scientists.

By later in this century, some of these glaciers could be little more than history.

Like rivers in slow motion, glaciers are fluid entities, with fresh water flowing through them.

Snow accumulates on top and gets pushed downward and outward by the snowfall of succeeding years.

For a healthy glacier, the change in its total ice mass per year—its **mass balance**—is either zero or positive.

Today, however, most of the world's tropical and midlatitude glaciers have a mass balance seriously in the red. In short, they're shrinking and thinning.

It isn't just melting that spells doom for ice.

Many glaciers erode largely through a process called **sublimation.**

When dry air flows over ice, the moisture in the ice can pass directly from its frozen state to its vapor state without any intervening liquid form.

Sublimation can occur even when the air is below freezing, and it's the dominant process of glacier loss in some areas.

In general, though, melting is the main culprit.

It takes eight times less energy for a molecule of ice to melt than it does for it to sublimate, so anything that pushes a glacier toward conditions favorable for melting can erode the glacier far more efficiently than sublimation.

Under the right weather conditions, direct sunshine can melt glacial ice at temperatures as cold as −10°C (14°F), and of course a reduction in snowfall can take its toll on a glacier even if average temperatures stay constant.

It's the unique mix of conditions at each location that shapes the behavior of tropical and midlatitude glaciers.

Those who study glaciers fear that this complex picture sometimes gets lost in the scramble to interpret glacial loss as a simple product of warmer temperatures.

Still, they agree on the main point: with a very few cold-climate exceptions, climate change is eroding lower-latitude glaciers at a dramatic pace.

Trouble toward the equator

Thanks to a century of visits by adventurers, photographers, and scientists, the ice cap on top of Africa's **Mount Kilimanjaro** is the most widely publicized of the tropics' disappearing glaciers, but others tell a similar tale.

Not far away from Kilimanjaro, **Mount Kenya** lost 7 of its 18 glaciers over the last century; its long-studied Lewis Glacier spread over a mere 0.1 km2 (0.04 sq mi) in 2010, only half of what it covered in 1993.

On the Indonesian island of Irian Jaya, the nation's highest peak—Puncak Jaya—lost two of its five glaciers since 1989, and ice now covers less than 20% of the area it did in 1950.

It's possible that both the Lewis Glacier and Puncak Jaya's ice will be virtually gone by 2050.

All of these are mere slivers next to the ice atop the **Andes** of South America.

With their string of peaks and high valleys, many extending above 5000 m (16,000 ft), the Andes hold more than 99% of the world's tropical ice.

In Peru, north of Lima, more than 700 glaciers dot Peru's **Cordillera Blanca**, or White Mountains (a name deriving mainly from the range's granodiorite rock, rather than the ice it sports).

Farther south along the Andes, not far from Bolivia, sits the vast, high **Quelccaya** ice cap, sprawling across a city-sized area yet continuing to erode at its margins.

Peru's tropical glaciers are far more than playgrounds for climbers and backdrops for photos: they carry both spiritual and practical significance.

The ice melt provides water and hydroelectric power that's vital during the dry season for much of the country, including up to 80% of the water supply for the perennially parched Pacific coast.

Lima spends months shrouded in clouds and drizzle, yet it averages less than 20 mm (0.8 in.) of rain a year, which makes it the world's second-largest desert city behind Cairo.

Lima’s 10 million residents now depend on glacial ice that is measurably receding.

So serious was Lima’s risk of losing power that the city shifted dramatically in the 2010s from hydropower to natural gas.

As for the potential water shortage, there’s no long-term solution in sight, though engineers have suggested a tunnel through the Andes.

What might be called “The Case of the Disappearing Tropical Glaciers” is the latest chapter in an ancient tale.

ice has waned and waxed since the end of the last ice age, sometimes shaping the destiny of whole civilizations.

The glaciers’ most recent decline began in the mid-1800s, toward the end of the Little Ice Age but before human-produced greenhouse gases started to rise substantially.

It’s likely that a blend of forces, possibly including a slight rise in solar output, got the trend started.

However, in recent decades greenhouse gases appear to have taken the reins, with assistance in the Andes from El Niño, the periodic warming of the eastern tropical Pacific.

When an El Niño episode is in progress, temperatures generally rise across the Andean glaciers, and summer rainfall and snowfall decrease.

El Niño events were unusually frequent from the 1970s through the 1990s, hastening melt across the Andes.

Today, even the most diehard climate change minimizers are hard-pressed to deny that tropical glaciers are vanishing before our eyes.

The Andes offer especially vivid examples of low-latitude ice loss, particularly in some of the most accessible parts of the ice (which often happen to be the lowest lying and the most tenuous to begin with).

For half a century, a ski lift took adventurers onto Bolivia's Chacaltaya Glacier, billed as the world's highest developed ski area (the summit is at 5345 m/17,530 ft).

But the lift wasn't used after 1998, and by 2009 the glacier had vanished.

The signs of glacier decline are no less dramatic as you move poleward. Patagonia is home to South America's largest ice fields, as well as glaciers that cascade all the way to sea level.

As with Antarctica and Greenland, the calving of glaciers that spill into the sea appears to be uncorking the flow of ice from higher up.

NASA teamed up with Chile's Centro de Estudios Científicos (Center for Scientific Studies) to analyze the Patagonian glaciers in 2003.

Combining space shuttle imagery with topographic data, they found that the rate of glacial thinning had more than doubled between 1995 and 2000, making it one of the fastest thinnings on the planet.

The melting has proceeded into the 2000s, with even the highest glaciers of southern Patagonia now showing signs of thinning.

So quickly is the Southern Patagonia Ice Field losing its ice that, from 2003 to 2006, the earth beneath it rose by nearly 12 cm (4.8 in.).

It's the most rapid rebound from the weight of glacial ice measured by GPS sensors anywhere on the planet.

Several factors may be involved in Patagonia's ice loss, including higher temperatures, reduced snowfall along the Andes' western flanks, and increasingly unstable glacier behavior, such as dynamic thinning and enhanced calving.

Things are also changing quickly in the midlatitudes of the Northern Hemisphere.

Glacier decline is obvious to millions in and near the **Alps,** whose ice has lost roughly half of its surface area and overall mass since 1850.

Glaciers across the Alps took a big hit as industrialization of the late 1800s deposited sunlight-absorbing soot on the ice surface.

This caused major retreat from 1850 to 1930 even as regional temperatures cooled.

Over the last several decades, warming temperatures and reduced winter snow have joined to foster a second round of major melting.

At heights of 1800 m (5900 ft), parts of the French Alps have experienced winters up to 3°C (5°F) warmer than those of the 1960s.

During this period, glacier coverage across the French Alps has decreased by more than 25%.

Some of the Alps' smaller glaciers lost as much as 10% of their ice mass in the warm summer of 2003 alone.

A stunning “warm wave” at the end of 2015 brought unprecedented mildness for early winter in the eastern Alps, with December temperatures averaging above freezing at elevations of up to 2500 m (8200 ft).

A linear extrapolation from recent trends indicates that some 75% of Switzerland’s glaciers could be gone by the year 2050, with perhaps 90% lost by 2090.

Any of these trends could accelerate or decelerate, of course, though there’s no particular reason to expect the latter.

In Montana, **Glacier National Park**, one of the jewels in the U.S. park system, has lost nearly 80% of the 150 or so glaciers that it had in 1850 (counting only those that span at least 25 acres each).

The others are expected to disappear this century, perhaps in the next several decades.

One observer has already suggested changing the park’s name to Glacier Moraine National Park, a title less likely to become obsolete.

The loss of both ice and permafrost are major issues for the central Asia’s highest mountains —the **Himalayas**, the vast **Tibetan Plateau**, and a number of nearby ranges.

Together, these are home to the largest nonpolar set of ice caps and glaciers on Earth: more than 100,000 in all, covering an area the size of Pennsylvania (about half of that on top of the Tibetan Plateau).

Across this region, sometimes dubbed the “third pole” of Earth, multiple threats to the ice are emerging.

Temperatures across the heart of the Himalayas (the Hindu Kush region) have risen at roughly the global rate since 1900, but with substantial multi-decade ups and downs, according to a 2017 analysis led by the China Meteorological Administration.

Most glaciers across the region are shrinking and losing mass, in some cases at an increasing pace.

Moreover, the amount of sun-absorbing black soot deposited on Tibetan glaciers has grown rapidly since the 1990s, a by-product of Asia’s growing population and burgeoning industry.

Because much of the Himalayan ice lies in thick swaths across broad valleys, it’s likely that the glaciers will thin vertically more quickly than they shrink horizontally, at least for the time being.

The possibility of increased glacial runoff poses problems for flood and water management downhill, especially toward the central and eastern Himalayas.

While lowlands may be able to benefit from stepped- up hydroelectric power, flood risks could grow in the next few decades.

The biggest threat is from what's known as glacier lake outburst floods, where natural moraine dams give way under the pressure of increased meltwater.

It's conceivable that the Himalayan glaciers might someday retreat enough to affect streamflows and related agriculture, but the annual monsoon is a more critical element in supplying water to much of South and East Asia.

Certainly, there's no real risk of losing these glaciers by the 2030s, an erroneous conclusion that crept into the 2007 IPCC assessment.

However, the timing of snow and glacier melt might change as the climate warms, and that could influence water supply during the dry premonsoonal months.

Even without such complicating factors, the sheer growth in population across this part of Asia looks set to strain water resources.

The picture is a bit more nuanced farther north, in the sub-Arctic lands of North America and Europe.

Certainly, **Alaska's** southern glaciers are eroding quickly, especially those that extend into the Pacific Ocean.

The vast **Columbia Glacier**, at the end of Prince William Sound, has lost roughly half its thickness and volume since 1980, and its front edge has moved more than 20 km (12 mi) upstream.

In 2007, the glacier's end point lost its grounding on the seafloor and became a floating ice tongue, the first such transformation observed in detail by scientists.

The glacier now sends huge icebergs into the sound (which, in turn, are convincing cruise ships to stay much farther from the glacier's edge).

The 40-year retreat of the Columbia's leading edge—now a matter of simple physics as much as climate change—is expected to slow in the 2020s, at least temporarily, once the front edge reaches shallower water near the shoreline.

In contrast, the sub-Arctic reaches of Europe hold some of the most publicized exceptions to global glacier retreat.

Several of **Norway's** coastal glaciers grew during the 1990s: for example, the leading edge of **Briksdalsbreen** advanced by more than 300 m (1000 ft).

However, these advances were easily explained by the strong westerlies and enhanced snowfall that prevailed during that decade along Norway's west coast, thanks largely to the positive mode of the North Atlantic Oscillation.

With negative NAO indices and drier conditions becoming more common in the 2000s, Norway's coastal glaciers again started to decline.

Briksdalsbreen retreated some 500 m (1600 ft) between 1999 and 2011.

Farther inland, virtually all of the glaciers in interior Norway have been shrinking for decades.

Many of those who deny human-produced warming have pointed to the growth of specific glaciers.

However, since climate change isn't a globally uniform process, it's easy to see why a few individual glaciers scattered across the globe have grown or held their own in recent decades, especially in places where warming hasn't yet pushed increased precipitation from snow toward rain.

The bigger picture becomes clear in global analyses published annually by the World Glacier Monitoring Service, which keeps track of 37 reference glaciers across 10 mountain ranges with data extending back to 1980.

Across the period 2015–2016, all but one of the 37 reference glaciers lost mass.

For all of those reference glaciers as a group, the mass balance has been negative in each year since 1980, with the deficit trending larger from decade to decade.

More than melting?

It's impossible to quantify just how much of the blame for the overall retreat in low- and midlatitude glaciers can be pinned on human activity, as glacier behavior today is partly a result of conditions present many years ago.

That said, tropical ice appears to be one of the most reliable of all witnesses to human-induced change.

For starters, the wild temperature swings often found at higher latitudes are muted near the equator, especially as you go up in height.

There's also little variation as you go from west to east around the globe.

As a result, the average readings vary only about 1.5°C (2.7°F) across the entire equatorial belt at glacial heights of about 6 km (3.7 mi), and there's little change across seasons as well.

This means that even a small trend should stand out more noticeably than it would in midlatitudes.

On the other hand, the tropics aren't the best-sampled part of the world, meteorologically as well as glaciologically.

A few weather stations are cropping up at the highest tropical altitudes, but observations remain sparse, both on the ground and aloft.

As a result, scientists are debating exactly how much temperatures across the higher altitudes of the tropics have risen.

Thus far, the jury seems to be settling on a warming somewhat less than the global surface average, but still enough to tip glaciers toward melting.

Across eastern Africa, the scanty data show little if any warming trend, which means that the glaciers atop Kilimanjaro may be a doomed relic of a wetter period rather than a victim of human-produced climate change.

Aside from temperature, there's a range of other processes that will help shape the future of tropical and midlatitude glaciers.

These can differ from region to region and even on the scale of individual mountains, which leads to a globe-spanning patchwork of varied ice loss sprinkled with local advances in a few midlatitude areas.

- **Sunlight** can directly melt glacial ice even when the air remains well below freezing (as noted previously). Local increases in cloud cover can either enhance or detract from ice loss, depending on how they affect sunlight as well as the longwave (infrared) radiation that clouds send toward the ground.
- **Moisture** in the air can also help or hinder glacier growth, depending on location. In normally dry parts of the tropics, where glaciers typically sit at heights well below freezing all year round, any increase in dry conditions can allow for more snow and ice to slowly leave glaciers via sublimation The melting process is different in the midlatitudes, where

temperatures often rise above the freezing point in summer. If those balmy temperatures came with enhanced moisture and cloudiness, the clouds could send more infrared energy toward glaciers, thus hastening their melting.

As glaciers continue to shrink, their predicament gets increasingly dire.

If a glacier is relatively flat, then the ratio of its surface area to its ice mass goes up.

This gives the atmosphere more area to work on, causing the glaciers to change state more quickly.

Perhaps the only way for glaciers outside the poles to avoid death by warming would be through a boost in snowfall.

That doesn’t appear to be happening on a large scale, despite a general increase in atmospheric water vapor worldwide.

The snowfall on glaciers atop midlatitude mountains is fed by strong, moist, upper-level winds, such as those near the west coasts of North America or Norway.

Glaciers in these areas may stand the best chance at survival, as the winds reaching them could bear more moisture in years to come.

Even then, temperatures would need to stay well below the freezing mark for the bulk of the year.

At the elevations where temperatures are marginal, a small change can make a big difference.

One study estimates that a 1°C rise in average temperature at the glacier’s equilibrium line would need to be counteracted by a 25% increase in average snowfall—a massive amount when it comes to long-term climatology.

7 - Oceans —- A PROBLEM ON THE RISE

Glaciers calving like there’s no tomorrow, sea ice dwindling to patches—it’s easy to find drama in the warming of our planet’s coldest waters.

But only a small part of Earth's oceans ever produces ice.

What about tropical and midlatitude waters, the ones that together span more than half the planet?

If appearances can deceive, then these oceans are master tricksters.

They might not look much different to the naked eye than they did a century ago, but they've gone through quite a transformation—warming up and growing more acidic, among other changes.

It appears they've hidden the true impact of our energy choices quite skillfully, absorbing much of the heat and CO_2 that would otherwise have gone toward warming the atmosphere.

Many scientists suspected as much for years, but the first solid numbers came from a NOAA team.

After assembling a variety of deep-ocean measurements for the period 1948–1998, they announced that every major ocean exhibited warming down to at least 1000 m (3300 ft) and that the top 300 m (1000 ft) had warmed globally by about 0.3°C (0.54°F).

You wouldn't be able to tell the difference in temperature by taking a quick dip, but the total heat added to the oceans represents as much as 90% of the energy trapped by human- produced greenhouse gases in those 50 years.

Water expands as it warms, and the temperature increase translates to a rise in sea level of about 25 mm (1 in.).

Again, that may not sound like much, but, as we will see, it's on top of the sea level rise from all other sources.

The oceans' heat content has only continued to increase since that landmark analysis, with the warming now confirmed at depths of at least 2000 m (6600 ft) and the warming rate increasing with depth.

The undersea storage of vast amounts of heat has serious implications for humanity's future.

Although the topmost layer of the ocean stays in balance with the atmosphere over timescales of a month or longer, the much deeper layer below the **thermocline** is more insulated.

This means the heat it slowly absorbs from above will take a long time to work its way back out.

Even if greenhouse gases magically returned to their preindustrial levels tomorrow, it would take many decades for the heat tucked away in the deep oceans to work its way out of the climate system.

This idea of “climate commitment” is a key theme in research on sea level rise.

Some studies have estimated that if global greenhouse gas emissions had leveled off in 2000 (which they didn’t), an additional warming of at least 0.5°C (0.8°F) beyond that year’s global average would be guaranteed from twentieth-century emissions alone.

Of course, as climate change unfolds, the oceans will do far more than absorb and release heat.

Ocean-driven storms, primarily tropical cyclones, are showing signs of strengthening as their fuel source heats up.

Sea levels are rising, too.

Much of this is due to the expansion of ocean water as it warms, as noted above, and over time the rise will be increasingly enhanced by glacial melting.

That melting, in turn, will shape ocean circulation in ways that push sea level even higher in some favored locations, as we’ll see below.

The uneven pattern of ocean warming may influence how, where, and when a variety of ocean–atmosphere cycles unfold.

This could spell the difference between a life of plenty and a life of poverty for many millions of people living close to the land in drought-or flood-prone areas.

The oceans are also gradually losing their natural alkalinity as they absorb carbon dioxide, a shift that threatens many forms of marine life.

The warming and acidification of the world's seas, together with reduced oxygen in some waters due to changes in circulation, has led to a bleak four-word summary of our oceans' future: hot, sour, and breathless.

All of this has made oceans one of the most compelling pitch points for climate activists.

The very real possibility of islands disappearing under the sea does make for a potent image, and yet the image isn't as crystal clear as one might think.

We still have limited data on the three-dimensional reality of oceans: how warm they are, where and how fast their currents are flowing, and so forth.

Although huge changes are clearly unfolding across the world's oceans, there remain big questions to be answered.

From sticks to satellites: Measuring sea level

Mean sea level (MSL), the average height of the ocean surface at a given point, sounds as if it should be an easy enough thing to calculate.

Subtract the astronomical tides, factor out the storms that come and go, and voilà—you've got your average.

In truth, however, MSL is enormously difficult to nail down.

For starters, the tides depend on the locations of the sun and moon within a set of cycles that takes 18.6 years to fully repeat itself.

Even when you factor this out, the underlying MSL isn't the same everywhere, but varies across different sections of the open ocean by as much as 50 cm (20 in.), depending on the surrounding topography and the warmth of the underlying waters.

The effects of global warming complicate things further.

The simplest and oldest method of measuring MSL relative to some point along the coast is through basic **tidal gauges**—essentially, sticks in the sand.

A technological leap came in the 1800s with **wells** that used floating gauges to factor out the effects of waves.

Similar techniques are still used at many coastal stations, though they've been joined lately by far more precise methods.

Modern gauges measure the time it takes sound waves to bounce off the water's surface and back, converting that into distance and then to sea level height.

Radar altimeters on satellites work in a similar way, typically sending microwave energy from space to the sea surface—the open ocean as well as coastal areas.

One key advantage of satellite-based sensors, when they're coupled with GPS data, is that they provide a measure of MSL that's independent of any land-based feature.

The **TOPEX/Poseidon** satellite, a collaboration between the U.S. and French space agencies, began gauging ocean heights across much of the world in 1992, ending its long run in 2006.

Its successor, *Jason-1*, scanned oceans from 2001 to 2013; *Jason-2* took to the skies in 2008, and *Jason-3* launched in 2016. *Jason-2* and *Jason-3* measure the height of virtually all of Earth's ice-free ocean surface with a vertical resolution as fine as 2.5 cm (1 in.).

Even sharper-eyed is **GRACE (Gravity Recovery and Climate Experiment.**

This U.S.–European satellite system, launched in 2002, delivered data through 2017, with a follow-on mission, GRACE-FO, launched in 2018.

The system tracks MSL with an error of less than 1 mm (0.04 in.).

Because it measures subtle changes in the gravitational tug exerted by the oceans as well as by Earth itself, GRACE helps scientists sort through the small-scale continental adjustments that may influence sea level.

Those geological tics can have a surprisingly big influence on sea level.

Coastlines rise and fall with the clashing of tectonic plates.

In some places, such as parts of Alaska and Japan, the motions are so pronounced that tidal gauges can't be used as a reliable index of long-term sea level change.

Other coastal locations, such as Venice and New Orleans, are plagued by gradual subsidence.

Still other regions are steadily rising, on the rebound after being squashed beneath kilometer-thick ice sheets for thousands of years.

In some regions, such as the Gulf of Bothnia (bordering Sweden and Finland), the rise in MSL due to ocean warming is obscured by a much larger rise in the coastline itself due to postglacial rebound, much like a piece of dough springing back slowly from the push of a baker's thumb.

The bounce-back from melting ice can be as much as 10 mm (0.4 in) per year, enough to more than cancel out sea level rise and produce a net drop in sea level from the perspective of some coastal towns.

Balancing the sea level budget

All of these various ups and downs have to be considered when evaluating changes in sea level.

With this in mind, the Intergovernmental Panel on Climate Change has done its best to pool the data and calculate global sea level rise.

The 2013 IPCC report estimated the average yearly rise in MSL from 1901 to 2010 at between 1.5 and 1.9 mm (0.059–0.075 in.) per year.

The middle of that range, 1.7 mm (0.067 in.), corresponds to a sea level rise over 109 years of about 19 cm (7.5 in.).

It may seem as if the IPCC set a fairly wide margin of uncertainty for something that's already happened.

This is in part because, for many years, the factors known to raise sea level didn't add up to the actual gauge-measured rise since 1900.

The Levitus study, noted above, acknowledged the problem.

It proved that the oceans had warmed deeply and dramatically, but it also concluded that the ocean warming was responsible for only 25%–33% of the gauge-observed rise in MSL, or roughly 0.5 mm per year.

Another 25%–33% was attributed to melting of land-based ice.

That left the remainder—between 34% and 50% of the observed century-scale rise in global MSL—unaccounted for.

It now appears that the twentieth-century contribution from the melting of glaciers and the Greenland ice sheet may be larger than earlier thought, but still not enough to close the gap.

That leaves Antarctica, where pre-satellite data are especially scarce, as the biggest question mark.

A team from the University of Reading/Met Office) concluded in a 2013 analysis that water from the Antarctic ice sheet could easily account for the remaining unexplained fraction of global MSL rise.

Still, the group noted that "[a] complete explanation remains to be achieved."

The pace of the sea's rise has increased since the 1800s, but with some noteworthy ups and downs along the way.

Roughly in line with global air temperature, the rate of MSL rise appears to have increased from the 1920s to the 1940s, then slackened until the 1980s before increasing once more.

Since the 1990s, scientists have been able to measure all of the key terms in the sea level budget, thanks to satellite monitoring of ice sheets and the help of Argo, a global network of ocean probes that sample conditions from the surface to depths of more than 1500 m (4900 ft) every few days.

The newly complete data suggest that global sea level has risen at almost twice the rate of its twentieth-century average, or around 3 mm (0.12 in.), since the early 1990s.

What's more, the pace of sea level rise showed a gradual acceleration in that period—enough to more than double the rise one would expect by 2100 if the rate were to hold steady, as confirmed in a 2018 analysis led by Steven Nerem (University of Colorado Boulder).

It's too soon to know how much of this acceleration might be tied to multi-decade variability.

Now that MSL is being tracked by the satellites mentioned above, it's possible to detect monthly and yearly trends with more precision, which has illuminated significant spikes and dips.

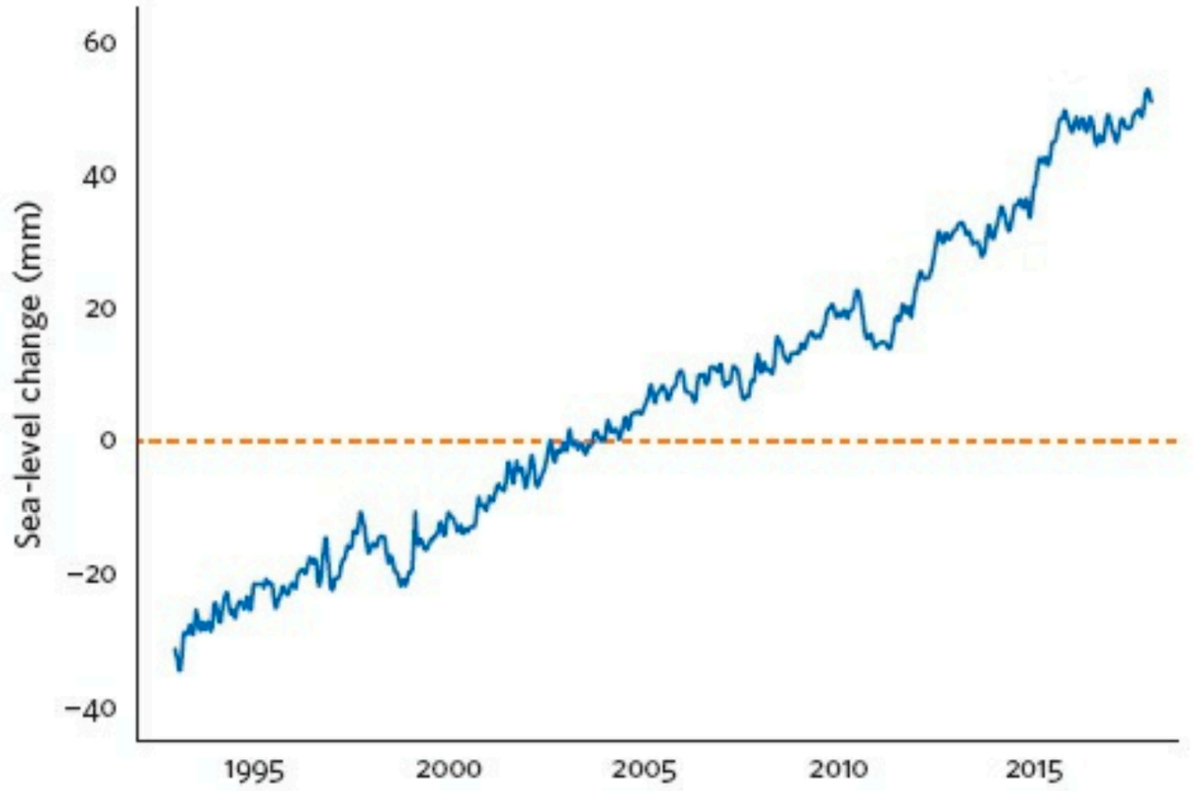
For example, the pace of rise slowed without explanation between about 2005 and 2008, then picked back up.

In 2010–2011, MSL actually dropped by 7 mm (0.3 in.), raising the eyebrows of experts.

An analysis from NCAR found the culprit: huge amounts of water vapor that had evaporated from the sea were squeezed out of the atmosphere by a rare confluence of weather patterns and dumped into Australia's vast, low-lying interior, where rainwater is partially blocked from flowing to the ocean.

This process alone was enough to bring MSL down temporarily.

Once those patterns shifted and Australia's record-setting torrents ended, global sea levels resumed their longer-term climb (see graphic).



Sea level rise varies from month to month and year to year, but the long-term trend since 1993—shown here from satellite-derived data—is unmistakable, averaging about 3.2 mm per year. Over that period, the rate has accelerated each year by about 0.08 mm per year. The normal seasonal cycle in sea level has been removed from this graphic.
(Sea Level Research Group, University of Colorado Boulder)

How high will the sea get?

Assuming humans continue to burn fossil fuels, it's a safe bet that sea level will continue to rise over the long term.

But by how much?

And what does this mean for people on or near the coast?

As we'll see below, the amount of sea level rise is liable to vary from region to region, though the globally averaged value is still a useful and important index.

Another thing to keep in mind is that such projections are typically calculated with the end of the twenty-first century as a benchmark, but some effects of sea level rise will be evident much sooner, and seas are expected to keep rising well into the twenty-second century (and probably far beyond it).

The 2013 IPCC report assigned likely ranges of sea level increase for each of four potential pathways of greenhouse gas increase.

The numbers below show how much MSL is projected to rise by 2081–2100 compared to 1986–2005.

- RCP2.6: 0.26–0.55 m (10–22 in.)
- RCP4.5: 0.32–0.63 m (13–25 in.)
- RCP6.0: 0.33–0.63 m (13–25 in.)
- RCP8.5: 0.45–0.82 m (18–32 in.)

Large as they are, these ranges actually suggest a gradual narrowing of what scientists believe could happen.

This reflects improvement in the models that depict temperature rise, ocean expansion, ice melt, glacier behavior, and other key factors.

The lowest-end projection is considerably higher than it was in the IPCC's 2001 assessment (0.09 m, or 3.6 in.), based in part on the increased rate of MSL rise since the 1990s.

Assuming this rate holds steady or grows, at least a few inches of global rise appear virtually certain.

At the other end of the spectrum, the top of the highest-end projection is lower than it was in 2001, but just slightly (0.88 m, or 35 in.).

Note that the differences between 2001 and 2013 also reflect changes in scenarios, so this isn't a totally apples-to-apples comparison.

There's been vigorous debate among climate scientists on whether MSL rise by 2100 could eclipse even the top-end IPCC estimates.

One of the key points of contention is which type of model produces the best results: the complex, **process-based models** that aim to directly simulate atmosphere, ocean, and ice, or an alternative set of **semi-empirical models**, which extend past behavior into the future by statistically relating increases in air temperature to sea level rise.

Process-based models have been the favored tool for years, and their performance has steadily improved, but there's an appealing simplicity to the semi-empirical approach.

After all, simply extrapolating recent trends would yield more than a foot of MSL rise this century, and it seems plausible that a warming world could accelerate those trends.

The schism between these two approaches broke open following the release of the 2007 IPCC assessment.

That report's lower-than-expected MSL ranges excluded some aspects of rapid dynamic thinning of ice sheets that weren't considered well enough understood to incorporate in the projections.

The IPCC did note that these factors could add another 100–200 mm to the top of the official range and stated that "larger values cannot be excluded."

During the same week the IPCC outlook went public, it was joined by a *Science* paper from the Potsdam Institute for Climate Impact Research.

Using a semi-empirical model, they found that if sea level rise in the twenty-first century stays proportional to the increase in global air temperature (a relation that held fairly firm in the twentieth century), then an MSL boost of 500– 1400 mm (20–55 in.) above 1990 levels could be on tap by century's end.

The IPCC explicitly acknowledged the ongoing debate in its 2013 assessment. "Many semi- empirical model projections of global mean sea level rise are higher than process-based model projections (up to about twice as large), but there is no consensus in the scientific community about their reliability and there is thus low confidence in their projections," stated the report.

Going forward, we can expect advances in both types of modeling, and the ocean's behavior itself should help reveal which approach might ultimately prove more accurate.

One of the main uncertainties that could yield higher-than-expected MSL rise this century is dynamic ice behavior, especially in the vast West Antarctic Ice Sheet (WAIS).

Apart from Greenland and Antarctica, it's widely accepted that other glaciers and ice caps will continue to melt at a steady or increasing pace.

A major 2011 analysis led by the University of British Columbia concluded that glaciers could add anywhere from 0.09 to 0.16 cm (3.5–6.3 in.) to MSL by 2100.

Meanwhile, Greenland will be adding an increasing amount of water through surface melting of its huge ice sheet.

However, this melting should increase at a pace more or less in line with regional temperature, so a catastrophic release of meltwater isn't expected over the next 100 years.

And although some East Antarctic glaciers are accelerating toward the sea, rising temperatures over the frigid interior should lead to increased snowfall, so this part of the continent may actually gain ice and thus trim a small amount from the expected global rise in MSL rise over the next few decades.

It’s the potential for sudden, large-scale ice loss from WAIS that’s the greatest concern for many researchers.

The land beneath much of WAIS actually slopes downward as you move inland; the sheet’s leading edge extends into the ocean via tongues and shelves that serve to buttress the ice behind it.

If the waters adjoining these extensions of oceanic ice warm up, or if shifts in ocean circulation happened to import warmer water, the buttresses might erode, which would potentially allow vast amounts of ice to flow seaward.

The IPCC cited this concern as the major asterisk in its 2013 review of sea level projections: “Based on current understanding, only the collapse of marine-based sectors of the Antarctic Ice Sheet, if initiated, could cause global mean sea level to rise substantially above the *likely* range during the 21st century.”

The panel added—but only with medium confidence—that such an event would add no more than several tenths of a meter to the twenty-first–century MSL rise.

Over the longer term, a full collapse of the WAIS could raise sea level by more than 3 m (10 ft).

A new and disturbing wrinkle emerged in the mid-2010s with research led by the University of Massachusetts Amherst and Penn State University.

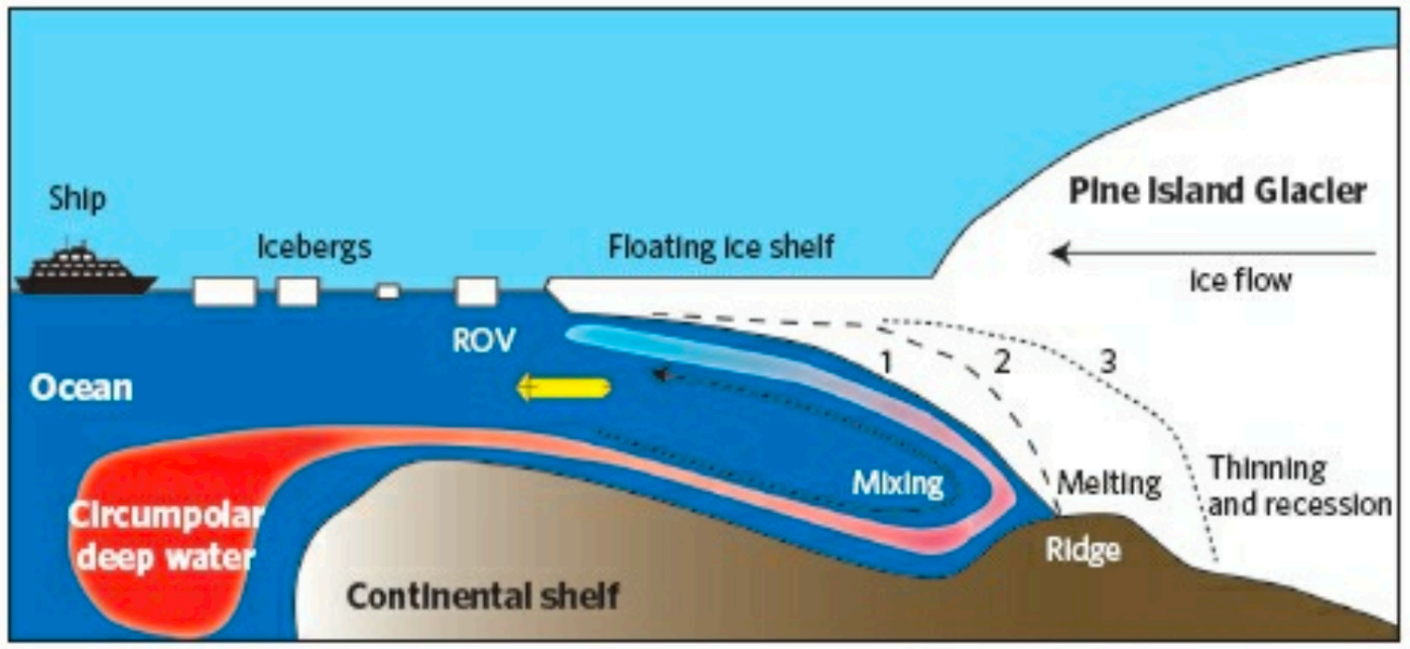
Incorporating new perspectives into simulations of past climate, the two found that the mammoth cliffs where West Antarctic glaciers meet the sea could play a much bigger role in sea level rise than earlier thought.

Because most of a marine ice cliff sits beneath water, it's prone to melting from below.

As a heat-stressed glacier retreats toward the Antarctic interior, the remaining cliff edge gets taller, adding to the sheer physical stress on its outward face.

If the marine ice cliffs at just one of these glacial outlets—Thwaites, about 150 km (90 mi) wide—were to collapse, it could set off a sequence of West Antarctic ice loss that leads to as much as 3 m (10 ft) of global sea level rise.

Building on this new paradigm, subsequent studies warned that society ought to prepare for sea level rise that could be on the order of 1–2 m (3–6 ft) by 2100 in a high-emissions scenario.



Vast amounts of meltwater are leaving the West Antarctic Ice Sheet via the Pine Island Glacier and nearby Thwaites Glacier. Warm waters at depth have been eroding the underwater edge of Pine Island Glacier since the early 1970s (location 1). The melting has led to a steeper undersea interface and faster glacial flow above (location 2). Measurements from a remote underwater vehicle (ROV) have documented the mixing of inflowing warm water and outflowing meltwater in the undersea cavity, as well as further thinning and recession of the glacier itself (location 3). (Bethan Davies, http://www.AntarcticGlaciers.org, after Schoof, 2010: Beneath a floating ice shelf. *Nature Geoscience*, **3**, 450–451)

There remain plenty of question marks around the fate of marine ice cliffs, though.

For instance, deformations within the ice could increase as the cliff edge approaches its critical height, perhaps limiting how long the cliff can be sustained.

On the other hand, a rapid cliff collapse might produce a vast field of icebergs that partially checks glacial advance.

All in all, research is continuing to point toward West Antarctica as the biggest single x-factor in sea level rise this century.

Risks of a rising ocean

What if, by the year 2100, sea level were to rise on the order of 0.6 m (24 in.)?

That's near the top of the 2013 IPCC midrange pathways and near the low end of semi-empirical projections.

It wouldn't be enough in itself to inundate the world's major coastal cities, but it could certainly lead to an array of serious problems, especially for populous, low-lying regions.

Some tropical islands could become virtually uninhabitable and, significantly, higher sea levels also mean a higher base on which to pile storm surges, tsunamis, and all other ocean disturbances.

These are likely to be the real troublemakers as sea level climbs over the next several decades.

One region at particular risk is the Ganges Delta of eastern **India and Bangladesh.**

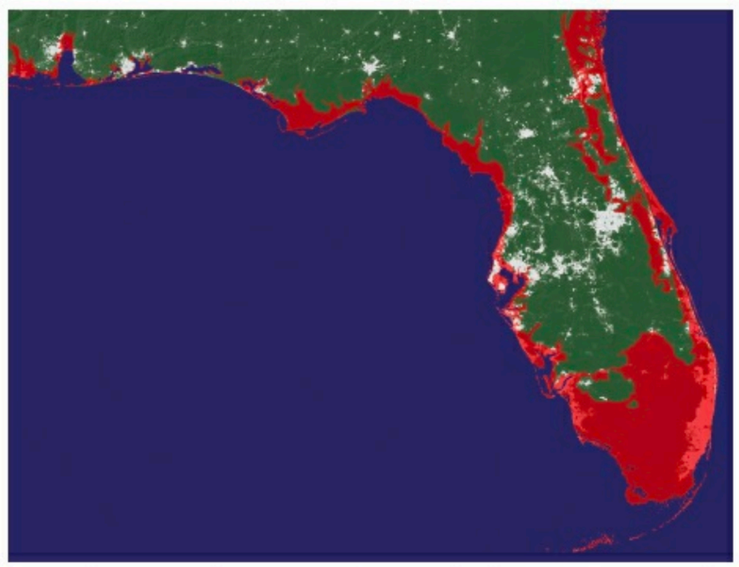
While most of Bangladesh is expected to remain above water even if the ocean rises several meters, such a rise could still displace millions of people in this extremely low-lying country.

Millions of other Bangladeshi people who remain near the coast will be vulnerable to catastrophic flooding.

Another place set for serious problems from even modest sea level rise is the glittering string of hotels, offices, and condominiums along Florida's southeast coast from Miami to Palm Beach, as well as the more rustic Florida Keys.

A sea level rise of 0.68 m (27 in.) would put some 70% of the Miami area—and nearly all of the Keys—underwater.

Building effective barriers against the rising ocean will be difficult, if not impossible, because seawater will tend to sneak through the porous limestone that undergirds south Florida.



Parts of Florida susceptible to a 6-m (20-ft) rise in sea level are shown in red, with paler areas showing high-population zones. (Jeremy L. Weiss and Jonathan T. Overpeck, University of Arizona)

One crucial aspect of sea level rise is that it won't be uniform across the world's oceans, a point hammered home by recent studies.

According to the 2013 IPCC report, close to a third of the world's coastlines may see an MSL rise that's at least 20% higher or lower than the global average.

How could this be?

As noted earlier, some coastlines are still rebounding from the last ice age, which would reduce their net sea level rise.

Others are experiencing subsidence, which can occur through natural geologic processes as well as when oil, gas, and water are drawn from beneath the surface.

Subsidence has allowed sea level to rise especially quickly in far southeastern Virginia, a critical nerve center for the U.S. military.

At Norfolk, the average ocean height since 1990 has climbed more than 12 cm (4.8 in.), well above the global average.

The northeastern U.S. coast is vulnerable to particularly large MSL rises due to potential shifts in ocean circulation.

Should the Atlantic's northward flow be impeded by meltwater streaming from Greenland, then colder, deep water wouldn't form as readily.

The net warming from this dynamic process would push sea levels even higher than projected across most of the North Atlantic.

Indeed, this dynamic process on top of global trends could end up producing some of the world's largest rises in sea level at long-settled coastlines from the Mid-Atlantic north to the Canadian Maritimes.

Looking centuries ahead, the geographic picture of MSL rise could be influenced by the timing of the Greenland and West Antarctic ice sheet melts, because of the loss of large gravitational tugs exerted by each sheet.

In both cases, the effect would be for water to move toward the tropics and away from the polar region where a given ice sheet is located.

With all of these changes unfolding, it's not out of the question—as odd as it seems to contemplate—that most of the world's coastal cities could be largely or completely underwater in a few hundred years.

The risk is seldom put in such stark terms, but it's a logical outcome of the amount of warming expected to occur over the next several centuries unless major emission reductions take place.

According to some estimates, sea levels ran 6–8 m (20–26 ft) above current MSL during the last interglacial period more than 100,000 years ago, enough to inundate every or nearly every low-lying coastal city.

Global temperatures during this period are believed to have been only about 1°–2°C (1.8°–3.6°F) higher than today's.

The warming back then was produced by changes in Earth's orbit around the sun, but greenhouse gases could bring as much or more warming in the next century and beyond.

Barring any unpleasant surprises, such as a collapse of key Antarctic marine ice cliffs, it'll probably take more than a few decades for sea level to begin approaching such catastrophic values.

The catch is that the greenhouse gases we're now adding to the atmosphere at unprecedented rates will continue to warm the climate—and raise the seas—for much more than a century.

One analysis led by the Potsdam Institute estimates that every degree Celsius of warming commits the planet to an eventual sea level rise of 2.3 m (7.5 ft) over the next 2000 years.

In short, the greenhouse ship is becoming progressively harder to turn around, and the seas are rising around it.

Climate change and El Niño

The oceans play a vast role in shaping the vagaries of weather and climate, and much of that influence comes through a set of **ocean–atmosphere cycles.**

Linked to arrangements of high and low pressure centers over various parts of the world, these cycles each alternate between two "modes," producing recognizable, repetitive weather patterns—drought, excessive rainfall, unusual warmth or cold, and so forth.

These can unfold half a world away from their oceanic triggers, and they're natural parts of climate.

It's possible that global warming will tamper with them, but right now, scientists have more questions than answers about whether and how this will happen.

Globally, the most important ocean–atmosphere cycle is the **El Niño–Southern Oscillation** (**ENSO**, whose two modes are known as **El Niño** and **La Niña.**

ENSO is based in the tropical Pacific Ocean, which spans a third of the globe from Ecuador to Indonesia.

Trade winds blow more or less continuously across this huge area from east to west, pushing warm water toward Indonesia, where the resulting balmy seas help generate persistent showers and thunderstorms.

The cold, upwelled water off Ecuador and Peru, meanwhile, stabilizes the air there and produces the region's legendary aridity (Peru's capital, Lima, gets about the same amount of rain each year as Cairo).

Every two to seven years, the trade winds weaken or reverse, the surface layer of warm water deepens and expands into the eastern tropical Pacific, and an El Niño sets in, typically lasting from northern autumn into northern spring.

The flip side, La Niña, occurs when trade winds are stronger than average, pushing cooler-than-usual water westward into the central tropical Pacific for one to three years.

About half of the time, neither El Niño nor La Niña is in progress and the Pacific is neutral.

El Niño increases the odds of drought across Indonesia, Australia, India, southeast Africa, and northern South America.

It tends to produce mild, dry winters in Canada and the northern United States and cool, moist winters in the U.S. South and Southwest.

It also lowers the chances of hurricanes in the North Atlantic and raises them for parts of the North Pacific.

La Niña, in general, has the opposite effects.

Many of the year-to-year spikes and dips in global temperature are a result of ENSO.

Because so much heat is stored in the planet's oceans, it takes only a slight change in ocean circulation to make a big change in how much of that heat enters the air.

Earth's atmosphere warms by as much as several tenths of a degree Celsius during a strong El Niño, as warmer surface waters allow heat to escape to the atmosphere.

Likewise, the atmosphere cools during a powerful La Niña, as cooler-than-average waters absorb heat from the air above.

Indeed, as global temperatures were rising sharply in the 1980s and 1990s, five El Niños occurred, including the two strongest ones of the last century (1982–1983 and 1997–1998).

However, La Niña began predominating after 2000, at the same time that the rate of global temperature rise dropped markedly.

The switch flipped again in the mid-2010s, when the planet's jump to three consecutive record-warm years (2014–2016) was boosted by a strong El Niño.

Importantly, the treble and bass notes of El Niño and La Niña play out atop the drumbeat of long-term warming.

When the ups and downs from ENSO are stripped out of the global temperature record, the long- term rise from greenhouse gases becomes even clearer.

Computer modelers are making progress in depicting El Niño and La Niña, and climate models can now simulate a variety of behaviors in the tropical Pacific's arrangement of ocean currents and rainfall patterns.

However, due in part to the natural climate variability that affects the size and frequency of El Niño and La Niña events in both models and observations, it's not yet clear how global warming will affect the mix.

Trade winds are expected to weaken, on average, and sea surface temperatures may warm more strongly near the equator than away from it.

Yet as the various processes controlling ENSO evolve, their combined effects could go in either direction.

The single most consistent trend among models that fed into the 2007 and 2013 IPCC reports is an eastward expansion of the western Pacific's warm pool, and its showers and thunderstorms, into the central Pacific.

According to work done at the National Center for Atmospheric Research, it's possible that this could move the region's typical state closer to that of El Niño, but with individual El Niño and La Niña events still occurring on top of that new configuration.

Will the Atlantic turn cold on Britain?

Even as global warming appears to be proceeding full steam ahead, there are hints that the United Kingdom and parts of the European continent could be in for a countervailing force—a cooling influence that might inhibit human-induced warming in some areas.

The storyline will be familiar to those who caught 2004's blockbuster *The Day After Tomorrow*, which took the idea to cartoonish extremes.

The nugget of truth in the film is that the northward flow of warm water into the far North Atlantic could slow or even shut down entirely, which would plunge much of northwest Europe into a climate more akin to that of Canada, Russia, or other countries at the same latitude.

In Hollywood's view, this occurs in a matter of days, flash- freezing almost everyone from Washington to Scotland.

In real life, any cooling would be mostly regional and far less dramatic, unfolding over at least a few years, more likely decades to centuries.

Even so, the possibility is worth taking seriously.

The Atlantic flow is part of the global loop of ocean circulation (see graphic above).

It's a system whose mechanics are well mapped but not fully understood.

One of the most important branches is often referred to as the **Atlantic thermohaline circulation**, because it's related to both temperature (thermo) and saltiness (haline).

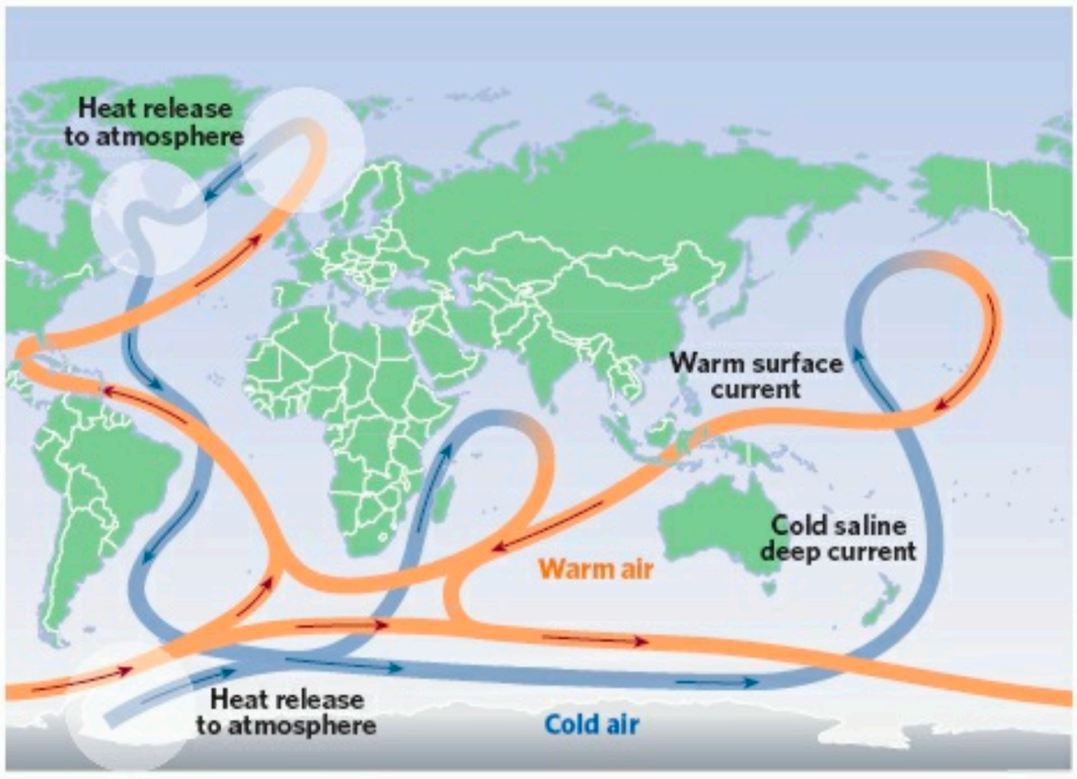
Many scientists now use a different name: the **Atlantic meridional overturning circulation** (AMOC). "Meridional" refers to its north–south orientation, and "overturning" refers to the somersault it performs in the far north.

Before it gets to that point, the AMOC is a vast conveyor belt of warm water that flows northward from the Brazilian coast.

In the North Atlantic, prevailing winds sculpt the sea surface flow into a clockwise gyre, sending warm water north along the U.S. coast in the Gulf Stream and bringing cool water south from Spain in the Canary Current.

An extension of the Gulf Stream called the North Atlantic Current pushes warm surface water even farther northeast toward the British Isles.

Journalists often use the term "Gulf Stream" to refer to the North Atlantic Current, but strictly speaking, that isn't correct.



The great ocean conveyor belt.

The flip side of the conveyor occurs when the warm surface water approaches Iceland and splits around it, heading for the Labrador and Norwegian Seas.

Along the way, it gradually chills and sinks.

Then, at several depths ranging from about 2 to 4 km (6600–13,100 ft), the frigid deep water begins the return trip to the South Atlantic, thus balancing the warm northward conveyor and completing the AMOC.

By any measure, the AMOC is a serious workhorse.

As it passes the latitudinal belt around 25°N, the circulation brings an average of roughly 1.3 petawatts of heat energy northward, day in and day out.

That's enough energy to run almost 100 trillion 14-watt light bulbs, or more than 60 times the energy being used by all the homes, offices, and factories on Earth at any moment.

Scientists have speculated for more than a century that this giant natural turbine might have more than one speed setting.

Since Earth began to warm up about 12,000 years ago, the AMOC has stayed fairly robust, except for a couple of dramatic weakenings around 11,500 and 8200 years ago, as glacial meltwater poured into the North Atlantic and suppressed the circulation.

These events were enough to cool the winter climate in and around the present-day United Kingdom on the order of 10°C (18°F) below present-day readings.

In other words, London in a typical January would have felt more like St. Petersburg in today's climate.

We're now in the midst of a century-plus AMOC slowdown that's less extreme, but still enough to put the speed setting at its lowest in at least 1600 years, according to a 2018 study led by the University College London.

The main worry for this century and beyond is the increasing presence of warmer, less-dense water in the far North Atlantic, which has more trouble sinking and forming the Atlantic's cold, deep return flow.

Increased melting from the Greenland ice sheet and glaciers adds to the freshening (i.e., reduced saltiness) of the waters, which reduces their density further.

Analyses led by the United Kingdom's Centre for Environment, Fisheries and Aquaculture Science show that the upper 1000 m (3300 ft) of the Nordic seas—the main source of sinking water for the Atlantic—became steadily fresher from the 1960s into the twenty-first century.

Likewise, in a 2018 analysis from the Potsdam Institute for Climate Impact Research found fingerprints in North Atlantic sea surface temperature that point to a drop of about 15% in the AMOC's strength since the mid-twentieth century.

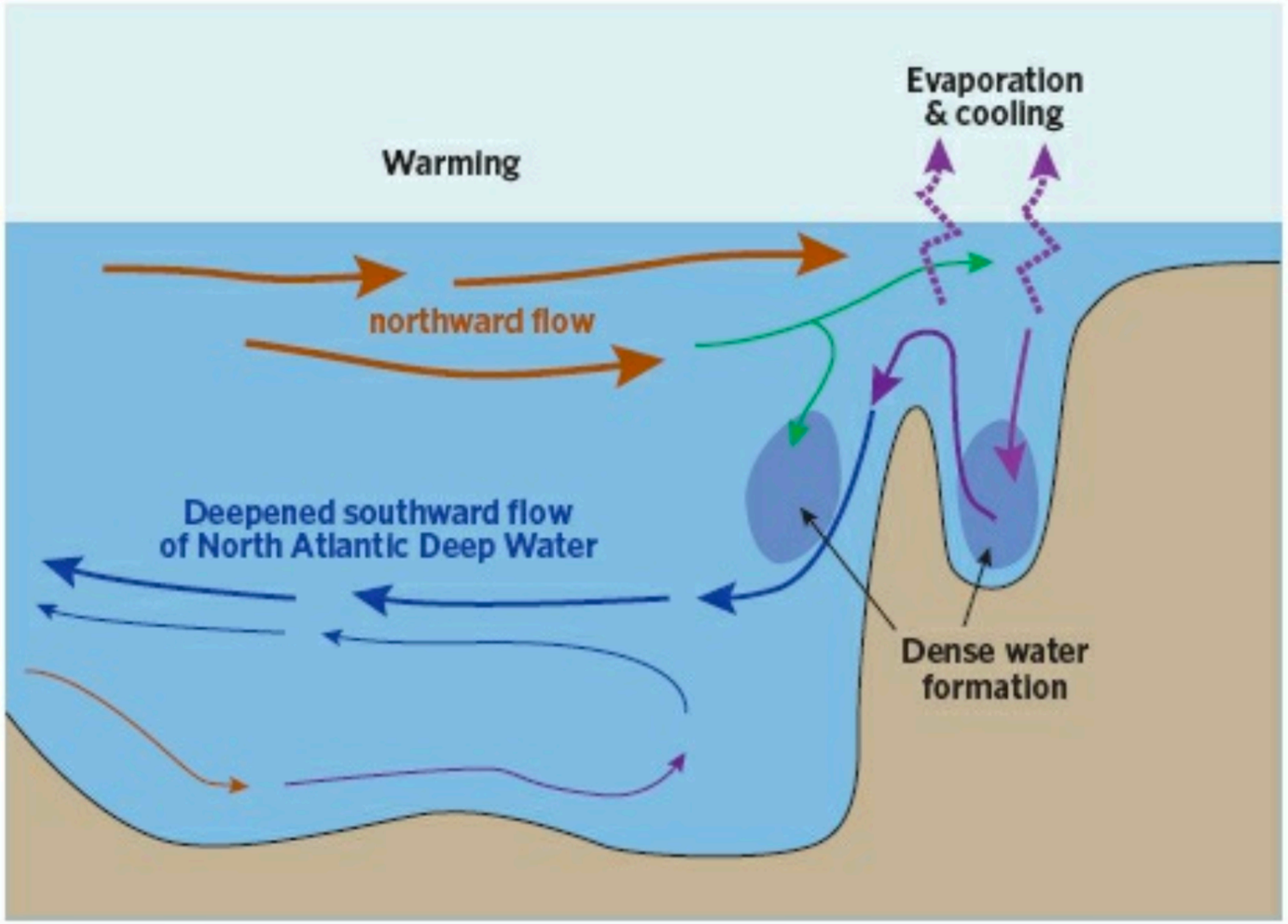
Computer models agree that evolving conditions in the Atlantic should put a substantial brake on the AMOC in the coming century.

In response to this concern, the United Kingdom's Natural Environment Research Council funded an observational program called RAPID that ran from 2001 to 2008, with a follow-up called RAPID-WATCH continuing through the 2010s.

At the time of RAPID's launch, the Met Office Hadley Centre had produced simulations in which the AMOC was halted in its tracks.

The results showed a net cooldown of around 1°–3°C (1.8°–5.4°F) by 2050 across most of the United Kingdom, Scandinavia, Greenland, and the adjacent North Atlantic, with warming elsewhere around the globe.

However, as the models became more sophisticated because of RAPID and other research, they produced a more modest slowdown of the AMOC



The Atlantic meridional overturning circulation shown in a vertical cross section extending from lower northern latitudes (left) to high latitudes (right).

Nearly all of the modeling generated for the 2013 IPCC assessment points toward at least a gradual decrease in AMOC intensity by the end of this century.

How much is still an open question: depending on how quickly you expect greenhouse gases to accumulate, the model consensus for the four IPCC emission pathways ranges from a slowdown of 11% to 34%.

Even with the AMOC slackening, the newest models still produce a net warming in the British Isles and nearby countries, with greenhouse gases more than compensating for the ocean slowdown.

On top of these big changes, there are signs of a natural speedup/slowdown cycle in the AMOC, with each phase peaking about every 60 or 70 years.

Measurements in the deep Atlantic are too sparse to show this trend, and there's not yet an agreed-upon physical explanation for it.

Nevertheless, sea surface temperatures across the North Atlantic show several distinct rises and falls since the late 1800s—a pattern referred to as the **Atlantic multidecadal oscillation.**

Interestingly, global air temperatures have risen and fallen over a roughly similar timeframe, which has led some scientists to investigate how the two processes might be intertwined.

If this 60- to 70-year cycle is robust, then the AMOC ought to undergo a slight natural slowdown from the late 2010s or 2020s into the 2040s or thereabouts.

The challenge will be to untangle this from any potentially larger human-induced slowdown.

With observations from RAPID showing that the AMOC can vary naturally over a variety of time frames (even a strong hurricane in the right spot can influence the AMOC), it might take a while for any trends related to climate change to show themselves clearly.

In the meantime, Britons might be best off planning for a warmer climate while learning to live with a bit of uncertainty about exactly how warm it might get.

Living in a changing sea

The creatures that make their home in the world's oceans might seem to have much more flexibility to deal with climate change than do their land-based cousins.

But do they?

Marine biologists are still sorting through the many interconnections among ocean species and are trying to map out how things might change as the sea warms up and as its chemistry changes.

As with life on land, there will be winners and losers, though not necessarily in balance.

Migratory fish should in principle be able to shift their ranges relatively easily, provided they can still access the food and habitat they need, but not all components of the oceanic food chain will change in unison.

A small rise in ocean temperature can affect the mix that exists among the tiniest marine plants and animals (including **phytoplankton**—which account for half of all the biomass produced on Earth—and the **zooplankton** that consume them).

Changes in these could lead to mass die-offs of seabirds and fish further up the chain.

Outflows of fertilizer-rich water can help produce huge blooms of phytoplankton near coastlines, but the picture may be different over the open ocean.

A 2006 study using NASA satellite data showed a global-scale link between warmer oceans and a drop in phytoplankton abundance and growth rates over the preceding decade.

This topic broke wide open in 2010 with the release of a bombshell study led by Dalhousie University.

Combining satellite data with nearly half a million measurements of ocean color taken by ships, they asserted that phytoplankton had declined in most of the world's oceans—especially across the tropics and subtropics—at an astounding average rate of about 1% per year since 1950, in sync with rising ocean temperatures (as well as vastly expanded fishing).

However, other work using model-based simulations has shown much smaller reductions in phytoplankton, and several critics have argued that Boyce's blending of data collected at different times and in different ways likely biased the results downward.

One of the world's largest and oldest phytoplankton surveys—the British-based Continuous Plankton Recorder, begun in 1931—has shown phytoplankton increasing into the twenty-first century across most of the North Atlantic.

The picture may vary by species, too.

Satellite-based analyses from NASA suggest that diatoms, the largest phytoplankton algae, declined by more than 15% globally from 1998 to 2012, while other work indicates smaller species may represent an increasing share of the global mix.

Scientists can zero in on shorter-term links between ocean chemistry and climate by observing what happens during sharp regional warmings.

Off the coast of Peru, the warm water of El Niño cuts off the upwelling of colder, nutrient-rich waters that support a variety of marine life.

Back in 1972, an El Niño intensified the effects of overfishing and decimated Peru's lucrative anchovy industry.

Many birds that fed on the anchovies died off, and in turn the local supply of guano— bird droppings used as fertilizer—plummeted.

Such chains of events aren't limited to the tropics.

In 1997, unusually warm waters developed in the eastern Bering Sea (which lies between Alaska and Russia), killing off the type of phytoplankton that normally dominate during the summer.

The disturbance worked its way up to devastate local zooplankton, seabirds, and salmon.

Such studies imply that, as global warming unfolds, some of the most dramatic impacts on ocean life may occur through intense seasonal and regional spikes in water temperature driven by El Niño and other ocean cycles.

It's somewhat analogous to the way climate change makes its presence most obvious and discomforting to humans during extreme heat waves.

And there's more to consider than warming alone.

Some ocean life is more dependent on seasonal changes in light than on temperature, especially at higher latitudes.

Other species are more purely temperature-driven.

Since cycles of sunlight aren't changing, the most light-dependent creatures haven't shifted their timing, whereas many species that are more temperature-dependent are starting their yearly growth cycle at earlier points in the warm season.

The differences between these groups may lead to a large-scale mismatch, threatening the seasonal synchrony in growth and feeding times that supports much of the sea's food chain.

Looming over this entire scene is the gargantuan threat posed by **ocean acidification**— perhaps the most underappreciated risk posed by our longtime love affair with fossil fuels.

This transformation is a direct result of the enormous amounts of carbon dioxide being soaked up by Earth's oceans.

Each year a net influx of roughly seven metric gigatons of CO_2—which includes close to 25% of all the carbon dioxide produced by human activity—goes into the sea.

In the well-mixed upper part of the ocean, where the absorption happens, the extra CO_2 is gradually making the water more acidic—or, to be more precise, the water is becoming less alkaline.

A 2010 report by the U.S. National Research Council notes that average pH values at the ocean surface have dropped from around 8.2 in preindustrial times to around 8.1 today.

That change may seem small, but it appears to be at least 10 times faster than anything the world's oceans have seen in the last few million years.

Because pH is measured on a logarithmic scale, every change of one point is equal to a tenfold increase or decrease.

The 0.1-point drop in pH already observed thus translates to a 30% increase in hydrogen ions throughout the upper ocean.

Depending on the pace of global greenhouse emissions, oceanic pH could drop to the range of 7.8–7.9 by 2100.

A pH value of seven is neutral, so the oceans would still be slightly alkaline, but less so than they've been for at least hundreds of thousands of years.

Scientists are only beginning to probe the influence of acidification on marine life.

As carbonate ions become scarcer, it affects the ability of many organisms, such as corals and oysters, to secrete their calcium carbonate shells.

Ocean acidification has also been shown to affect the ability of young clownfish to sense chemicals in their environment that help them detect and avoid predators.

A growing body of work provides an unsettling preview of what acidification might have in store for marine life.

Reefs located near natural oceanic vents for carbon dioxide, including sites in the Mediterranean and near Papua New Guinea, show dramatic declines in calcifying organisms and far less diversity in foraminifera when pH drops below 7.8.

Among the winners in these regions are sea grasses, which take advantage of the additional CO_2 for photosynthesis.

The regions where upwelling currents already bring cold, low-pH water to the surface are especially telling.

Billions of oyster larvae have died off since 2005 during periods of upwelling at hatcheries along the Oregon and Washington coasts.

The creatures' shell-building was hindered by pH that was unusually low, even for these locations.

And in the Southern Ocean off Antarctica, a team of scientists found evidence in 2008 that pteropods—tiny creatures also known as "sea butterflies"—were suffering major damage to their shells in regions of upwelling.

Ocean acidification also affects non-calcifying organisms such as fish, octopus and squid.

Many marine animals rely heavily on their chemosensory ability to detect the presence of food, predators, and mates for reproduction, and to communicate with other organisms.

Several studies now show that ocean acidification can affect the nature of the chemicals these animals sense, as well as their ability to detect those chemical cues.

Even if humans can turn the tide on greenhouse gas emissions, it will take tens of thousands of years more before the ocean returns to its preindustrial pH levels.

Some engineering fixes have been proposed—for instance, adding mammoth amounts of limestone to the ocean to reduce its acidity—but these are likely to succeed only on a local level, if even then.

Any such technical fix is laden with ecological question marks: for instance, could an attempt to de-acidify the oceans cause something even more harmful to occur?

Similarly, it's important to keep in mind that any scheme to tamp down global warming by blocking sunlight, as proposed in several geoengineering strategies, can't directly address ocean acidification, as these schemes don't reduce the amount of CO_2 entering the atmosphere and, by extension, the oceans.

Coral reefs at risk

The most famous oceanic victims of climate change may be the magnificent **coral reefs** that lace the edges of the world's subtropical and tropical waters.

More than half of the planet's coral reefs have already been destroyed or extensively damaged by various types of human activity and by spikes in water temperature pushed higher by overall ocean warming.

Climate change is unlikely to completely eliminate the world's reefs, but many individual coral species may face extinction, and some of the world's most spectacular reefs are in big trouble.

It's amazing that coral reefs manage to produce the color, texture, and overall splendor that they do.

After all, they grow in the oceanic equivalent of a desert: the nutrient-poor waters within 30 m (100 ft) of the surface.

The reefs' success in this harsh environment is because of their unique status as animal, vegetable, and mineral.

Coral reefs consist of **polyps**—tentacled animals that resemble sea anemones—connected atop the reef by a layer of living tissue. Inside the tissue are microscopic **algae** that live in symbiosis with the polyps.

The algae produce the reefs' vivid colors and feed the polyps through photosynthesis, while the polyps provide the algae with nutrients in their waste.

The polyps also secrete **limestone** to build the skeletons that make up the reef itself.

Coral reefs are durable life forms—as a group, they've existed for at least 40 million years— but they're also quite selective about where they choose to live.

The microalgae need access to light in order to photosynthesize, and the skeleton needs a foundation, so coral reefs form in shallow, clear water on top of a base that's at least moderately firm, if not bedrock.

Finally, the waters around coral reefs need to be warm, but not too warm.

Generally, the average water temperature should be above 27°C (81°F), with annual lows no cooler than 18°C (62°F).

If the waters spend much time at temperatures more than 1°–2°C (1.8°–3.6°F) above their typical summertime readings, the algae can begin to photosynthesize at a rate too fast for the coral to handle.

To protect its own tissue, the coral may expel the microalgae, allowing the white skeletons to show through the translucent polyps.

When this occurs on a widespread basis, it's known as **mass bleaching** (although chlorine has nothing to do with it).



Rich, diverse, and vulnerable—abundant life on the coral reef. (UCAR Digital Image Library, photo by Kathy Krucker)

The record-setting El Niño of 1997–1998 led to unusually warm water in much of the tropical Pacific, triggering a bleaching disaster unlike anything ever recorded.

Some 16% of the world's coral reefs were damaged in that event alone.

A weaker event in 2002 bleached up to 95% of individual reefs across parts of the western Pacific, and another strong El Niño in 2010 produced record-warm oceans and a second round of global-scale bleaching.

This time the victims included some groups of coral near Indonesia that had managed to survive the 2004 tsunami.

The screws tightened further in the mid-2010s, as sustained global warmth across large swaths of tropical ocean led to a devastating round of bleaching.

More than half of a global sample of reefs saw at least 30% bleaching in 2015–2016, according to a 2018 survey from James Cook University.

Among the hardest-hit areas was the iconic Great Barrier Reef, the world's largest reef complex, where a catastrophic, months-long oceanic heat wave in 2016 killed some 30% of the reef's corals.

Some coral species can bounce back from bleaching, although it can take decades to heal an entire reef.

Other species—especially those with lavishly branched forms and thin tissues—don't appear to be so lucky.

Experiments hint that some coral may be able to adapt to bleaching over time by teaming up with different microalgae more suited to warmer temperatures.

The upshot over the long term may be a thinning of the herd, as it were, leaving behind a hardier but less diverse and less spectacular set of coral reefs.

Such a process may already be under way in parts of the Great Barrier Reef in the wake of 2016's mass bleaching, but a major concern is that the corals that survived may be too few and too far between to support successful reproduction and recruitment of new corals.

To complicate the picture, global warming isn't the only threat the reefs are facing:

- **Agricultural fertilizer** that washes offshore can deposit nitrogen and phosphorus that encourage the formation of seaweed or phytoplankton at the expense of corals.
- **Overfishing** can increase seaweed growth on top of reefs by removing many of the herbivorous fish that would otherwise keep the growth in line.
- **Sediment from rivers** can block needed sunlight, interfere with the polyps' feeding, or even bury a reef whole. Logging and other land-based activities have produced enough sediment downstream to put more than 20% of the coral reefs off Southeast Asia at risk, according to the World Resources Institute.

The ultimate impact of climate change on reefs depends in part on how extensive these other stressors become.

Many older coral reefs have built upward to the maximum height that sea levels allow, so a higher sea surface would give these reefs more vertical room to grow.

Meanwhile, the lowest-lying coral reefs could probably grow upward quickly enough to keep up with modest sea level rises—a typical vertical growth rate is about 1 m (3 ft) in a thousand years.

As long as temperatures increase slowly enough to avoid coral bleaching, warmer temperatures appear to speed up the calcification process that builds reefs.

And the global spread of warmer waters will add a sliver of territory around the current margins of coral reef viability, though that extra real estate is limited by other requirements such as a firm foundation, clean water, and sufficient light.

Together, these modest pluses will face an array of negatives produced by the environmental stresses noted above, not to mention global warming itself.

Ocean temperatures are already rising at a rate that exceeds the pace at which corals can adapt.

Along with causing bleaching, warmer waters may nourish the enemies of reefs as much or more than they help the reef itself.

One study from Cornell University found that some pathogens tend to thrive at temperatures as little as 0.6°C (1.0°F) above the optimal readings for coral themselves.

Another risk is the ocean acidification discussed above, which can both limit the amount of carbonate available for reef building and perhaps erode the reef skeletons directly, in both cases slowing the rate of reef growth.

Also, tropical cyclones may become more intense with time, although it appears that a healthy reef can recover from hurricane or typhoon damage fairly quickly.

In short, we can expect at least a few coral reefs to be around in the world of our grandchildren, but they may not be in the same locations or glow with the same brilliance that snorkelers and scuba divers now savor.

8 - Hurricanes and Other Storms ROUGH WATERS

Spinning their way across the warmer parts of the globe, tropical cyclones throw fear into millions of coastal dwellers each year.

Depending on which ocean they're in, these huge weather systems are known as hurricanes, cyclones, or typhoons.

Whatever you call them, they've made their presence known across the tropics and midlatitudes, particularly in the new century.

A record 10 typhoons raked Japan in 2004. Australia was hammered in early 2006 by a trio of cyclones, each packing winds above 250 kph (155 mph either offshore or on the coast—the first time the region has recorded three storms of that caliber in a single season.

Cyclone Nargis, which slammed into Burma in 2008, killed more than 100,000 people, making it the world's deadliest tropical cyclone in at least three decades.

The super-sized, late-season storm known as Sandy, classified as a hurricane until just before it came onshore, ravaged large sections of the New Jersey and New York coastlines in 2013.

And in 2017, a trio of intense Atlantic hurricanes—Harvey, Irma, and Maria—caused massive destruction from Texas to Florida and led to horrific suffering across the Caribbean, including a prolonged aftermath in Puerto Rico that took thousands of lives on its own.

The Atlantic onslaught of 2017 summed up vivid memories of 2005, which produced three of the six most powerful Atlantic hurricanes on record up to that point: **Wilma** (the strongest), **Rita** (fourth strongest), and **Katrina** (sixth strongest).

All three made devastating strikes on coastal areas.

Wilma slammed Cancún and south Florida, Rita pounded the Texas–Louisiana border, and Katrina swept through Miami before bringing mind-boggling misery and more than 1800 deaths to the New Orleans area and the Mississippi coast.

Does global warming have anything to do with all this havoc?

Several major studies have made a strong case that tropical cyclones worldwide are indeed getting more powerful and will continue to do so.

At the same time, a number of hurricane experts (many of them forecasters) have taken the position that climate change plays only a minor role at best in hurricane activity and that it will be far outweighed by other factors in the future.

The legitimate science and policy questions at hand make this a lively arena of debate.

To the north and south of the tropics, people tend to be more concerned with **coastal storms.**

These wintry blasts of rain, snow, and wind sweep through places such as New England, British Columbia, and western Europe, packing less punch than hurricanes but often affecting much larger areas.

Coastal storms appear to be shifting gradually poleward over time.

Though the trends aren't highly dramatic, they could have major importance for the densely populated coastlines of North America and Eurasia, especially when coupled with rising sea levels.

Well away from the coasts, the most intense bouts of wind and rain are likely to occur with severe **thunderstorms**, which occasionally pack the tiny but powerful windstorms known as **tornadoes.**

Many hurricanes generate tornadic thunderstorms as well.

Although tornadoes are being reported more often in many parts of the world, it looks as if that can be pinned on greater awareness of them.

Thankfully, there's no solid evidence that the most violent tornadoes have become more intense or frequent, although they do seem to be clustering into active periods separated by longer stretches of quietude.

A taste of things to come?

Two main questions loom over any discussion of future tropical cyclones: will there be more of them in a warmer world, and will they be more destructive?

The 2007 IPCC report noted that the number of intense hurricanes and typhoons had increased during recent decades in some regions.

This shift remains difficult to nail down, and it isn't fully global, as reiterated by the panel's 2013 report.

However, it's especially robust in the North Atlantic, where a distinct jump in frequency of the most intense hurricanes has taken shape since the 1970s.

The notion of stronger hurricanes and typhoons in a warmer world makes physical sense.

Warm ocean waters help give birth to tropical cyclones and provide the fuel they need to grow.

Generally, the cyclones can't survive for long over sea surfaces that are cooler than about 26°C (79°F).

The larger the area of ocean with waters above that threshold, and the longer those waters stay warm in a given year, the more opportunity there is for a tropical cyclone to blossom if other conditions allow for it.

Not all tropical waters are conducive to hurricanes.

Within about 5° of the equator, the winds can't easily gather into a rotating pattern.

Moreover, the South Atlantic doesn't seem to have the right combination of upper-level winds and warm surface water to produce hurricanes, although 2004 brought a major surprise in Brazil.

That leaves the prime breeding waters for tropical cyclones between about 10° and 30° across the North Atlantic, North Pacific, South Pacific, and Indian Oceans.

The terrible 2005 trio of Katrina, Rita, and Wilma shows how important warm water is for hurricanes.

Katrina and Rita both surged to their peak intensity—Category 5—just as they passed over a patch of the central Gulf of Mexico heated by an infusion of deep, warm water from the Caribbean.

This channel, called the **Loop Current** for its arcing path into and back out of the Gulf, is often a hurricane booster.

In 2005 it featured some of the warmest waters ever observed in the open Gulf, topping 32°C (90°F) in places.

Similarly toasty waters had also overspread much of the Caribbean, where Wilma intensified in an astonishing 12 hours from Category 1 status to attain the lowest barometric pressure ever observed at sea level in the Western Hemisphere: 882 hectopascals (26.05 inches of mercury).

Unusually warm water also fueled 2017's onslaught of hurricanes.

Heat content in the topmost 160 m (525 ft) of the northwest Gulf of Mexico was at record-warm levels just before Harvey intensified there, according to an analysis led by the National Center for Atmospheric Research.

To keep a truly intense hurricane at top strength, the ocean's warmth needs to extend deep—ideally, at least 100 m (330ft).

Otherwise, the powerful winds and waves can churn up enough cold water from below to dilute the fuel source and diminish a slow-moving cyclone in a matter of hours.

Both Rita and Katrina began weakening after they left the Loop Current; they began to pull up cooler subsurface waters, while drier air and stronger upper-level winds from the United States started to infiltrate.

By the time they struck the **Gulf Coast**, both storms were down to Category 3 strength.

That didn't matter much, though, because the storms had already piled up fearsome amounts of water that slammed into the coast in the form of punishing storm surges.

Katrina's surge was a U.S. record: at least 8.2 m (27 ft) above sea level at its worst, with waves on top of that, destroying the homes of thousands who thought they lived at safe elevations.

The surge also helped lead to the failure of several levees in and around **New Orleans**, which in turn produced the cataclysmic flood that left most of the city submerged and paralyzed.

The wreckage of the 2005 Atlantic hurricane season extended far beyond U.S. shores.

Ocean temperatures were markedly above average across vast stretches of the North Atlantic tropics and subtropics.

All that long-lasting warmth, coupled with nearly ideal upper-level winds, led to 27 named Atlantic storms in 2005, which smashed the record of 21 set in 1933.

Still another system was upgraded to tropical-storm strength posthumously, bringing the season's final total to 28.

This swarm of storms was noteworthy not just for its sheer size but for its unusual geographic spread as well.

Even Spain got its first-ever tropical storm, **Vince**, which made landfall on the southwest coast near Huelva.

Keeping count: Will there be more cyclones in the future?

The sheer number of Atlantic hurricanes in 2005 led many to ask whether tropical cyclone frequency might already be on the rise due to climate change.

One might think that a larger, longer-lasting zone of warm-enough water would be sufficient to spawn more and more tropical cyclones.

And indeed, both 2005 and 2017 saw Atlantic waters that were unusually warm over vast areas for long periods, which helped seasonal forecasters accurately peg the action months ahead of time.

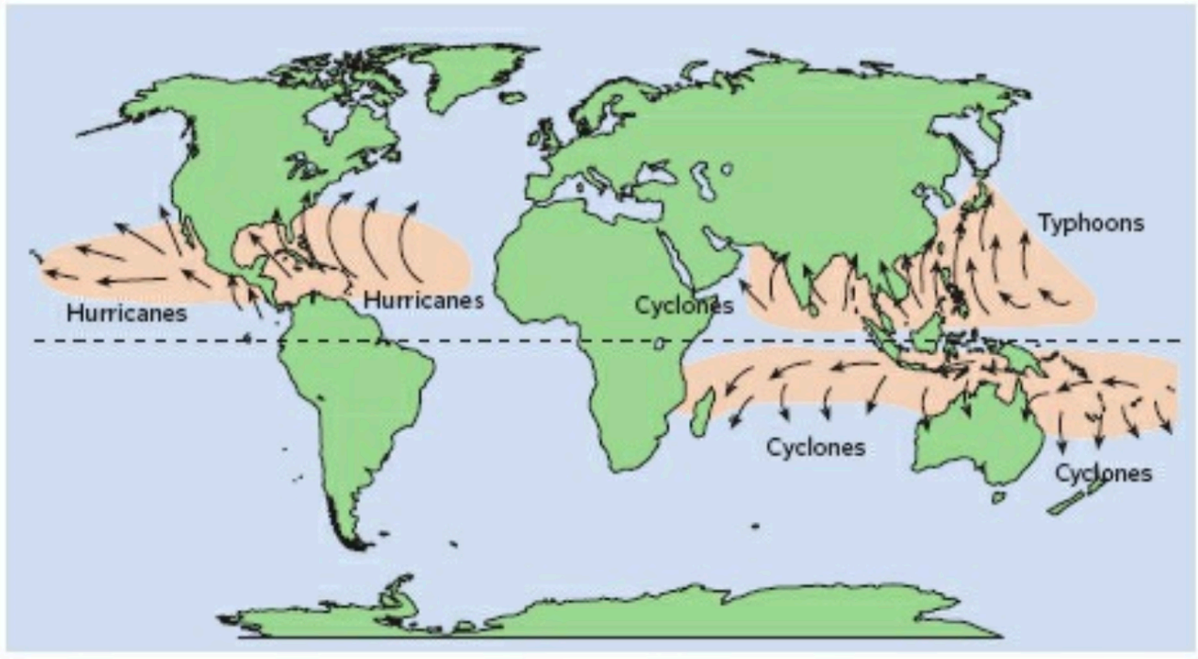
Yet other seasons, such as 2013, weren't able to take as much advantage of relatively warm water.

That year, the Atlantic produced a paltry total of two hurricanes—the smallest total in 31 years—and neither of these systems made it even to Category 2 status.

Both the 2007 and 2013 IPCC assessments concluded that, over the last century, the average number of tropical cyclones worldwide hasn't been significantly boosted by greenhouse gases.

Making such an assessment of the frequency of past tropical cyclones, and projecting their future, is a complex task for several reasons:

- **Competing factors.** Even over sufficiently warm water, only a small fraction of the weak circulations that form in a given year spin up to cyclone status. A number of factors can lead to their early demise: unfavorable upper winds, an influx of dry air, or competition from nearby systems, to name a few. Warm water is critical, but it isn't enough to guarantee that a hurricane will form.
- **Variable records.** It's hard to know exactly how many tropical cyclones roamed Earth before satellite monitoring became routine in the 1970s. Although ships did a good job of reporting stronger tropical cyclones in the pre-satellite era—at least across major shipping lanes—it's likely that some of the weaker, smaller systems over the open ocean were missed then but are being caught now. Better observations also help explain an apparent rise in the number of short-lived Atlantic cyclones, ones that last for two days or less. If these factors are taken into account (including an estimate of the cyclones that once went uncounted), the total number of Atlantic systems may not have risen significantly over the last century, even though the strongest have become more prevalent in recent decades.
- **Complex global interrelations.** For reasons not fully understood, the planet normally does a good job of conserving the total number of tropical cyclones. About 40–60 per year attain hurricane strength, and there has been no significant long-term global trend up or down during the few decades of reliable data. The frequency stays fairly constant because the planet's ocean basins appear to trade off in their periods of peak production. Typically, when the Atlantic is producing hurricanes aplenty, parts of the Pacific are relatively quiet, and vice versa. Some of this seesaw effect is due to **El Niño** and **La Niña,** the cyclic warmings and coolings of the eastern tropical Pacific . The large-scale air circulations driven by La Niña encourage Atlantic hurricanes but suppress them in some parts of the Pacific; the opposite is true of El Niño.



World regions prone to hurricanes, typhoons, and cyclones: different names for the same phenomenon.

Until relatively recently, computers weren't of much help in assessing whether we'll see more tropical cyclones in a warmer world.

That's because most hurricanes and typhoons are small enough to fall in between the grid points of the comprehensive global models that map climate out to a century or more.

It's also tough for a model to pick up on the meteorological subtleties that allow one system to become a hurricane and prohibit the next one from doing so—the same unknowns that plague tropical weather forecasters on an everyday basis.

But since the mid-2000s, researchers have made headway in using models to envision the long-term future of tropical cyclones.

This is in part due to faster computers that allow new models to depict the entire globe in sufficient detail.

An early sign of progress came in 2006, when a group of Japanese scientists published results from a globe-spanning model with 20-km (12-mi) resolution, enough to follow individual tropical cyclones in a broad-brush fashion.

The results suggested that in a midrange emissions scenario, the late twenty-first-century Earth could see fewer tropical cyclones but a higher fraction of intense ones.

Other scientists have nested fine-scale regional models inside coarser global-scale ones, or blended statistical and physical modeling.

The results from these varied approaches have largely been in sync with each other on the main take-home message: we may see fewer tropical cyclones worldwide over time—anywhere from 6% to 34% fewer by late this century, according to a 2010 roundup in *Nature Geoscience* by hurricane researchers from NOAA.

The 2013 IPCC report echoed this conclusion, deeming it likely that global tropical cyclone counts will either drop or hold steady this century.

It’s not entirely clear why many models are tending to spin up fewer storms in the future.

One possibility is a predicted drop in relative humidity aloft, which would threaten the cocoon of moisture in which a hurricane thrives.

Another is a projected weakening of the global circulation through which air rises across the tropics and sinks over the subtropics.

It's also possible in theory that the SST threshold needed for tropical cyclone formation could rise hand in hand with overall ocean warming.

Some work with high-resolution modeling, including a 2013 analysis by the Massachusetts Institute of Technology, suggests that tropical cyclones might actually become more prevalent in many regions, rather than less.

However, such findings hadn't yet produced a wholesale shift in the consensus viewpoint as of the late 2010s.

Even if tropical cyclones become a bit less prevalent on a global scale, as most experts expect, they could still increase over some oceans while decreasing over others.

Models haven't yet agreed on a collective portrait of region-by-region change, but a few intriguing threads have emerged.

This could lead to a stronger warming of tropical waters in the Pacific than in the Atlantic, which in turn might trigger El Niño–like circulation patterns in the atmosphere.

By examining 18 model simulations that fed into the 2007 IPCC report, the University of Miami and NOAA found that upper-level winds became stronger (i.e., more hostile for hurricanes) in the northeast Pacific and Atlantic while relaxing in the Indian and western Pacific Oceans.

However, given that global and regional models still struggle to depict the behavior of El Niño and La Niña, there remains a good deal of uncertainty in how hurricane counts will unfold from one basin to the next.

Regardless of whether or not we see more tropical cyclones, we can still count on most of them never reaching land.

Even if a given ocean began to spawn more cyclones, it doesn't necessarily mean that more of them will hit the coastline, as evidenced by the busy Atlantic seasons of 2010 and 2011 that largely spared the United States.

By the same token, even a slow season can bring destructive landfalls.

In 1992, the Atlantic was far quieter than average, yet it still produced Florida's massively destructive **Hurricane Andrew.**

Is hurricane season getting longer?

One might expect warmer waters to prolong the period when tropical cyclones might spin up.

Several analyses have found that the span between the Atlantic's first and last named storm in a given year has increased by about one day per year since the 1970s.

That's largely due to a recent increase in the number of early-season tropical cyclones—those that develop before the season's official start date of June 1.

Fortunately, pre-season tropical cyclones are seldom very strong, and climate models to date have given mixed signals as to whether we can expect any sustained lengthening of hurricane season.

Packing more power

What about strength?

Here, the tune is more ominous.

Though the IPCC saw little evidence of change in its 2001 report, subsequent studies found a jump in the destructive power of tropical cyclones on a global basis, as noted in the 2007 and 2013 reports.

An increase in hurricane strength is both plausible on a physical basis and apparent in some of the most recent modeling.

Two key studies pushed this topic into the headlines in 2005, where it shared space with that year's bumper crop of Atlantic superstorms.

MIT examined tropical cyclones since the 1970s, when satellite monitoring became widespread.

They looked not just at the raw wind speeds but at the destructive force they packed over their lifespan.

This force, the **dissipation of power**, is proportional to the wind speed cubed.

Thus, a wind of 100 kph would be twice as damaging as a wind of 80 kph.

Small increases in wind speed, then, can produce major ramp-ups in destructive force.

From this vantage point, they discovered that tropical cyclones in the Atlantic and northwest Pacific were packing nearly twice the power they did in the 1970s—the result of the strongest storms getting even stronger as well as lasting longer.

A team from the Georgia Institute of Technology found similar results when they ranked tropical cyclones since the 1970s by **intensity category.**

As expected, they found little change in the total number of tropical cyclones in all categories—a sign of the global balance noted above.

The eye-opener was their discovery that the strongest cyclones, those at Category 4 or 5, were almost 50% more frequent globally in 1990–2004 than they were in 1975–1989.

Put another way, Category 4 and 5 storms went from comprising about 25% of all tropical cyclones globally to comprising about 35% of them.

A follow-up analysis in 2010 showed the trend continuing.

Predictably, these studies provoked a storm of controversy.

They also exposed a schism between climate change researchers, few of whom had done detailed studies of tropical cyclone behavior, and hurricane specialists, who knew the cyclones and the data by heart but weren't always experts on the larger-scale processes related to global warming.

In his 2007 book *Storm World*, science writer Chris Mooney cast the battle as only the most recent example of a tiff between observationalists and theoreticians, with roots extending back to the 1800s.

One of the most vocal critics of the new findings was the late William Gray (Colorado State University), who founded the field of seasonal hurricane prediction.

A long-time skeptic of climate change, Gray assailed the studies on various technical grounds.

Also dismissive of a climate change link were several researchers at the **National Oceanic and Atmospheric Administration** (NOAA), whose **National Hurricane Center** (NHC) forecasts Atlantic storms.

Individually, some NOAA scientists agreed that human-induced climate change has the potential for "slightly increasing the intensity of tropical cyclones".

However, they suggested that the apparent surge in top-strength hurricanes could be at least in part due to inconsistencies in the official record.

For instance, lower-resolution satellites used in the 1970s may not have sensed the tiny areas of high, cold cloud tops associated with the strongest cyclones.

Other observational tools, such as the instrument packages dropped from aircraft into hurricanes, have also evolved over the years, as have the analysis techniques that classify the storms.

Over the past few years, they have been carrying out a long-term effort to reanalyze several centuries of Atlantic hurricanes, which has already resulted in some hurricanes being upgraded and others downgraded.

Even by restricting one's view only to years with the best data, it's apparent that change is afoot, as shown by Florida State University and NOAA/University of Wisconsin–Madison in studies published in 2008 and 2009.

Acknowledging that observational techniques had changed, they set out to compare apples to apples by analyzing only those cyclones that occurred during the period of highest-quality satellite data, which began in 1981.

Through this clarifying lens, the scientists found that the peak winds produced by the globe's strongest tropical cyclones had indeed grown stronger, at a rate of about 1–2 meters per second (2–4 mph) per decade.

Even such a seemingly modest value adds to a storm's destructive power, as noted above.

Over several decades, such a trend might be enough to push a substantial fraction of storms from one intensity category to the next.

The plot thickens even further when you consider the Atlantic multidecadal oscillation (AMO).

It has long been known that, over the past century-plus, sea surface temperatures in the North Atlantic alternated between warming and cooling for periods of about 30–40 years each.

Many researchers have tied this waxing and waning of SSTs to presumed cycles in the Atlantic thermohaline circulation, also known as the meridional overturning circulation.

This circulation is only now being measured with some precision, and its dynamics are poorly understood.

It's also possible that post–World War II pollution helped shade and cool the North Atlantic for decades.

Nevertheless, they showed that the frequency and average intensity of Atlantic hurricanes has risen and fallen every few decades in sync with the AMO—lower from the 1890s to the 1920s, higher from that point through the 1950s, lower again from the 1960s to the 1990s, and higher from the mid-1990s into the 2010s.

Even if there is a strong physical basis for the AMO, it doesn't mean climate change can be ignored.

The AMO only affects hurricane formation in the Atlantic, whereas the studies mentioned above suggest a strengthening trend across several other ocean regions as well.

Moreover, the Atlantic uptick since the 1990s could be a combined product of the AMO and general ocean warming.

For example, a 2006 study from NCAR found that global warming had more than twice the influence of the AMO in producing 2005's extra hurricane-fueling warmth in the waters of the tropical North Atlantic.

Thus, if and when the AMO cycle finally turns toward a quieter Atlantic—perhaps by the early 2020s—things may not calm down quite as much as one would otherwise expect.

Despite their limited ability to project tropical cyclones into the future, computer models tend to agree that overall warming ought to help induce a modest increase in strength.

The IPCC's special 2012 report on climate extremes noted that the most recent models at that point were consistently pushing typical winds in the strongest storms upward this century by anywhere from 2% to 11%.

The latter would be nearly enough to bring a major hurricane from one Saffir–Simpson category to the next.

This doesn't mean that more of these intensified storms will necessarily make landfall, but those that do could pack more of a wallop.

Subsequent studies have largely confirmed this picture.

One question that's just beginning to be explored is whether the size of tropical cyclones might be changing.

The biggest cyclones can bring hurricane-force winds to 10 or more times greater area than the smallest ones, and they can drive enormous storm surges.

Cyclone diameter isn't part of official global databases, though, and most climate change studies to date have omitted this important dimension.

One analysis found no significant change in tropical cyclone size in satellite imagery going back to 1981.

Several recent modeling experiments found hints that tropical cyclones might get larger over time in some areas, but the results varied sharply by region and study.

Model resolutions may need to get sharper in order to capture tropical cyclone size at the detail needed to nail down any future changes with confidence.

However, it's long been known that hurricanes tend to get larger as they move poleward, so any shift toward higher-latitude cyclones—already a consistent trend in the western Pacific—could tip the balance toward larger circulations as well.

Surges and downpours

Extreme **storm surges**—such as the one that swept across parts of the Philippines during Supertyphoon Haiyan in November 2013, killing thousands—will continue to loom as a catastrophic possibility from tropical cyclones in the century to come.

However, the shape of that threat will shift as sea levels rise and the coast itself evolves.

Subsidence and the loss of protective **wetlands** are already making many areas more vulnerable.

More than any place on Earth, the **Bay of Bengal** exemplifies the risks of tropical cyclone surge.

Fully half of the world's deadliest surge tragedies have taken place in eastern India, present-day Bangladesh, and Burma.

Half a million Bangladeshi people died as a result of a catastrophic 1970 landfall (the event that prompted a famous relief concert by ex-Beatle George Harrison).

In 2008, **Cyclone Nargis** took a track unusually far to the south, becoming the first major cyclone ever recorded in Burma's populous Irrawaddy Delta.

A surge 3 m (12 ft) high struck at high tide and swept as far as 40 km (25 mi) inland.

With no experience of such a storm in living memory—and with state-controlled media downplaying the storm's threat—delta residents had no inkling of what they would face.

The official death toll topped 130,000, though many thousands more may have died.

Rising sea level only adds to the Bay of Bengal's flood risk, while soaring coastal populations make even more people vulnerable to a given flood.

The intricate delta-style coastline would be extremely difficult to protect through engineering, even if the nations could afford such a fix.

Landfalling tropical cyclones also produce a great deal of rain, on coastlines as well as far inland.

That may get worse as atmospheric moisture and precipitation increase on a global scale.

Inland flooding—as experienced in spectacular fashion around Houston during Harvey and in North Carolina from 2018's Hurricane Florence—is already a serious threat in many hurricane- prone areas.

With Katrina and Sandy as major exceptions, inland flooding has been the leading cause of U.S. hurricane deaths in recent decades. It's also a major problem for agriculture.

Rainfall in the average tropical cyclone may have already increased by 6%–8%, according to estimates by NCAR.

Modeling associated with the 2013 IPCC report suggests that rainfall rates near the core of a typical storm could rise by 30% or more by later this century.

There's also evidence that tropical cyclones may increasingly drag their heels as climate change unfolds.

All else being equal, a slower-moving storm tends to produce bigger rainfall totals at a given spot.

In a study led by NCAR, 22 recent U.S. hurricanes were implanted in a model depicting a high-emission, late-twenty-first- century climate.

In that warmer environment, the hurricanes' forward motion dropped by an average of 9% while rainfall rates jumped by 24%—a potential double whammy.

With all this in mind, it would seem sensible to prepare for the worst, despite the remaining gaps in knowledge.

One major problem for planners to take on board is the explosive population growth and development taking place on many cyclone-prone coasts.

In 2006, hurricane researchers from across the spectrum of climate change viewpoints crafted a landmark joint statement, agreeing that the "urgent problem of our lemming-like march to the sea requires immediate and sustained attention."

Coastal concerns beyond the tropics

Hurricanes and typhoons may cause horrific damage, but coastal storms threaten the well-being of millions of people who live well away from the tropics.

Unlike tropical cyclones, which need an envelope of warm, humid air in order to grow, coastal storms—part of the larger group known as **extratropical storms**—thrive on the contrasts between cold continental air and warm marine air.

Moving with the polar jet stream, in which they're typically embedded, coastal storms strike many of the major coastlines at mid- and high latitudes, mostly from autumn through spring.

Along the U.S. Atlantic coast, they're often dubbed **nor'easters**, a reference to the strong onshore winds they bring.

The lack of a mountain barrier in western Europe allows coastal storms to plow well inland with heavy rain and high wind.

That was the case over the last week of 1999, when an especially vicious pair of post-Christmas storms dubbed **Lothar** and **Martin** pounded much of the region with winds as strong as 219 kph (136 mph).

The storms killed more than 140 people and ruined centuries-old treescapes from France's Versailles to Germany's Black Forest.

As with hurricanes, the challenge in projecting coastal storms is twofold: will there be more of them, and will they become more intense?

A few studies show that extratropical cyclones have tended to shifted poleward in both the Northern and Southern Hemispheres, a sign of the jet stream shifting in response to a warming atmosphere.

As for intensity and frequency, the trends aren't as robust, and they vary by region.

While extratropical storms appear to have become stronger and more numerous toward the top end of the North Atlantic, there's some evidence the opposite has occurred over the eastern Pacific and North America, according to the 2013 IPCC report.

What about the future?

Here again, as the IPCC stresses, the outcome seems likely to vary by region, especially in the Northern Hemisphere.

If the warm currents of the North Atlantic weaken as projected, then the resulting rearrangement of ocean currents may send stronger and more frequent midlatitude storms into western Europe.

The latest round of high-resolution models also suggests a potential increase in the strongest extratropical cyclones along and near the U.S. East Coast.

Globally, however, the models actually point toward a small decrease in midlatitude storm frequency, especially toward the subtropics.

A multimodel review published in 2013 by Stony Brook University finds a reduction of 6%–32% (depending on the time of year) in strong extratropical storms across North America, given a high-emissions scenario.

This could spell trouble in the form of less rainfall for parts of the United States that straddle the zones between expected subtropical drying and higher-latitude moistening.

Lower-latitude coastal regions such as southern California and the Mediterranean may find themselves experiencing more prolonged and/or intense dry spells in winter, as windy, rainy, fast-moving storms track farther north over time.

The fierce drought that gripped California from 2011 to 2016 developed in large part from a northward-displaced storm track. Even a powerful El Niño event in 2015–2016 saw the jet stream persistently hanging too far north to give Southern California the moist winter that seasonal outlooks had deemed highly probable.

Any regional projections of extratropical storm behavior are framed by substantial uncertainty at this point.

Smaller-scale effects that aren't well modeled or understood could overpower a global trend in particular locations, and although climate change models tend to make the NAO increasingly positive in the decades to come, that's hardly a done deal.

There's plenty of natural variation in the NAO, both on a yearly basis (the index can slide back and forth several times in a single winter) and over longer time frames.

Another wild card is the Arctic, where temperature rises have been especially sharp in recent decades.

A group of scientists at Atmospheric and Environmental Research and Rutgers, The State University of New Jersey has been analyzing possible links between depleted Arctic sea ice, warm Arctic winters, and midlatitude winter weather.

The basic idea is that disruptions of the Arctic's high-altitude polar vortex—triggered either directly or indirectly by a warming climate —may be related to high-profile periods of unusual cold and/or snow across midlatitudes, including eastern North America.

The concepts are being actively debated by researchers, and it's still too soon to know whether such a process might lead to consistently stormier winters near the U.S. East Coast.

Coastal-storm flooding: A deepening problem

Even if coastal storms don't change drastically, the extreme high-water levels they generate will still be pushed upward by the long-term rise in sea levels.

One region where high population meets up with a high risk for surge from coastal storms is the **North Sea.**

The Netherlands have spent centuries defending themselves from the sea with a massive series of dikes.

Much of coastal England, which sits slightly higher on average, gets along with a variety of dams and other flood-control devices.

Both countries found themselves unprepared for the history-making North Sea storm of February 1, 1953, which sent huge surges into the British, Dutch, and Belgian coasts.

Over 1800 people died in the Netherlands, where a powerful surge overpowered dikes and levees in the lower Rhine Valley.

England saw more than 300 deaths on its east coast, including 58 on Canvey Island, from where more than 10,000 residents were safely evacuated.

The 1953 storm triggered a new age of massive efforts to tackle coastal flooding.

The Netherlands spent more than US$5 billion on its **Delta Project,** completed over four decades.

Its centerpiece is a pair of doors, each about 20 m (66 ft) high, that can close off the Rhine if a major coastal storm looms.

Along the same lines, the 1953 storm inspired Britain to enhance seawalls along its east coast and to spend over half a billion pounds on the **Thames Barrier**, which was built between 1972 and 1982.

Spanning the Thames in eastern London at a width of 520 m (1700 ft), it includes 10 massive gates, 6 of which rise from concrete sills in the riverbed and swing to form a single wall.

In 1999, the operating rules of the barrier were changed, allowing for more frequent closures in order to keep tidal waters from blocking river runoff produced by heavy rain.

This change makes it hard to assess how much climate change may have affected the closure rate.

For whatever reason, the gates are swinging shut more frequently—a total of 64 times from 2009– 2010 through 2016–2017, and 81 times in the prior decade, compared to 29 times in the 1990s and only four times in the 1980s.

All by itself, the record-wet British winter of 2013–2014 triggered an astounding 50 closures.

It's been estimated that the barrier could be closing an average of 30 times per year by the 2030s.

Such concerns prompted the development of a 100-year plan to protect the Thames Valley from the increased surge risk related to climate change.

One of the many options put forth is a far wider barrier across the Thames Estuary that would span 16 km (10 mi).

However, the Thames Estuary 2100 plan, released in 2012, recommended sticking with the current barrier, making repairs and improvements as needed, until at least the 2070s.

In another well-publicized initiative, Italy has spent more than a decade working on a set of 78 hinged, inflatable steel gates to protect **Venice** from rising waters.

The US$7 billion project was dubbed **MOSE (Modulo Sperimentale Elettromeccanico** in a tip of the hat to Moses' parting of the Red Sea.

Normally lying flat on the seafloor, the gates will be pulled up as needed in a 30- minute process that would block three inlets from the Adriatic and help keep the city safe from increasingly threatening surges triggered by Mediterranean storms.

Originally due to be completed by 2011, MOSE has fallen prey to a series of technical snafus and management scandals that pushed the expected opening day back until at least 2022.

Minor floods are now so frequent in Venice that visitors and locals alike often find themselves facing streets ankle-deep with water.

Although MOSE won't stop these nuisance floods, it would be employed for any surge expected to exceed 110 cm (3.6 ft).

The gates were designed to handle surges up to 200 cm (6.6 ft).

In 1996 alone, over 100 floods topped the 1-m mark, and a severe 1966 flood reached 180 cm (6 ft).

Alas, it wouldn't take much of an increase in global sea level rise to push a repeat of that 1966 flood over the new barriers.

The city also continues to subside by as much as 0.2 cm (0.08 in.) per year, which isn't helping.

Some researchers have already pondered the idea of a "Great Wall of Venice" that might eventually surround the city in order to preserve it.

9 - Ecosystems and Agriculture THE FUTURE OF FLORA, FAUNA, AND FARMING

If all that global warming did was to make life a bit steamier for the people who consume the most fossil fuels, then there'd be a karmic neatness to it.

Alas, climate change doesn't keep its multitude of effects so nicely focused.

A warming planet is liable to produce a cascade of repercussions for millions of people who have never started up a car or taken a cross-country flight.

Many animal and plant species will also pay a price for our prolific burning of fossil fuels.

Food, and the lack of it, could be where a changing climate exerts some of its most troublesome impacts for society.

While the changes could affect ranching and grazing as well as arable farming, much of the research to date has focused on croplands.

Because of longer dry spells, hotter temperatures, and more climatic uncertainty, the twenty-first century is likely to see major shifts in the crops sown and grown in various regions.

Well-off countries might break even or even benefit from the changes early on, if they can keep a close eye on the process and adapt their agriculture efficiently.

This is especially the case for large midlatitude countries, where crop zones can shift yet remain within national borders.

Sadly, the same may not be true of the tropics, where most of the world's food crops are grown.

The most problematic impacts on agriculture may wind up occurring in the poorest countries, those with the least flexibility and the most potential for catastrophic famine.

On top of all this, there's global population growth and the accompanying increase in food demand.

Only about a third of Earth's land surface is used for farming and other purposes.

What global warming does to parts of the other two-thirds—the world's natural ecosystems—could be even more wrenching than the effects on managed lands.

Pollution, development, and other side effects of civilization have already put uncounted species at risk, and now climate change threatens to make the situation far worse.

These ecosystems provide a surprisingly wide range of goods and services to society, an often underappreciated bounty that may be at more risk than we realize.

As with agriculture, the tropics appear to be particularly vulnerable when it comes to wildlife.

The normally muted variations in temperature at low latitudes imply that it would take only a small warming to push organisms into conditions they've never faced before.

This concept was put in stark terms by a 2013 study from the University of Hawai‘i at Mānoa.

Scrutinizing a blend of 39 models used for the latest IPCC assessment, they calculated the point at which annual global temperatures would stay completely above the yearly range observed from 1860 to 2005.

Globally, in a high-emissions scenario, the estimated date for this transition is 2047, but many tropical regions would pass the threshold as soon as the 2030s.

With large areas entering uncharted atmospheric territory, many creatures would have to venture far in order to stay within a familiar climate regime.

Along these lines, a 2009 study by the Smithsonian Tropical Research Institute studied tropical mammals whose present range is no more than a few dozen kilometers.

They found that more than 20% of those species would have to travel at least 1000 km (600 mi) in order to get to a place where temperatures by the year 2100 will be comparable to readings they experienced in their habitats in the 1960s.

How many creatures could simply evolve in place?

Many forms of life are quite adaptable in their behavior, and when faced with a pressing crisis, such as new competition for food, some species can even go through physical evolution remarkably quickly—as fast as 22 years, in the case of one type of Darwin's finch whose beak size changed to improve its access to seeds.

However, most adaptation to past climate change unfolded over much longer time scales, and it's not yet clear which species could keep up with today's comparatively rapid warming.

In a 2013 survey of more than 500 birds, mammals, amphibians, and reptiles, Yale University and University of Arizona found that the average rate of climate-driven adaptation for most groups corresponded to a temperature rise of less than 1°C per million years —which is thousands of times slower than this century's climate may demand.

Some populations have already lost the race.

A 2016 study in *PLOS Biology* documented more than 450 cases of "local extinction," where species have disappeared from the warmest reaches of their domains as temperatures rise.

A fitting poster child for species loss might be the Bramble Cay melomys, a rodent now believed to be extinct after floods boosted by sea level rise swamped its only domain, a tiny island in Australia's Great Barrier Reef.

It's the first mammal on Earth whose loss has been directly tied to the effects of global warming.

One landmark study of the extinction risks from climate change, led by the University of Leeds and published in 2004, looked at regions that together span a fifth of the planet's land surface.

The report found that, by the year 2050, 15%–37% of plant and animal species across these areas would be committed to eventual extinction should a middle-of-the-road scenario for increased emissions come to pass.

If emissions are on the high side, the range jumps to 21%–52%.

The IPCC's 2007 and 2014 reports reinforced the grim the picture painted, with the 2014 report noting that the risks of species extinction rises under all of the scenarios examined.

As models of species risk have incorporated more variables, the range of possibilities has grown.

This means an even wider swath of species could be in jeopardy—or, on the other hand, if we're lucky, perhaps a smaller number than previously thought.

While those who deny or downplay climate change may note that species have come and gone in vast numbers long before humans entered the picture, that's not exactly a reassuring thought, especially since climate change appears to have played a role in some of the biggest mass extinctions in Earth's history.

The canaries of climate change

Miners once brought canaries into their dank workplaces to test whether the air was safe to breathe; if the canary died, it was time to get out.

Similarly, some creatures can't help but reveal the mark of a shifting climate through their very ability—or inability—to survive.

Amphibians and **reptiles** offer some of the most poignant examples of species at risk.

Because these are ectothermic (cold-blooded) creatures, they must keep themselves within a fairly restricted range of air temperature, and they've got to crawl wherever they go.

There's no doubt that amphibians are in serious trouble.

More than 30% of amphibian species were found to be vulnerable, endangered, or critically endangered in the most recent comprehensive survey, the 2004 Global Amphibian Assessment.

The most recent update, in 2008, reiterated that diagnosis, and it found populations declining in more than 40% of existing species.

At least 159 amphibian species may already be extinct.

What's been challenging is to pin down the role of a changing climate amid many other threats to amphibians.

Temperatures in many key habitats haven't yet risen to dangerous levels, but there's some evidence that even modest warming can have significant impacts when it's accompanied by other risk factors, such as the presence of pesticides.

Frogs are among the species most at risk at many locations around the globe.

One of the biggest emerging threats to amphibians is the skin-damaging, reflex-slowing fungus known as **chytridiomycosis.**

It's been implicated in amphibian loss over the last couple of decades in many locations around the world, including the **Monteverde Cloud Forest Reserve** in Costa Rica.

Twenty out of 50 species of frogs at the preserve vanished after the unusually warm, dry spring of 1987.

One special species found nowhere else, the exotically colored **golden toad**, numbered at least 1500 during that La Niña spring.



Golden toads were wiped out by a fungal disease during a La Niña spring in Costa Rica. (Charles H. Smith, U.S. Fish and Wildlife Service)

Within two years, it could no longer be found.

Researchers were puzzled: chytridiomycosis tends to prosper in cool, moist weather, yet the Monteverde Cloud Forest Reserve was warming overall.

However, it turned out that the bulk of the warming at Monteverde was occurring at night, whereas days had actually cooled slightly (probably due to increased cloud cover).

The reduced day-to-night range could have kept conditions closer to optimal for the villainous fungus, although the timing might simply have happened to coincide with the natural course of the fungal epidemic.

A broader look at these relationships, led by Jason Rohr (University of South Florida), suggests that the main climatic culprit may actually be increased temperature variability—as during a shift from El Niño to La Niña years—which could act to weaken the frogs' defenses against fungal invasions.

In 2018, genomic analysis showed that chytridiomycosis first appeared in the early to mid-twentieth century in East Asia.

This finding may help clarify how our changing climate could have intersected with other factors—including the growth of international trade in amphibians—to help trigger the global die-off.

Butterflies are an excellent indicator of climate change on the march, partly because these widespread and aesthetically pleasing creatures have been carefully studied for centuries.

Both drought and flood can cause a butterfly population to crash, as happened to 5 of 21 species studied during a dry spell in the mid-1970s across California.

The severe drought and record heat across the U.S. Midwest in 2012 took a major toll on the iconic swarms of monarch butterflies that grow there before migrating south to Mexico each fall.

The number of acres occupied by monarchs in Mexican reserves dropped by 49% in the winter of 2012–2013 compared to the previous year, with only a modest rebound later in the 2010s.

Dry, hot conditions not only impede the summertime growth of the butterflies, but they also reduce the prevalence of milkweed, where monarchs lay their eggs.

Milkweed is also attacked directly by the herbicide glyphosate, which became common across the U.S. Midwest in the 1990s.

Since then, monarch production across the Midwest has reportedly dropped by more than 80%.

In the late 2010s, the U.S. Fish and Wildlife Service was evaluating whether to protect the monarch butterfly under the Endangered Species Act.

Because climate zones are concentrated in mountainous regions, the effects on butterflies are especially evident there.

A 2010 study found that many California butterflies are moving upward as the climate warms, with an average elevation gain per species of nearly 100 m (330 ft) over a 20-to-30-year period along the west slope of the Sierra Nevada.

The Apollo butterfly, whose translucent wings are often seen across high-altitude areas in Europe, is no longer found at elevations below 838 m (2749 ft) in France's Jura Mountains.

Although they're skilled fliers, the Apollo butterflies are finding themselves with less and less livable terrain as they're forced to climb in elevation.



The Apollo, one of many butterfly species being pushed to higher altitudes by warming temperatures. (Wenkbrauwalbatros/Wikimedia Commons)

A major survey of European butterflies in 2009 found that dozens of species will be unsustainable in their present locations if the warming expected by late this century comes to pass.

“Most species will have to shift their distribution radically to keep pace with the changes,” said a report from the Helmholtz Centre for Environmental Research.

Even a creature as humble as the **dormouse** can tell us something about global warming.

Made famous by Lewis Carroll’s *Alice in Wonderland,* this tiny, nocturnal mammal now ranges only about half as far across Britain as it did in Carroll’s time, and more than half of roughly 3000 British species of dormouse are on the decline.

Part of this is due to the fragmentation of the trees and hedgerows where it likes to live.

However, mild winters interfere with its hibernation patterns, and warm summers cause additional stress.

Were it not for so much civilization in the way, the dormouse might easily be able to migrate north to escape the warm-up.

A program to reintroduce the creature in various parts of Britain has found some success.

Trouble brews when interdependent species respond in differing ways to a warming atmosphere.

Those who use temperature as their cue to breed or migrate may shift their timing as warming unfolds, while those whose behavior is driven more by the seasonal waxing and waning of sunlight may not.

At the Netherlands Institute for Ecology, they hav been studying disrupted synchrony—the mismatched timing that can occur when creatures find that a food source has departed from its usual seasonal schedule.

In studying a bird called the Dutch great tit, for example, they found that the birds' egg-laying behavior hasn't kept up with the earlier springtime emergence of its main food source, caterpillars.

Similarly, in California, a Stanford University team found that the progressively earlier seasonal die-off of plantains on a nearby ridge left the Bay checkerspot caterpillar without its favorite food.

The butterfly eventually disappeared from the ridge.

Such timing mismatches could also threaten the relationship between plants and the insects that pollinate them. NASA is working to relate the amount of nectar in beehives (easily measured by weighing the hives) to the satellite-detected timing of vegetation.

The agency's HoneyBeeNet reports that the peak nectar flow in Maryland hives now occurs about a month earlier than in the 1970s.

It's not only insects that are threatened by asynchrony.

Pennsylvania State University has found that the plants eaten by **caribou** calves in Greenland are emerging earlier than they did in the 1990s, as depleted sea ice allows air temperatures to warm across the region.

However, the seasonally driven birth dates of the calves aren't changing, so the plants may now be past their nutritional peak by the time the caribou start foraging.

As they put it, "The animals show up expecting a food bonanza, but they find that the cafeteria already has closed."

For other animals, it's camouflage—or the lack of it—that puts them at risk.

North Carolina State University has described the effect climate change may have on a number of mammal species across the globe that seasonally molt from brown to white to camouflage themselves against their background.

They found that **snowshoe hares** have little ability to shift the timing of their coat color from brown to white or white to brown when the winters are short.

Since fall and spring snowfall is decreasing throughout the globe in temperate regions, the work demonstrates that the odds are rising that hares will be caught with their white coats mismatched against bare ground during these transition seasons, a time when their mortality rates are already the highest.

Do examples like the ones above add up to a **global shift?**

An extensive survey published by the University of Texas at Austin and Wesleyan University found that 279 out of 677 species examined in large-scale, long-term, multi-species research showed the telltale signs of human-induced climate change.

On average, these species had moved roughly 6 km (3.7 mi) poleward, or about 6 m (20 ft) up in elevation, per decade.

Further evidence came with an exhaustive global survey in 2008, led by NASA of more than 29,500 datasets that each showed a significant physical or biological change related to temperature.

About 90% of the biological changes were in the direction one would expect from a warming climate.

In an 2016 meta-study University of Florida showed that climate change is affecting more than 80% of a set of core ecological processes, ranging in scale from genetic to physiological to ecosystem-wide.

While the movement of species and the risks of extinction have gotten much attention, climate change is affecting some animals in a less-publicized way: they're getting smaller.

A 2011 survey by the National University of Singapore assessed dozens of studies that found shrinkage among various types of plants, fish, amphibians, birds, and mammals.

Smaller size can be a direct result of stressors such as drought or ocean acidification, but it can also serve as an adaptation that helps some heat-stressed creatures survive.

A lower weight generally means more surface area per pound, which helps animals dispel heat more readily.

The big squeeze

Simply keeping up with a changing climate will be a challenge for many animals and plants.

It's hard to overstate how quickly the chemical makeup of the atmosphere is changing relative to Earth's history.

The unprecedented increase in greenhouse gases translates into a warming projected for the next century that may seem gradual to us but will be warp speed by geological standards.

In the Arctic, where the range of life forms that can survive the harsh environment is relatively narrow, it's easy to identify the animals most at risk, chief among them **polar bears**.

Elsewhere around the planet, ecologists are keeping a close eye on **biodiversity hotspots,** regions where climate and geography team up to nurture a vast variety of species—far more than one might expect in such small areas.

The 35 hotspots identified to date by researchers are scattered throughout the tropics and midlatitudes, largely on coastlines and mountain ridges.

The Monteverde Cloud Forest Reserve mentioned earlier is part of one of the largest hotspots, the Mesoamerican forests that cover much of Central America.

Most hotspots feature large contrasts in temperature and moisture across small areas, which enables many different types of flora and fauna to find their niches.

The flip side of this concentrated contrast is that a seemingly small shift in climate could have major effects on plants and animals accustomed to a very specific environment.

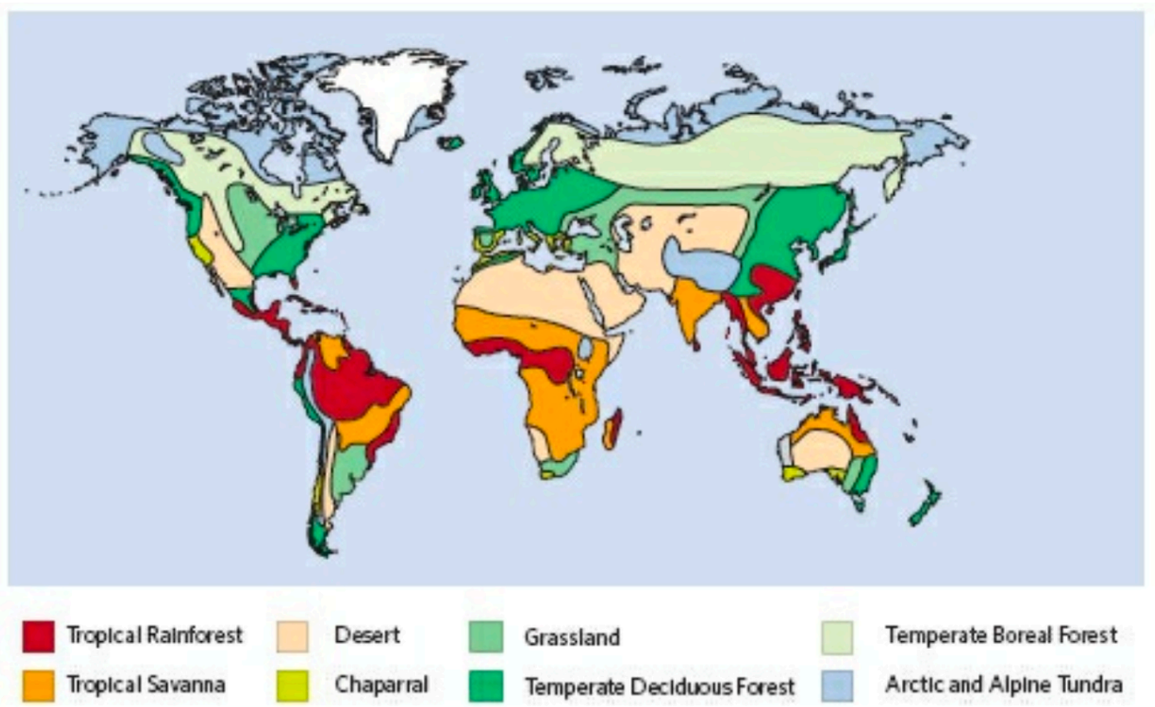
Moreover, because many hotspots are localized—often surrounded by more uniform, human-dominated landscapes—plants and animals may not be able to easily shift their ranges outward.

What about the rest of the tropics and midlatitudes?

These areas are often classified by geographers into **biomes**, regions that share similar traits of climate and landscape.

Among the commonly used groupings for land areas are the following:

- **Forests.** Covering roughly a third of the planet, these include the great boreal forests of Canada, Alaska, and Russia; the evergreen and deciduous forests of midlatitudes; and the tropical rainforests that support amazing biodiversity despite nutrient-poor soil.
- **Grasslands.** The savannah that covers much of Africa, Australia, and southern South America, and the temperate grasses that sprawl across the mid-sections of North America and Eurasia as well as Argentina.
- **Deserts.** About a fifth of Earth's land surface. Many of the largest deserts are clustered in the zones around 30°N and 30°S, where air tends to sink and dry out. Most of Antarctica is also considered desert.
- **Tundra.** The cold lands of high altitudes and the Arctic that support only shallow-rooted, low-rising plants.



The edges of many of these biomes are shifting already.

Some studies in Europe have shown a northward push of up to 120 km (75 mi) in the north edge of certain biomes over the last century.

The tropics themselves—as demarcated by high-level jet streams—widened by roughly 180 km (110 mi) between 1979 and 2009, according to satellite data analyzed by scientists at the University of Washington and China's Lanzhou University.

Other work suggests an even greater widening may have occurred.

Implied in this expansion is a poleward push of subtropical deserts, although the greatest expansion may be occurring over the oceans, according to a 2017 analysis by the University of Virginia.

Farther north, one of the largest-scale shifts expected over the next 100 years is the spread of **boreal forest** poleward into fast-thawing tundra and, in turn, the transformation of the southern edge of some boreal forests into deciduous forests or grasslands.

But forests take time to establish themselves, and climate may outrun the process, leaving vast areas in transition until soils and plants can adapt to a new climate regime.

To make matters worse, any climatic state we envision for the future is itself a moving target, since the greenhouse gases we're adding to the air now will be warming the climate for a century or more.

Will new forests be able to take root in these ever-shifting conditions?

As for land-based creatures in the tropics and midlatitudes, it might seem as if they can easily migrate poleward and upward as needed.

The problem is that most of the planet's terrain outside hotspots and other protected areas is pockmarked with towns and cities and laced with highways, cultivated fields, and other potential barriers to migration.

Those creatures adapted to higher altitudes, like the pika, may find themselves stranded on island-like patches of terrain, where climate will only constrict their range further over time.

A growing focus of ecological research is the need for **refugia**—a term originally applied to locations where warmth-loving creatures survive ice ages and where those who prefer cold make it through interglacial periods.

While such havens will remain critical for species over the very long term, there's a more immediate need for protection from human-produced climate change.

This might be found in larger areas where species can migrate (macrorefugia) or in favorable pockets (microrefugia)—perhaps some as small as a few acres—that allow creatures to hunker down in place.

Whether it's ecological hotspots or the Arctic, if the exotic animals that favor these rarefied environments are potentially the biggest losers in climate change, the biggest winners may be the most mundane.

Animals that are common and robust could spread even farther, such as the various species of deer now thriving across rural and semirural parts of North America and Eurasia.

A landscape of shifting climate may also offer a nearly ideal platform from which nonnative plants, animals, and insects can spread like wildfire.

In this era of mass global travel, it's easier than ever for **invasive species** to tag along and then move poleward with warming temperatures.

Kudzu, a vine that grows as quickly as 22 cm (9 in.) per day, is a classic example.

It was brought from Asia to Philadelphia in 1876 for the nation's centennial exposition, then used for decades across the U.S. South to stabilize roadsides and feed animals before its invasive qualities were recognized.

Long since out of control, kudzu has been spreading steadily northward: in 2009 Canada's first-ever kudzu colony was found on the north shore of Lake Erie.

The most successful invasive species are often less picky than native species, able to tolerate a wide range of conditions, including heat and/or drought.

One famous example is **tumbleweed**, or Russian thistle.

Introduced to the U.S. Great Plains in the 1870s, tumbleweed quickly spread across the U.S. frontier, eventually becoming a symbol of its new home.

Similarly, **rabbits** multiplied like, well, rabbits after 29 of them were brought to Australia in 1859 by rancher Thomas Austin.

Today, the United States is dealing with the destructive and sometimes deadly **fire ant**, a major invasive pest that's ravaged both the countryside and populated areas and made its way to the Eastern Hemisphere.

Where invasive flora and fauna are limited mainly by cold temperatures, a warming climate should only enhance their ability to spread.

Marine life is also likely to undergo major reshaping as the planet warms.

Bugged by a changing climate

Many **insects** respond strongly to small shifts in temperature, rainfall, and humidity.

As a rule, warmer temperatures should allow a poleward spread of many insect species.

That doesn't necessarily mean each type of bug will be found over ever-larger chunks of the globe.

In fact, the first comprehensive analysis of parasites (including insects as well as other types) found in 2017 that habitat loss from climate change could render up to 10% of all parasite species extinct by 2070, even as the most successful adaptors flourish.

One big challenge is to determine how climate change that favors infectious agents will intersect with changes in where and how people live.

History gives us a glimpse of what can happen when climate heats up on a large scale.

The so- called Medieval Warm Period—the several centuries of relatively mild weather across much of the Northern Hemisphere between about AD 950 and 1300—may have helped rats and fleas to thrive across Europe, paving the way for the **Black Death** that killed a third of Europeans between 1347 and 1352.

More than two billion people live within the range of mosquitoes that transmit the parasites that cause **malaria.**

Happily, protective measures such as insecticidal bed nets and mosquito spraying helped bring down the global death toll by nearly 50% from 2000 to 2016, according to the World Health Organization.

Still, malaria was affecting more than 200 million people and taking more than 400,000 lives each year.

It's obvious that any future rebound in infection rates would carry a huge toll in terms of human suffering, yet it's been difficult for researchers to nail down exactly how climate change will affect malaria, given the variety of other factors that shape the course of the disease's spread.

For more than 30 years, malaria has been moving into previously untouched parts of Africa's western highlands, possibly due to a scarcity of cool, mosquito-killing temperatures.

It was long assumed that warming temperatures would expand the range of malaria in such a way, thus widening the swath of humanity threatened.

However, recent models suggest that the range of malaria-carrying mosquitoes might simply shift rather than expand.

Intensified drought, for example, may render some areas unable to sustain malaria-carrying mosquitoes even as the disease moves into other areas, such as African highlands, that were once malaria-free.

A 2013 modeling study from University of Cologne suggests that malarial risk should decrease by midcentury across much of tropical Africa while increasing in some other parts of the continent, including the East African highlands (especially at altitudes above 2000 m or 6600 ft) as well as the southern Sahel, where the population is growing rapidly.

In places like Europe and the United States, where malaria was endemic for centuries, the disease is considered unlikely to make a comeback—even with warmer temperatures—as long as control measures and public-health systems remain strong.

Reassuring as this progress might seem, mosquito populations are booming across many of the developing regions where malaria is most prevalent.

And as noted above, warming temperatures may be helping the virus spread into areas where people aren’t accustomed to it and may not have the resources to deal with it.

Together, these factors pose continuing challenges to public health efforts in chronically poor areas.

Unlike malaria, **dengue fever** is on the upswing in many parts of the globe.

It's caused by a group of five potentially fatal mosquito-borne viruses, one of them just discovered in 2013.

Close to half of the world's population now lives in affected areas, according to the World Health Organization.

Anywhere from 50 to 100 million people are infected with dengue each year, of whom around 500,000 need to be hospitalized.

Children account for a large proportion of these serious cases, including most of the fatalities.

One early study estimated that a 1°C (1.8°F) rise in average global temperature could raise the number of people at risk of dengue by up to 47%, but the complex factors that influence transmission make it hard to know if the disease would actually spread that quickly.

A more recent study found that increased economic development has the potential to more than counteract any climate-related boost in dengue's spread by 2050.

Still, the disease has been pushing its way into developed countries: a few cases have been reported from Texas to Florida since 2009, and the first European outbreak in almost a century struck the Portugal islands of Madeira in 2012.

Researchers have been investigating whether climate change and other factors might help dengue to spread from lower elevations east of Mexico City into the high-altitude urban core, where it would threaten close to 20 million people.

The project has already found *Aedes aegypti*, the primary dengue-carrying mosquito, at altitudes of 2430 m (7970 ft), the highest on record for its presence in Mexico.

Unusual summertime warmth and drought helped another mosquito-borne illness make quick inroads into North America.

The **West Nile virus** first showed up around New York City in 1999.

Hot summers can be favorable for proliferation of mosquitoes that transmit the virus, which is hosted by birds in the wild and can spill over into humans.

The virus spread regionally at first and then, during the record-breaking heat of 2002, swept across much of the United States and into Canada. Up to 2005 it had killed more than a thousand North Americans.

By the end of the decade, the annual human toll had dropped rapidly, perhaps because birds were becoming immune to the virus or possibly because fewer humans with low-grade infections were being tested and counted.

However, the disease again spiked during the torrid summer of 2012, with 286 U.S. deaths reported—the largest annual toll since the virus first reached American shores.

Insects can also cause trouble for people indirectly, such as by attacking the food we eat.

The Colorado potato beetle, a striped, centimeter-long U.S. native, is a serious threat to potato crops in Europe.

Each year a few of these beetles hitchhike into the United Kingdom, but to date the bug has failed to establish a foothold, thanks to vigilant inspections, public awareness efforts, and eradication campaigns by the U.K. government.

However, a study by the U.K. Central Science Laboratory found that if Europe were to warm by an average of 2.3°C (4.1°F) by 2050, this pest could more than double its potential range in Great Britain, putting virtually all potato-farming areas at risk.

Shrinking forests

As we've seen, rainforests and their loss can have a huge impact on the global climate.

Russian scientists theorize that a large tropical rainforest such as the Amazon serves as a **biotic pump**, with the trees themselves helping to perpetuate the atmospheric flow that sustains rainfall.

On a seasonal scale, it's precipitation more than temperature that drives the rainforest, so anything that affects when and how hard it rains can have a major impact.

El Niño tends to produce drought across much of the Amazon, so a key factor will be how El Niño evolves in the future, which is still an open question.

In addition, it appears that unusually warm Atlantic waters northeast of Brazil can help strengthen Amazonian drought.

The Atlantic may also be supplying enhanced moisture to the Amazon during non-drought years, which could help explain why the region has become wetter on average since 1990, despite two intense droughts in 2005 and 2010.

The various global climate models considered in the 2013 IPCC report tend to support this trend toward a beefed-up hydrologic cycle across the Amazon.

Under the highest-emissions scenario for the 2080s and 2090s, the maximum five-day rainfall expected in a given year rises by 10% or more, while each year's longest dry spell is extended by a week or more.

Should climate change end up pushing the Amazon toward more bona fide drought, the trend could be exacerbated by land clearing, which causes local-scale drying due to loss of the moisture- trapping forest canopy.

The fear is that these and other positive feedbacks could lead to massive rainforest loss across the Amazon, although that scenario is far from a given.

Models that track climate together with the interactions of species within a rainforest canopy are still early in development, but they may eventually clarify how tropical forests as a complete ecosystem stand to deal with drought.

There's more certainty about what climate is doing to forests farther toward the poles.

Mountain and boreal forests are one of the largest ecosystems already being strained by a rapidly changing climate.

Permafrost melt is one big source of high-latitude forest stress, but there are other major factors.

Across western North America, the intersection of pests and a warming atmosphere has led to one of the world's largest **forest die-offs.**

A series of major droughts and warm winters since the 1990s from Mexico to Alaska has transformed the landscape across huge swaths, destroying millions of hectares of forest through **beetle invasions** and **forest fires.**

The extent of both is unprecedented in modern records. In 2002 alone, fire and disease killed enough lodgepole pines across the Canadian province of British Columbia to cover the entire area of Belgium and the Netherlands.

As of 2012, the total beetle-affected area had reached 180,000 km^2 (69,000 sq mi).

Roughly half of British Columbia's pines are now gone, and the total may approach 60% by the 2020s.

Even as the epidemic slows in that province, it's been gaining ground toward the northeast into Alberta.

The main insect behind this devastation is the **bark beetle**, which comes in various species across North America.

This hardy creature burrows into trees, lays its eggs, then girdles the tree from within, blocking the nutrient flows between roots and branches.

Climate in recent decades has given North American bark beetles carte blanche for explosive growth.

The spruce bark beetle, one of the most common varieties in colder forests, normally requires two years to reach maturity, but an extremely warm summer can allow a generation to mature in a single year.

Conversely, it takes two bitterly cold winters in a row to keep the population in check.

The combination of milder winters and warmer summers since the 1990s has allowed the beetle to run rampant.

Another variety, the **mountain pine beetle**—once limited to lower-elevation tree species such as lodgepole and ponderosa—has moved uphill as temperatures warm up.

It has invaded whitebark pine in Utah at elevations up to 3000 m (10,000 ft).

In Colorado and Wyoming, the beetles' territory doubled between 2007 and 2010, encompassing 14,000 km^2 (5400 sq mi).

Though it's now peaked across the region, the epidemic has left swaths of dead or dying forest covering some of the nation's most famous scenery, with pine needles turning a sickly red before they fall off.

Hikers must now keep a wary eye for dead trees that can tumble into campgrounds and onto trails.

In Canada, the beetles are moving into boreal forests that were long thought to be less vulnerable than their southern counterparts.

They've also begun to reproduce within **jack pine**, a critical move that could allow the insects to cross the continent and lead to massive infestations across the forests of eastern North America.

At the same time, the southern pine beetle, endemic from the U.S. South into Central America, is starting to make its own name a misnomer.

Once limited to areas south of the Mason–Dixon line, the beetle has established a recent foothold in the pine forests of southern New Jersey, and by 2015 it was being observed as far north as Connecticut.

As the coldest nights of each winter gradually warm up, the beetle's range may extend into southern Canada by midcentury, according to a 2017 study from Columbia University.

Drought weakens the ability of trees to fight off these aggressive bugs, because the dryness means a weaker sap flow that's easier for beetles to push through.

Across parts of New Mexico, Arizona, Colorado, and Utah, where a record-setting drought took hold in 2002, pine beetles and other drought-related stresses killed up to 90% of the native **piñon** trees.

A drought in the 1950s had already taken many century-old piñons, but the more recent drought produced "nearly complete tree mortality across many size and age classes," according to a 2005 study from the University of Arizona.

Recent research suggests that piñon could manage to hang on in regional pockets that are cool and moist enough, although continued warming and drying will threaten even those sanctuaries.

Certainly, both insects and fire are part of the natural ecosystems of North America.

Some species of trees actually rely on fire to propagate.

Jack pine seeds are tucked into tightly sealed resinous cones that only open when a fire comes along and melts the resin, which then allows the seeds to spread and sprout in the newly cleared landscape.

The question is how climate change might amplify the natural process of forest fires, especially across the high latitudes of North America and Asia.

For example, some studies project an increase of up to 100% by the end of this century in the amount of land consumed by forest fire each year across Canada.

These fires are a major source of pollution, and they pack a double whammy on the carbon budget: the fires themselves add significant amounts of carbon dioxide to the air (not to mention carbon monoxide, methane, and ozone-producing oxides of nitrogen), and the loss of trees means less CO_2 being soaked up from the atmosphere (at least until new growth occurs).

As for the beetles and budworms, most major infestations tend to run their course after a few years, followed by new epidemics a few decades later as the next crop of tree victims matures.

No one knows exactly how that natural ebb and flow will intersect with the climate-driven shifting of the insects' potential range across the U.S. and Canadian West.

It’s also not out of the question that a decades-long **megadrought** could set in.

It’s happened before, at least across the U.S. Southwest.

Lamont–Doherty Earth Observatory warns that by 2050 that region may be in the grip of a warming-driven drought that’s more or less permanent.

Already, a large swath from New Mexico to California has been struggling with on-and-off, but mostly “on,” drought.

Most winters from 2000 to 2018 fell short of average rainfall in Los Angeles, and the period from July 2006 through June 2007 was L.A.’s driest in more than a century of record keeping, with a mere 82 mm (3.21 in.) of rain.

Five years of grinding drought kicked in from 2011 to 2016 across most of California: in 2013, both Los Angeles and San Francisco smashed all-time calendar-year records for low precipitation.

Meanwhile, Lake Mead, which slakes the thirst of the growing Las Vegas area, dipped to water levels in the early 2010s unseen since the lake was created in the late 1930s.

The long-term risk to Las Vegas became dire enough to trigger construction of a third intake valve—a new "straw," opened in 2015, that will allow the city to continue drawing water even as lake levels reach new lows.

All of these issues complicate long-running debates on how best to manage ecosystems across western North America.

The overarching question is how to allow natural processes to play out while fending off the kind of large-scale, fire- or bug-produced devastation that could permanently alter the landscape.

Population growth at the urban–wildland interface is another key issue.

The more homes that pop up at the edge of forests, the greater the odds that humans will trigger wildfires—either inadvertently or intentionally—and the higher the stakes for both people and ecosystems once those fires take off.

Crops and climate: A growing concern

In recent years, a variety of factors (no doubt including the influence of factory farming) has led to an enormous global reliance on a tiny number of staple crops.

Out of some 50,000 edible plant species, a mere three—wheat, rice, and maize (corn)—make up more than half of the world's current food intake by calorie.

To deal with the demands of climate change, especially over the long haul, food producers around the world will need to demonstrate flexibility in the timing, location, and nature of their planting.

That won't be easy when there is no single, fixed "new" climate to plan for, but rather a climate in never-ending flux.

The challenges will be especially great for subsistence farmers whose options are restricted even in the best of times.

The complex effects of climate change on plants are often reduced to a simplistic, and misleading, equation: plants take in carbon dioxide and use it to grow, so increasing the amount of CO_2 in the air can only be a good thing.

One of the more prominent U.S. organizations stressing the potential benefits of a world with more carbon dioxide was the now-defunct Greening Earth Society.

From the 1980s into the 2000s, this group promoted the view that the extra CO_2 in the air would help crops and other vegetation to grow more vigorously.

As they pointed out, the tendency of global warming to be most pronounced in the winter and at night should only lengthen growing seasons on average across much of the world.

By itself, the fertilization effect of carbon dioxide isn't controversial; it's been supported by research in the laboratory and in the field, albeit with a few caveats, and underscored in many climate reviews, including the IPCC's.

To cite just one prominent study, the Maryland-based Smithsonian Environmental Research Center reported in 2010 that a set of 55 plots of trees monitored there for 22 years grew at up to twice the expected rate.

Among other things, the enhanced CO_2 triggers a shrinkage of the pores on plant leaves, allowing them to retain moisture more easily.

A 2013 analysis headed from Harvard University found that trees at more than a dozen spots across the Northern Hemisphere had become increasingly efficient in their use of water over the last two decades—even more than models had predicted.

Again, enhanced CO_2 and its effect on plant pores turned out to be the best explanation for the trend.

Increased carbon dioxide also speeds up photosynthesis rates for so-called C_3 plants, including wheat, rice, and soybeans.

In one experiment, the yields of C_3 crops increased by an average of 14% when carbon dioxide was boosted to 590 ppm compared to 370 ppm.

Livestock that graze on C_3 grasses, which are mainly at higher latitudes, could also benefit.

In contrast, C_4 crops, including maize and sorghum, gain relatively little benefit from the extra CO_2, as they don't get the boost in photosynthesis that C_3 plants do.

Among nonagricultural plants, woody vines such as kudzu appear to thrive on the extra CO_2.

This is in part because their clingy mode of growth allows the plant to devote more energy to photosynthesis as opposed to building a structure to keep it standing.

A 2006 study at Duke University found that one noxious vine—poison ivy, which sends over 350,000 Americans to the doctor each year—grew at three times its usual pace with doubled CO_2.

To make matters worse, the ivy's itch-producing chemical became even more toxic than before.

The most recent analyses have brought down the overall agricultural benefit we can expect from carbon dioxide increases, in part because the negative influence from **near-surface ozone** will be larger than scientists once thought.

One of the open-air studies carried out as part of the Free Air CO_2 Enrichment program, or FACE (see sidebar "Will greenhouse gases boost plant growth?"), tested the effects of enhanced carbon dioxide and ozone on soybean fields.

A doubling of atmospheric CO_2 boosted soy yields by about 15%, but that benefit went away when ozone levels were increased by as little as 20%.

Another open-air study, this one in Wisconsin, obtained similar results for aspen, maple, and birch trees.

Ozone may also cut the nutritional value of some crops, although research is still in its early days.

Global warming itself will have an impact on crops, of course, some of it favorable.

Modeling that fed into the 2013 IPCC assessment shows how **frost days** (the number of 24-hour periods when temperatures drop below freezing) might evolve over the next century.

The number of such days drops by more than a dozen by the end of this century under the midrange RCP4.5 emissions pathway, with the largest changes across far western North America, eastern Europe, and central Asia.

That, in turn, should lead to somewhat longer growing seasons across much of the planet's midlatitudes.

Already, plant hardiness zones across the contiguous United States have shifted north in many areas, according to maps produced in 1990 and 2012 by the U.S. Department of Agriculture.

Across the globe, scientists expect a hotter climate to enhance the overall potential for cereals and some other crops across large parts of northern Europe, Russia, China, Canada, and other higher-latitude regions.

This should lead to an overall expansion of the world's land areas conducive to agriculture.

However, that alone doesn't guarantee tomorrow's plants will be just as healthy as today's.

The protein level in grain crops can decrease if nitrogen doesn't increase in tandem with carbon dioxide, and a number of studies to date show that CO_2-enhanced crops can be significantly depleted in zinc, magnesium, or other micronutrients, perhaps because there aren't enough trace elements from the soil entering the plant to keep up with the photosynthesis boost from CO_2.

Winners and losers in farming

A battle royal is setting up for the coming decades across the world's farmlands, where the benefits of extra CO_2 and longer growing seasons will fight it out with intensified drought, spikes of extreme heat, invasive species, and other negatives.

There are both encouraging and worrisome signals.

On the down side, the fertilization effect of CO_2 for crops like wheat and rice tends to decrease once CO_2 is boosted beyond a certain point.

While some developed areas may benefit from climate change, at least for a while, these gains could be offset by thirsty soils and a resulting spike in irrigation demand.

Even if the world is lucky enough to get a few decades where agricultural winners match or even outnumber losers, the resulting imbalance could put huge stresses on agricultural trade and food security, including relations between developed and developing nations—and that's without considering the impact of converting food crops to fuel, or the relentless growth in meat consumption, which increases the amount of land and water devoted to food production.

On the plus side, the world's food production system has done an impressive job in recent decades in keeping up with the ever-increasing number of people on Earth.

From 1960 to 2015, world population more than doubled (from 3.0 to 7.3 billion), yet global harvests more than tripled, even though the amount of land being cultivated increased by only about 10%–15%.

The UN's Food and Agriculture Organization estimates that harvests will need to grow another 60% beyond 2005–2007 levels by 2050 in order to meet worldwide food demand.

Technology and innovation should continue to help boost total global yields for the foreseeable future, and the effects of global warming aren't expected to cause a literal drop in productivity anytime soon.

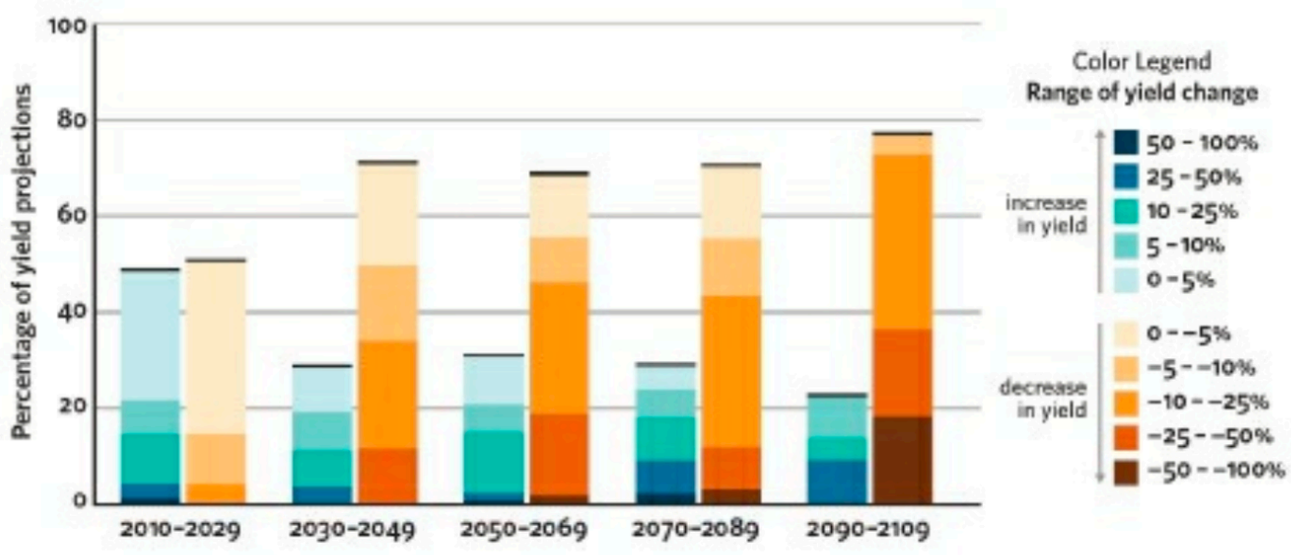
The main concern is that even a small climate-related dent in expected yields (especially if other problematic factors come into play at the same time) could have an outsized impact on a growing world that's come to expect a bountiful supply of food that increases year over year.

The question then becomes: how much will tomorrow's agriculture be held short of its potential as the planet warms up?

It now looks as if the net global effects of carbon dioxide and climate change may shift from positive to negative sooner than we once expected.

The 2014 IPCC report on impacts, adaptation, and vulnerability (Working Group II) brought this uneasy message home.

It found that the lion's share of studies now indicate that climate change will have an overall negative effect on crop yields as soon as the 2030s or 2040s (see graphic), with decreases of 25% or more appearing by that time period in some studies.



Projected changes in crop yield as a function of time, with and without adaptation, incorporating more than 1000 sstudies from across the world. For each time interval, each colored slice indicates the percentage of studies for that interval showing a given change in crop yields. (IPCC)

The verdict right now is hardly unanimous: as can be seen in the graphic, a substantial percentage of studies show that climate change may actually help *increase* global crop productivity.

Taking all this into account, the panel's consensus outlook is that median global yields can be expected to drop by 0%–2% per decade for the rest of this century, compared to the yields that would have occurred without a changing climate.

The panel notes that adaptation could make a real difference.

If farmers can adjust what they grow and when they grow it, working to match the evolving climate in their location, then some could actually come out ahead—at least in some areas, and assuming that local temperatures don't rise much more than 2°C over preindustrial values.

The 2014 IPCC report also took a region-by-region look at where climate change is likely to have the biggest impacts on agriculture.

Rising temperatures and changes in rainfall may trim the potential productivity of rice crops in Asia and cereal crops across Africa, possibly dealing a major blow to African food security.

Europe is a mixed bag, with climate change likely serving as a net boost to grains across northern Europe and perhaps eastward into parts of central Asia, while cereal crops toward the Mediterranean take an overall hit.

Much like Europe, North America could see net benefits to crops toward the north and detriments toward the south, although farmers across much of the continent will need to adapt to an enhanced risk of heat waves and precipitation swings.

That's even truer for Australia, where recurring floods and droughts already exert a huge impact on agriculture.

Regional projections of future agriculture must reckon with the potentially destructive role of extremes, such as the damage produced by hot, dry summers.

For example, the **groundnut**—a staple across western and southern India—produces far less fruit when temperatures consistently top 35°C (95°F) over a few days.

Many other crops have similar temperature thresholds.

Wheat that's exposed to 30°C (86°F) for more than eight hours produces less grain, and **rice pollen** become sterile after only an hour of 35°C heat.

Scientists are increasingly incorporating the mix of crops and climate into computer models.

Most studies of individual crops have been relatively small scale, while analyses of future climate are at their strongest in depicting patterns across large sections of the globe.

A natural intersection point is regional climate models, which zero in from global models to portray climate across a specific area in finer detail.

The expanding set of global and regional efforts is now being evaluated through the Agricultural Model Intercomparison and Improvement Project, which involves participants from around the globe active in crop, climate, and economic modeling.

As of 2018, the effort had already entrained dozens of models, including 15 for rice, 21 for maize, and 38 for wheat.

The project's ultimate goal is to improve the collective understanding of world food security and to enhance the ability of both developing and developed nations to adapt to changes down the road.

Meanwhile, agricultural specialists are working on hardier strains of key crops that can deal with potential climate shifts.

Half the world's population eats rice as a staple, so its fate on a warming planet is critical to world food stocks.

In the Philippines, the International Rice Research Institute (IRRI)—famed for its work in bolstering yields in the 1960s Green Revolution —is using non–GMO-style crossbreeding to develop strains of rice that can resist heat, drought, and flood.

The institute recently unveiled a variant that can produce a decent crop even after being fully submerged for two weeks.

One challenge in developing heat- and drought-suited crop strains is that they may prove less productive than traditional strains during good years.

Another critical step in formulating crops for a warming world will be to ensure that they stay as nutritious as their predecessors, given the effects of CO_2 on micronutrients noted above.

Otherwise, the silent threat of "hidden hunger"—nutrient deficiencies that are now estimated to affect as many as two billion people—could only get worse.

Technological gains may amount to little unless social and financial arrangements are in place to help even the most humble farmer take advantage of the progress.

Many subsistence farmers stick to tried-and-true crops and techniques, preferring to minimize risk given their lack of safety cushions.

International alliances such as the CGIAR (Consultative Group on International Agricultural Research) are looking into policy options such as new forms of crop insurance that could give risk-averse farmers more room to adapt to a changing climate.

Farmers throughout the tropics could also benefit from locally tailored information on climatic shifts.

Part 3 - The Science

HOW WE KNOW WHAT WE KNOW ABOUT CLIMATE CHANGE

10 - Keeping Track TAKING THE PLANET'S PULSE

It's customary to see polar bears, forest fires, and hurricane wreckage when you watch a video about climate change.

What you're unlikely to see are the dutiful meteorologist launching a weather balloon, the grandmother checking the afternoon's high temperature in her backyard, or a satellite quietly scanning Earth's horizon from space.

These are among the thousands of players—human and mechanical—who work behind the scenes to help monitor temperature and other variables that reveal the state of our planet.

There's no substitute for this backstage action.

Day by day, month by month, year by year, the people and instruments who take Earth's vital signs provide bedrock data for the tough decisions that individuals, states, and companies face as a result of climate change.

The rise of climate science

The practice of assigning a number to the warmth or coolness of the air has royal roots.

Galileo experimented with temperature measurements, but it was Ferdinand II, the Grand Duke of Tuscany, who invented the first sealed, liquid-in-glass thermometer in 1660.

The Italian Renaissance also gave us the first barometers, used for measuring air pressure.

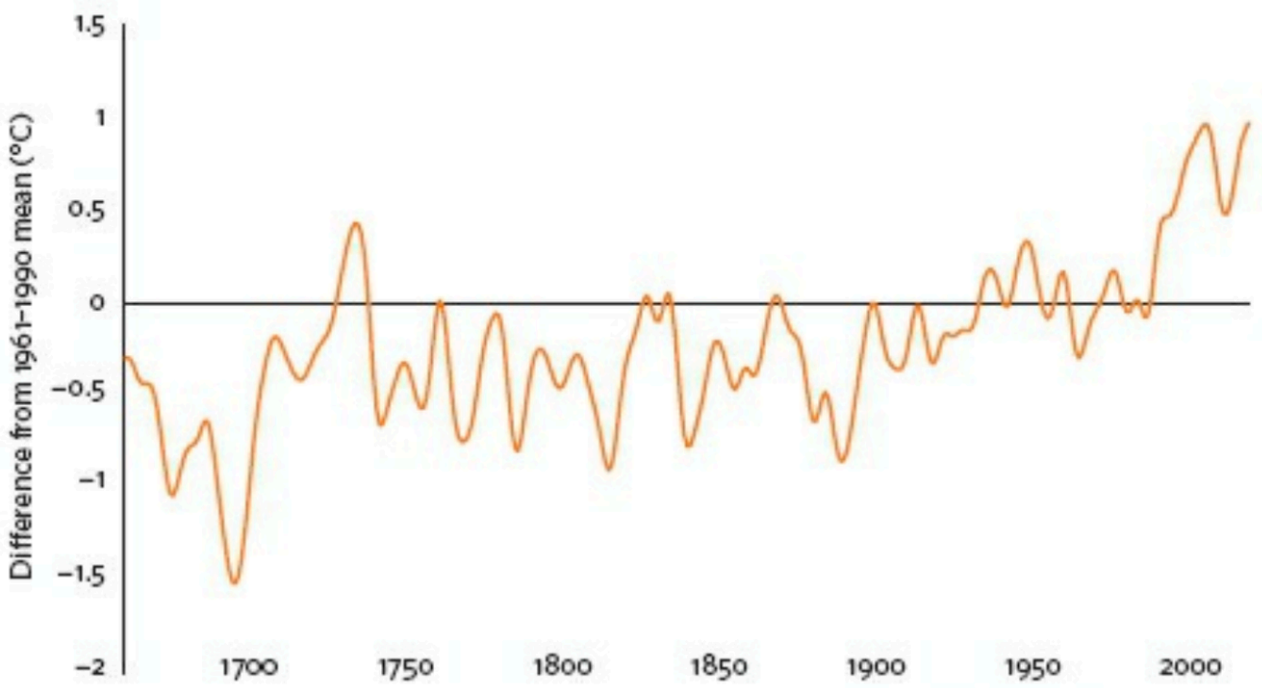
By the 1700s, weather observing was all the rage across the newly enlightened upper classes of Europe and the U.S. colonies.

We know it was a relatively mild 22.5°C (72.5°F) in Philadelphia at 1:00 p.m. on the day that the Declaration of Independence was signed—July 4, 1776—because of the meticulous records kept by U.S. cofounder and president-to-be Thomas Jefferson.

Some of the first sites to begin measuring temperature more than 300 years ago continue to host weather stations today.

Most reporting sites of this vintage are located in Europe.

A cluster of time-tested stations across central England provides an unbroken trace of monthly temperatures starting in 1659—the longest such record in the world—as well as daily temperatures from 1772 (see graph for annual averages).



Average annual temperature across central England, 1660–2017. (Met Office Hadley Centre)

Like the remnants of a Greek temple or a Mayan ruin, these early readings are irreplaceable traces of a climate long gone.

Such stations represent only a tiny piece of the globe, so on their own they don't tell us much about global-scale changes.

However, they do help scientists to calibrate other methods of looking at past climate, such as tree-ring analysis.

They also shed light on the dank depths of the **Little Ice Age**, when volcanoes and a relatively weak sun teamed up to chill the climate across parts of the world, especially the Northern Hemisphere continents.

A few regional networks were set up in Europe during those cold years, starting in the 1780s with Elector Karl Theodor of Bavaria, whose meteorological society in Mannheim, Germany, provided thermometers and barometers to volunteers across Europe and collected and analyzed the annual data.

The more thorough assembling of temperature data began in earnest after the arrival of the telegraph in the mid-nineteenth century, which made possible the rapid-fire sharing of information.

Overnight, it seemed, weather mapping changed from a historical exercise to a practical method of tracking the atmosphere from day to day.

The embryonic art of weather forecasting, and the sheer novelty of weather maps themselves, helped feed the demand for reliable daily observations.

Weather services were established in the United States and Britain by the 1870s and across much of the world by the start of the twentieth century.

Beginning in the 1930s, the monitoring took on a new dimension, as countries began launching weather balloons that radioed back information on temperatures and winds far above ground level.

Most of these data, however, weren't shared among nations until well after World War II.

One of the products of the postwar drive for global unity was the **World Meteorological Organization**, founded in 1951 to link national weather agencies and later folded into the United Nations infrastructure.

In 1961, U.S. President John F. Kennedy urged the UN General Assembly to consider "further cooperative efforts between all nations in weather prediction and weather control."

services to agree on observing practices and data protocols.

Before long, a global weather observing system was in place, a model of international collaboration and data exchange that stood firm against the tensions of the Cold War and continues to this day.

During this midcentury whirlwind of progress, the focus was on improving daily weather forecasts.

What about the temperature of the planet as a whole?

Amazingly, hardly anyone was tracking it.

The pioneering climatologist **Wladimir Köppen** took an early stab at it in 1881.

There was a brief flurry of activity in the 1930s when, in separate studies, the U.S. Weather Bureau and British engineer Guy Stewart Callendar drew on the limited data available—mostly from North America and Europe—and found that Earth had warmed since the late 1800s.

Little did they know that the warming itself happened to be focused in North America and Europe, a fact that emerged only decades later.

“But for this accident, people might have paid little attention to the idea of global warming for another generation,” notes climate biographer Spencer Weart.

Interest in global temperature had cooled off—as had the planet itself—by 1961, when Murray Mitchell of the U.S. Weather Bureau found a cooling trend that dated back to the 1940s.

His was the most complete attempt up to that point at compiling a global temperature average.

However, like his predecessors, Murray was tricked by the wealth of U.S. and European weather stations.

In the sparsely observed Southern Hemisphere, the midcentury cooldown was far weaker than it was in the north.

Still, the trend was evident enough by the 1970s to prompt widespread attention.

It took until the late 1970s for temperatures to begin inching back up and concern about global warming to resurface.

Finally, in the early 1980s, several groups began to carefully monitor the annual ups and downs of global air temperature across the entire planet, eventually including the oceans.

These datasets are now among the most important pieces of evidence brought into the courtroom of public and scientific opinion on climate change.

Measuring the global warm-up

There's plenty of physical evidence beyond temperature measurements to indicate that a warm-up is under way, from sea level rise to melting glaciers.

Still, there are those who consider the amount of warming measured in recent decades not much to worry about.

This begs an interesting question: how do the experts come up with a single global surface temperature out of the thousands of weather stations around the world?

Doing so is a far more involved process than simply adding up the numbers and averaging them out.

For starters, you need to include the best-quality observations available.

Modern weather stations are expected to fulfill a set of criteria established by the WMO.

For example, thermometers should be located between 1.25 and 2 m (49–79 in.) above ground level, because temperature can vary strongly with height.

On a cold, calm morning, you might find frost on the ground while the air temperature at station height is 2°C (36°F) or even warmer.

Also, an official thermometer should be housed in a proper instrument shelter, typically one painted white to reflect sunlight and with louvers or some other device that allows air to circulate.

Anyone who's been in a car on a hot day knows how quickly a closed space can heat up relative to the outdoors.

Consistency is also critical.

Ideally, a weather station should remain in the same spot for its entire lifetime, but as towns grow into cities, many of the oldest stations get moved.

That means the station is sampling a different microclimate, which can inject bias into the long-term trend.

One of the biggest agents of station change in the mid-twentieth century was the growth of air travel.

Countless reporting stations were moved from downtown areas to outlying airports, where regular observations were needed for flight purposes and where trees and buildings wouldn't interfere with wind and rainfall measurement.

Even when stations stay rooted, the landscape may change around them and alter their readings.

Many urban locations are vulnerable to the heat-island effect.

Large buildings help keep heat from radiating out to space, especially on summer nights, and this can make temperatures in a downtown area as much as 6°C (10°F) warmer than those in the neighboring countryside.

On top of all this, gaps can emerge in the records of even the best-sited, longest-term stations.

Natural disasters, equipment or power failures, and wars can interrupt records for days, months, or even years, as was the case at some locations during World War II.

Problems may also emerge after the data are sent, as electronic or software glitches may corrupt the numbers in ways that aren't immediately obvious.

Even seemingly innocuous changes, such as moving to a different time of day for once-daily measurements, can have a real impact.

If highs and lows are are both recorded each day in the late afternoon, the long-term averages will end up higher than if the measurements are taken at midnight, since an especially warm afternoon and evening could then spread across two 24-hour observing periods instead of just one.

So, what is the Earth's temperature, anyway?

If you live in Rome, Melbourne, or Shanghai, you might not realize how close your climatic experience is to that of the Earth as a whole, at least min the average.

Each of these cities has an annual mean air temperature within 2ºC of 14.6ºC (58.3ºF), which is roughly the planet-wide average.

A hundred years ago, the global average was closer to 13.6ºC (56.5ºF).

In a perfect world, every temperature record would be free of the blemishes noted above.

That's not the case, of course, but climate scientists have found clever ways to work with what they've got.

For instance, they make use of the fact that temperature departures have large spatial scales: if Washington, D.C., is unusually warm on a given day, then nearby Baltimore should be unusually warm as well.

This allows biases at individual sites to be teased out through station-to- station comparisons.

By using stations with long records, and weighting each station by the area it represents, researchers can produce a reasonable estimate of long-term worldwide trends in the form of a **time series**—an accounting of global average temperature from one year to the next.

One major task in building such a time series is to determine how best to come up with a spatial average of global temperature, given that some parts of the globe are much better monitored than others.

Another is to find out how much the global reading for a given year has departed from the recent average (which is typically calculated from data covering a 30-year period, long enough to transcend shorter-term dips or peaks).

While individual scientists have taken stabs at time series of global temperature for more than a century, several major climate centers now carry out yearly analyses of global temperature, each with its own methods of tackling the challenge.

- **University of East Anglia.** The Climatic Research Unit of the University of East Anglia (UEA) applies a technique called the "climate anomaly method." As the name implies, this method starts with the monthly anomalies observed at each station in a given year—how much warmer or cooler it is relative to a 30-year base period, such as 1961–1990. UEA does this calculation for about 4800 stations worldwide. It averages those station anomalies across grid boxes that span 5° each of latitude and longitude, then by hemisphere, and finally for the globe as a whole, with the Met Office Hadley Centre providing the ocean analyses. The **Japan Meteorological Agency** employs a similar approach, but it uses a 1971–2000 baseline and then adjusts the final result so it reflects the anomaly against the most recent 30-year base period (currently 1981–2010).

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- **The National Aeronautics and Space Administration.** A group at NASA originally led by James Hansen devised the “reference station method.” Using temperatures observed at about 6000 stations, it calculates averages across a much smaller number of much larger grid boxes than UEA uses (about 80 in all). The longest-term record in each grid box is dubbed the reference station and considered the gold standard for that box. NASA’s anomalies have long been carried out relative to the period 1951–1980, just before human-produced global warming took off in earnest.
- **The National Oceanic and Atmospheric Administration.** NOAA uses a “pairwise” method that automatically identifies and compares temperatures at nearby stations. The technique hunts for red flags such as a gradual drift or an abrupt jump at one site that’s not reflected by its neighbor, then adjusts the data for these effects. Among the three groups, NOAA calls on the largest number of stations: about 7200, with that number expected to rise severalfold in the mid-2010s. Similar to UEA, it calculates anomalies for each station relative to a 30-year base period (typically the most recent, such as 1981–2010), with a separate calculation done for each month of the year. Data-sparse areas are filled in on the basis of the long-term patterns of temperature variation observed within grid boxes that span 5° of latitude and longitude.

Each one of these techniques contains a lot of redundancy—on purpose.

According to some experts, it takes no more than about a hundred strategically placed points around the world to create a decent first cut at global temperature.

Everything beyond that refines and sculpts the estimate; important work, to be sure, but nothing that would change the result drastically.

Which of the four groups' methods is best?

That's an impossible question to answer, because there's no absolutely perfect source to compare the results against.

A better question might be, how sensitive is the final result to different approaches?

If only one approach were used, we really wouldn't know, but with several techniques in play, we can begin to assess how solid the results are.

Each center's approach is designed to address the issues it believes are most critical.

Trends among the groups can vary depending on what years you start and end with, but reassuringly, there doesn't seem to be all that much difference among the results.

What may look like important contrasts—for instance, NASA ranked 2013 as the seventh-warmest year on record at the time, while NOAA put it in fourth place—relate more to tiny differences in particular years.

There's no disagreement among the groups on the big picture of long-term warming of the atmosphere at Earth's surface.

As reflected in the 2013 IPCC report, the consensus shows a global temperature rise, considering both land and ocean areas, of **roughly 0.89°C (1.60°F from 1901 to 2012**, with 90% confidence that the actual value is in the range of 0.69°–1.08°C (1.24°–1.94°F).

Most of that net warming has occurred in the last few decades—**about 0.72°C (1.30°F from 1951 to 2012)**, with a 90% confidence interval of 0.49°–0.89°C (0.88°–1.60°F).

If we extend the IPCC analysis to include the record-demolishing years of 2014–2017—which outpaced all prior years—the warming since 1901 is **closer to 1.0°C (1.8°F)**.

The solid nature of this consensus was reinforced in a highly touted independent study from the University of California, Berkeley.

The work garnered support from the Charles Koch Foundation, whose stated mission is to "advance social progress and well-being through the study and advancement of economic freedom."

Given the well-publicized conservative leanings of oil magnate Koch, some climate contrarians assumed the Berkeley Earth Surface Temperature (BEST) project would find less global temperature change than the four groups above and thus provide the vindication they'd so long sought.

They drew on the same 7200-station network as NOAA but added several thousand stations (many with shorter records) from other databases.

They also approached local discontinuities in a way that differed from the groups above, separating before-and-after climate datasets as if they'd come from independent stations.

In a series of peer-reviewed papers in 2012 and 2013, the BEST team burst one of the biggest collective bubbles of climate change denial.

With a confidence level of 95%, they estimated land-only temperatures in the decade 2000–2009 to be 0.85°–0.95°C (1.53°–1.71°F) warmer than the 1950s.

As it happens, this closely paralleled the trends found by the other four groups for the last century.

Although the Berkeley group focused on land-based readings, most estimates of full-globe temperature include temperatures over the **oceans**, which are a bit tricky to incorporate.

Since oceans represent 70% of the planet's surface, it's critical that surface temperature is measured over the sea as well as over land.

There's no fleet of weather stations conveniently suspended above the sea, unfortunately, but there are many years of sea surface data gathered by ships and, more recently, by satellites.

Because the sea surface and the adjacent atmosphere are constantly mingling, their temperatures tend to stick close to each other.

This means that, over periods of a few weeks or longer, a record of sea surface temperature (SST) serves as a reasonable proxy for air temperature.

Until the mid-twentieth century, SSTs were mainly measured by dipping uninsulated buckets (originally made of wood, then canvas) into the ocean and hauling them back to deck, with the water often cooling slightly on the trip up.

Around the time of World War II, it became more common for large ships—especially U.S. ones—to measure the temperature of water at the engine intake.

This tended to produce readings averaging about 0.4°C (0.7°F) warmer than the canvas-bucket method.

Some ships measured air temperature directly on deck, but these readings had their own biases: the average deck height increased through the last century, and daytime readings tend to run overly high due to the sun heating up metal surfaces on the ship.

It wasn't until the 1990s that routine measurements from ships and buoys were gathered promptly, processed for quality control, and coupled with newer estimates derived from satellites, which can infer SSTs by sensing the upward radiation from the sea surface, especially where it's not cloudy.

Together, all these sources now help scientists to calculate temperatures for the entire globe, oceans included.

Researchers adjust the SST data so that the pre–World War II bucket observations are consistent with the more recent engine-intake readings.

As with the land readings, there are differences in how the leading groups compile and process ocean data, but these don't appear to influence the long-term trends for global temperature too much.

Heat at a height

When people refer to global warming, they're usually talking about the air in which we live and breathe —in other words, the air near ground or sea level.

But the atmosphere is a three-dimensional entity.

Conditions at the surface are affected profoundly by what's happening throughout the **troposphere**—the lowest layer of the atmosphere, often called the "weather layer."

The troposphere extends from the surface to about 8–16 km (5–10 mi) above sea level.

It's tallest during the warmer months and toward the tropics, wherever expanding air is pushing the top of the troposphere upward.

There are two ways to measure temperatures high above Earth's surface.

The first is to use **radiosondes**, the balloon-borne instrument packages launched from hundreds of points around the globe twice each day.

Sending data via radio as they rise, the devices measure high-altitude winds, moisture, and temperatures.

As useful as they are, radiosondes have their limits for tracking climate change.

They're launched mainly over continents, with large distances separating their paths; they're produced by more than one company and thus don't behave uniformly; and they're prone to small biases that can make a big difference when assessing long-term trends.

The second technique is to use **satellites.**

These provide a more comprehensive global picture, but they were at the center of a now-resolved issue that was, for years, the subject of one of the most heated scientific debates in climate change science.

Shortly after global warming burst into political and media prominence in 1988, the University of Alabama in Huntsville began using data from a series of satellites to infer global temperature in three dimensions.

Throughout the 1990s and into the new millennium, their findings roiled the world of climate research and lent ammunition to global warming skeptics.

The satellite data suggested that the troposphere wasn't warming much if at all, despite the warming at ground level that was evident in observations and in physical markers such as glacial retreat.

Unless the fundamental understanding of lower atmosphere was off base, then one set of measurements had to be wrong.

The data are gathered primarily by instruments called **microwave sounding units** (MSUs), stationed on board NOAA satellites.

The MSUs measure the temperature of gases at various altitudes below them by measuring the energy that they emit.

Unlike a surface station or weather balloon, the MSUs provide data that's inherently averaged over large areas.

The formulas are complicated, but in general, each individual temperature measurement covers anywhere from about 8000 to 38,000 km^2 (3100–14,700 sq mi) through a depth of a few kilometers and a time period of a few hours.

All told, the satellites cover more than 90% of the planet's surface.

The fireworks began with a paper, published in the journal *Science* in 1990.

It showed no discernible temperature trend since 1979 in the lowest few kilometers of the atmosphere.

By comparison, surface temperatures had warmed roughly 0.2°C (0.36°F) during the 1980s.

Updated through the 1990s, the data continued to show virtually no warming.

What's more, radiosonde analyses tended to agree with the satellites, with relatively minor warming in northern midlatitudes and a slight cooling in the tropics.

Many people inclined to disbelieve global warming cited these reports as proof that the world wasn't heating up, but most climate scientists weren't convinced.

In 1998, two scientists in California, discovered that atmospheric drag might be affecting the MSU readings.

They claimed that air molecules were slowing down the satellites, increasing the effect of gravity on their paths and causing them to drop slightly, hence changing their viewing angle and skewing their measurements.

The data were adjusted for this error but they also made a separate correction in the opposite direction for another newly discovered bias, so the result was little net change.

Meanwhile, the global surface temperature continued to climb, so the gulf between surface- and satellite-measured upper-air trends widened even further.

The persistent disagreement left policymakers and the public wondering what to believe.

In 2000, the U.S. National Academies weighed in with a high-profile special report.

It agreed that the overall surface warming trend since 1979 was “undoubtedly real” but noted the possibility that high-altitude air had warmed more slowly than the surface, thanks to factors such as sulfate pollution, cool air filtering down from the record-cold stratosphere, and volcanic emissions from 1991’s eruption of Mount Pinatubo.

Since then, events have conspired to further diminish the satellite data as an argument against global warming.

For one thing, the years since 2000 have been notably warm compared to the 1980s and 1990s, both at the surface and aloft.

Meanwhile, several other groups gained access to full MSU data, and their analyses (including new techniques for separating stratospheric cooling from tropospheric warming) tended to show even stronger satellite-based warming trends, closer to surface observations.

Radiosonde trends have also crept upward in some studies.

Overall, these and other corrections brought the surface and satellite estimates into much closer agreement with each other.

The 2013 IPCC report examined seven different analyses of temperature trends in the lower stratosphere from 1979 to 2012, each drawing on radiosonde or satellite data.

The best estimates all fall into the range of 0.13°–0.16°C (0.23°–0.29°F) per decade.

That's in the same ballpark as surface-based trends for this period reported by UEA, NASA, and NOAA, which range from about 0.15° to 0.16°C (0.27°–0.29°F) per decade.

With the dust now settled, it's clear that the atmosphere is warming up miles above us, as well as on the streets and farms where we live.

How the greenhouse turns

The troposphere's name stems from the Greek tropos, meaning to turn-which is exactly what the troposphere does.

Heated largely from below, as sunlight strikes Earth, this layer of air churns and bubbles like a boiling pot of water, with calm conditions the exception more than the rule.

Usually the air is blowing faster horizontally than vertically, but no matter how you look at it, the troposphere is an ever-changing domain.

All this motion-up, down, and sideways- helps give rise to the variety of weather we experience at ground level.

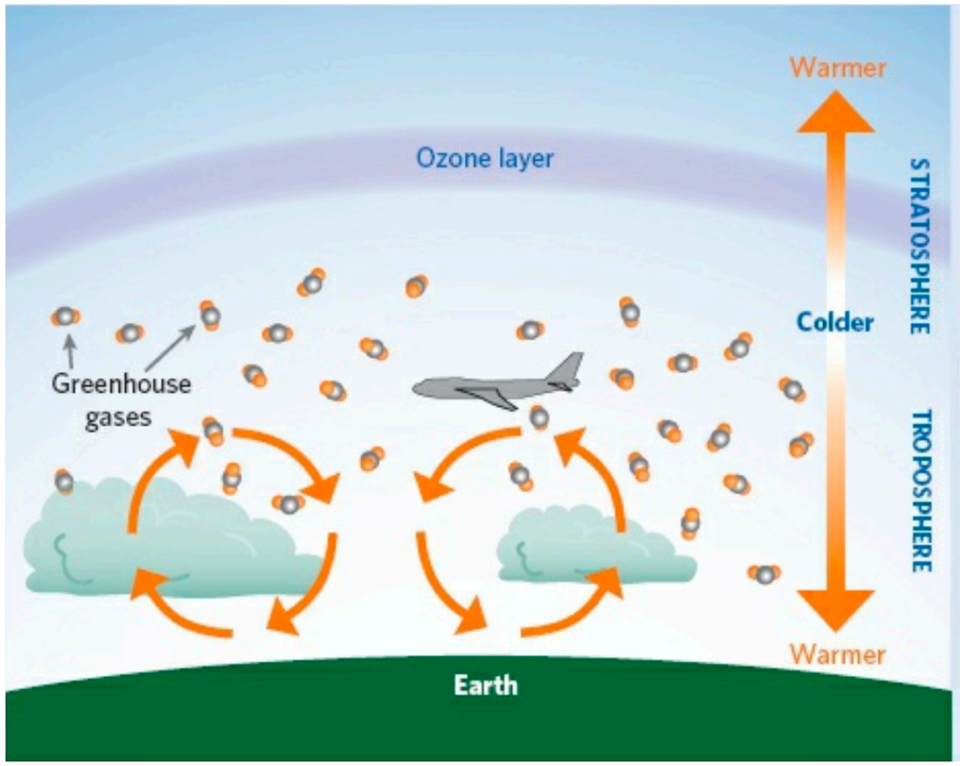
Because the troposphere is so well mixed, there should be a close connection between how much it warms near the surface and higher up.

There's not a one-to-one relationship, however, because greenhouse gases aren't spread through the troposphere as evenly as you might expect, and their effects vary with height.

Shorter-lived greenhouse gases, such as low-level ozone, are chemically transformed or washed out by rain or snow within a few days. Thus, they tend to stay within the same general latitudes as their sources-often rid- ing the jet stream from Russia to Canada or from eastern North America to Europe during their short time aloft.

Water vapor is most prevalent at lower altitudes, especially above the oceans, but it's been increasing higher up, thanks ni part to high-flying aircraft. (Planes also add fossil fuel emissions to the thin, cold air.)

Longer-lived greenhouse gases, such as carbon dioxide, are thoroughly mixed across the troposphere, both horizontally and vertically. That's why a single station high atop Hawaii's Mauna Loa can tell us how much CO_2 the global atmosphere holds and how that's changing.



In general, the higher in the troposphere a greenhouse gas is, the more powerful its effect on the troposphere as a whole.

This is because the amount of energy a greenhouse molecule emits is directly related to its temperature.

At high altitudes, where temperatures are low, the molecules emit less of the heat energy they receive from Earth into space, resulting in more heat remaining in the troposphere.

This picture reverses in the stratosphere, where - oddly enough - human actions have led to cooling rather than warming.

Carbon dioxide doesn't absorb much heat energy from Earth at these heights, but it continues to radiate heat to space, thus acting to cool the stratosphere.

Meanwhile, the partial loss of sunlight-absorbing ozone over the last 40 or so years exerted a cooling influence.

As a result of these and other factors, temperatures in the lower stratosphere plummeted to record low-levels in the early 21st century, and they remain colder than their pre-1980s averages.

Global dimming?

To get a clear idea of how our climate is changing, you've got to measure more than temperature.

Rain and snow are an important part of the story, and they're uniquely challenging to assess.

Other variables, such as **water vapor** and the nature of **cloud cover**, have their own complexities.

Satellites can measure both, but only with strict limitations.

For example, some can't sense lower clouds if high clouds are in the way.

A set of globe-spanning satellites called the Constellation Observing System for Meteorology, Ionosphere, and Climate (COSMIC)— launched in 2006 by a U.S.–Taiwan partnership, with a follow-up planned for the late 2010s— has provided fuller three-dimensional portraits of temperature and water vapor.

Other missions are keying in more specifically on cloud cover.

Even so, it's likely to be years before we can assess with confidence how these features are changing.

The droplets and particles in our atmosphere have everything to do with how much sunlight reaches Earth.

Sunshine and its absence weren't part of the dialogue on global warming until relatively recently.

Global dimming is real—or at least it *was* real.

From the 1950s through the 1980s, the amount of visible light reaching Earth appears to have tailed off by a few percent.

In terms of climate, that's quite significant, though it would be difficult for the average Jane or Joe to perceive the dimming against the much more dramatic on-and-off sunlight cycles that occur naturally each day.

The dimming also wasn't enough to counteract the overall global warming produced by greenhouse gases.

In fact, the timing of global dimming's rise to fame was a bit paradoxical, since the phenomenon began to reverse in the 1990s (and as we will see, the recent brightening brings its own set of worries).

Global dimming offers a good example of how a set of little-noticed measurements, collected diligently over a long period, can yield surprising results and can help answer seemingly unrelated questions.

Back in 1957, as part of the worldwide International Geophysical Year, a number of weather stations across the globe installed **pyranometers.**

These devices measure the amount of shortwave energy reaching them, including direct sunshine as well as sunlight reflected downward from clouds.

Slowly but surely, the amount of energy reaching the pyranometers dropped off.

A drop of 22% was discovered across Israel, and similar reductions were noted in Estonia.

However, there was hardly a rush to study the problem.

With the spotlight on global warming by that point, scientists might have found it counterintuitive to investigate whether less sunlight was reaching the globe as a whole.

Climate models, by and large, didn't specify such dramatic losses in radiation.

Yet the data cried out for attention, especially once they were brought together on a global scale.

Using a set of long-term readings from the highest-quality instruments, it was found that solar radiation at the surface had dropped 2.7% per decade worldwide from the late 1950s to early 1990s.

Two other studies examining much larger datasets found a smaller global drop— around 1.3% per decade from 1961 to 1990.

Even these studies failed to jolt the climate research community until 2002, when the dimming was linked to a steady multi-decade reduction in the amount of water evaporating from standardized measuring pans around the world.

Warmer temperatures ought to have evaporated more water, not less, they reasoned (as suggested by global measurements of water vapor).

However, it was realized that a reduction in sunlight could more than offset this effect.

It's possible that global evaporation actually rose in spite of the pan-based data, as noted in the 2007 IPCC report.

Pan-based sensors can only show how much water might evaporate from a completely wet surface, but a warmer, wetter, and dimmer world could still send more total moisture into the air—through the leaves of plants, for example.

The lifetime of moisture in the atmosphere, which could change with the climate, is another factor that makes the evaporation issue especially hard to unravel.

Still, the notion of dimming through much of the latter twentieth century is now fairly well accepted, and it's a small but significant part of the picture when scientists try to balance the global energy budget of incoming and outgoing radiation at Earth's surface over the last few decades.

Dimming might even help explain why some Arctic tree-ring records show a divergence between temperature and tree growth since the 1950s.

If these high-latitude trees are especially light-sensitive, then even a small dimming might be enough to limit their growth.

Some areas—especially urban, industrialized parts of the Northern Hemisphere—showed several times more dimming than other parts of the globe.

This regional patchwork makes sense given what we know about the causes of dimming.

The prime suspects are soot and other airborne particles, or aerosols, produced when fossil fuels are burned.

For hundreds of years, people in London had first-hand experience of the sky-darkening effects of soot.

Aerosol pollution does this type of darkening in several ways:

- **Attracting water.** Many aerosols are hydrophilic, or water-attracting. When conditions are right for clouds to form, a highly polluted air mass can produce many times more cloud droplets. The moisture is spread more thinly, though, so each droplet is smaller on average. The upshot is that this larger array of droplet surfaces can reflect more sunlight upward before it has a chance to reach the ground. From space, the clouds look brighter; from Earth, they're darker.
- **Extending cloud lifespans.** In general, it takes longer to get rain or snow out of a cloud full of many small particles, as opposed to fewer, larger ones. For this reason, the addition of aerosols seems to lengthen the lifespan of clouds and thus increase their reflective power over time.
- **Reflecting light.** Even on a cloudless day, aerosols themselves reflect sunlight to space. A major study of pollution across India and the adjacent ocean found that the subcontinent's thick veil of winter pollution can trim the amount of sunlight reaching the surface by as much as 10%.

Global warming itself may have been responsible for some of the dimming—perhaps around a third—by increasing the amount of water vapor in the atmosphere (which makes more moisture available for cloud formation) and perhaps rearranging the distribution of clouds (lower and/or thicker clouds tend to reflect more than higher, thinner ones).

As noted above, it's not entirely clear how cloud cover is evolving on a global scale, though water vapor increased on the order of 1% per decade during the strong global warm-up of the 1980s and 1990s.

It's also unclear whether global dimming will resurge as a major player in the decades to come.

Aerosols are more visible and noxious than greenhouse gases but more practical to control, so their reduction can be an easier political sell.

Most of the world's highly industrialized nations began cleaning up their smokestacks and exhaust pipes by the 1970s, and the economic downturn of the 1990s across the former Eastern Bloc further reduced aerosol pollution.

Meanwhile, the breakneck growth across southern and eastern Asia has put aerosol emissions on the rise in those areas since the 1990s.

The result is a more complex patchwork, with the net global effect appearing to be an overall drop in aerosol emissions in the 1990s, a slight rebound from about 2000 to 2005, and a slight drop later in the decade.

The amount of sunlight reaching Earth's surface has evolved in line with these trends, combined with other factors.

At some locations, solar radiation jumped anywhere from 1% to 4% from the early 1990s to the early 2000s.

Lamont–Doherty Earth Observatory suspects this may have reflected a recovery from the sun-shielding effects of the 1991 Mount Pinatubo eruption and the extensive cloudiness generated by the frequent El Niño events of the early and mid-1990s.

Slight dimming resumed in the early 2000s and has continued since.

Our cleanup of aerosols comes at an ironic price.

The transition away from global dimming helps explain why temperatures rose as sharply as they did in the 1990s.

Few would complain about a world that's brighter and less fouled by aerosol pollution, yet if we manage to reduce the global prevalence of aerosols in a sustained way, we'll also lose some of the overall cooling they provide as a bulwark against greenhouse warming.

In gaining a cleaner atmosphere, the world thus stands to lose one of its stronger buffers against climate change.

That may be one reason why the idea of consciously dimming the atmosphere through geoengineering has attracted attention.

11 - The Long View A WALK THROUGH CLIMATE HISTORY

During the four and a half billion years of our solar system, Earth has played host to an astounding array of life forms.

Much of the credit goes to our world's climate, which is uniquely equable compared to conditions on the other major planets in our solar system.

Even if they had water and other necessities, Venus would be too hot to sustain life as we know it, and Mars and the planets beyond it too cold.

However, there's evidence that both planets once harbored oceans and may have been mild enough to support life before their climates turned more extreme.

Likewise, during its lively history, Earth's climate has lurched from regimes warm enough to support palm-like trees near the poles to cold spells that encased the present-day locations of Chicago and Berlin under ice sheets more than half a mile thick.

The changes unfolding in our present-day climate make it all the more important to understand climates of the past.

Thankfully, over the last 50 years, technology and ingenuity in this field of study have brought dramatic leaps forward in our knowledge.

The clues drawn from trees, pollen, ice, silt, rocks, and other sources answer many questions, while leaving many others still open.

Taken as a whole, though, these data paint a remarkably coherent picture.

We now have a firm grasp of the changes that unfolded over the past three million years—a period ruled mainly by ice ages, with the last 10,000 years of interglacial mildness being one of the few exceptions.

We also know a surprising amount about what transpired even earlier, including many intervals when Earth was a substantially warmer place than it is now.

Could those ancient warm spells serve as a sneak preview of what we might expect in a greenhouse-driven future?

We know that there's already more carbon dioxide in our atmosphere than at any time in at least 2 million years.

Carbon dioxide has increased by more than 30% in only a century, an astonishing rate by the standards of previous eras.

At the rate we're pumping greenhouse gases into the atmosphere, they could be more prevalent by the year 2050 than they've been in at least 10 million years.

That's one reason why **paleoclimatologists** (those who study past climates) are taking a close look at how climate and greenhouse gases interacted during previous warm intervals.

The analogies are admittedly less than perfect.

As we'll see, Earth's continents were positioned quite differently 200 million years ago than they are now.

Even 10 million years ago, India was still drifting across the Indian Ocean, and North and South America had yet to join up.

As a result, the ocean circulations were very different as well.

Also, most of the species that were then swimming, crawling, or flying across our planet are now extinct.

Many other aspects of the environment at that time are hard to discern today.

In spite of our less-than-ideal hindsight, we've gained enormous ground in our understanding of how climate works by piecing together the past, and major advances continue to arrive in the still-young field of paleoclimatology.

In 2017, a new ice core from Antarctica provided a record that spanned 2.7 million years, more than a million more than the previous one.

This latest core extends all the way back to the point when a crucial switch occurred in the pace of the on-and-off tempo of ice ages.

If there's one message that's become increasingly clear from the ice cores, it's that natural forces have produced intense and rapid climate change in the past.

That’s not a comfort so much as it is a warning, as we continue to add carbon dioxide to the air at a rate far beyond those natural processes.

What controls our climate?

The great detective story that is paleoclimatology hinges on a few key plot devices.

These are the forces that shape climate across intervals as brief as a few thousand years and as long as millions of years.

Greenhouse gases are a critical part of the drama, of course, but other processes and events can lead to huge swings in climate over greatly varying timespans.

The sun-blocking particles spewed into the stratosphere by an unusually large volcano may cool the planet sharply, but most of those impacts are gone after a couple of years.

n contrast, climate change due to slight shifts in orbital patterns plays out over tens or hundreds of thousands of years.

That’s still relatively fast in geological terms.

The slow drift of continents, another major player in climate variation, can operate over hundreds of millions of years.

The overlap among these various processes creates a patchwork of warmings and coolings at different time intervals, some lengthy and some relatively brief.

It’s a bit like the mix of factors that shape sunlight in your neighborhood.

On top of the dramatic but dependable shifts in sunshine you experience due to the 24-hour day and the transition from summer to winter, there are much more irregular shifts produced as cloudiness comes and goes.

Below is a quick summary of the biggest climate shapers.

We’ll come back to each of these later in the chapter.

- **The sun is literally the star of the story.** At its outset, the sun shone at only about three-quarters of its current strength; it took several billions of years for its output to approach current levels. One fascinating question (discussed later) is how Earth could be warm enough to sustain its earliest life forms while the sun was still relatively weak. In another five or six billion years the sun—often referred to as an “ordinary middle-aged star”—will reach the end of its life. At that point it should warm and expand dramatically before cooling to white-dwarf status.

- **Variations in Earth's orbit around the sun are hugely important.** These small deviations are responsible for some of the most persistent and far-reaching cycles in climate history. Each of these orbital variations affects climate by changing when and how much sunlight reaches Earth. As shown in the graphics on the opposite page, there are three main ways in which Earth's orbital parameters vary. Several visionaries in the mid-1800s, including James Croll—a self-taught scientist in Scotland—suspected that these orbital variations might control the comings and goings of ice ages. In the early 1900s, Serbian mathematician Milutin Milankovitch quantified the idea, producing the first numerical estimates of the impact of orbital variations on climate. While his exact calculations have since been refined, and there's still vigorous debate about what other factors feed into the timing of climate cycles, the basic tenets of Milankovitch's theory have held firm.



The Earth, circa 250 million years ago, showing the supercontinent Pangaea.

- **The changing locations of Earth's continents also play a huge role in determining climate.** Francis Bacon observed in 1620 how easily South America and Africa would fit together if they were puzzle pieces. Many schoolchildren notice the same thing long before they're taught about **continental drift**, Alfred Wegener's bold 1915 hypothesis that Earth's continents moved slowly into their present positions. Today, satellites can monitor continental drift by tracking minuscule land movements, but scientists are also able to deduce the long-ago locations of continents thanks to various techniques—including analyzing traces of magnetism generated by Earth's geomagnetic field over the millennia and preserved in volcanic magma. These and other clues tell us that much of Earth's landmass once existed as **supercontinents.** The first (and most speculative) is **Rodinia**, a clump that bears little resemblance to today's world map. Formed more than a billion years ago, Rodinia apparently broke up into several pieces. Most of these were reassembled as **Gondwana**, which drifted across the South Pole between about 400

and 300 million years ago. Gondwana became the southern part of **Pangaea**, an equator-straddling giant. In turn, Pangaea eventually fragmented into the continents we know now. The locations of these ancient continents are important not only because they shaped ocean currents but because they laid the literal groundwork for ice ages. Only when large landmasses are positioned close to the poles (such as Gondwana in the past, or Russia, Canada, Greenland, and Antarctica today) can major ice sheets develop. However, high-latitude land isn't sufficient in itself to produce an ice age. The southern poles harbored land for most of the stretch from 250 to 50 million years ago, yet no glaciation occurred, perhaps because CO_2 concentrations were far higher than today's.

- **Volcanoes** can spew enormous amounts of ash, soot, and other particles and gases—some of them greenhouse gases—into the atmosphere. A single large eruption can be powerful enough to loft sun-shading particles into the stratosphere, as was the case when Mount Pinatubo blew its top in 1991. When this happens, the high-altitude debris can cool the planet by more than 1°C (1.8°F) for a year or more, until gravity pulls the particles back to Earth. Tropical volcanoes are especially effective at this cooling, since their stratospheric debris can easily spread across both Northern and Southern Hemispheres. Smaller volcanoes have less of an individual impact, but they can make a difference collectively. The 2013 IPCC report concluded that a series of minor eruptions after 2000 played at least some role in the relative slowdown in atmospheric warming observed from the late 1990s into the early 2010s. Over the much longer term, volcanoes act as a warming influence, as they add perhaps 0.1–0.3 metric gigatons of carbon to the atmosphere per year on average. That's a substantial amount, though it's less than 1% of what humans now add through fossil fuel burning. At present, it appears that the volcanic addition is balanced by other natural forces and feedbacks that remove carbon at roughly the same pace. However, it's possible that a frenzy of volcanism early in Earth's history pumped the atmosphere's greenhouse gas levels far above anything that's occurred since. Later on, undersea eruptions associated with the spreading of continental plates may have been the main source of the carbon dioxide that led to Earth's remarkable warmth from about 250 to 50 million years ago

Round and round we go (and so does our climate)

Everyone who lives outside the tropics knows the power of the four seasons, but a surprising number of people-including most of the Harvard graduates interviewed at random in the classic documentary *A Private Universe*-can't explain why the seasons occur.

Many assume it's because Earth is closer to the sun in summer and more distant in winter, but that doesn't wash.

It's true that Earth's orbit is slightly asymmetric, but we're actually closest to the sun each year between January 2 and 5, when it's summer ni Sydney but winter in London.

The real reason for the seasons is that Earth is tilted about 23.4° relative to its orbit around the sun (see graphic).

This tilt aims each hemisphere toward the sun in an alternating pattern that produces summers, winters, springs, and autumns.

What does this have to do with climate change?

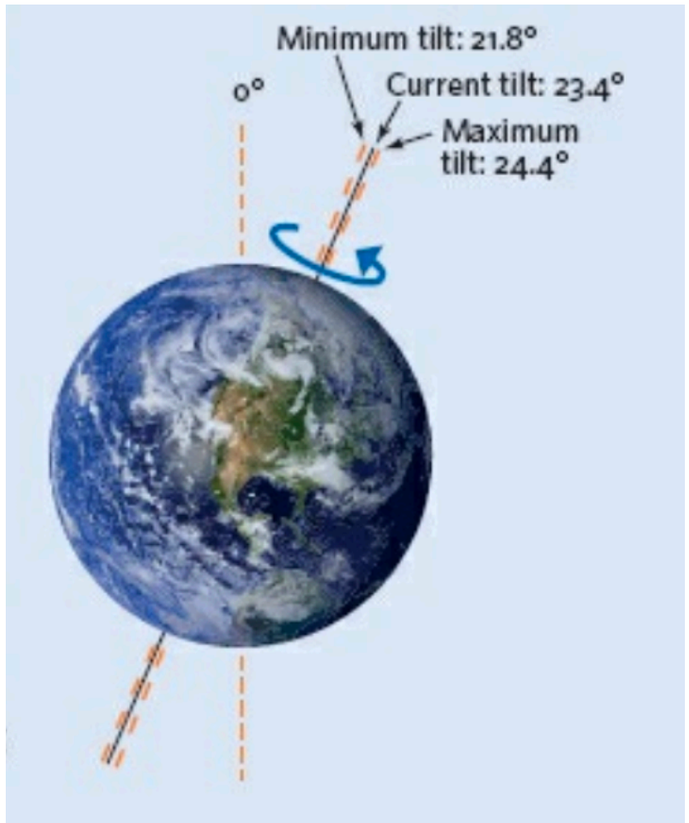
Each of the factors just alluded to-Earth's tilt, the eccentricity of its orbit, and the timing of its closest approach to the sun-change slightly over time in a cyclical manner.

This is due mainly to the intersecting gravitational effects of the sun, the moon, and the other planets ni our solar system.

Together, the orbital cycles produce dra- matic swings in climate over tens of thousands, or even hundreds of thousands, of years, with carbon dioxide greatly boosting the atmosphere's response.

By studying how these natural climate shifts come and go, scientists also find clues about how Earth might respond to change that's human induced.

Earth's tilt goes up and down, ranging Earth's tilt goes up and down, ranging from about 21.8° to 24.49 and back over about 41,000 years (the range widens or narrows a bit with each cycle).



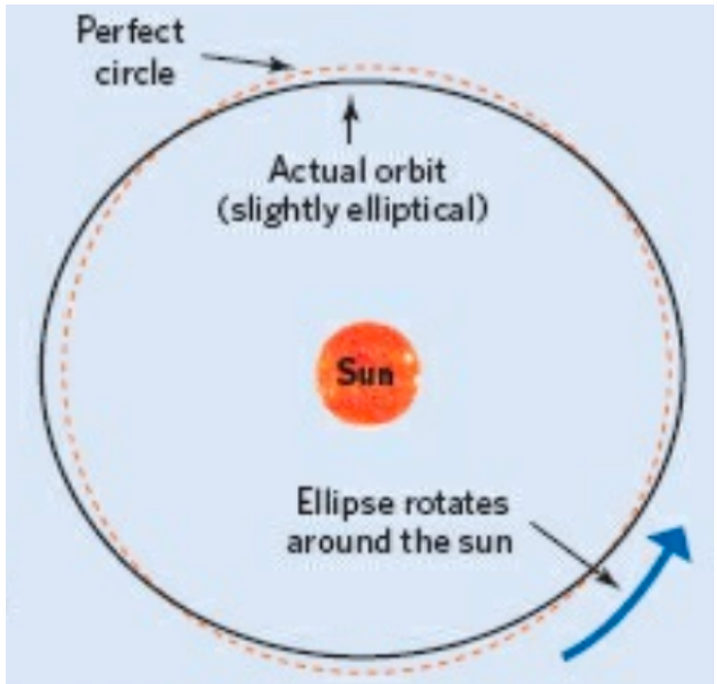
The tilt is now just over 23.4° and on the decrease.

When the tilt is most pronounced, it allows for stronger summer sun and weaker winter sun, especially at high latitudes.

Ice ages often set in as the tilt decreases, because the progressively cooler summers can't melt the past winter's snow.

At the other extreme, increasing tilt produces warmer summers that can help end an ice age.

Earth's orbit around the sun is not precisely circular, but elliptical, with the sun positioned slightly to one side of the center point.



Currently this brings Earth about 3% closer to the sun in early January (perihelion) than in early June (aphelion), with about 7% more solar energy reaching Earth at perihelion.

The **eccentricity**, or "off-centeredness", of the orbit varies over time in a complicated way.

The net result is two main cycles, one that averages about 100,000 years and another that runs about 400,000 years.

While the eccentricity is low, there's little change through the year in the3 Earth-sun distance.

When eccentricity is high, the sunlight reaching Earth can be more than 20% stronger at perihelion than at aphelion. This either intensifies or counteracts seasonality, depending on the hemisphere and the time of the year that perihelion and aphelion occur.

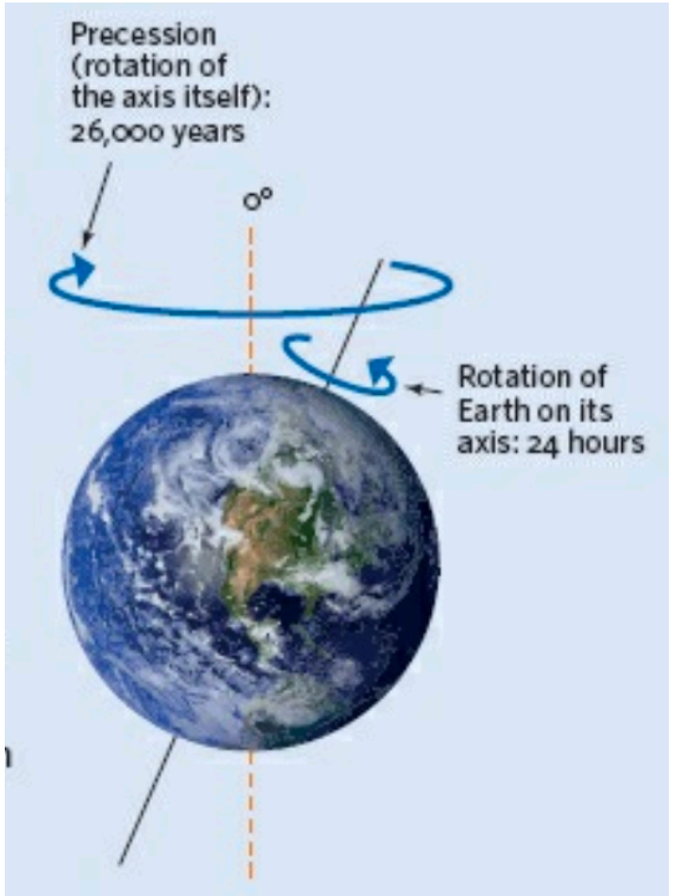
The last eight ice ages have come and gone roughly in sync with the 100,000-year eccentricity cycle, but it's not yet clear how strongly the two are related.

Earth's axis rotates slowly around an imaginary center line, like a wobble in a spinning top.

The main cycle of this process - known as **precession** - takes about 26,000 years, and it shifts the dates of perihelion and aphelion forward by one day every 70 years.Thus, in about 13,000 years, Earth will be closest in July instead of January.

This will intensify the seasonal changes in solar energy across the Northern Hemisphere and weaken them in the south.

The stronger summer sun is likely to bolster African and Asian monsoons, as was the case 10,000 years ago.



- **Celestial bodies** such as **asteroids** make cameo appearances in Earth's climate saga. Most are minor. Only the very rarest of giant asteroids, which typically arrive millions of years apart, can produce climatic effects comparable to those from a major volcanic eruption. As with volcanoes, a huge asteroid might cool climate for a year or two by blasting dust and soot into the stratosphere. However, a big enough asteroid could also produce an explosive wave of heat as it enters the atmosphere, enough to set much of Earth's vegetation ablaze. Such fires would send enough carbon dioxide into the atmosphere to cause a substantial global warming that could last centuries, far longer than the brief post-collision cooling. This lurid scenario apparently played out around 66 million years ago, when an enormous asteroid struck Earth. The most likely point of impact is an area on the north end of Mexico's Yucatán Peninsula now covered by a giant crater of pulverized rock. Scientists combing the globe have found widespread deposits of iridium and other substances that are far more prevalent in asteroids than on Earth, all apparently deposited at the same time. Also dating from the same period are clumps of charcoal-like carbon, a hint that widespread fires occurred. More than half of all species are thought to have perished in this cataclysm and its aftermath, including most dinosaur species. (A few that survived may be the predecessors of modern-day birds and crocodiles.)

How scientists read the clues

The technology used to analyze materials and glean climate clues from them has advanced at a phenomenal rate in recent decades.

The traces of past climates can show up in animal, vegetable, and mineral forms, providing a wealth of evidence for paleoclimatic detective work.

Below are the four main types of paleoclimate **proxy data**—records that can be used to infer atmospheric properties such as temperature or precipitation:

- **Biological:** tree rings; pollen; corals; and fossils of plants, animals, insects, and microscopic creatures such as diatoms. Their locations, concentrations, and conditions point toward the atmospheric and oceanic conditions under which various life forms prospered. Many biological materials can be dated using radioactive decay rates or the presence of stable isotopes (more on these later). The chemical makeup of these samples can also yield important information.
- **Cryological:** cylindrical samples called ice cores, collected by drilling through dense, deep ice sheets or glaciers where air bubbles and dust are trapped within yearly layers of snowfall. The bubbles preserve the greenhouse gases and other atmospheric constituents present at a given time, and the ice cores can be analyzed through radiometric dating (see p. 280). The dust layers and the character of the ice itself also reveal climate processes at work.
- **Geological:** rock, sand dunes, ocean sediments, glacial debris, stalagmites, and other materials sampled from sea and lake beds and dry land. Again, radioactive decay within these materials helps establish their age. Volcanic rock can also be dated by the traces of Earth's geomagnetic field. The location and condition of telltale geological features helps establish when sea levels rose and fell or when glaciers scoured the land surface. Many ocean sediments are laden with the remnants of shells built by creatures the size of a grain of sand, including bottom-dwelling **foraminifera** as well as **plankton** that prefer near-surface waters. The prevalence and type of these shells can reveal the amount of carbon and other elements that entered the sea at various points.

- **Historical:** written records of atmospheric conditions and biological phenomena tied to climate. The latter might include the timing of spring blooms for a given type of tree or the poleward extent of a particular species of insect. In some areas, including England and Japan, such records go back hundreds of years with almost daily precision.

Establishing dates for climate proxies can be a challenge.

Trees are among the easiest: each ring indicates a year's worth of growth, and its width hints at the state of that year's climate.

For trees that are still alive, or ice cores drawn from the surface downward, it's a simple matter to count backward from the present and establish dates in the past, using multiple samples to ensure accuracy.

For many other proxies, such as fossilized trees, you also need a benchmark, something you can use to establish the time period in which the proxy was created.

A typical starting point is **radiometric dating**, which revolutionized the paleoclimatic dating game in the 1950s.

Radiometric techniques rely on the fact that elements such as **carbon** and **uranium** exist in several different forms, or **isotopes**, each with a different number of neutrons.

Some of these isotopes are unstable, and they decay over time at measurable and predictable rates.

For instance, neutrons hitting the upper atmosphere are continually converting nitrogen into an unstable carbon isotope known as ^{14}C. This sifts down into Earth's atmosphere and gets absorbed by living things, decaying over time as new ^{14}C isotopes are created higher up.

Given a certain amount of ^{14}C, half of it will have decayed back into nitrogen after about 5700 years (its **half- life**).

Knowing this, scientists can measure how much ^{14}C has decayed in a given substance and use that information to determine how old the substance is.

Although these unstable isotopes are rare compared to stable carbon atoms, they're well distributed throughout Earth and its ecosystems.

This allows geochemists to date a wide variety of substances, from carbon dioxide in an ice-core air bubble to the shells of marine creatures.

One limit to using ^{14}C is its relatively short half-life.

Because there's very little carbon left to measure after a few half-life cycles, most carbon dating only goes back about 50,000 years.

For older substances, **mass spectrometry** dating is used.

Uranium is the material of choice for this, because it decays into lead at a very slow rate. Uranium's half-life is about 700 million years for the isotope ^{235}U and an astounding 4.5 billion years—the age of Earth itself—for ^{238}U.

As with any measurement technique, errors can creep into radiometric dating.

This happens mainly when a substance is contaminated by other materials of a different age through chemical reactions, erosion, or some other process.

Careful cross-calibration with similar materials from other locations can help alleviate this problem.

Isotopes don't have to be unstable to be useful in dating. Stable isotopes provide another window on the climate's past. The most extensively studied is ^{18}O, a form of oxygen.

Because ^{18}O is heavier than standard oxygen, it condenses more readily and evaporates more slowly.

The difference between the two rates depends on temperature, ocean salinity, and other factors.

For example, coral reefs hold differing amounts of ^{18}O depending on the sea surface temperature present when they formed.

In ice cores (and other places where ancient water is trapped) the relative amount of ^{18}O can help reveal the temperature at which the moisture condensed.

This can help delineate annual layers in the ice as well as reflecting longer-term temperature trends.

Computer models are another important part of the paleoclimate toolbox.

In theory, any model that simulates future climate ought to be able to reproduce the past as well, given enough information to start with.

Indeed, many of the most sophisticated global models designed to project future climate are tested by seeing how well they reproduce the last century.

Another useful test is to see how well the models handle modern-day seasonal transitions: in midlatitudes, at least, going from summer to winter can serve as a microcosm of a long-term overall climate shift.

For prehistoric times, there's far less data to check a model against, but even in these cases, simple models can be helpful in assessing what climates might be plausible.

The really big picture: From Earth's origins to the expansion of life

It takes quite a mental stretch to fully comprehend the total lifespan of Earth's atmosphere— unless you're a biblical literalist, that is.

Most of our planet's history unfolded long before the appearance of life forms that we're now familiar with.

To get a sense of this, picture Earth's 4.5-billion-year lifespan as if it were compressed into a mere 24 hours.

In this imaginary day, dinosaurs wouldn't come on the scene until after 10:00 p.m.

Even more impressive, the period since the last ice age ended—the time when most human civilizations emerged—would occupy only the last fifth of a second before midnight.

Deducing the complete history of the atmosphere is thus a major challenge.

Earth itself offers few clues to help us figure out what climate was up to during the planet's first two billion years or so.

In fact, a good deal of our knowledge about Earth's earliest climate comes from space.

By studying the behavior of stars similar to the sun, astronomers have deduced that the solar furnace (a continuous nuclear reaction that forms helium out of hydrogen) took a long time to reach its modern-day strength.

It appears the sun's luminosity started out only about 70%–75% as strong as it is now, gradually increasing ever since.

With the sun relatively weak for so long, you might envision early Earth as a frozen, lifeless wasteland.

Amazingly, this wasn't the case.

Fossils of tiny bacteria show that life existed as far back as 3.5 billion years.

Rocks from around that time show the telltale signs of water-driven erosion, another sign that Earth wasn't encased in ice.

How did Earth stay warm enough for life during those early years?

That's the crux of what paleoclimatologists call the **faint young sun paradox**, discovered by astronomers Carl Sagan and George Mullen in 1972.

Explanations for the paradox have been floated ever since.

In 2010, Danish scientist Minik Rosing and colleagues proposed that widespread oceans and relatively thin clouds allowed the early Earth to absorb much more heat than it would today.

However, greenhouse gases are the most commonly proposed solution to the paradox.

Put enough of these gases in the air, and Earth can stay warm enough to sustain life even with a weaker sun.

The amount of carbon now stored in fossil fuels and in the oceans is more than enough to have provided such a greenhouse boost if it were in the air during the weak-sun period.

Such a boost might not have come from carbon dioxide alone, though, as recent geochemical analysis of ancient soil suggests that there was far too little CO_2 in the air to do the trick by itself.

Sagan and Mullen suggested that ammonia might be a lead player, but it doesn't hold up well under the ultraviolet radiation of the sun.

Methane is now becoming a key piece of the picture.

In a 2014 study published in *Astrobiology*, Eric Wolf and Brian Toon (University of Colorado Boulder) simulated Earth's early atmosphere with a relatively complex three-dimensional model.

By setting carbon dioxide at about 50 times its current concentration, and methane at about 1000 times the values observed today, Wolf and Toon obtained global temperatures consistent with geological indicators suggesting that the early Earth atmosphere was at least as warm as today's.

Volcanoes can inject huge amounts of greenhouse gas into the atmosphere in a short amount of time, so the intense volcanism of Earth's earliest days might have added enough to keep the planet from freezing.

However, that begs another question: how did the greenhouse gas levels gradually decrease over billions of years at just the right tempo to keep Earth warm, but not too warm?

It's possible, but unlikely, that volcanic activity diminished at exactly the pace needed.

Another possibility is that some other climate factor kicked into gear to keep greenhouse gases within a range tolerable for life on Earth.

One candidate is the steady expansion and elevation of continents during much of Earth's history.

Landmasses reduce the amount of carbon dioxide in the air through a multimillion-year process known as **chemical weathering**, which occurs when rain or snow fall on rocks that contain silicate minerals.

The moisture and silicates react with carbon dioxide, pulling the CO_2 out of the air.

Carbon and water then flow seaward in various forms, and most of the carbon ends up stored in ocean sediments that gradually harden into rocks.

Over billions of years, this process has been responsible for some 80% of the carbon now stored underground. (Rain and snow also draw CO_2 out of the air when they leach limestone, but in that case, the carbon heads seaward in a form used by marine creatures to build shells, a process that returns the carbon to the atmosphere.)

Chemical weathering is such a powerful process that the growth of the Himalayas over the last 55 million years might have helped to pitch Earth into its recent series of ice ages, as we'll see later.

Apart from these vast changes, chemical weathering can also influence the atmosphere through a negative feedback process, one that tends to counterbalance climate change in either direction.

In general, the warmer it gets, the more rapid chemical weathering becomes and the more carbon dioxide is pulled from the air.

This is in part because the chemical reactions behind weathering operate more rapidly when it's warmer.

Another reason is that warmer oceans send more water vapor into the air, and more rain and snow falls onto continents to stimulate weathering.

Thus, the warmer the air gets, the more carbon dioxide is drawn out of the atmosphere through weathering, and that should put a cooling effect into the mix.

Conversely, if ice starts to overspread the land, weathering should decrease and carbon dioxide levels ought to increase, thus working against the glaciation.

Earth's profusion of **plant life**, which appears to have begun around 500 million years ago, serves as another brake on greenhouse gases.

Plants and trees take in carbon dioxide through photosynthesis and release it when they die.

Generally, plant growth increases on a warmer, wetter Earth—so the hotter it gets, the more plants there are to help cool things down.

Indeed, one line of thought positions life itself as the main negative, or balancing, feedback on the entire Earth system.

Of course, there are also many **positive feedbacks** between CO_2 and climate, which tend to amplify rather than dampen a change.

For instance, as carbon dioxide increases and warms the planet, more water vapor—itself a greenhouse gas—should evaporate into the atmosphere and further increase the warmth.

Similarly, oceans absorb less carbon dioxide the warmer they get, which leaves more CO_2 in the air to stimulate further warming.

It's the balance between negative and positive feedbacks that generally determines which way climate will turn at any given point.

A big concern about the twenty-first century is that positive feedbacks, such as those now taking place in the Arctic, may overwhelm negative feedbacks, some of which (such as chemical weathering) take much longer to play out.

Whether they're looking a thousand or a billion years into the past, paleoclimatologists find that greenhouse gas concentrations are intimately linked with the rise and fall of global temperature.

Ice cores and other records bear out this tight linkage over the last million years, and there's no reason to think it should be absent earlier on.

There's a wrinkle in this relationship, though—one often pointed out by global warming contrarians in support of a misleading conclusion.

Because the CO_2 changes are so closely coupled in time to the growth and decay of ice sheets, it's hard to tell exactly which came first and when.

For example, there appear to be periods in which undersea volcanism produced vast amounts of CO_2, after which temperatures rose dramatically, and other periods where orbital cycles triggered ice-sheet growth and colder temperatures, with the process greatly enhanced by CO_2 decreases.

Paleoclimatologists are still sorting out these chains of events.

It appears that orbital changes play a major role in kicking off and ending ice ages, but during the growth and decay stages, there's a web of positive feedbacks at work among CO_2, ice, and other climate elements.

For instance, cooler oceans absorb more carbon dioxide, which acts to support further cooling.

The bottom line is this: increasing CO_2 goes hand in hand with warming temperatures, whether the climate shift is kicked off by some other factor or by CO_2 itself—and it's the latter situation we find ourselves in now.

Hot in here: Earth's warm periods

What will the world look like if the more dire projections of climate change by the year 2100 come to pass?

Computer models and basic physical theory tell us a lot about what to expect, but we can also gaze back into climate history for clues.

A good place to start is the **Mesozoic era.**

It began after the world's greatest extinction episode, which occurred about 250 million years ago, and ended with another vast extinction 66 million years ago.

In between are three periods well known to dinosaur lovers: the Triassic, Jurassic, and Cretaceous.

All three intervals were notably warm, with no signs of any major glaciations or even much ice at all.

This is in part because Earth's landmasses were located away from the North Pole, mostly clustered in the warm supercontinent of Pangaea.

However, parts of Pangaea extended to the South Pole, and the warmth continued even as Pangaea broke up and the Northern Hemisphere continents shifted closer to their present-day locations.

Two of the warmest intervals were the mid- to late Cretaceous and the early Eocene (part of the Tertiary epoch), which fall on either side of the asteroid-driven extinction that occurred 66 million years ago.

In both periods, average global readings soared to 5°C (9°F) or more above present values.

That's only a little beyond the IPCC's latest high-end projections of global temperature rise up to the year 2100.

Proxy data for these past periods indicate that carbon dioxide was several times more prevalent in the atmosphere than it is today.

This enhances scientists' confidence that the extra CO_2 stoked the ancient warmth and that the greenhouse gases we're now adding could do something similar in our future.

Where did the extra CO_2 come from?

The most likely candidate is an active period of **undersea volcanism**.

Earth's tectonic plates clash most visibly along the world's volcanic coastlines, such as the Pacific's Ring of Fire.

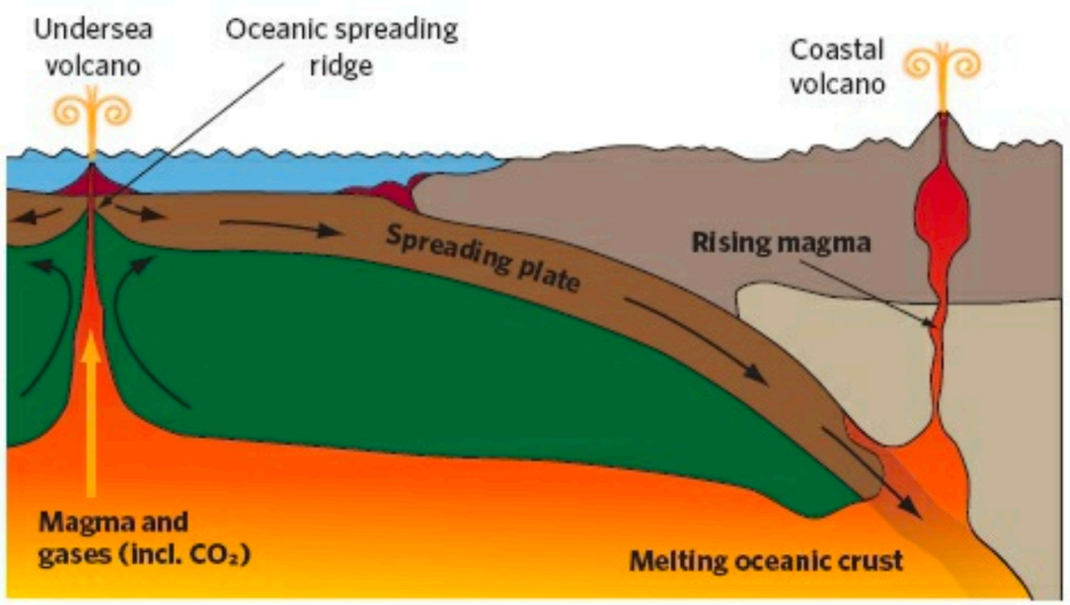
However, there's additional volcanic activity far beneath the oceans.

Most major basins have a mid-ocean ridge, where tectonic plates separate and vast amounts of magma and carbon dioxide surge upward to fill the void.

The magma eventually solidifies to form fresh crust, slowly elevating the center of the ridge as the older material spreads outward.

It appears that, on average, the plates were separating up to 50% more quickly about 100 million years ago than they are now.

This would have allowed more CO_2 to bubble upward and would have pushed the plates toward continents more vigorously, stimulating more activity (and more CO_2 emissions) at coastal volcanoes.



The process involved in undersea volcanism and the surface and atmospheric effects.

Another source of greenhouse gas might have provided even more of an influx at key times.

Hundreds of billions of tons of carbon are believed to be locked in cold ocean seabeds in the form of **methane clathrates**, ice crystals that hold methane in their molecular structure (also known as methane hydrates).

Methane clathrates remain stable if conditions are cold enough and/or if enough pressure is exerted on them (as is the case at the bottom of the sea).

But a change in ocean temperature or circulation during a warm era might be enough to destabilize the methane and send it pouring into the atmosphere.

This is a long-term concern for our future climate, and it may have played a role in at least two massive warmings during already warm times.

One was the temperature spike and mass extinction that kicked off the Mesozoic era 248 million years ago.

The other is a distinct blip early in the already toasty Eocene epoch called the Paleocene–Eocene Thermal Maximum (PETM), which occurred about 56 million years ago.

During the PETM, global ocean temperatures soared by as much as 5°C (9°F) over a few thousand years and remained high for almost 100,000 years.

Isotope studies by NASA indicate that a vast amount of greenhouse gas was pumped into the atmosphere during the PETM, though over a longer period and at a slower rate than our current emissions.

Among the possible sources is a burst of volcanism in the North Atlantic, as suggested by recently analyzed igneous rock formations off the coast of Norway.

Close calls for life on Earth

In a classic 1920 poem, Robert Frost observed that some people expect the world to go out in fire, while others think the end will be an icy one. Our planet edged toward those extremes at several different points in its history.

Two examples brought vastly different consequences for ecosystems.

One of the coldest periods in early Earth history ran from about 750 to 580 million years ago.

There's evidence that glaciers scoured most of Earth's large landmasses, including some within 10° of the equator.

The implication is that much or all of the planet was covered with ice for millions of years—a scenario dubbed Snowball Earth in 1992 by geologist Joseph Kirschvink of the California Institute of Technology.

There isn't yet a firm consensus on what kicked off the icing, but since Earth's land was concentrated in the tropics, where precipitation is heaviest, chemical weathering could have pulled large amounts of carbon dioxide from the air, cooling the global atmosphere.

If the planet actually did freeze over entirely, the powerful positive feedback of a "white Earth" —especially with continents and land ice concentrated at lower, high-sun latitudes- could have reflected most of the incoming sunlight and helped preserve the ice.

How such a state would have ended isn't known, but a slow accumulation of carbon dioxide from volcanoes—on top of a cold-induced slowdown in chemical weathering- seems the most likely possibility.

Whether or not ice actually covered the entire Earth, this period of widespread glaciation was followed by a sudden profusion of multicellular organisms.

Some theorists speculate that a Snowball Earth and a subsequent warm-up could have prodded the rapid evolution of primitive life into more complex

At the other end of the spectrum, an especially intense warming around 250 million years ago spelled doom for most of the life forms present at the time.

The Permian-Triassic extinction drew the curtain on more than 90% of marine species and more than two-thirds of land-based creatures.

The die-off unfolded quickly in geological terms, over fewer than a million years.

Greenhouse gases likely soared to many times their present-day amounts, and high-latitude oceans warmed to as much as 8°C (14°F) above present-day readings.

The warmer oceans likely enhanced the separation of surface waters from cooler deep oceans, reducing the usual mixing between the layers that distributes oxygen and nourishes many marine organisms.

Researchers haven't settled on a single cause for the extinction.

A modeling study led by Jeffrey Kiehl (NCAR) suggested that enhanced volcanic activity over almost a million years could produce enough atmospheric warming through added carbon dioxide to slow the ocean currents that bring oxygen to deep-water marine life.

Other possibilities include a giant meteorite (although most evidence of such an ancient impact would be gone by now), an outpouring of the greenhouse gas methane in the form of methane hydrates released from the ocean floor, a massive release of hydrogen sulfide produced as a result of oxygen depletion in the oceans, or some blend of these and/or other factors.

One thing that’s fairly clear about past warm periods is that mid- and high-latitude areas were downrigh balmy as compared to today.

Coral reefs grew as far poleward as 40°, close to the latitudes of present-day Philadelphia and Melbourne.

Crocodiles, dinosaurs, and leafy trees flourished in polar regions, while other areas featured vast deserts covered with sand dunes.

A 2006 study of sediments beneath the Arctic Ocean found that sea surface temperatures (SSTs) during the PETM soared to as high as 23°C (73°F), which is far beyond earlier estimates.

In the tropics, SSTs may not have been much warmer than they are now, though some intriguing evidence of readings at or above 33°C (91°F) has been collected off the coasts of present-day South America and East Africa, and some climate models suggest that SSTs may have even topped 35°C (95°F).

The oceans were at a high-water mark during these warm times.

For millions of years, **sea level** was more than 200 m (660 ft) higher than it is now.

This put vast stretches of the present-day continents under water, including the central third of North America and all of eastern Europe from Poland southward.

The oceans were so high and vast for several reasons.

One is that there wasn't any water locked up in long-term ice sheets.

Another is that water, like any other fluid, expands as it warms up, just as the oceans are doing now (albeit on a much more modest scale).

A third reason is the shape of the oceans themselves.

The faster separation of tectonic plates noted above would have produced larger mid-ocean ridges as new crustal material surged upward.

This, in turn, would have raised the height of the seafloor and the resulting sea level by surprisingly large amounts—perhaps enough to account for most of the difference between the Triassic–Eocene sea levels and today's.

From greenhouse to icehouse: The planet cools down

The most recent 1% of our atmosphere's history is mainly a story of cooling.

Starting in the mid- Eocene, around 50 million years ago, sea levels began to drop, continents shifted ever closer to their current locations, and global temperatures began to fall.

This long process wasn't linear— there were sharp warmings and coolings in between, each lasting many thousands of years—but overall, Earth's temperature took a tumble of perhaps 4°C (7.2°F) or more.

This eventually pushed the planet into the glacial–interglacial sequence that began fewer than three million years ago and continues today.

What caused the cooldown?

Once again, it appears that carbon dioxide was the most direct culprit. CO_2 in the atmosphere went from several times its modern-day amounts in the mid-Eocene to barely half of its present-day concentration during the last few ice ages.

The question of what drove down CO_2 levels is far more challenging.

Mid-ocean ridges were separating more slowly when the cooldown began, a sign that volcanic activity was pumping less CO_2 into the atmosphere.

The biggest decreases in volcanism apparently occurred prior to 40 million years ago.

That would correspond to the first pulse of global cooling, but the spreading rate had increased again by the time the second pulse began, leaving that one unexplained.

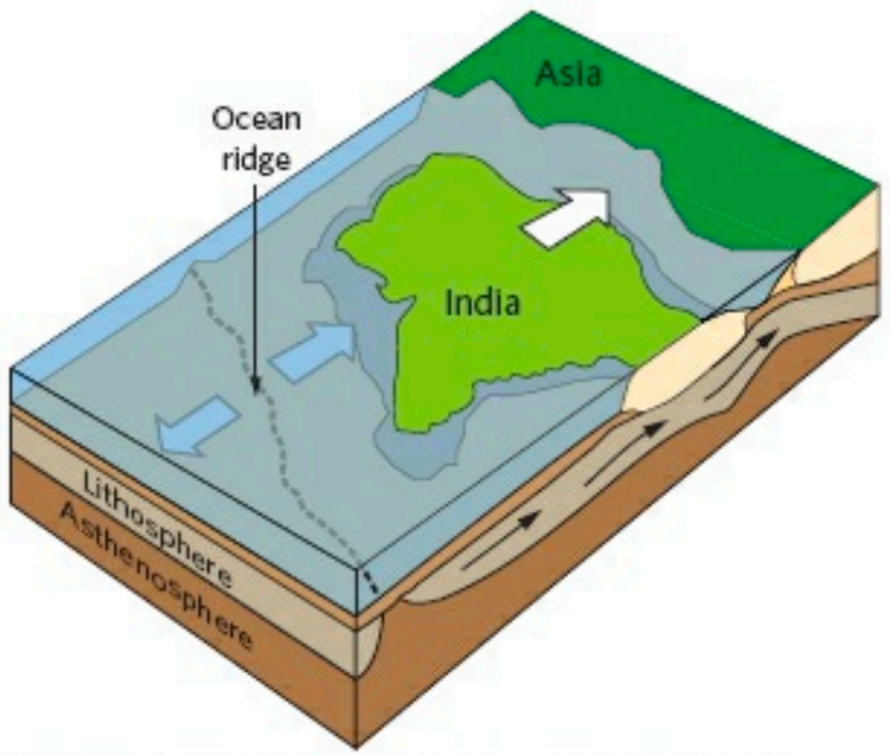
In their search for a solution to this riddle, one group of U.S. scientists cast their eyes toward the **Tibetan Plateau.**

This gigantic feature began to take shape around 55 million years ago as the Indian subcontinent joined Asia, pushing the Himalayas and the plateau upward.

The pace quickened about 30 million years ago, and today most of the Alaska-sized Tibetan Plateau sits at elevations greater than 3700 m (12,000 ft).

It's bordered by mountains that soar more than twice as high.

In the late 1980s, a combined Woods Hole Oceanographic Institution, Florida State University, and University of Virginia study proposed that the intense chemical weathering produced by the Tibetan Plateau could have pulled enough carbon dioxide from the air to trigger the global temperature drop that led to the most recent ice ages.



The Indian subcontinent about to crash into Asia 60 million years ago.

Normally, a drop in global temperatures would be expected to act as a negative feedback, reducing precipitation and slowing down weathering.

But in this case, we're faced with a change in the playing field (Earth's geography), and that could control the rate of weathering entirely on its own, overpowering any feedbacks.

If ever there was a geological event conducive to weathering, it's the uplift of the Tibetan Plateau—perhaps the most massive above-ground feature in Earth's history.

In the absence of enough data to confirm or disprove the study, the debate over the advent of Earth's ice ages goes on.

There were other tectonic developments of interest during the long cooldown. **Antarctica** separated from South America between 20 and 25 million years ago.

With air and water now able to circle around Antarctica unimpeded, the resulting tight circulation might have fostered the growth of the first major Antarctic ice sheet about 13 million years ago (though modeling studies have been inconclusive on this point).

Meanwhile, North and South America slowly approached each other, finally joining at the Isthmus of Panama roughly 3 million years ago.

This closed the long-standing oceanic connection through which warm, salty water had flowed beneath trade winds from the tropical Atlantic into the Pacific.

In a major readjustment, the Atlantic shifted into a mode much like the one we see today, with the tropical waters flowing northward along an intensified Gulf Stream and on into the far North Atlantic.

This influx of warm, saline water into the Arctic would have cut down on sea-ice formation.

Ironically, it could have also furnished the moisture needed to build ice sheets over land and thus hastened glaciation.

Thanks to some combination of the factors above—plus, perhaps, others yet to be discovered —Earth inexorably slid into what some experts call a “deteriorating” climate.

Antarctic ice grew in a start-and-stop fashion.

Glaciers began to appear about seven million years ago in the Andes, Alaska, and Greenland.

Finally, **ice sheets** began spreading across North America and Eurasia about 2.7 million years ago.

Since then, more than two dozen glaciations have ground their way across much of present-day Canada, northern Europe, and northwest Russia.

In between each glaciation was a shorter, milder period, with ice extent and temperatures closer to those at present and even warmer in some cases.

At first, these **interglacial** periods occurred in between ice ages about every 41,000 years.

This corresponds remarkably well to the 41,000-year cycle in the tilt of Earth's axis.

When Earth's tilt is maximized, the extra solar energy received at high latitudes in summer melts ice sheets more easily (even though winters are getting less sun than usual).

When the tilt is lower, summers are cooler, and ice sheets get a running start with unmelted snow from the previous year.

About a million years ago, however, the 41,000-year rhythm of ice ages ended.

The ice ages got stronger and began to last much longer—about 100,000 years each, with interglacial periods on the order of 10,000–15,000 years long.

What was behind this shift isn't yet clear.

Earth's orbit also has a 100,000-year cycle in its off-centeredness, or eccentricity, but most paleoclimatologists don't believe that this relatively weak cycle was enough in itself to wrench a change in the ice-age frequency.

However, it could have interacted with some other cyclic process to provoke the timing.

Along with what produced the current tempo of ice ages, scientists are also keenly interested in what's been maintaining it.

One of the more exotic hypotheses is that Earth's orbit may be passing through a disk of interplanetary dust about every 100,000 years.

Another intriguing idea, advanced by Harvard University, is that the 41,000-year cycle in Earth's tilt is still the main driver but that ice sheets respond in an alternating fashion every second or third peak (about every 80,000 or 120,000 years), which would lead to an average frequency of about every 100,000 years.

And a 2013 study by the University of Tokyo elaborated on a long-standing hypothesis that the vast ice sheets may overextend themselves in North America, eventually becoming so large that a cyclic solar peak can melt more of the ice's surface than snowfall can replenish.

There's much still to be learned.

The period just before the last ice age began, from about 130,000 to 115,000 years ago, provides a sobering illustration of what climate change could bring to our planet over the next few centuries.

That interglacial kicked off with a sharp warming most likely driven by orbital cycles.

The 2013 IPCC report concluded that sea level during the interglacial likely topped out between 5 and 10 m (16–33 ft) above today's global average, even though greenhouse gases were less abundant than today.

Computer modeling by the University of Arizona and NCAR showed that major ice-sheet melting at both poles may have fed the sea level rise.

The same modeling found that Arctic summer temperatures by 2100 will be close to those achieved in the last interglacial warming.

Although it could take decades or even centuries for that warming to translate into a 6-m (20-ft) sea level rise—which would swamp major coastal areas around the world—there are worries that the process of ice- sheet melt could prove irreversible sooner than we think.

Emerging from the ice

By the time of its peak about 20,000 years ago, the last ice age had carved out a very different- looking world than the one we know now.

North America and Europe were half encased in ice, with global temperatures anywhere from 3° to 8°C (5°–14°F) colder than today's averages.

South of these ice sheets, it was largely dry and windy, especially over continental areas.

Tundra covered most of southern Europe, and spruce trees characteristic of high latitudes extended into the mid-Atlantic area of the present-day United States.

One place where rainfall exceeded today's norm is the Great Basin: Nevada and Utah were freckled with dozens of large lakes that now exist only as salt flats.

Sea levels were so low—more than 100 m (330 ft) below today's— that Australia extended northward into New Guinea, and Japan and Korea were separated only by shallow water.

The ice age may not have looked exactly like this through its entire 100,000-year lifespan.

Until ice cores were plucked from Greenland and the Antarctic, many scientists had pictured ice-age climates as being relatively static.

Instead, the cores revealed that Earth swung through a number of distinct warmings and coolings during the start of the last ice age, about 115,000 years ago, up to and beyond its conclusion around 15,000 years ago.

As the clarity and completeness of ice-core data improved in the 1990s, the findings were accompanied by analyses of ocean sediments from several dispersed basins and stalagmites from caves in China.

The new evidence made it clear that warm and cold swings during the ice age were far more widespread, and could unfold far more rapidly, than scientists had previously thought.

Two of the orbital **cycles** discussed earlier involving Earth's tilt and its precession, synchronized to produce a strong peak in the summertime input of solar energy across the Northern Hemisphere about 15,000 years ago.

Ice sheets began to retreat dramatically, but the path to a warmer climate wasn't a smooth one.

As the melting intensified, water occasionally poured out of glacial lakes in spectacular bursts, some of which were large enough to influence ocean circulation and trigger rapid climate shifts.

Core value: Getting information out of ice

Our understanding of the last few ice ages has been boosted immeasurably by the rich vein of data obtained through ice cores from Greenland, Antarctica, and several other glacier-studded areas.

The task is laborious: each ice sheet is more than 3 km (1.9 mi) thick, but each cylindrical core is only about 10 cm (4 in.) in diameter, and each segment of the core must be drilled and removed a few meters at a time.

Weather only permits a few short months of work each year.

Even in midsummer, daytime temperatures can be colder than-20°C (-4°F) on top of the Greenland ice sheet; in central Antarctica, that's a mild summer day.

The first major ice cores were retrieved from both Antarctica and Greenland as part of the International Geophysical Year (IGY) of 1957-1958.

In Greenland, the first cores were taken at camps far from the thickest part of the ice sheet.

As interest in climate change intensified and researchers clamored for more of the ice, two parallel projects set out to gather definitive cores from the top of the sheet.

By the early 1990s, these U.S. and European teams-separated by only about 32 km (20 mi)-had drilled from the summit of Greenland's ice sheet to bedrock, a depth of just over 3 km (9900 ft).

The two cores they obtained each spanned a little more than 100,000 years of climate.

In 2010 a new coring project hit bedrock in northern Greenland; it has led to data extending back almost 130,000 years, enough to capture the warmth of the last interglacial period.

Meanwhile, Russian scientists at Antarctica's Vostok Station, which was founded during the IGY, were engaged in their own decades-long drilling effort.

By 1995, they'd reached a depth of around 3600 m (11,800 ft) before they halted to avoid contaminating a vast lake beneath the ice.

The Vostok core is only about 20% longer than those from Greenland.

However, since the highest reaches of the Antarctic ice sheet get so much less snowfall per year than Greenland's summit, each meter of the Vostok core details a much longer stretch of climate.

The total Vostok record covers about 440,000 years.

One of the most striking findings from both poles is that carbon dioxide and methane, two key greenhouse gases, rose and fell in near lockstep with the comings and goings of ice ages.

Typically, the concentrations dropped by 20%-50% during a glaciation, then quickly recovered afterward.

For more than a decade, the oldest core on Earth was the one completed in 2004 by the EU-sponsored European Project for Ice Coring in Antarctica (EPICA).

Retrieved at a site atop Dome C, about 500 km (300 mi) from Vostok, the EPICA core yielded climate analyses extending back 800,000 years.

More recently, a team from Princeton University cracked the code of "blue ice" —ancient ice pushed sideways by glacial stresses, making it difficult to date.

In 2017, the group announced they'd found a core with ice up to 2.7 million years old, providing a new window on the advent of the first ice age.

How fast can climate flip?

Evidence as far back as the 1970s hinted that climate during glacial periods might not have been as static as many believed.

The pieces came together in the 1990s, when scientists confirmed a sequence of major warmings and coolings that unfolded against the cold backdrop of the last ice age.

Instead of furnishing a constant chill, it seems the 100,000-year ice age unfolded in a much more irregular fashion.

Lamont-Doherty Earth Observatory drew on data from sediments in the North Atlantic to trace the evolution of the ice age in eye-opening detail.

The kind of upward and downward spikes they found don't seem to occur in warm regimes like our present-day, postglacial climate, but they serve as a reminder that climate can switch from one mode to another in the geologic equivalent of a heartbeat.

The great ice sheets that coated much of Europe and North America never disappeared during these warmings and coolings, but conditions across many ocean and continental areas varied sharply as the temperature swings came and went.

(1) **Going up.** On 25 different occasions during the last ice age, global air temperatures are believed to have climbed about half of the way back to their interglacial levels, then sunk back to more typical ice-age readings.

The warm-ups pushed average temperatures in Greenland up by as much as 16°C (29°F) in as little as 40 years, while the much slower return to glacial cold took about a thousand years.

These warmings-which appear to have been concentrated in the Northern Hemisphere-are called Dansgaard-Oeschger (D-O) events.

(2) **Going down.** Less frequently-at six points during the ice age-climate lurched in the other direction.

Vast fields of icebergs poured from North America into the North Atlantic, disrupting the ocean circulation and cooling the climate for several hundred years.

These so-called Heinrich events.

Together, the Heinrich and D-O events explain much of the variability in global climate that shows up in sediments and ice cores from the last 100,000 years.

Scientists are still hunting for what might lie behind the timing and occurrence of both types of climate swings.

For instance, each Heinrich event occurs after three to five D-O events, implying that the extended warm spells might have progressively destabilized the growing ice sheets and led to a Heinrich-style iceberg armada.

Another interesting facet is that many of the D-O events are separated by around 1500 years, with a few spaced at about 3000 and 4500 years.

Some paleoclimatologists say this suggests that the events are triggered by an underlying cycle... but that sometimes a beat or two is skipped.

Other paleoclimatologists believe these sub-beats may be little more than random variations.

Perhaps the most dramatic was the **Younger Dryas** period, a frigid encore that brought a return to cold nearly as extreme as the ice age itself.

The period is named after a tenacious alpine flower—*Dryas octopetala*—that thrives in conditions too austere for most other plant life.

The flower's pollen, preserved in Scandinavian bogs, provides evidence of extremely cold periods.

The most recent (youngest) such period began nearly 13,000 years ago and persisted for some 1300 years.

Ice cores from Greenland indicate that the Younger Dryas took as little as a decade both to set in and to release its grip.

During the interim, average temperatures across the British Isles sank to as low as −5°C (23°F), and snowfall increased across many mountainous parts of the world.

Drought and cold swept across the Fertile Crescent of the Middle East.

This may even be what prodded the ancient Natufian culture to pull back from wide-ranging hunting and gathering and shift toward what ranks as the world's earliest documented agriculture.

As for the cause of the Younger Dryas, a few researchers argue that a possible meteor impact around 12,800 years ago is the most likely trigger, with recent work finding hints of widespread dust deposits and wildfires.

However, most paleoclimatologists point to the massive draining of **Lake Agassiz**, a vast glacial remnant that covered parts of present-day Manitoba, Ontario, Minnesota, and North Dakota.

Such an influx of fresh water could disrupt the thermohaline circulation that sends warm waters far into the North Atlantic.

There's spirited debate about whether the Agassiz melt flowed in the right direction to affect the North Atlantic.

An even more perplexing piece of the puzzle comes from the Southern Hemisphere, where records from Antarctica and other areas show that the cooldown there began a thousand years earlier—closer to 14,000 years ago.

The Younger Dryas was followed by a much shorter, less intense cooling some 8200 years ago, lasting about 150 years.

The next few thousand years were, by and large, a nourishing time for civilizations.

Scientists who map out the climate for this period can draw on human history as well as other biological, physical, and chemical clues.

Across northern Africa and southern Asia, the peak in summertime sunshine stimulated by orbital cycles during the postglacial era produced monsoons far more intense and widespread than today's.

From about 14,000 to 6000 years ago, much of the Sahara, including the drought- prone Sahel, was covered in grasses and studded with lakes that played host to crocodiles and hippopotami.

Traces of ancient streams have been detected by satellite beneath the sands of Sudan, while the Nile flowed at perhaps three times its present volume.

Meanwhile, from about 9000 to 5000 years ago, most of North America and Europe were ice-free, and summer temperatures there likely reached values similar to today, perhaps a touch warmer in some areas and at some times.

These mild times weren't destined to last forever.

Starting about 5000 years ago, global temperatures began a gradual cooldown, interspersed with warmings lasting a few hundred years each.

This pattern is consistent with ice-core records from toward the end of previous interglacial periods, although it's far from clear when the next ice age will actually arrive.

One of the earliest and sharpest signs of this global cooling is the prolonged drought that crippled Near East civilizations about 4200 years ago.

Orbital cycles had already weakened the summer sunlight that powered bountiful monsoons across Africa and Asia.

At some point, the gradual drying killed off the grasses that covered much of the Sahara, which quickly pushed the region from savannah toward the desert that still prevails.

From AD 1000 to today (and beyond

Because of their context-setting importance for what lies ahead, the last thousand years of climate have drawn intense scrutiny.

The last millennium began near the peak of a 300-year stretch of warmth centered around the North Atlantic called the **Medieval Warm Period** (also called the Medieval Climatic Optimum or Medieval Climate Anomaly).

Temperatures in some parts of the world—especially present-day North America—were close to modern-day levels.

It may indeed have been an optimal climate for some Europeans, or at least a livable one, but people in parts of the Americas suffered from intense drought that ravaged entire cultures.

One example is the **Mayans**, whose empire extended from present-day Mexico across much of Central America.

After nearly 2000 years of development, many Mayan cities were abandoned between about AD 800 and AD 950.

Sediments in the nearby Caribbean reveal several strong multiyear droughts during this period and on into the next century.

The Mayan collapse in eastern Mexico apparently coincided with that region's most intense and prolonged drought in over a millennium.

To the north, the region near the current Four Corners area settled by Ancestral Pueblo peoples experienced periods of erratic rain and snow during the Medieval Warm Period.

Tree-ring studies indicate an especially intense drought near the end of the 1200s, when the region's emblematic cliff dwellings were abandoned.

Scholars have warned against pinning the Mayan and Ancestral Pueblo declines solely on drought, pointing to signs of other factors such as migration, a high demand on trees and other resources, and power struggles within each culture and with neighboring peoples.

Even so, climatic stresses may have been the last straws—and certainly they can't have helped matters.

After the Medieval Warm Period faded, the long-term cooling resumed, intensifying during the 1400s and continuing into the mid-1800s.

This was the Little Ice Age, which sent the global average temperature down by as much as 0.5°C (0.9°F) to its lowest century-scale levels in thousands of years.

It wasn't a uniform cooling: some parts of the world appear to have seen little effect, while temperatures plummeted across much of North America and Eurasia.

Periodic plagues and famines ravaged Europe, and glaciers descended from the Alps to engulf a number of villages.

The chill wasn't uniform in time, either.

It was marked by an irregular seesaw of rapid climatic shifts, driven by complex and still little understood interactions between the atmosphere and the ocean.

One influence may have been a drop in **solar energy.**

Isotopes of carbon in tree rings and beryllium in ice cores show a drop-off in solar radiation during much of the Little Ice Age.

Moreover, sunspot observations that began around 1610 show a near-absence of reported sunspots between 1645 and 1715 (the so-called Maunder Minimum).

However, recent studies have brought down the relative importance of this solar effect on the Little Ice Age.

Debate continues, but one current view is that the solar slowdown acted mainly to help reinforce the direct effects of volcanoes, which appear to have erupted more frequently from the late 1200s onward than during the Medieval Warm Period.

The 1815 eruption of Indonesia's **Mount Tambora**—the most violent ever recorded on Earth—led to a disastrously cold summer across much of the globe in 1816.

That "year without a summer" brought crop failures to northern Europe as well as snows in Vermont as late as early June.

After the mid-1800s, Earth's climate took a decided turn for the warmer.

The overall climb in temperatures over the twentieth century, however, was punctuated by an apparent flattening— perhaps even a slight cooling—from around the mid-1940s to the late 1970s.

The cooling was sharpest in the Northern Hemisphere, and its causes are still being debated.

The leading hypothesis is that it's related mainly to sun-reflecting aerosol pollution spewed out from North America and Europe in the industrial boom after World War II.

Modeling carried out for the 2013 IPCC report confirms this view, especially when the indirect effect of aerosols on clouds is taken into account, according to a study in *Environmental Research Letters* led by the U.K. National Centre for Atmospheric Science.

In an alternate view, much of the midcentury hemispheric cooling can be traced back to two ocean-related factors, according to a pair of studies in 2008 and 2010 led by the Colorado State University.

One of those factors is an apparent drop in SSTs in 1945 that proved to be spurious, the result of a temporary increase in canvas-bucket measurements at the end of World War II.

The other is a major drop in Northern Hemisphere SSTs from 1968 to 1972, focused in the North Atlantic and perhaps linked to a large influx of fresh water observed around the same time.

Global warming resumed with gusto in the 1980s, and by the end of the twentieth century it was becoming clear that temperatures were at least on a par with those of the Medieval Warm Period.

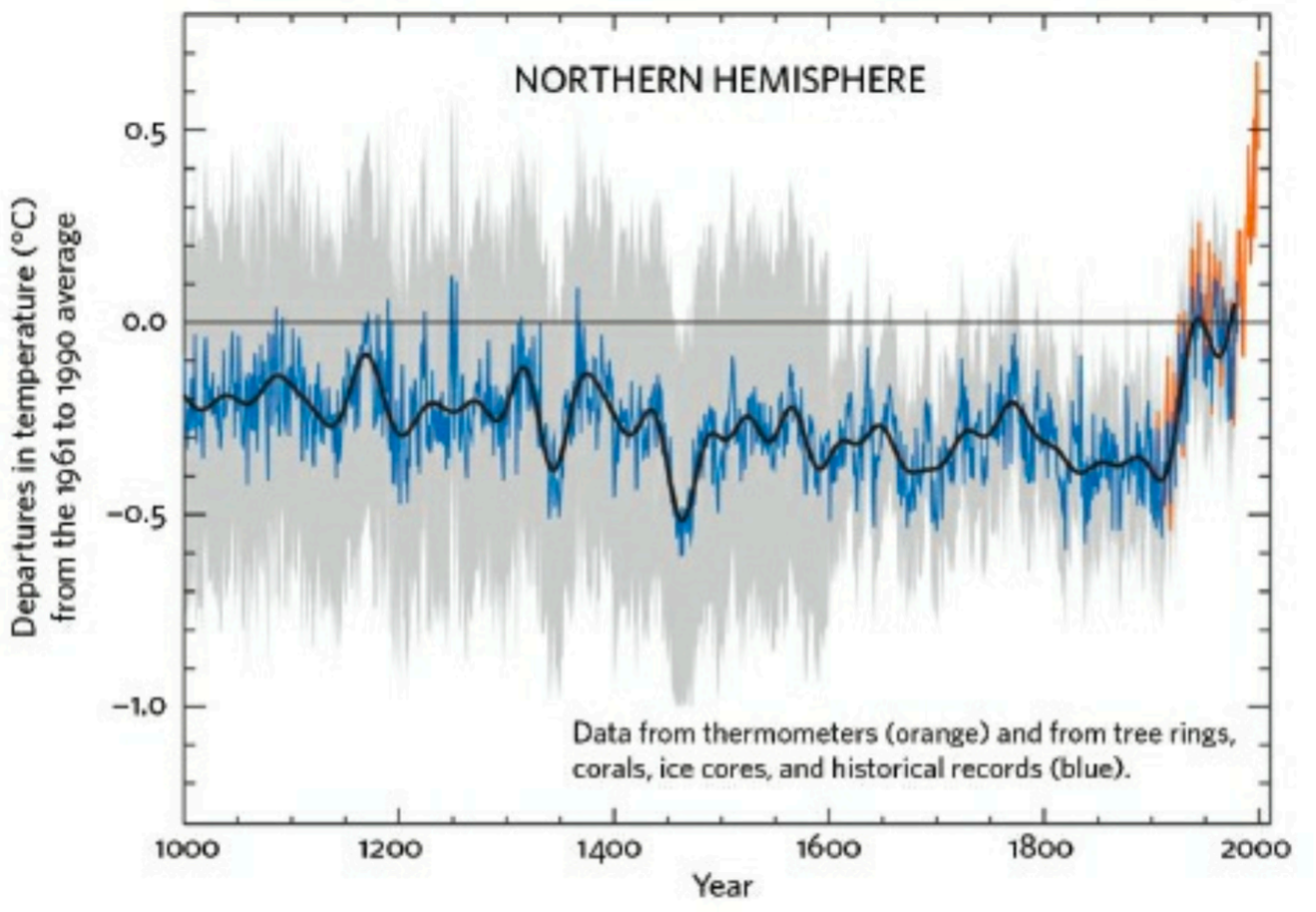
That recognition triggered a high-profile debate that brought unprecedented attention to the fine points of paleoclimate analysis.

In the end, an important finding about the last 1000-plus years of climate was confirmed.

In the late 1990s, the University of Virginia, the University of Massachusetts Amherst and the Laboratory of Tree-Ring Research carried out the first attempt to reconstruct global temperature from the past millennium based on varied proxy data from around the world.

Initial results reported that the 1990s had likely been the warmest decade of the last 1000 years.

The key graphic produced, nicknamed the “hockey stick” for its distinctive shape, was prominently featured in the IPCC’s 2001 report.



The “hockey stick” graph featured in the 2001 IPCC assessment, showing temperature (departures from the 1961–1990 average) in the Northern Hemisphere during the past 1000 years. (IPCC)

One thing remains crystal clear: the greenhouse gas content of our atmosphere has entered territory never before explored in human history.

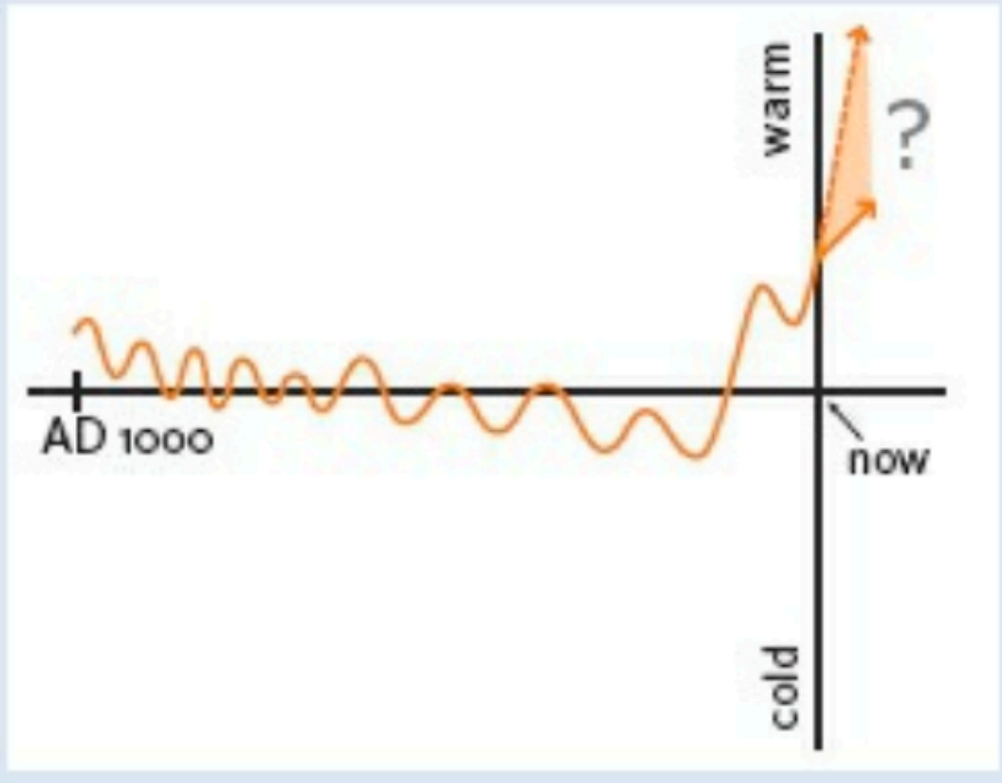
Given what we know about these gases, the implications for our climate are profound.

Four ways to think of past and future climate

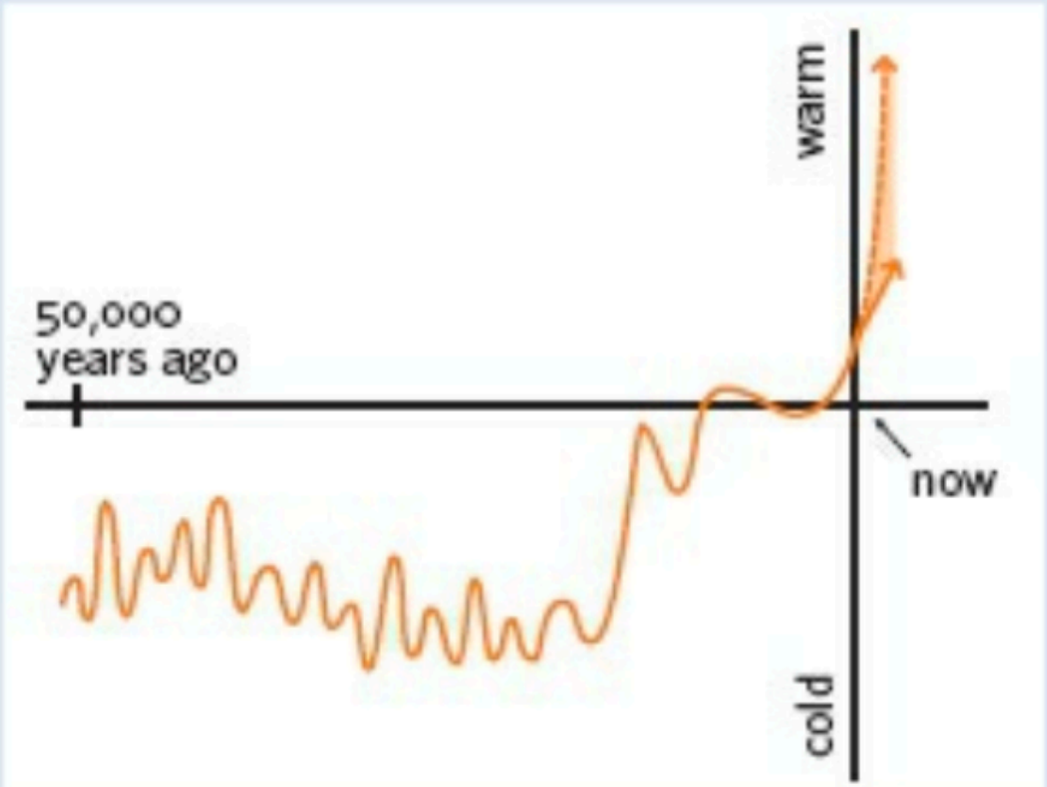
The answer to whether Earth is warming or cooling depends in large part on what time frame you're considering.

Each graph below includes a very rough schematic of past (and plausible future) temperatures.

- **Over hundreds of years.** Around AD 1000, Earth was nearly as warm as it is now. After that, temperatures over much of the planet cooled by around 0.2°–0.5°C (0.4°–0.8°F), mostly during the period known as the Little Ice Age, from the 1400s to the mid-1800s. The global average has since rebounded by roughly 1°C (1.8°F) as human-produced greenhouse gases have increased. According to the 2013 IPCC report, further warming this century could run anywhere from 0.3° to 4.8°C (0.5°–8.6°F), depending on the rate of emissions. Further warming after 2100 could be substantial if greenhouse emissions continue to increase through this century, and the delayed response of ice sheets and deep oceans to warming could produce major sea level rise over several centuries.

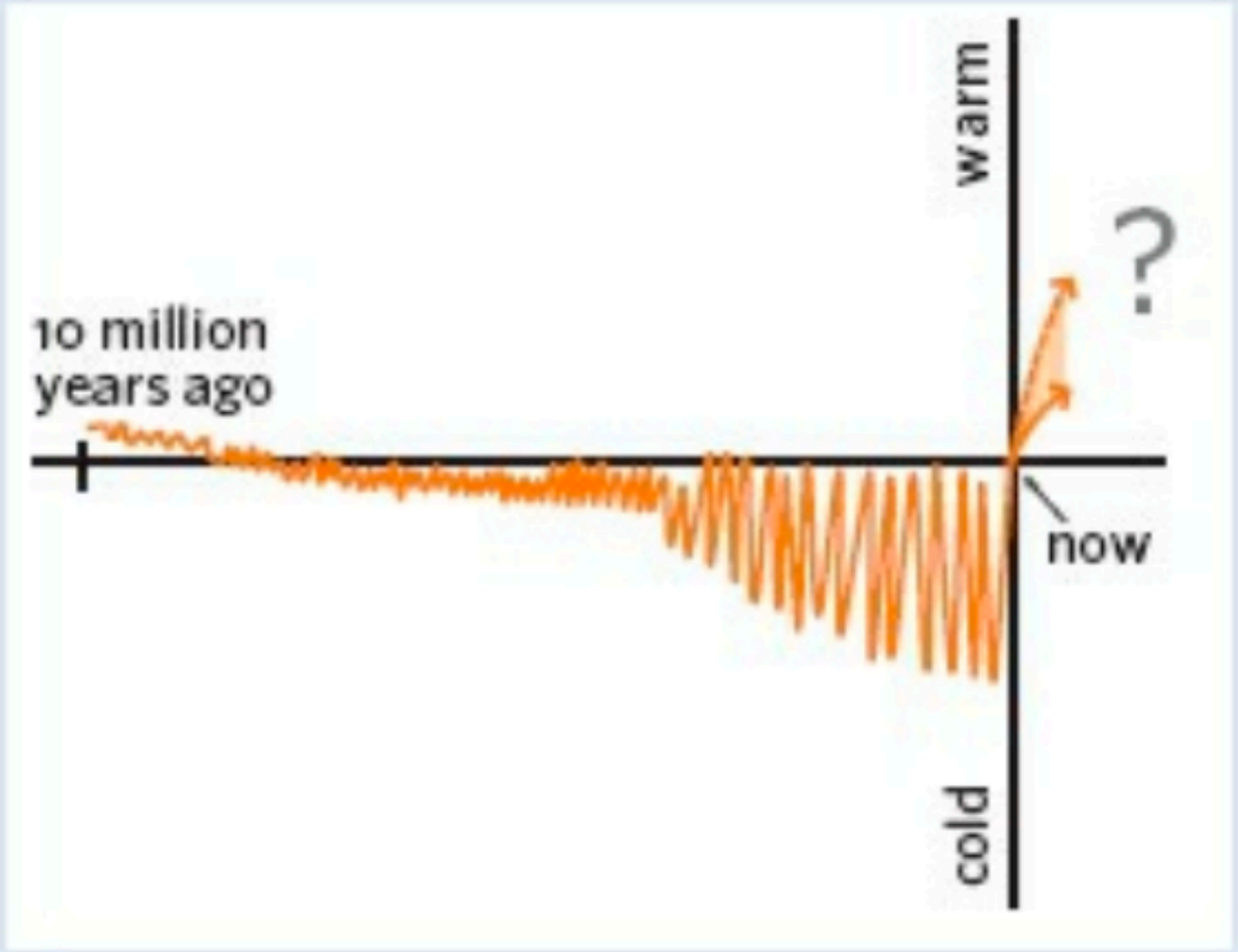


- **Over tens of thousands of years.** The most recent ice age began about 115,000 years ago and ended about 11,500 years ago. Then came a dramatic warm-up, which lasted until about 3000 BC. Since then, Earth's temperature has changed relatively little, with a very slight cooling interrupted by warmer periods and punctuated by the last century's sharp temperature rise.

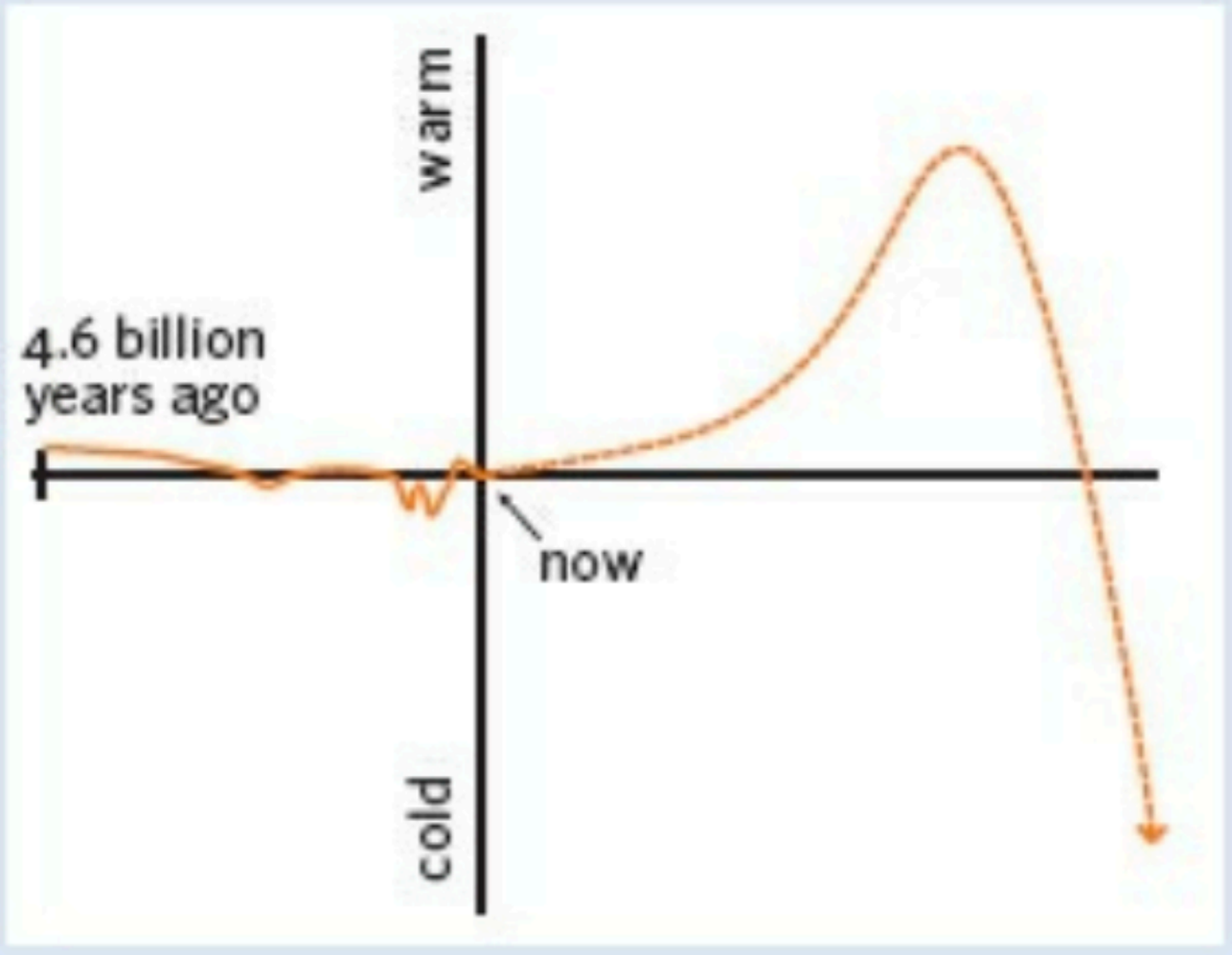


More than a thousand years from now, after humans have exhausted (or stopped burning) fossil fuels and much of the resulting greenhouse gas has left the atmosphere through natural processes (mostly through slow absorption by the ocean), we may return to cooler times. If the length of the last interglacial is any guide, we're due for another ice age in the next few thousand years, although some scientists believe it's more likely to take 20,000 years or more. It's even possible that a mix of polar melting, land-use changes, and slowly waning greenhouse gas concentrations could postpone the next ice age for an undetermined period.

- **Over tens of millions of years.** A gradual, though sometimes erratic, cooling trend has been under way for at least 55 million years, perhaps due in part to carbon dioxide removed by weathering on top of the growing Himalayas. Northern Hemisphere glaciations began about 2.7 million years ago. Warm and cool periods come and go on top of this overall cooling. Simple extrapolation would keep Earth on this extremely slow cooling trend, although there's no evidence to tell us how long that might last.



Over billions of years. On this scale, the details are blurry, but the overall picture is clear. The sun will continue to heat up and eventually undergo changes to its size and structure, producing a climate on Earth hot enough to evaporate the oceans and make life impossible. Eventually, the sun will shrink and start to fade, with the solar system reaching its cold "final configuration" in around 12 billion years.



Are you ready?

10 Big Findings from the 2023 IPCC Report on Climate Change

The report provides the most comprehensive, best available scientific assessment of climate change.

It also makes for grim reading.

Across nearly 8,000 pages, the report details the devastating consequences of rising greenhouse gas (GHG) emissions around the world — the destruction of homes, the loss of livelihoods and the fragmentation of communities, for example — as well as the increasingly dangerous and irreversible risks should we fail to change course.

But the IPCC also offers hope, highlighting pathways to avoid these intensifying risks.

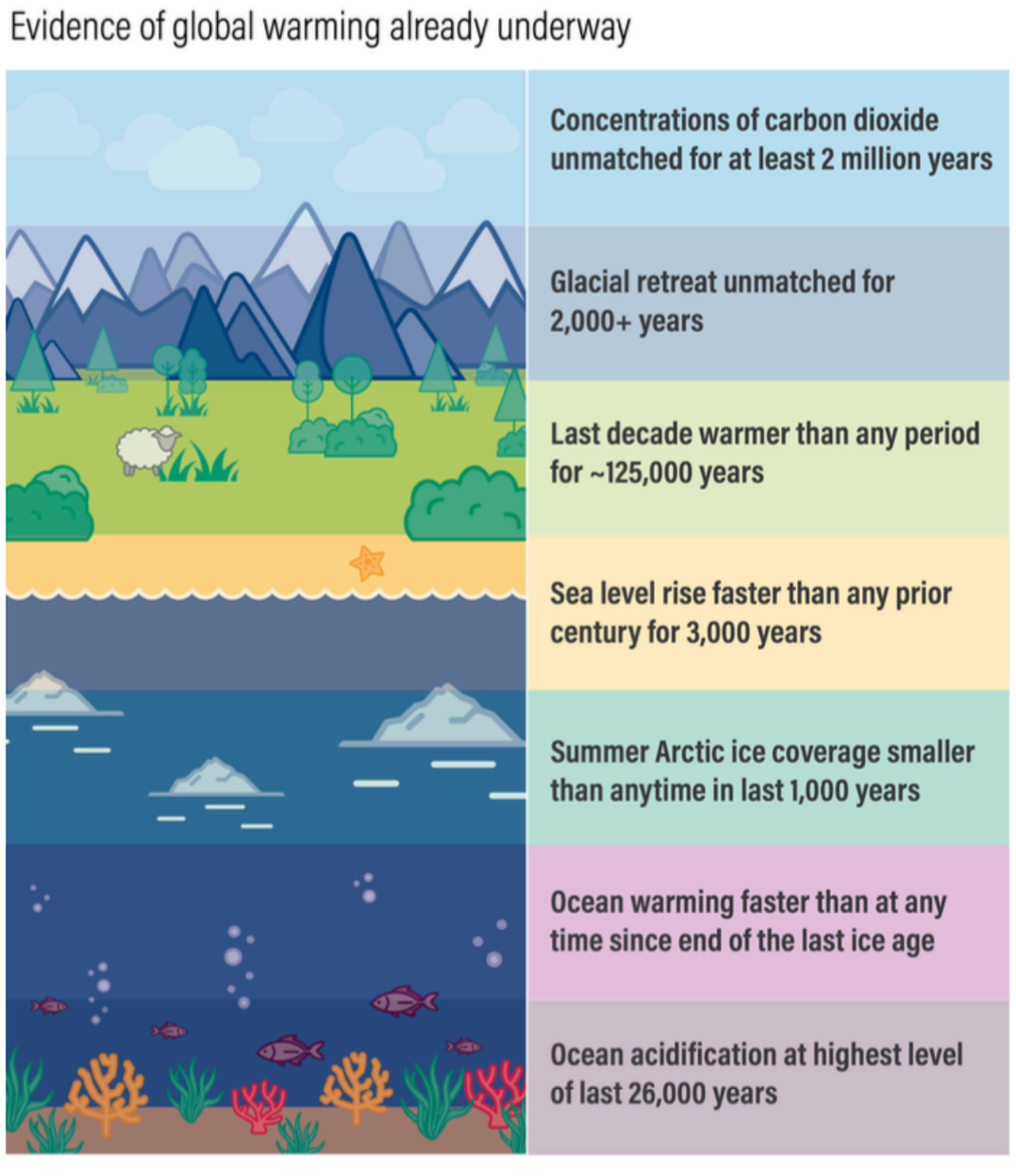
It identities readily available, and in some cases, highly cost-effective actions that can be undertaken now to reduce GHG emissions, scale up carbon removal and build resilience.

While the window to address the climate crisis is rapidly closing, the IPCC affirms that we can still secure a safe, livable future.

Here are 10 key findings you need to know:

1. Human-induced global warming of 1.1 degrees C has spurred changes to the Earth's climate that are unprecedented in recent human history.

Already, with 1.1 degrees C (2 degrees F) of global temperature rise, changes to the climate system that are unparalleled over centuries to millennia are now occurring in every region of the world, from rising sea levels to more extreme weather events to rapidly disappearing sea ice.



Additional warming will increase the magnitude of these changes.

Every 0.5 degree C (0.9 degrees F) of global temperature rise, for example, will cause clearly discernible increases in the frequency and severity of heat extremes, heavy rainfall events and regional droughts.

Similarly, heatwaves that, on average, arose once every 10 years in a climate with little human influence will likely occur 4.1 times more frequently with 1.5 degrees C (2.7 degrees F) of warming, 5.6 times with 2 degrees C (3.6 degrees F) and 9.4 times with 4 degrees C (7.2 degrees F) — and the intensity of these heatwaves will also increase by 1.9 degrees C (3.4 degrees F), 2.6 degrees C (4.7 degrees F) and 5.1 degrees C (9.2 degrees F) respectively.

Rising global temperatures also heighten the probability of reaching dangerous tipping points in the climate system that, once crossed, can trigger self-amplifying feedbacks that further increase global warming, such as thawing permafrost or massive forest dieback.

Setting such reinforcing feedbacks in motion can also lead to other substantial, abrupt and irreversible changes to the climate system.

Should warming reach between 2 degrees C (3.6 degrees F) and 3 degrees C (5.4 degrees F), for example, the West Antarctic and Greenland ice sheets could melt almost completely and irreversibly over many thousands of years, causing sea levels to rise by several meters.

2. Climate impacts on people and ecosystems are more widespread and severe than expected, and future risks will escalate rapidly with every fraction of a degree of warming.

Described as an "an atlas of human suffering and a damning indictment of failed climate leadership" by United Nations Secretary-General António Guterres, one of the report's most alarming conclusions is that adverse climate impacts are already more far-reaching and extreme than anticipated.

About half of the global population currently contends with severe water scarcity for at least one month per year, while higher temperatures are enabling the spread of vector-borne diseases, such as malaria, West Nile virus and Lyme disease.

Climate change has also slowed improvements in agricultural productivity in middle and low latitudes, with crop productivity growth shrinking by a third in Africa since 1961.

And since 2008, extreme floods and storms have forced over 20 million people from their homes every year.

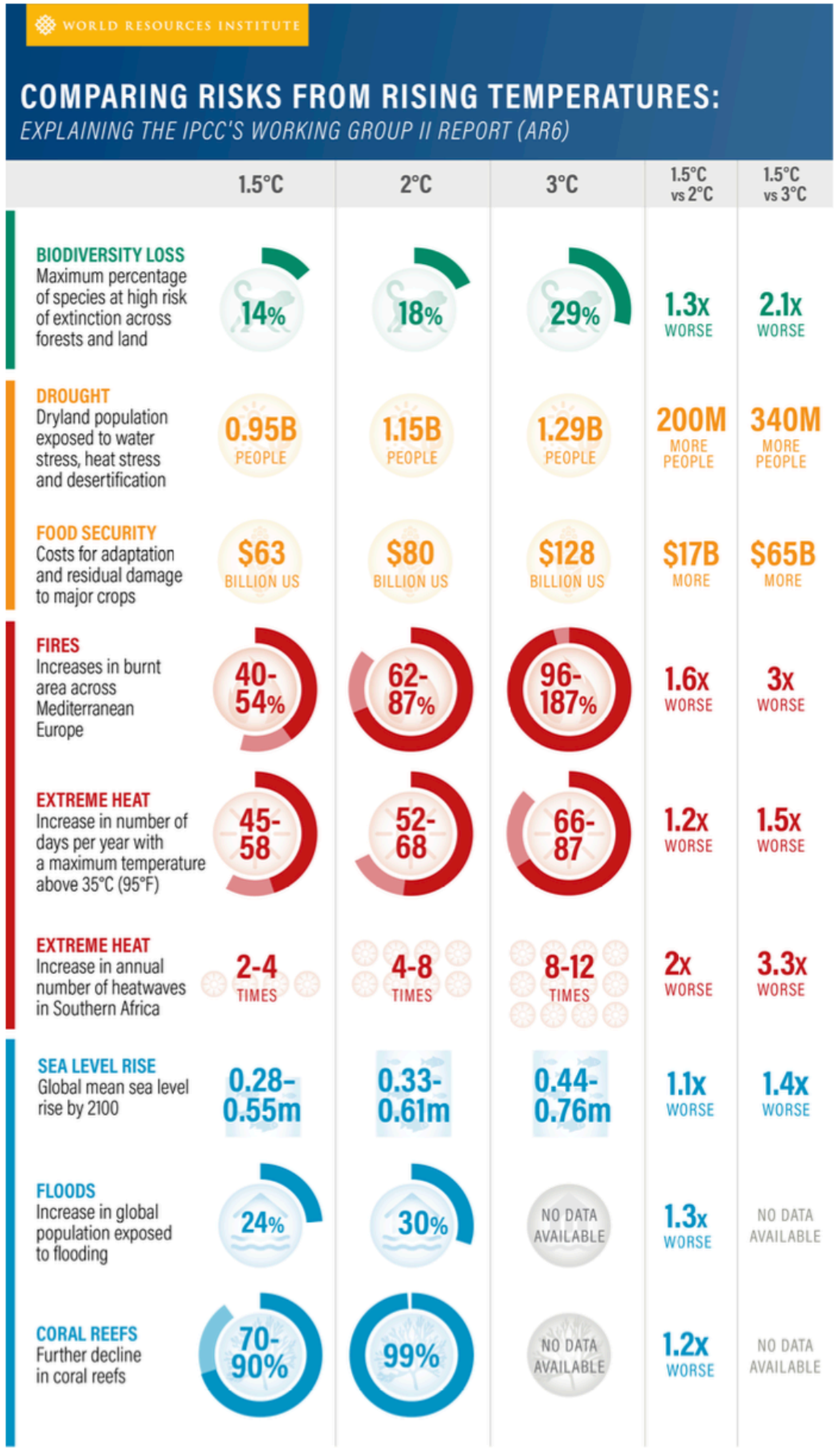
Every fraction of a degree of warming will intensify these threats, and even limiting global temperature rise to 1.5 degree C is not safe for all.

At this level of warming, for example, 950 million people across the world's drylands will experience water stress, heat stress and desertification, while the share of the global population exposed to flooding will rise by 24%.

Similarly, overshooting 1.5 degrees C (2.7 degrees F), even temporarily, will lead to much more severe, oftentimes irreversible impacts, from local species extinctions to the complete drowning of salt marshes to loss of human lives from increased heat stress.

Limiting the magnitude and duration of overshooting 1.5 degrees C (2.7 degrees F), then, will prove critical in ensuring a safe, livable future, as will holding warming to as close to 1.5 degrees C (2.7 degrees F) or below as possible.

Even if this temperature limit is exceeded by the end of the century, the imperative to rapidly curb GHG emissions to avoid higher levels of warming and associated impacts remains unchanged.



3. Adaptation measures can effectively build resilience, but more finance is needed to scale solutions.

Climate policies in at least 170 countries now consider adaptation, but in many nations, these efforts have yet to progress from planning to implementation.

Measures to build resilience are still largely small-scale, reactive and incremental, with most focusing on immediate impacts or near-term risks.

This disparity between today's levels of adaptation and those required persists in large part due to limited finance.

According to the IPCC, developing countries alone will need $127 billion per year by 2030 and $295 billion per year by 2050 to adapt to climate change.

But funds for adaptation reached just $23 billion to $46 billion from 2017 to 2018, accounting for only 4% to 8% of tracked climate finance.

The good news is that the IPCC finds that, with sufficient support, proven and readily available adaptation solutions can build resilience to climate risks and, in many cases, simultaneously deliver broader sustainable development benefits.

Ecosystem-based adaptation can protect lives and livelihoods

HAZARD: Loss of life and assets due to intense wildfires
SOLUTION: Forest management to reduce risk of super-fires

HAZARD: Asset loss, yield reduction and contamination due to flooding
SOLUTION: Restore wetlands to absorb and filter flood waters

HAZARD: Crop failures and live-stock loss due to drought
SOLUTION: Agro-forestry to make better use of soil moisture and reduce evaporation

HAZARD: Urban flooding due to intense rainfall
SOLUTION: Restore water catchment areas and expand green spaces to reduce flood risks

HAZARD: Loss of land, livelihoods and assets due to rising sea levels and coastal erosion
SOLUTION: Protect and restore coastal wetlands to combat coastal erosion and keep pace with rising seas

HAZARD: Landslides, soil loss and siltation due to intense rainfall
SOLUTION: Protect and restore forests to stabilize soils and slow water runoff

HAZARD: Reduced or intermittent river flow due to drought
SOLUTION: Protect and restore forests and watersheds to regulate flow

HAZARD: Asset loss, yield reduction and transport disruption due to flooding
SOLUTION: Protect and restore forests to slow water run-off

HAZARD: Heat stress due to urban heat islands
SOLUTION: Expand green spaces and plant trees to help cool cities

HAZARD: Loss of life and assets due to storm surges and inundation
SOLUTION: Protect and restore mangroves, marshes, and reefs to buffer coasts and absorb floodwaters



Source: Global Commission on Adaptation 2019.

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Ecosystem-based adaptation, for example, can help communities adapt to impacts that are already devastating their lives and livelihoods, while also safeguarding biodiversity, improving health outcomes, bolstering food security, delivering economic benefits and enhancing carbon sequestration.

Many ecosystem-based adaptation measures — including the protection, restoration and sustainable management of ecosystems, as well as more sustainable agricultural practices like integrating trees into farmlands and increasing crop diversity — can be implemented at relatively low costs today.

Meaningful collaboration with Indigenous Peoples and local communities is critical to the success of this approach, as is ensuring that ecosystem-based adaptation strategies are designed to account for how future global temperature rise will impact ecosystems.

4. Some climate impacts are already so severe they cannot be adapted to, leading to losses and damages.

Around the world, highly vulnerable people and ecosystems are already struggling to adapt to climate change impacts.

For some, these limits are “soft” — effective adaptation measures exist, but economic, political and social obstacles constrain implementation, such as lack of technical support or inadequate funding that does not reach the communities where it’s needed most.

But in other regions, people and ecosystems already face or are fast approaching “hard” limits to adaptation, where climate impacts from 1.1 degrees C (2 degrees F) of global warming are becoming so frequent and severe that no existing adaptation strategies can fully avoid losses and damages.

Coastal communities in the tropics, for example, have seen entire coral reef systems that once supported their livelihoods and food security experience widespread mortality, while rising sea levels have forced other low-lying neighborhoods to move to higher ground and abandon cultural sites.

Whether grappling with soft or hard limits to adaptation, the result for vulnerable communities is oftentimes irreversible and devastating.



Coral has turned into rubble in the shallow waters off the coast of Indonesia. Increased temperatures from climate change means that mortality has increased in coral reef systems in coastal communities. Photo by Velvetfish/iStock.

Such losses and damages will only escalate as the world warms.

Beyond 1.5 degrees C (2.7 degrees F) of global temperature rise, for example, regions reliant on snow and glacial melt will likely experience water shortages to which they cannot adapt.

At 2 degrees C (3.6 degrees F), the risk of concurrent maize production failures across important growing regions will rise dramatically.

And above 3 degrees C (5.4 degrees F), dangerously high summertime heat will threaten the health of communities in parts of southern Europe.

Urgent action is needed to avert, minimize and address these losses and damages.

At COP27, countries took a critical step forward by agreeing to establish funding arrangements for loss and damage, including a dedicated fund.

While this represents a historic breakthrough in the climate negotiations, countries must now figure out the details of what these funding arrangements, as well as the new fund, will look like in practice — and it's these details that will ultimately determine the flows to those experiencing loss and damage.

5. Global GHG emissions peak before 2025 in 1.5 degrees C-aligned pathways.

The IPCC finds that there is a more than 50% chance that global temperature rise will reach or surpass 1.5 degrees C (2.7 degrees F) between 2021 and 2040 across studied scenarios, and under a high-emissions pathway, specifically, the world may hit this threshold even sooner — between 2018 and 2037.

Global temperature rise in such a carbon-intensive scenario could also increase to 3.3 degrees C to 5.7 degrees C (5.9 degrees F to 10.3 degrees F) by 2100.

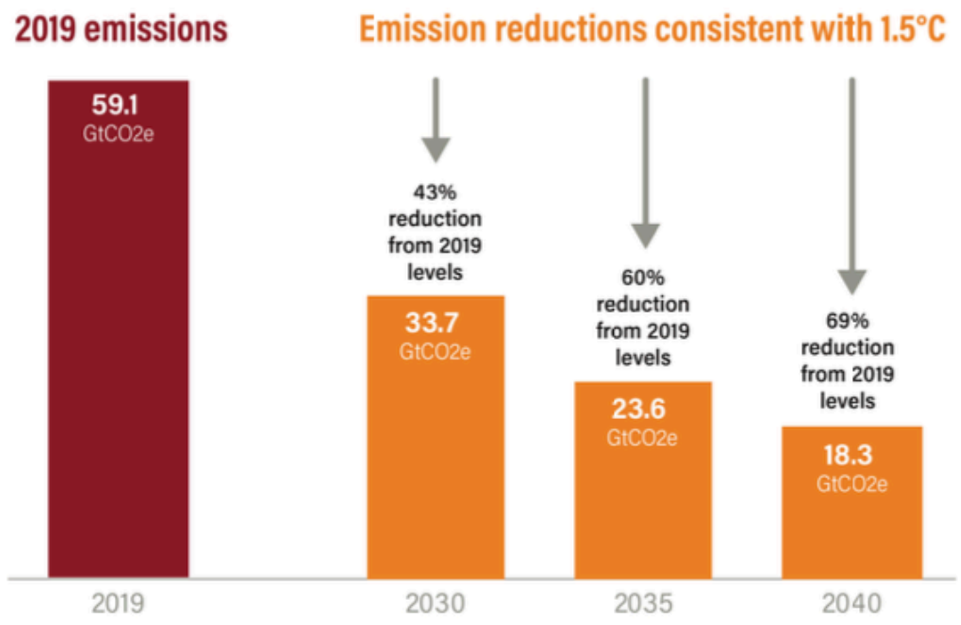
To put this projected amount of warming into perspective, the last time global temperatures exceeded 2.5 degrees C (4.5 degrees F) above pre-industrial levels was more than 3 million years ago.

Changing course to limit global warming to 1.5 degrees C (2.7 degrees F) — with no or limited overshoot — will instead require deep GHG emissions reductions in the near-term.

In modelled pathways that limit global warming to this goal, GHG emissions peak immediately and before 2025 at the latest.

They then drop rapidly, declining 43% by 2030 and 60% by 2035, relative to 2019 levels.

GHG emission reductions needed to keep 1.5°C within reach



Note: Analysis of pathways that limit warming to 1.5 degrees C with no or limited overshoot.
Source: IPCC AR6.
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While there are some bright spots — the annual growth rate of GHG emissions slowed from an average of 2.1% per year between 2000 and 2009 to 1.3% per year between 2010 and 2019, for example — global progress in mitigating climate change remains woefully off track.

GHG emissions have climbed steadily over the past decade, reaching 59 gigatonnes of carbon dioxide equivalent (GtCO2e) in 2019 — approximately 12% higher than in 2010 and 54% greater than in 1990.

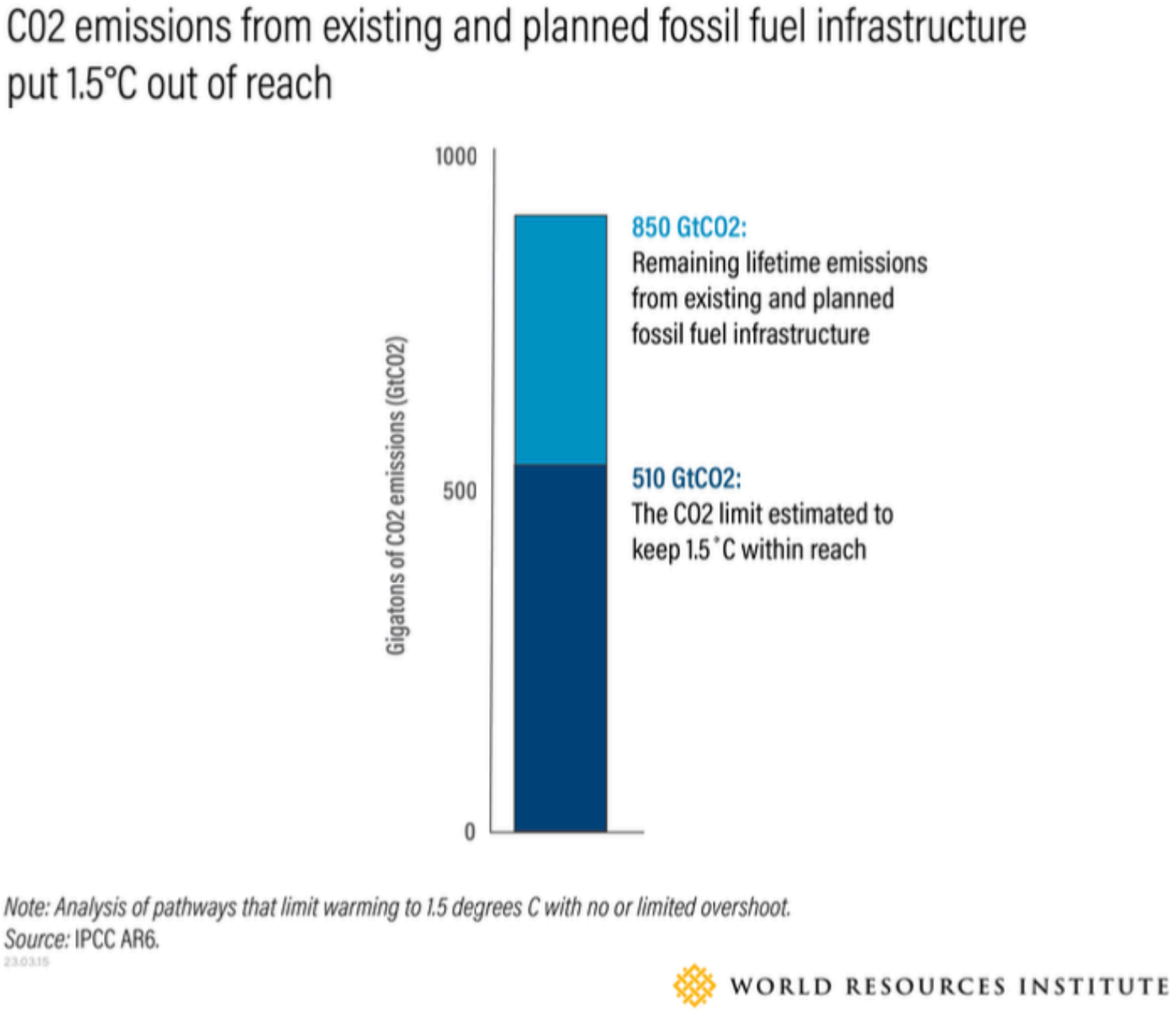
Even if countries achieved their climate pledges (also known as nationally determined contributions or NDCs), WRI research finds that they would reduce GHG emissions by just 7% from 2019 levels by 2030, in contrast to the 43% associated with limiting temperature rise to 1.5 degrees C (2.7 degrees F).

And while handful of countries have submitted new or enhanced NDCs since the IPCC's cut-off date, more recent analysis that takes these submissions into account finds that these commitments collectively still fall short of closing this emissions gap.

6. The world must rapidly shift away from burning fossil fuels — the number one cause of the climate crisis.

In pathways limiting warming to 1.5 degrees C (2.7 degrees F) with no or limited overshoot just a net 510 GtCO2 can be emitted before carbon dioxide emissions reach net zero in the early 2050s.

Yet future carbon dioxide emissions from existing and planned fossil fuel infrastructure alone could surpass that limit by 340 GtCO2, reaching 850 GtCO2.



A mix of strategies can help avoid locking in these emissions, including retiring existing fossil fuel infrastructure, canceling new projects, retrofitting fossil-fueled power plants with carbon capture and storage (CCS) technologies and scaling up renewable energy sources like solar and wind (which are now cheaper than fossil fuels in many regions).

In pathways that limit warming to 1.5 degrees C (2.7 degrees F) — with no or limited overshoot — for example, global use of coal falls by 95% by 2050, oil declines by about 60% and gas by about 45%.

These figures assume significant use of abatement technologies like CCS, and without them, these same pathways show much steeper declines by mid-century.

Global use of coal without CCS, for example, is virtually phased out by 2050. Although coal-fired power plants are starting to be retired across Europe and the United States, some multilateral development banks continue to invest in new coal capacity.

Failure to change course risks stranding assets worth trillions of dollars.

7. We also need urgent, systemwide transformations to secure a net-zero, climate- resilient future.

While fossil fuels are the number one source of GHG emissions, deep emission cuts are necessary across all of society to combat the climate crisis.

Power generation, buildings, industry, and transport are responsible for close to 80% of global emissions while agriculture, forestry and other land uses account for the remainder.

Take the transport system, for instance.

Drastically cutting emissions will require urban planning that minimizes the need for travel, as well as the build-out of shared, public and nonmotorized transport, such as rapid transit and bicycling in cities.

Such a transformation will also entail increasing the supply of electric passenger vehicles, commercial vehicles and buses, coupled with wide-scale installation of rapid-charging infrastructure, investments in zero-carbon fuels for shipping and aviation and more.

Policy measures that make these changes less disruptive can help accelerate needed transitions, such as subsidizing zero-carbon technologies and taxing high- emissions technologies like fossil- fueled cars. Infrastructure design — like reallocating street space for sidewalks or bike lanes — can help people transition to lower- emissions lifestyles.

It is important to note there are many co-benefits that accompany these transformations, too.

Minimizing the number of passenger vehicles on the road, in this example, reduces harmful local air pollution and cuts traffic-related crashes and deaths.

Transformative adaptation measures, too, are critical for securing a more prosperous future.

The IPCC emphasizes the importance of ensuring that adaptation measures drive systemic change, cut across sectors and are distributed equitably across at-risk regions.

The good news is that there are oftentimes strong synergies between transformational mitigation and adaptation.

For example, in the global food system, climate-smart agriculture practices like shifting to agroforestry can improve resilience to climate impacts, while simultaneously advancing mitigation.



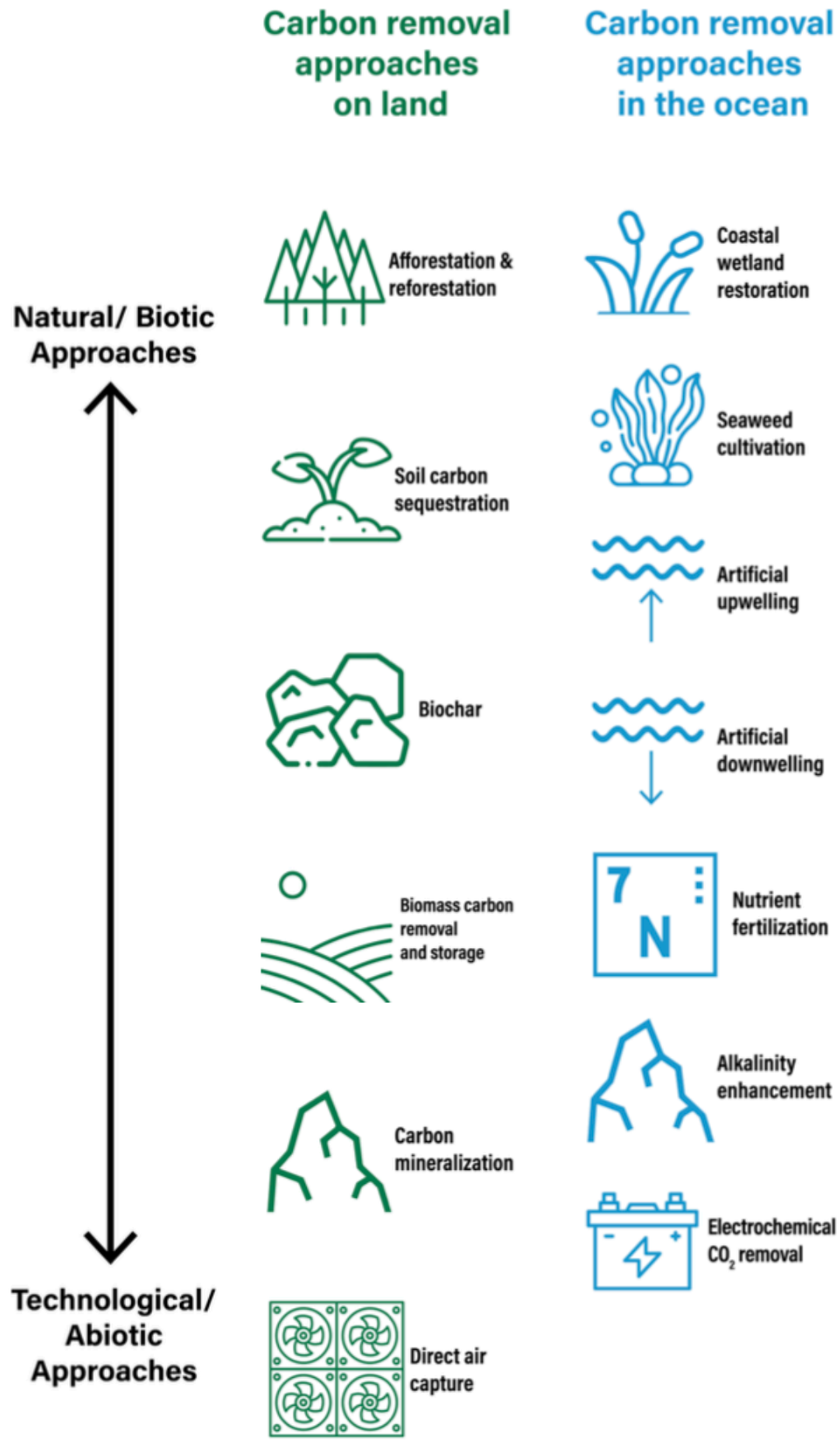
8. Carbon removal is now essential to limit global temperature rise to 1.5 degrees C.

Deep decarbonization across all systems while building resilience won't be enough to achieve global climate goals, though.

The IPCC finds that all pathways that limit warming to 1.5 degrees C (2.7 degrees F) — with no or limited overshoot — depend on some quantity of carbon removal.

These approaches encompass both natural solutions, such as sequestering and storing carbon in trees and soil, as well as more nascent technologies that pull carbon dioxide directly from the air.

The amount of carbon removal required depends on how quickly we reduce GHG emissions across other systems and the extent to which climate targets are overshot, with estimates ranging from between 5 GtCO2 to 16 GtCO2 per year needed by mid-century.



All carbon removal approaches have merits and drawbacks.

Reforestation, for instance, represents a readily available, relatively low-cost strategy that, when implemented appropriately, can deliver a wide range of benefits to communities.

Yet the carbon stored within these ecosystems is also vulnerable to disturbances like wildfires, which may increase in frequency and severity with additional warming.

And, while technologies like bioenergy with carbon capture and storage (BECCS) may offer a more permanent solution, such approaches also risk displacing croplands, and in doing so, threatening food security.

Responsibly researching, developing and deploying emerging carbon removal technologies, alongside existing natural approaches, will therefore require careful understanding of each solution's unique benefits, costs and risks.

9. Climate finance for both mitigation and adaptation must increase dramatically this decade.

The IPCC finds that public and private finance flows for fossil fuels today far surpass those directed toward climate mitigation and adaptation.

Thus, while annual public and private climate !nance has risen by upwards of 60% since the IPCC's Fifth Assessment Report, much more is still required to achieve global climate change goals.

For instance, climate finance will need to increase between 3 and 6 times by 2030 to achieve mitigation goals, alone.

This gap is widest in developing countries, particularly those already struggling with debt, poor credit ratings and economic burdens from the COVID-19 pandemic.

Recent mitigation investments, for example, need to increase by at least sixfold in Southeast Asia and developing countries in the Pacific, fivefold in Africa and fourteen-fold in the Middle East by 2030 to hold warming below 2 degrees C (3.6 degrees F).

And across sectors, this shortfall is most pronounced for agriculture, forestry and other land use, where recent financial flows are 10 to 31 times below what is required to achieve the Paris Agreement's goals.

Finance for adaptation, as well as loss and damage, will also need to rise dramatically.

Developing countries, for example, will need $127 billion per year by 2030 and $295 billion per year by 2050.

While the report does not assess countries' needs for finance to avert, minimize and address losses and damages, recent estimates suggest that they will be substantial in the coming decades.

Current funds for both fall well below estimated needs, with the highest estimates of adaptation finance totaling under $50 billion per year.



Young mangroves ready for planting in Watamu, Kenya. Ecosystem-based adaptation measures like restoration can be low-cost ways to help communities adapt and protect biodiversity. Photo by MariusLtu/iStock.

10. Climate change — as well as our collective efforts to adapt to and mitigate it — will exacerbate inequity should we fail to ensure a just transition.

Households with incomes in the top 10%, including a relatively large share in developed countries, emit upwards of 45% of the world's GHGs, while those families earning in the bottom 50% account for 15% at most.

Yet the effects of climate change already — and will continue to — hit poorer, historically marginalized communities the hardest.

Today, between 3.3 billion and 3.6 billion people live in countries that are highly vulnerable to climate impacts, with global hotspots concentrated in the Arctic, Central and South America, Small Island Developing states, South Asia and much of sub-Saharan Africa.

Across many countries in these regions, conflict, existing inequalities and development challenges (e.g., poverty and limited access to basic services like clean water) not only heighten sensitivity to climate hazards, but also limit communities' capacity to adapt.

Mortality from storms, floods and droughts, for instance, was 15 times higher in countries with high vulnerability to climate change than in those with very low vulnerability from 2010 to 2020.

At the same time, efforts to mitigate climate change also risk disruptive changes and exacerbating inequity.

Retiring coal-fired power plants, for instance, may displace workers, harm local economies and reconfigure the social fabric of communities, while inappropriately implemented efforts to halt deforestation could heighten poverty and intensify food insecurity.

And certain climate policies, such as carbon taxes that raise the cost of emissions-intensive goods like gasoline, can also prove to be regressive, absent of efforts to recycle the revenues raised from these taxes back into programs that benefit low-income communities.

Fortunately, the IPCC identifies a range of measures that can support a just transition and help ensure that no one is left behind as the world moves toward a net-zero- emissions, climate-resilient future.

Reconfiguring social protection programs (e.g., cash transfers, public works programs and social safety nets) to include adaptation, for example, can reduce communities' vulnerability to a wide range of future climate impacts, while strengthening justice and equity.

Such programs are particularly effective when paired with efforts to expand access to infrastructure and basic services.

Similarly, policymakers can design mitigation strategies to better distribute the costs and benefits of reducing GHG emissions.

Governments can pair efforts to phase out coal-fired electricity generation, for instance, with subsidized job retraining programs that support workers in developing the skills needed to secure new, high-quality jobs.

Or, in another example, officials can couple policy interventions dedicated to expanding access to public transit with interventions to improve access to nearby, affordable housing.

Across both mitigation and adaptation measures, inclusive, transparent and participatory decision-making processes will play a central role in ensuring a just transition.

More specifically, these forums can help cultivate public trust, deepen public support for transformative climate action and avoid unintended consequences.

Looking Ahead

The IPCC's report makes clear that risks of inaction on climate are immense and the way ahead requires change at a scale not seen before.

However, this report also serves as a reminder that we have never had more information about the gravity of the climate emergency and its cascading impacts — or about what needs to be done to reduce intensifying risks.

Limiting global temperature rise to 1.5 degrees C (2.7 degrees F) is still possible, but only if we act immediately.

As the IPCC makes clear, the world needs to peak GHG emissions before 2025 at the very latest, nearly halve GHG emissions by 2030 and reach net-zero CO_2 emissions around mid-century, while also ensuring a just and equitable transition.

We'll also need an all-hands-on-deck approach to guarantee that communities experiencing increasingly harmful impacts of the climate crisis have the resources they need to adapt to this new world.

Governments, the private sector, civil society and individuals must all step up to keep the future we desire in sight.

A narrow window of opportunity is still open, but there's not one second to waste.

WHAT WOULD IT TAKE TO FIX GLOBAL WARMING?

The global warming problem isn't going to be solved tomorrow, next week, or next year: we're in this one for the long haul, and there clearly isn't any single solution.

A truly multifaceted approach would include governments agreeing to and enforcing targets, innovators developing low-carbon energy sources and improving efficiency, and individuals and organizations doing their best to reduce their own carbon footprints (and motivating others to do so).

Let's take a look at the problem—and what we need to do to solve it—as a whole.

Looking at the global-scale challenge before us, the most obvious worry is the sheer momentum that needs to be overcome.

We've already added a great deal of greenhouse gas to the atmosphere, and as of this writing we're still adding at least as much each year as the year before.

Even if we can start to reduce global emissions in a sustained way, there's enough inertia in the physical drivers of climate to keep us rolling toward an ever-warmer future for many years to come.

In short, then, if we're to reduce the chance of the most serious long-term climate trouble, we need to take real action on multiple levels—and do it sooner rather than later.

Understanding the goals

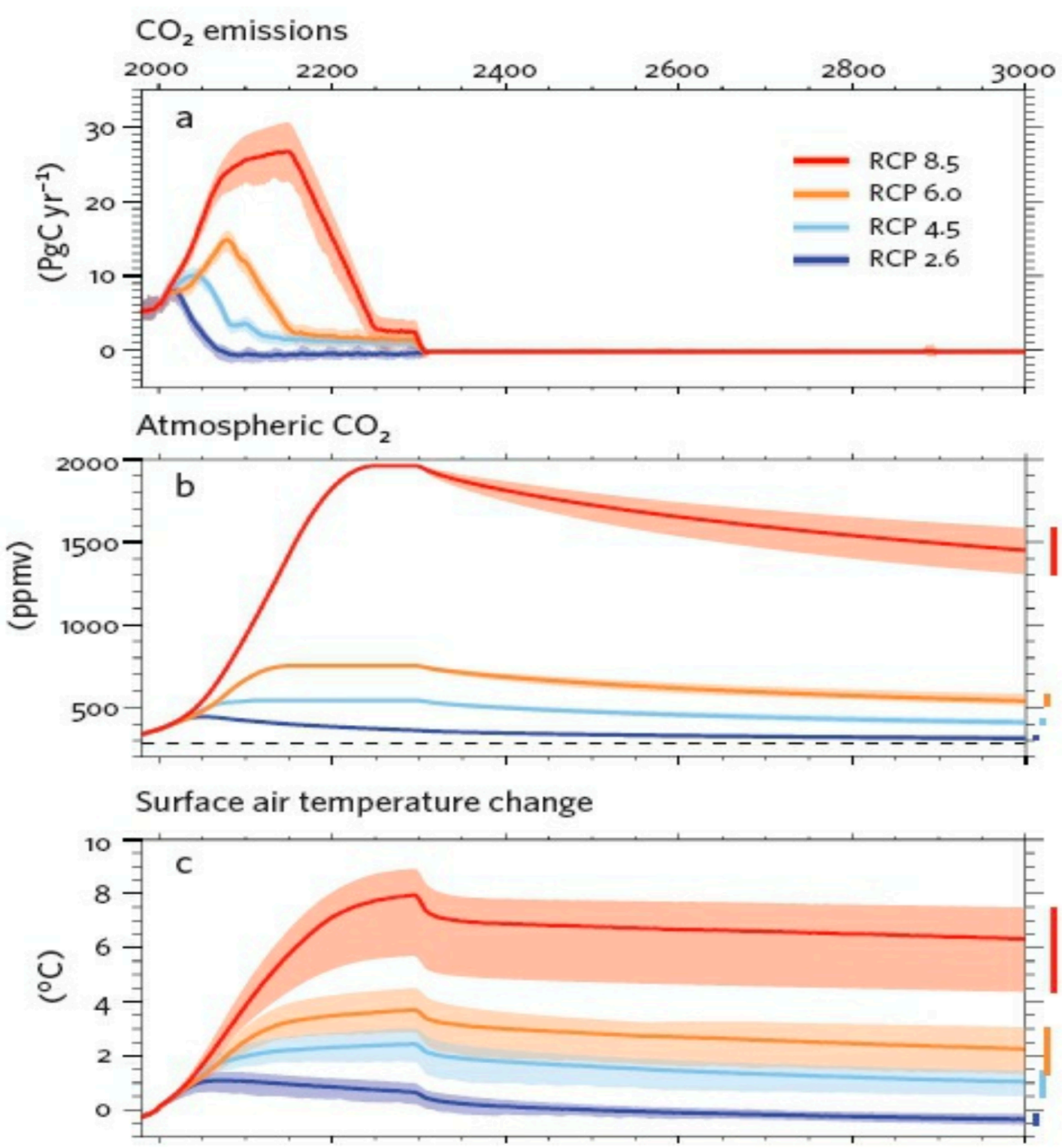
In terms of reducing emissions to mitigate climate change, at least three types of goals are commonly discussed by policymakers and activists working in the field of climate change.

- **Stabilizing emissions.** Making sure that each year we emit no more than the year before, and ideally less.
- **Stabilizing concentrations.** Reducing emissions enough so that the amount of greenhouse gas in the atmosphere levels off at a target and stays there (or falls back down).
- **Stabilizing temperature.** Keeping the atmosphere from warming beyond a certain point.

Obviously, these three types of goals overlap.

Our emissions build up in the atmosphere, changing the concentration of greenhouse gases, which in turn alters the temperature.

The schematic shows how these three quantities evolve along the four pathways outlined in the most recent IPCC report.



Earth system models of intermediate complexity (EMICs) are detailed enough to capture important processes, yet simplified enough so that very long-term simulations are feasible. This graphic shows how EMIC runs carried out in support of the 2013 IPCC assessment depict three key variables over the next thousand years—carbon dioxide emissions (top), CO_2 concentrations (middle), and global surface air temperature (bottom)—for each of the four IPCC emission pathways, or RCPs. Temperature change is relative to the period 1986–2005. To determine the total temperature change since preindustrial times (1850–1900), add roughly 0.6°C (1.1°F). See p. 16 for RCP details. Shaded bands show the range for each model ensemble, with each solid line denoting an ensemble mean. The abrupt temperature drop in 2300 is a result of eliminating all CO_2 emissions and non-CO_2 forcing factors. (IPCC)

Packed inside this complex-looking graphic are a few distinct messages.

First off, you can see that simply leveling off emissions won't be enough to stabilize concentration and therefore temperature, because elevated levels of carbon dioxide persist in the atmosphere for many years, even centuries.

After all, if water's flowing into a bathtub faster than it can drain out, you need to *reduce* the flow—not just keep it constant.

And even though stabilizing global emissions would be an enormous short-term achievement—one that actually occurred for several years in the mid- 2010s—we need to go much further and make significant *cuts* in emissions (the "downhill" side of the four emission curves shown in the top graphic) in order to keep the world from getting significantly hotter.

As we'll see later, some countries have already managed to reduce their emissions, but globally, there's a long way to go.

Moreover, there's the **time-lag** factor to consider.

If warmer temperatures bring lasting changes to the Earth system, such as the loss of major ice sheets or the release of trapped methane, then substantial climate change could be locked in for centuries to come.

As the graphic shows, even if we do manage to make significant emissions cuts, it will take decades for the concentration to begin leveling off, and more than a century for the temperature to stop rising —that is, unless we can implement major year-over-year reductions in global emissions starting in the 2020s and continuing thereafter (the RCP2.6 scenario).

To help simplify this picture, many researchers and policymakers have been focusing on the **cumulative carbon** that's put into the atmosphere, rather than the year-by-year emissions.

There's quite a wide range of estimates on how much total carbon we can afford to emit, but surprisingly, the ultimate temperature jump doesn't appear to hinge on the timing of emissions so much as the total amount.

Thus, no matter what temperature rise we're prepared to risk, the longer we wait to start cutting emissions, the more stringent the later cuts will have to be.

In its 2013 assessment, the IPCC considered this type of global carbon budgeting in detail.

The panel tallied how much CO_2 we could burn and still keep the temperature rise at no more than 2°C over preindustrial levels (using the average temperature in the period 1861–1880 as the preindustrial benchmark).

If we agree that a two-out-of-three chance (66%) of staying within the 2°C margin is an acceptable margin of risk, then humanity's total carbon emissions from CO_2— past, present, and future—must stay below about 1000 metric gigatons (and that's not even considering the added impact of other greenhouse gases, such as methane).

However, as of 2017, we've already emitted roughly 575 metric gigatons of carbon in the form of CO_2, which means we've used up more than half of our long-term budget.

To retain that desired two-out-of-three chance of not exceeding the 2°C target, we'd need to keep our entire future emissions less than our total past ones.

Selecting a target

The IPCC's consideration of the 2°C benchmark raises a critical question: how much global warming would be truly dangerous?

There's no single bright line that separates mild effects from serious threats, but researchers have done their best to identify points beyond which particular risks become more likely.

Many studies suggest that a rise of 3°C (5.4°F) relative to preindustrial levels would be well past the edge of the comfort zone.

It may be enough to trigger unstoppable melting of the Greenland ice sheet, for example, or a net drop in world food production.

Any further warming could jeopardize large parts of the Antarctic ice sheet and cause other dire consequences.

With these and many other factors in mind, the most commonly cited temperature target for climate stabilization is the familiar **2°C (3.6°F** benchmark above preindustrial levels, or around 1°C (1.8°F) above the global temperature of the mid-2010s.

This value, the "do not exceed" goal of the Paris Agreement of 2015, was adopted by the European Union as far back as 1996.

It's also shared by many researchers and activist groups.

As noted by one researcher in 2013,

"Despite the many uncertainties inherent in it, 2°C has been able to prevail as the global temperature threshold. It functions as the central point of reference in the climate debate, and as the one concrete objective on which key actors from policy, media, and research have been able to reach at least interim agreement."

Some parties have long called for more ambitious goals.

At the Copenhagen meeting in 2009, a group of more than 40 small island nations and like-minded activists pushed for a target of **1.5°C (2.7°F).**

The idea was to minimize the risk that rising sea levels could swamp low-lying countries.

The Paris Agreement included the 2.0°C target, but it also urged that the world's nations strive to keep the global temperature rise below 1.5°C.

Even more stringent was the goal proposed by the grassroots effort **350.org**, which has organized more than 10,000 demonstrations and events since 2009.

The group's name comes from its central goal of bringing global CO_2 concentrations down from their current values of around 410 ppm to **350 ppm**, which is deemed by the group (and by NASA) to be the value consistent with a reasonable shot of keeping the eventual global temperature rise to 1.5°C.

Even if it's beyond the reach of current technology and political will, the 350-ppm goal serves as an icon of urgency and the need for action, as well as a reminder that we're already on a path toward much higher concentrations.

2°C—or beyond?

Given physical and political realities, the world may not be able to avoid 1.5°C at this point.

After all, as we sawearlier, Earth has warmed roughly 1°C thus far, and some research has indicated there's an additional 0.5°C (0.9°F) to come as the planet adjusts to emissions already in the atmosphere.

Together, these bring us right up to the 1.5°C threshold.

The implications of going past 1.5°C to 2.0°C were brought home in an IPCC special report, produced in 2018, that assessed a growing body of research showing that impacts were liable to be substantially greater for 2.0°C of warming as opposed to 1.5°C.

Even a brief "overshoot" of 1.5°C—where global temperature nudges past that threshold for a period of years before dropping back—could have long-lasting impacts, the report warned.

At least in theory, we might still have a chance of staying below 2°C, but the odds of reaching that goal are diminishing quickly.

Global emissions have soared more than 60% since the **UN Framework Convention on Climate Change** (**UNFCCC** process began in 1990).

Although emissions held more or less constant while the world's economy grew robustly from 2014 through 2016—a very encouraging sign—the apparent trend was nicked all too soon, as emissions rose more than 1% in 2017.

Even as the Paris Agreement raised hopes worldwide, it was evident that the accord's starting-point pledges were inadequate to meet even the 2.0°C target.

At the same time, abandoning the 1.5°C target would not only signal acceptance of the greater risks of a 2.0°C warming, but it could also lead to cynicism and fatalism about the global effort to keep greenhouse gases in check.

This puts even more emphasis on the need for national, regional, and local initiatives—plus action by companies and individuals—to help stem the tide of emission growth.

All this effort could be for naught unless the act of burning carbon carries a price tag that's valid in some form or fashion around the entire world.

Otherwise, there's a major risk of what's been dubbed **carbon leakage.**

Should one nation reduce its emissions voluntarily, the supply of unburned fuel will rise and its price on the open market will go down, all else being equal.

In turn, that would open the door for some other entity to purchase and burn the fuel that was conserved in the name of climate.

A similarly thorny problem is **rebound effects**, which occur when the benefits of increased efficiency are directed into activities that burn more energy.

Paradoxically, the result can sometimes be *more* carbon emission rather than less.

For example, a family might decide to save money on utilities by insulating their home, but then spend that newfound cash on an extra plane trip each year—thus expanding their carbon footprint even more than they'd reduced it by sealing their drafty house.

This isn't to denigrate the importance of personal as well as national emission reductions, both of which are absolutely crucial in any effort to keep our planet from overheating.

It’s simply to recognize that our global economic system is hard-wired to consume whatever energy is available on the free market.

If the cost to our global commons isn’t somehow factored in, then achieving major emissions cuts will be a truly daunting effort.

The true scope of this challenge was highlighted in 2012 through a hugely popular *Rolling Stone* article by Bill McKibben entitled “Do the Math,” accompanied by a standing-room-only lecture tour.

McKibben pointed out that, according to the Carbon Tracker Initiative, the world’s proven reserves of fossil fuel in 2011—the oil, gas, and coal held by the planet’s private and state-owned energy firms—represent a staggering 2795 billion metric tons of carbon.

That’s more than six times the total that can be burned from this point onward while still retaining a two-thirds chance of meeting the 2°C target, on the basis of the IPCC calculations noted above.

It’s difficult to imagine the world’s big energy firms, which are among the largest and most profitable companies on the planet, voluntarily writing off more than 80% of their holdings (trillions of dollars’ worth) and leaving them in the ground.

Yet if even half of those proven reserves are burned, the chance of avoiding what the IPCC terms "dangerous interference with the climate system" will become slender indeed.

Overall, this looks like quite the discouraging picture.

However, we have little choice but to face the predicament head on.

That includes working in a variety of ways to ensure that any eventual multifaceted solution can benefit as much as possible from progress achieved on many fronts.

Policy experts are looking at various ways to approach the carbon problem from an international standpoint, and plenty of new technologies are on tap to help reduce emissions substantially wherever there's an incentive for doing so.

When one considers greenhouse gases other than CO_2, the picture brightens just a bit.

Most of those gases are less prevalent and shorter-lived than CO_2, but far more powerful.

Together, they account for more than 30% of the global warming impact from gases we're adding to the atmosphere each year.

The good news is that some of them could be reduced more easily and affordably than CO_2, and political agreements that can smooth the way are already in place, such as the Montreal Protocol and regional air pollution laws.

In the EU, for instance, diesel-powered cars and light trucks are allowed to emit only 16% as much nitrogen oxide as they were permitted to belch as recently as 2004.

There is some hope among policy experts and climate scientists that a two-step approach—cutting the emission rates of non-CO_2 greenhouse gases right away, plus reducing CO_2 over a somewhat longer time frame—might prove fruitful.

New approaches to keeping carbon in Earth's ecosystem, including stronger **deforestation** limits, could also have a big impact relatively quickly.

Whether any particular temperature, emission, or concentration goal can be met depends on other factors as well, such as how fast **new technology** is developed and adopted, how the demand for energy evolves, and how seriously we take **energy efficiency**, which in turn depends partly on the political will to prioritize climate change.

Another important factor is the state of the world **economy.**

Global emissions of CO_2 from fossil fuels actually fell more than 4% between 1980 and 1983, a period of high oil prices and widespread recession, and they dropped by about 2% in 1992 and 1999, when the economies of eastern Europe and Russia were struggling.

Although a major global recession might produce similar drops again, the years 2014–2016 proved that the world's economy can now thrive even with emissions held constant.

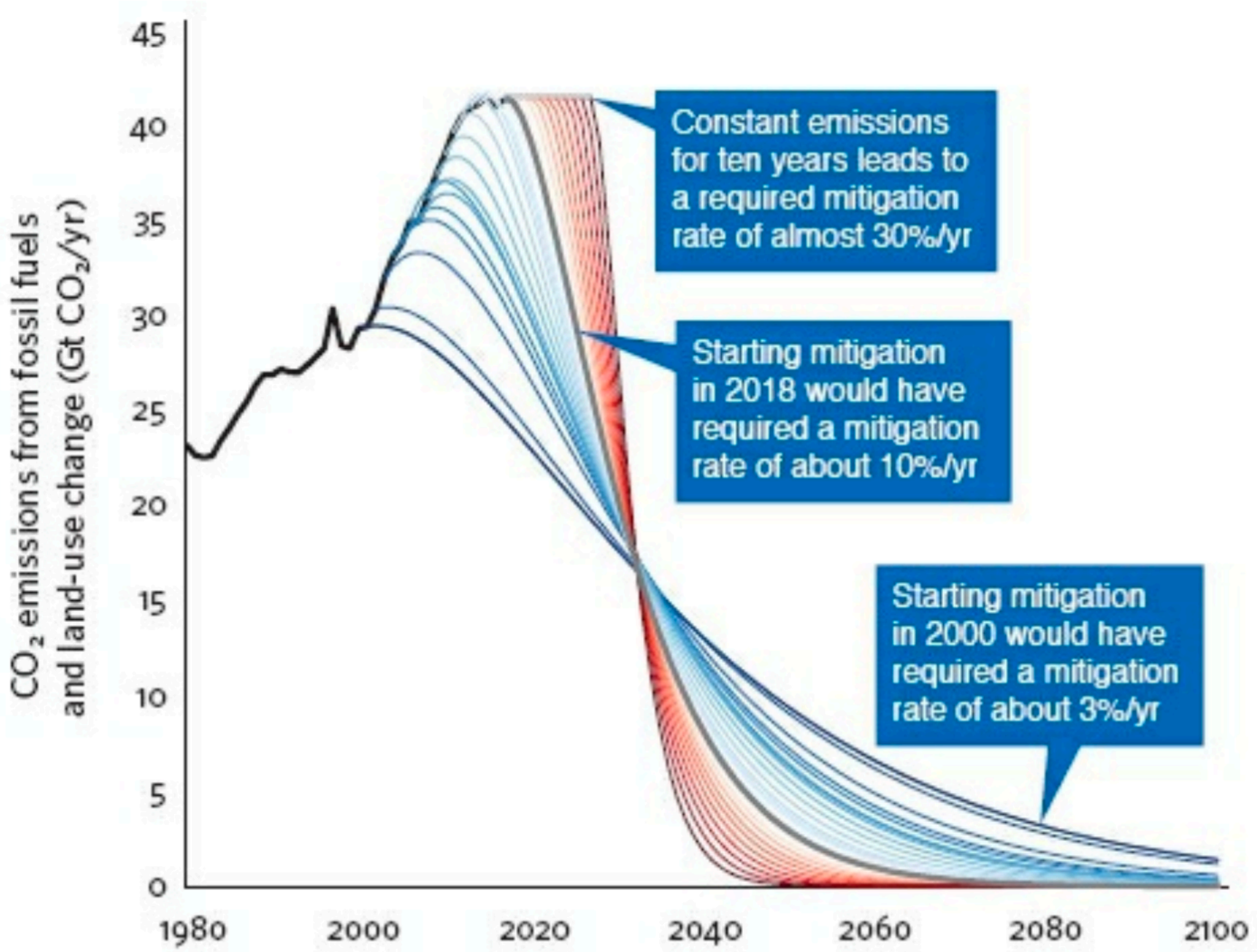
Naturally, all these factors are interrelated in complex ways.

For example, high oil prices can simultaneously dampen economic growth, encourage efficiency, and stimulate investment in alternative energy sources—all of which are likely to reduce emissions.

But if the economy suffers too much, governments may feel pressure from voters to prioritize short-term growth over long-term environmental issues.

Despite all these complications, it seems at least theoretically possible that we could manage long-term net global emission cuts of a few percent within a decade or two, assuming there are incentives for reducing carbon use that carry weight in the global marketplace.

And, though implementing such deep cuts would be liable to cause short-term fiscal pain (perhaps significant in some quarters), the long-term economic and environmental gains from energy efficiency and renewable energy could be enormous and widespread.



If we want to maintain a two-thirds chance of keeping the global temperature rise over preindustrial levels at no more than 2°C, the required mitigation rate per year goes up dramatically the longer we wait to get started. (Robbie Andrew/CICERO Center for International Climate Research, http://folk.uio.no/roberan/t/global_mitigation_curves.shtml, after Michael R. Raupach and colleagues, 2014: "Sharing a quota on cumulative carbon emissions," *Nature Climate Change*, **4**, 873–879.)

The wedge strategy

Since almost all human activity contributes to greenhouse gas emissions on some level, the task of reducing global emissions can seem somewhat overwhelming.

What if we thought of it as a series of simultaneous smaller goals?

That's the philosophy behind the "wedge" approach to climate protection.

The idea is to break down the enormous carbon reductions needed by midcentury into more manageable bits, or wedges, each of which can be assigned to a particular technology or strategy.

When it was introduced, the wedge concept triggered widespread interest and excitement, in part because it was claimed that the needed emission cuts through 2050 (later redefined as cuts over the next 50 years) could be handled entirely through existing technologies.

The concept has proven hugely influential, and it's a handy way to compare and contrast various pieces of the overall emission-reduction puzzle.

The wedge concept originates from the projection of historical CO_2 emissions into the future, starting from 2004.

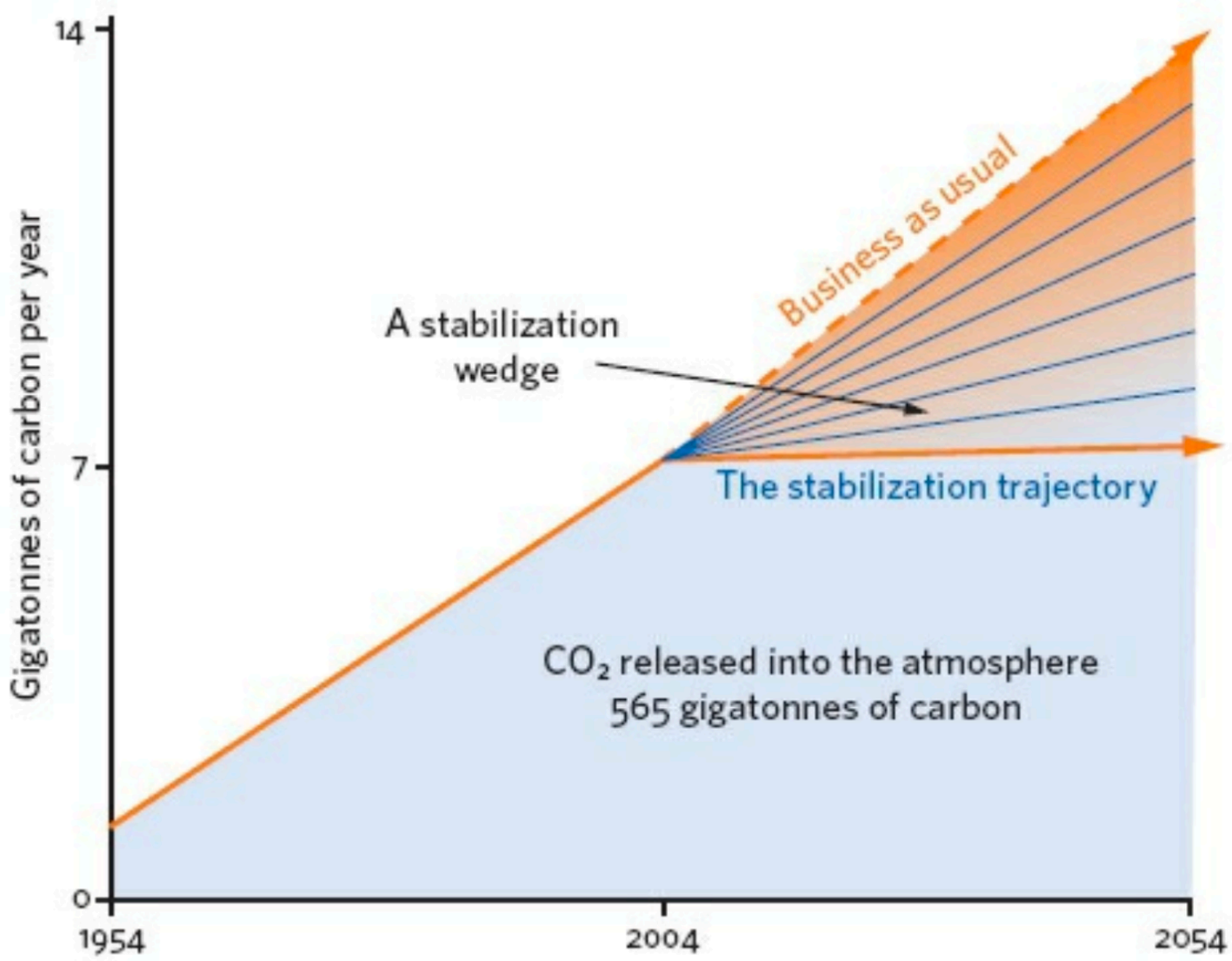
Let's assume it is 2004 right now, and let's also assume that emissions can be instantly stabilized—in other words, the yearly increases in CO_2 all go to zero right now—and remain that way until at least the 2050s.

This is represented by the flat black line on the diagram.

It was argued that this would correspond to an eventual CO_2 concentration of about 500 ppm.

Such a route falls somewhere between the two most optimistic emissions pathways (RCP2.6 and RCP4.5) in the 2013 IPCC report.

This would provide us with a good chance of staying below 3°C by century's end, and at least a shot of remaining below 2°C, depending on how sensitive the climate is to CO_2.



The original wedge-strategy diagram, as conceived by Stephen Pacala and Robert Socolow.

However, if emissions were to increase as they have in most recent years, at more than 1% per year—the red line on the diagram—then by 2054 we'll be adding twice as much CO_2 to the atmosphere as compared to 2004 (see the red "business as usual" line on the diagram).

The result would be a warming far more severe than 2°C, in line with the most pessimistic of the IPCC pathways.

The triangle between the black and red lines shows the difference between the desired path of steady emissions and the dangerous uphill path.

To get rid of the triangle, we'd need to come up with at least seven wedges, each of which would eliminate a billion metric tons (a metric gigaton) of annual carbon emission by 2054.

Further wedges would probably be needed after 2054 to stay below the 2°C target. Originally, the researchers identified 15 examples of potential wedges (see descriptions), each of which was already being implemented on a reasonably large scale somewhere in the world.

That's with the concentration goal now bumped up to 550 ppm, which adds perhaps 0.5°C to the eventual temperature peak and makes reaching the 2.0°C goal far more problematic.

Taking into account the stark realities of our current path as well as the latest science, researchers form University of California, Irvine mapped out a far more ambitious variation in 2013 paper called "Rethinking wedges."

The wedges (expressed in terms of the year 2054 versus 2004):

Doubling vehicle fuel economy and **cutting the distance driven per car in half** (one wedge each). Fuel economy is on the rise in many wealthy nations, including the United States, but auto use is also increasing rapidly in the developed world.

Installing lights and appliances with state-of-the-art energy efficiency in all new and existing residential and commercial buildings, thus reducing emissions by about 25%. There's been incremental progress in recent years, especially with the advent of compact fluorescent (CFL) and LED lighting.

Improving the efficiency of coal-fired power plants from 40% to 60%, plus cutting in half the energy lost when fossil fuels are extracted, processed, and delivered to those plants. Progress has been relatively slow in this area.

A 50-fold expansion in wind energy, replacing coal. As of 2012, the global total of installed wind-power capacity was more than five times the 2004 value. A second wedge could be obtained by adding another four million turbines to generate hydrogen for fuel-cell vehicles.

A 700-fold expansion of photovoltaic (PV) solar energy, again replacing coal. This would require enough panels to blanke every inch of an area the size of New Jersey. However, many of the panels could be mounted on roofs and walls. The total capacity of solar PV panels installed around the world grew roughly 20-fold from 2004 to 2012, with no end in sight to the rapid growth.

A 50-fold expansion in ethanol, displacing petroleum. Even with plants that yield far more energy per acre than the corn an sugar cane used now for ethanol, these fuel crops would still require a region about the size of India. Biofuels production more than tripled from 2004 to 2010, but the trend then flat-lined over the following two years.

A halting of current deforestation, coupled with plantations on non-forested land, together covering a total area about 40% the size of Australia by 2050. There's been major progress on deforestation, due in part to agreements made at the 2009 Copenhagen summit.

Employing conservation tillage on all cropland, in which farmers avoid plowing and thus reduce the amount of CO_2 escaping from tilled soil. It's now used on only a small fraction of cropland globally, but interest in these techniques is spreading

Tripling the amount of energy now generated by nuclear sources, adding about 700 one-gigawatt plants as well as maintaining or replacing all nuclear plants now in use. Nuclear power has been on the decline globally in recent years, with global electricity production from nuclear plants dropping more than 3% from 2009 to 2011. It plummeted another 7% in 2012, largely due to the Fukushima disaster and the power-down of other nuclear plants in Japan.

Quadrupling the use of natural gas in power plants, replacing an equal number of coal-fired plants. Electricity production from natural gas grew by more than 10% from 2004 to 2010, though it dipped slightly in 2011.

Sequestration—capturing carbon emitted by large fossil fuel power plants and storing it underground—with one wedge each coming from standard coal or natural-gas plants, synfuel plants (which generate synthetic fuel from coal), and hydrogen plants that draw on coal or natural gas. Although research continues, even experimental deployments have been few and far between.

They maintained that a 50-year goal was easier to grasp than a century-long one.

Among other things, it's comparable to the length of a career or the lifetime of a coal-based power plant.

As they put it, "Humanity can solve the carbon and climate problem in the first half of this century simply by scaling up what we already know how to do."

How was this 2004-era picture looking as the 2010s wound up?

Apart from occasional downturn-produced dips and the flattening of the mid-2010s, global emissions have tended to rise a bit faster than the 1.5% annual increase assumed in the graph.

In 2011, seven years after the idea originated, they took a fresh look at the wedge strategy and concluded that, given the unabated emissions up to that point, we now needed nine rather than seven wedges.

They proposed a total of 19 wedges, each as large as those in the original 2004 set, that could bring CO_2 emissions close to zero within 50 years.

Doing so, they hasten to add, would require "an integrated and aggressive set of policies and programs to support energy technology innovation across all stages of research, development, demonstration, and commercialization.

No matter the number required, wedges can still simplify and quantify the challenge. But the problem was never easy."

Indeed, while the wedge strategy is a helpful way to assess which directions we might want to go in, it certainly doesn't make the job a cakewalk, as the researchers acknowledged from the start.

"Each of these wedges represents a staggering amount of effort from the public and private sector".

It's obvious that some proposed wedges are far more practical and/or politically palatable than others.

At least one of the original wedges—the ramp-up of ethanol—turned out to be fraught with unforeseen problems.

Even if alternatives are deployed, there's no guarantee that they will supplant traditional fuels.

And if the history of energy development is any guide, the economies of scale tend to push each sector toward a strongly preferred fuel type: gasoline for vehicles, coal for electric power plants, and so on.

This would work against, say, the parallel large-scale growth of electricity generation from wind, solar, and nuclear power.

Moreover, key interest groups— whether it be fossil fuel companies or hard-green activists—are often hesitant to relax long-held positions that end up favoring only a small number of wedges rather than the full smorgasbord.

There's also the obvious fact that we can't stabilize emissions today or tomorrow as if we were flipping a switch.

Even using current technologies, such massive change would take years to implement.

This is why many scenarios constructed by the IPCC and others include a ramp-up phase of continued growth, followed by substantial cuts in emissions—ultimately bringing us well below today's emission levels—in contrast to the flat-line "stabilization trajectory" shown on the above graph.

One other concern: bolstering the clean-energy wedges may not reduce the global use of dirty energy if there aren't economic incentives that transcend national boundaries, as noted above.

Looming in the background, and not considered directly in the wedges, is the inexorable rise in global **population**, which threatens to swamp the emission cuts made in efficiency and technology.

There are hopeful signs of an eventual population peak late this century, but before this occurs, more than 3 billion people could be added to the current total of close to 8 billion.

Because the idea of stabilizing global population has proven to be such a political minefield, it's sometimes omitted from climate change discussions.

However, there's no getting around the fact that more people on the planet will—all else being equal—lead to more greenhouse emissions.

Despite all these various qualms, the wedge concept remains useful as a rough framework for examining the total emissions picture.

Each wedge is technically feasible—it's just a matter of society choosing which ones to emphasize.

There's been real progress on some of the wedges, including wind and solar energy.

Government subsidies or other incentives can further shape this course, and it's possible that different countries will opt for different technologies, thus providing a well-rounded portfolio on the global scale.

The wedges also allow one to work backward and set even shorter-term goals, taking into account that some wedges might grow more rapidly at first while others would take a while to get going.

Now to get more technical.

We will now develop equations and make predictions.

GLOBAL WARMING SCIENCE

A Quantitative Introduction to Climate Change and Its Consequences

1 - OVERVIEW

The first three sections of these notes address the basics:

What are greenhouse gases and how do they lead to warming?

How and why does the atmospheric warming vary in time and space (both as a function of latitude and height)?

And finally, sea level rise.

Beginning with greenhouse gases, the blue line in Figure 1.1a shows the iconic Mauna Loa CO_2 record collected since 1958, preceded by an ice-core-based reconstruction.

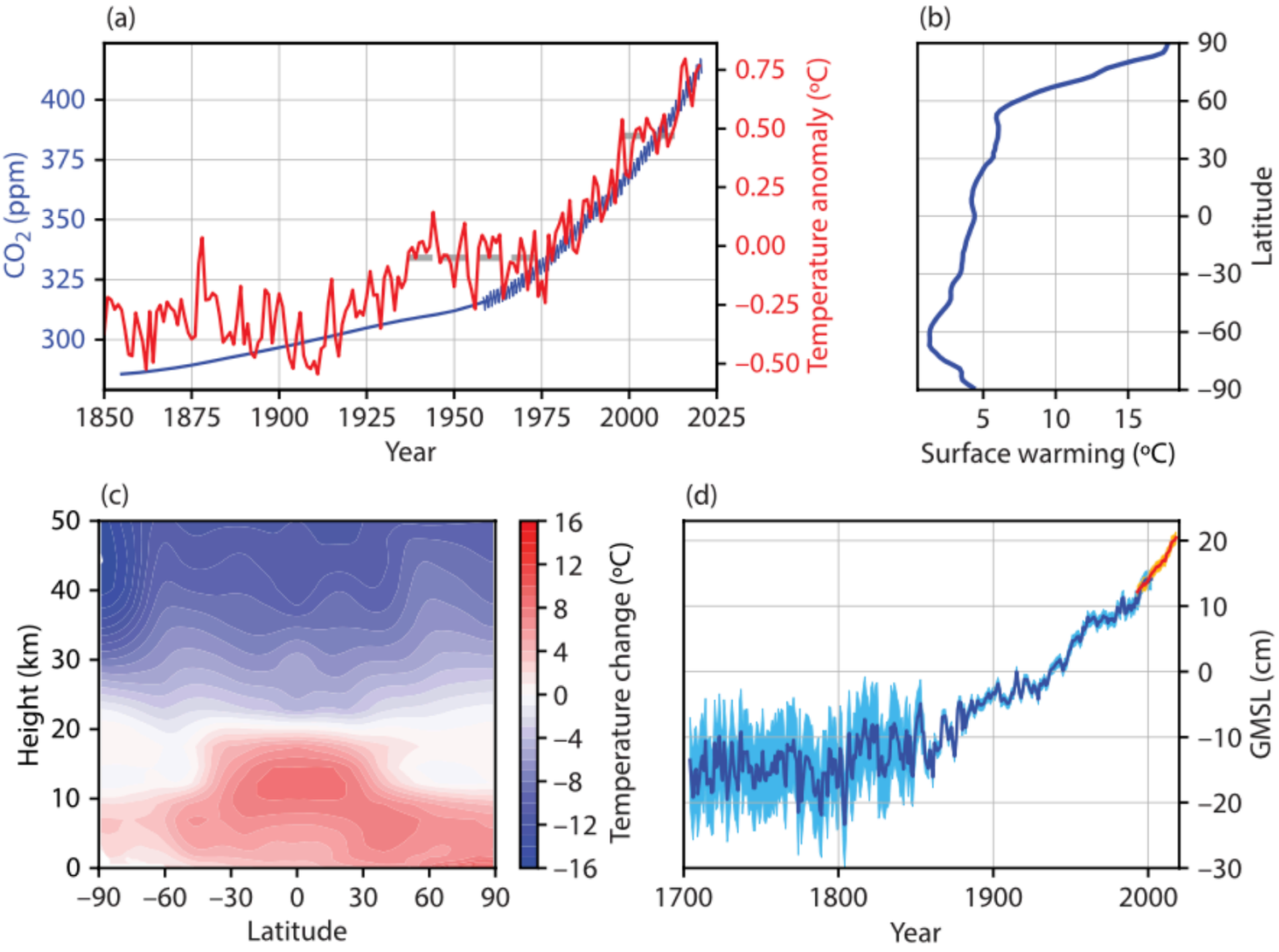


Figure 1.1: Greenhouse effect, warming, and sea level rise.

(a) Atmospheric CO_2 concentration (blue) and global mean surface temperature anomaly (in red, defined as the deviation from the mean over 1961–1990) since 1850. Also shown by gray dashed lines are two "hiatus" periods of *seemingly* reduced rates of warming. (b) The projected surface warming over the 21st century in an RCP8.5 scenario in a climate model, as a function of latitude, showing a very pronounced polar (mostly Arctic) amplification. (c) Projected atmospheric temperature changes over the 21st century as a function of altitude and latitude, showing a tropospheric warming and stratospheric cooling in the RCP8.5 scenario. (d) Estimated global-mean sea level anomaly since 1700 (blue line) and the estimated uncertainty (light-blue shading). In red and orange: the satellite record since 1993.

CO_2 concentration has been at 280 ppm for over 10,000 yr, since the last ice age, and has therefore increased by about 50% so far, at an unprecedented speed.

There is, of course, no doubt that CO_2 is increasing and that the increase is attributable to the anthropogenic burning of fossil fuel.

Once in the atmosphere, we will see that it will take thousands of years for the CO_2 to naturally decline after anthropogenic emissions are eliminated.

Later we will addresses the question of how greenhouse gases trap heat and lead to warming, both on a molecular level and by examining the Earth global energy balance.

These are ideas that have been well understood for a while now.

The warming due to greenhouse gas increase is explored later.

The red line in Figure 1.1a shows that the globally averaged surface temperature anomaly (defined as the deviation from a reference value, in this case the mean over 1961–1990) has warmed so far by over 1 C.

Future climate projections rely on estimates of future greenhouse gas concentrations, referred to as Representative Concentration Pathways (RCPs) and followed by a number indicating the expected enhancement in radiative heating.

We discuss these scenarios in later and note briefly now that RCP8.5 is what one might think of as a business- as-usual scenario in which CO_2 concentration increases to a very high value of over 1000 ppm by year 2100.

While one hopes that such a high future greenhouse gas concentration is not a realistic scenario, it allows us to clearly understand climate change trends and mechanisms that may be more difficult to identify in less severe scenarios.

Figure 1.1b shows that the projected surface warming under RCP8.5 is strongly amplified toward the poles (especially the Arctic), while Figure 1.1c shows that the stratosphere is projected under the same scenario to cool significantly above an altitude of about 20 km, while the troposphere warms.

Polar amplification and stratospheric cooling are already observed today, and we will examine several mechanisms that are responsible for these signals.

We will also see that even at the present level of CO_2, additional warming would have occurred if not for the cool, deep ocean, which takes hundreds of years to warm in response to the enhanced greenhouse forcing.

While CO_2 has increased monotonically, the warming of the global mean surface temperature seems to have paused during 1940–1970 or so and during 1998– 2013, as shown by the horizontal dashed lines in Figure 1.1a.

We will show that such seeming "hiatus" periods in the increase in global mean surface temperature are an expected consequence of anthropogenic warming in the presence of natural climate variability.

One of the most consequential results of ocean warming and of the expected melting of land-based ice is sea level rise, as analyzed in Section 4 and shown in Figure 1.1d.

Global mean sea level has increased by some 30 cm over the past 150 yr, is currently increasing at about 3.5 mm/yr, and is projected to rise by up to a meter by 2100.

Many processes are responsible for sea level rise, from the expansion of warming ocean water to land-based ice melting.

Furthermore, sea level rise is expected to vary from region to region, and we will discuss the many mechanisms involved, from wind and atmospheric pressure changes, to the gravitational effect of melting ice over Greenland and Antarctica, and more.

The timescales of the processes involved vary from a near-instantaneous response (e.g., to changes in atmospheric pressure or ocean currents) to hundreds of years (warming of the deep ocean, melting in Greenland), to many thousands (significant melting of East Antarctica), with uncertainty levels typically higher for processes with longer timescales.

Two more ocean-centered issues are examined after that, beginning with ocean acidification and then the possible collapse of the meridional overturning ocean circulation.

Ocean acidification is referred to as “the other global warming problem.”

The absorption by the ocean of about a quarter of the anthropogenically emitted CO_2 (with another quarter absorbed by the land biosphere) has significantly reduced the warming experienced so far.

Yet Figure 1.2a shows that as a result, ocean pH, the measure of seawater acidity, already decreased from 8.16 to 8.06, which implies a significant increase of 25%(!) in the concentration of H+ ions in the ocean.

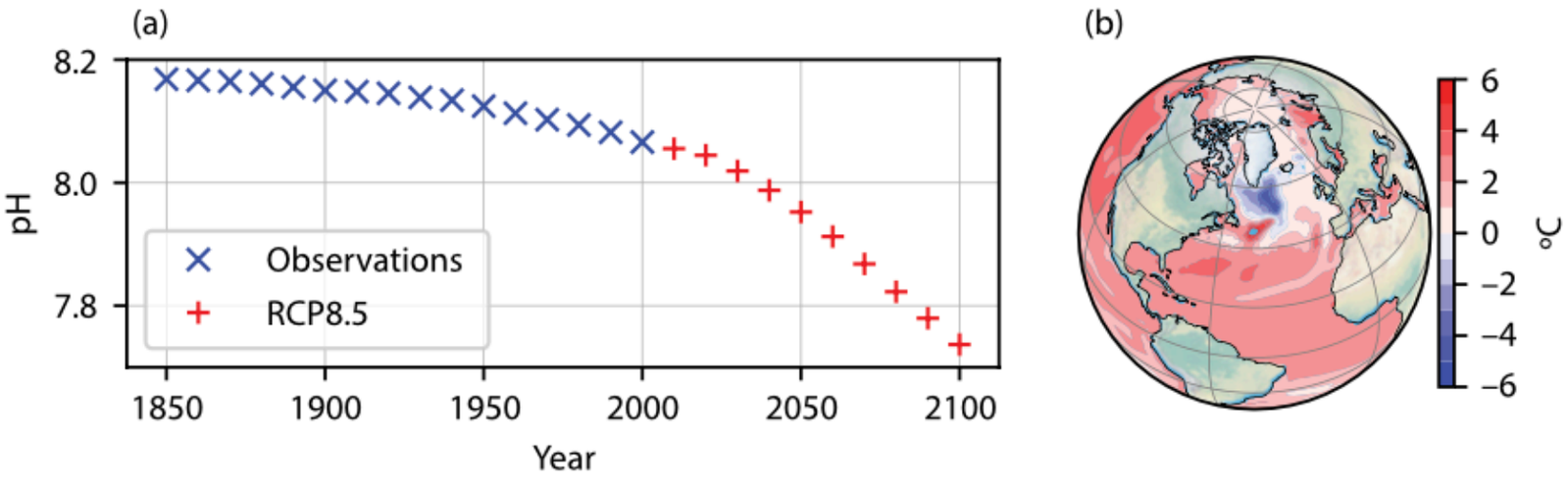


Figure 1.2: Ocean acidification and circulation.

(a) Observed mean surface ocean pH from 1850 to 2000 (blue), followed by projected pH to 2100 in the RCP8.5 scenario (red). (b) Projected sea surface temperature change over the 21st century in the RCP8.5 scenario, showing an overall warming, yet with a local cooling in the northern North Atlantic due to the projected collapse of the ocean overturning circulation.

We will examine the basic carbonate chemistry behind ocean acidification, how it may affect the deposition of calcium carbonate structures by oceanic organisms, and how atmospheric CO_2 can eventually decline once emissions are significantly reduced, on a timescale of thousands of years.

While there is little uncertainty involved in assessing expected ocean pH for a given atmospheric CO_2 level, the response of ocean biology is complex and is still being studied.

Section 6 discusses how the oceanic meridional circulation, which carries heat poleward and contributes to the warmth of the high-latitude North Atlantic, may collapse in a global warming scenario over the next century or so.

The circulation collapse is expected to contribute to a regionally reduced warming and even cooling in the northern North Atlantic, as shown in the model projection of Figure 1.2b, as well as to other disruptions to the current state of the ocean.

We will analyze how a gradual CO_2 increase may lead to an abrupt ocean circulation response, explaining in the process the concept of climate *tipping points*.

2 - GREENHOUSE

We begin with the basics:

What are greenhouse gases, how do they absorb radiation, and why and how does this lead to warming?

Observations clearly show that CO_2 has increased over the past century and a half due to the burning of fossil fuels, at a rate much faster than has occurred naturally in the past and some projections suggest that this rise may not stop or slow down dramatically in the near future.

Figure 2.1 shows the observed CO_2 concentration time series (blue), the—hopefully—unrealistic high-end greenhouse gas concentration scenario known as the Representative Concentration Pathway (RCP) 8.5 used by the Intergovernmental Panel on Climate Change (IPCC), and some more moderate scenarios.

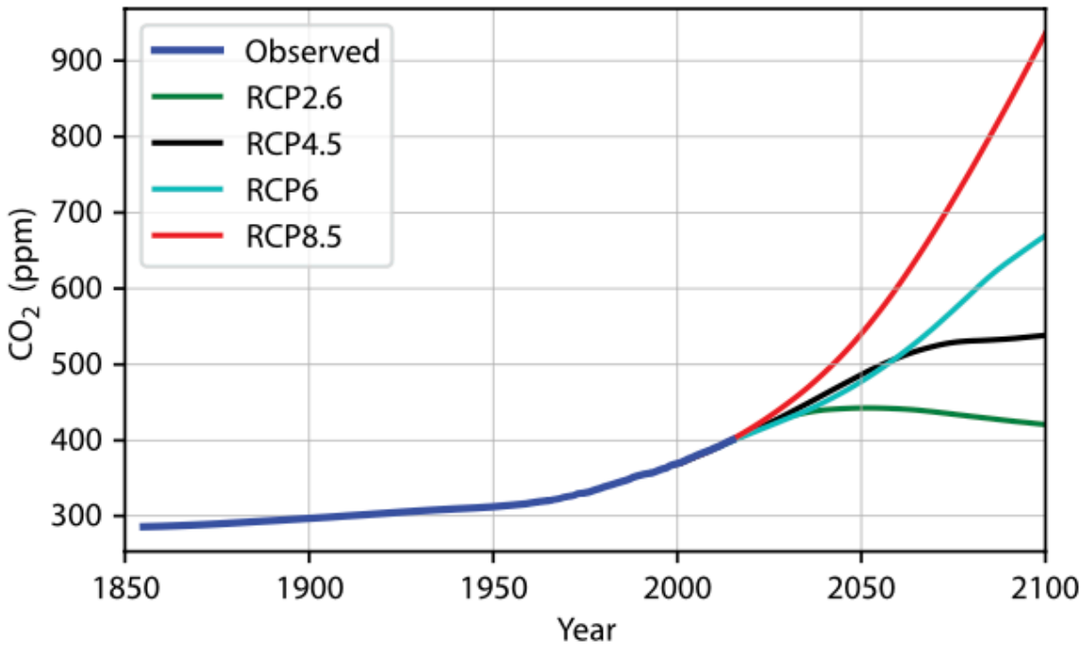


Figure 2.1: CO_2 timeseries.

Annually averaged CO_2 concentration, observed and projected according to different RCP scenarios.

The concentrations of other anthropogenic greenhouse gases have also increased, including methane (CH_4), nitrous oxide (N_2O), CFC12, and CFC11, which further increase the anthropogenic greenhouse effect.

The discussion starts with simple representations of the greenhouse effect, allowing us to understand how it leads to warming.

We then consider how greenhouse gases interact with radiation and the role of the wavelength- dependence of the absorption of electromagnetic radiation (heat) by greenhouse gases.

conclude with a comparison of different greenhouse gases (2.2.5) and with the water vapor feedback that further amplifies the warming effect of anthropogenic greenhouse gases (2.2.6).

2.1 - THE GREENHOUSE EFFECT

We can estimate the globally averaged Earth surface temperature based on the balance of incoming radiation from the Sun with the outgoing heat escaping to outer space.

Consider first what the averaged Earth surface temperature would have been in the absence of the radiative effects of the atmosphere.

The incoming solar radiation at the top of the atmosphere is given by the solar constant,
$S_0 = 1361\ W/m^2$.

The total solar radiation encountered by Earth is therefore the solar constant times the cross-section area of Earth (orange area in Figure 2.2), $S_0 \pi R_E^2$, where R_E is the Earth radius.

Over a day, this radiation is distributed over the entire Earth surface area, $4 \pi R_E^2$, so that the radiation per unit area, averaged over the Earth surface and over a day, is the ratio of the total radiation and the total area, or $S_0/4$.

The fraction of the incoming solar radiation reflected (by ice, clouds, land, and so on), or the *albedo*, is denoted by $\alpha \approx 0.3$.

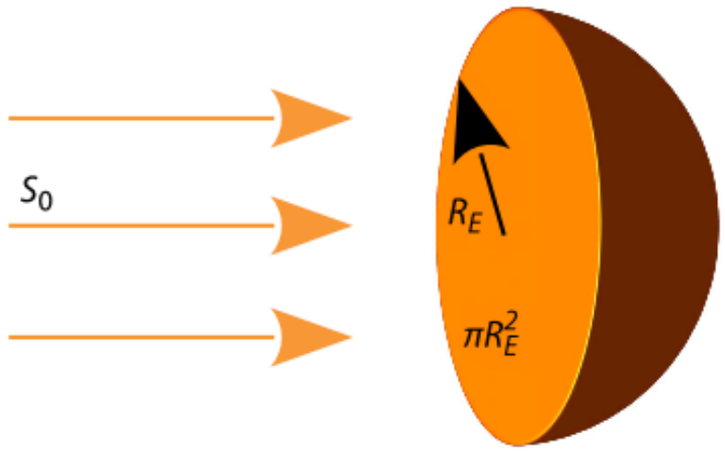


Figure 2.2: Solar forcing.

The total solar radiation distributed over the entire Earth surface is equal to the incoming solar radiation flux per unit area at the top of the atmosphere, (W/m), times the cross-section of the Earth, (m), which is shown as the orange cross-section area, for a total of W. This radiation is then distributed over the entire Earth area during one day due to the Earth rotation.

Both the Earth and the Sun emit radiation based on their corresponding surface temperatures.

This emitted radiation is close to that of a *black body* at these temperatures.

A black body is one that absorbs all incident radiation, and when it is in thermal equilibrium at a temperature *T*, it emits a total radiation per unit area of σT^4 , where $\sigma = 5.670 \times 10^{-8}$ W m^{-2} K^{-4} is the Stefan-Boltzmann constant.

Assume the outgoing radiation from Earth (escaping heat) to be given by the black body radiation formula, with *T* being the globally averaged surface temperature.

The electromagnetic radiation from the Sun has wavelengths of about 0.25–2 micrometers (μm) and is referred to as *shortwave* (SW) radiation, which includes visible light that is characterized by wavelengths of 0.4–0.7 μm.

The thermal radiation emitted by Earth is characterized by wavelengths of roughly 5–35 μm and is therefore referred to as *longwave* (LW) radiation.

Assuming the Earth to be in thermal equilibrium, the incoming shortwave radiation from the Sun that is not reflected must be equal to (i.e., balance) the outgoing longwave radiation (Figure 2.3a),

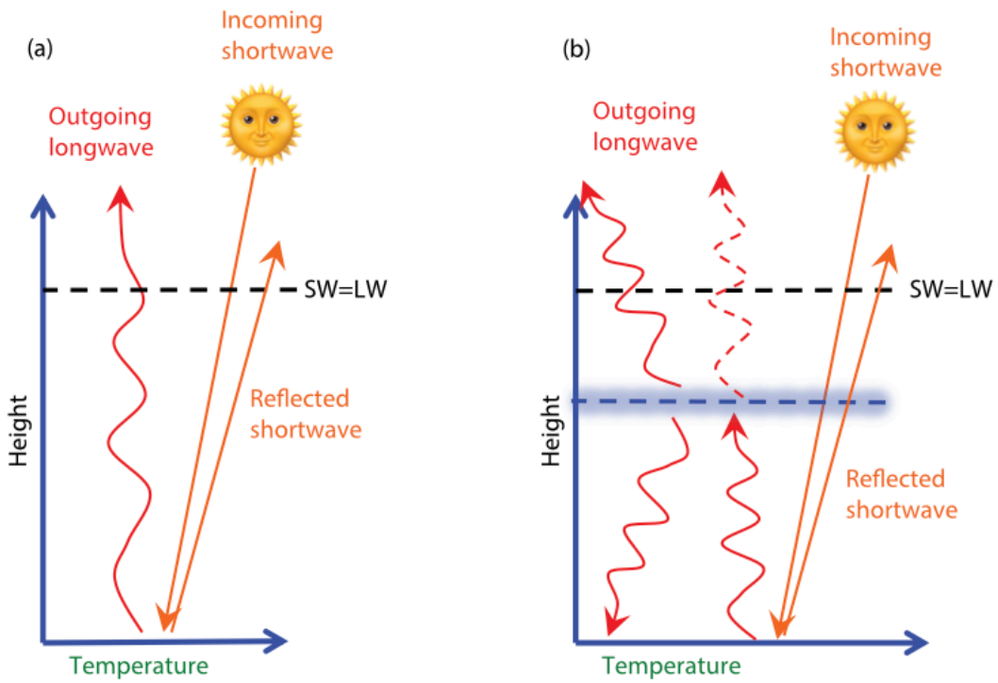


Figure 2.3: Two models of Earth's energy balance.

(a) A one-layer model ignoring the radiative effects of the atmosphere. (b) A two-layer model including the greenhouse effect of the atmosphere. Straight arrows indicate received and reflected SW radiation. Solid wiggly red lines denote LW radiation emitted by the surface and atmosphere; the dashed wiggly line denotes the part of the surface-emitted LW radiation not absorbed by the atmosphere

$$\frac{S_0}{4}(1-\alpha) = \sigma T^4, \qquad (2.1)$$

which gives

$$T = \left(\frac{(S_0/4)(1-\alpha)}{\sigma}\right)^{1/4} = 255\text{K} = -18^\circ\text{C} \equiv T_0. \qquad (2.2)$$

This is too cold, as at such a temperature the Earth surface would have been frozen, and the actual globally average surface temperature of the Earth is about 14 C (287 K), so something is clearly missing.

That missing factor is the greenhouse effect of the atmosphere.

2.1.2 The greenhouse effect: a two-layer model

We now add the greenhouse radiative effect of the atmosphere, whose temperature is denoted by T_a.

To begin, we treat the atmosphere as a single layer and assume that it absorbs heat (longwave radiation) escaping from the surface and then re-emits it both up and down at a rate depending on the atmospheric temperature (Figure 2.3b).

We write separate energy balance equations for the surface and for the atmosphere.

The atmosphere is not a perfect black body and therefore only emits a fraction ϵ of the radiation of a black body with the same temperature.

Similarly, it also absorbs a fraction ϵ of the LW radiation from the surface.

This fraction is referred to as the LW *emissivity*, which is smaller than but close to one.

The emissivity, also equal to the absorptivity (the part of incident radiation absorbed by a surface divided by that absorbed by a black body), is a function of the water vapor and CO_2 concentrations, among other things, and can be set for preindustrial climate to, say, 0.75.

Thus the atmosphere emits LW radiation at a rate of $\epsilon\sigma T_a^4$ W/m² both upward and downward.

As the surface (whose emissivity is closer to one) radiates σT^4 W/m² , the atmosphere absorbs only a fraction ϵ of this radiation, with the rest continuing to outer space (wiggly dashed upward arrow in Figure 2.3b).

Given these assumptions, the energy balances for the surface and atmosphere may be written as

$$\begin{aligned} \frac{S_0}{4}(1-\alpha) + \epsilon\sigma T_a^4 &= \sigma T^4 \\ \epsilon\sigma T^4 &= 2\epsilon\sigma T_a^4. \end{aligned} \quad (2.3)$$

The first equation is the energy balance for the surface, showing the input from the Sun and from the atmosphere on the LHS and the output on the RHS.

The second equation represents the energy balance for the atmosphere, with input LW radiation from the surface on the LHS and output due to the LW radiation from the atmosphere on the RHS.

The factor 2 on the RHS represents the upward and downward LW emission from the upper and lower surfaces of the assumed single-layer atmosphere.

Note that the atmosphere is assumed transparent to the shortwave radiation from the Sun, a reasonable assumption to first order, so that the solar radiation warms the surface, and the radiation from the surface then affects the atmosphere.

Substitute the second equation in the first,

$$\frac{S_0}{4}(1-\alpha)+\frac{1}{2}\epsilon\sigma T^4=\sigma T^4,$$

so that

$$T=\left(\frac{(S_0/4)(1-\alpha)}{\sigma(1-\epsilon/2)}\right)^{1/4}=T_0(1-\epsilon/2)^{-1/4}=284\mathrm{K}=13^{\circ}\mathrm{C}, \qquad (2.4)$$

which is reasonably close to the observed global mean surface temperature.

Had we assumed the atmosphere to be a perfect black body, the corresponding solution for $\epsilon=1$ is $T=T_0 2^{1/4}=303K=30^oC$, which is too warm.

The difference between the energy balances with and without an atmospheric layer taken into account (panel a vs panel b in Figure 2.3, and eqn 2.2 vs eqn 2.4) demonstrates the greenhouse effect of the atmosphere.

It should be noted that water vapor is the main atmospheric greenhouse gas that contributes to the LW emissivity/absorptivity of the atmosphere and accounts for most of the greenhouse effect, although with a critical contribution from CO_2.

Finally, getting to the anthropogenic greenhouse effect, we note that an increased concentration of greenhouse gases increases the LW absorptivity/emissivity of the atmosphere, ϵ, and the solution (eqn 2.4) shows that the surface temperature will increase accordingly.

It turns out, though, that while the explanation of the natural greenhouse effect of the atmosphere using this two-layer model is helpful, this is too much of an oversimplification when it comes to the atmospheric response to the anthropogenic increase in greenhouse gas concentrations.

This picture is accordingly refined in the next subsection.

It is worth noting, perhaps, that an actual greenhouse operates differently from the above atmospheric greenhouse effect.

The glass allows sunlight in, and while it prevents much of the infrared (LW) radiation from the surfaces inside the greenhouse from escaping due to its optical properties, this is not the main mechanism by which it keeps the air in the greenhouse warm.

The glass temperature is not significantly colder than that of the air inside the greenhouse, and thus the LW radiation it emits is not much lower than that of the air in the greenhouse.

Similarly, while glass is a relatively good insulator, the typical thickness of about 3 mm used for constructing greenhouses cannot significantly reduce conductive (diffusive) cooling through the glass.

Instead, the glass keeps the greenhouse warm by preventing air exchanges with the surroundings: surfaces under the glass absorb solar radiation and warm, also warming the interior air that is in contact with these surfaces.

The glass, then, prevents the warm air from rising and being replaced by (or mixed with) cooler air, as would have happened outside the greenhouse, keeping the interior warm.

In the atmosphere, greenhouses gases trap longwave radiation rather than restricting the movement of warm air.

2.1.3 The emission height and lapse rate

The absorption by the atmosphere of longwave radiation emitted by the surface depends on the wavelength of the radiation, as further discussed later.

This differential absorption depends on the concentration of different greenhouse gases in the atmosphere, each of which absorbs at different wavelengths.

It turns out that the atmosphere absorbs nearly all photons of longwave radiation emitted from the surface in the wavelength range corresponding to CO_2 absorption.

Why is it, then, that increasing CO_2 concentration is still expected to lead to further warming?

One of the main reasons is that the temperature declines with height in the lower few kilometers of the atmosphere, rather than the atmosphere being a single layer with a single temperature, as assumed in earlier.

The atmospheric temperature decreases linearly with height in the troposphere at a *lapse rate* of approximately 5–9.8 C per km and then increases in the stratosphere (Figure 3.8a below) due to absorption of SW solar radiation by the ozone layer there.

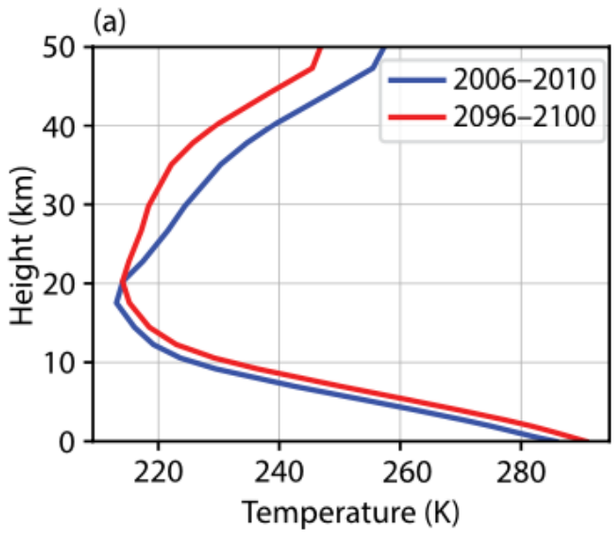


Figure 3.8: Stratospheric cooling.

(a) Mid-latitude (30 N–50 N) zonally averaged temperature profiles for an RCP8.5 projection at the beginning and end of the 21st century.

Most of the radiation emitted from Earth's surface is absorbed by the air above it, which re-emits the radiation both downward and upward to the air above, and so on.

At some level in the atmosphere, the overlying air is thin enough that most of the longwave radiation emitted from that level is able to make it to outer space without being absorbed again.

This is the *emission height*, and the air temperature at this height determines the flux of outgoing longwave radiation from the atmosphere.

Increasing the concentration of greenhouse gases moves the emission height farther up, as some of the radiation from the previous emission height is trapped by the newly added greenhouse gas molecules above this height.

Doubling the CO_2 concentration, for example, raises the emission height by about 150 m.

Given that there is only weak LW absorption above it, the temperature of the emission height at equilibrium, once the climate system adjusts to the changed greenhouse gas concentration, is determined by the balance between incoming shortwave radiation and outgoing longwave radiation from this height.

Figure 2.4 shows a schematic of the temperature profile and emission height (a) before an increase of the CO_2 concentration, (b) immediately after an assumed abrupt increase, and (c) after the atmospheric temperature adjusted.

Before the increase in greenhouse gas concentration (a), the atmosphere emits LW radiation from the emission height to space at a temperature for which the outgoing longwave radiation (OLR) is equal to the incoming shortwave radiation (as in eqn 2.1).

After the CO_2 increase, the emission height moves upward, and because of the lapse rate, the temperature that radiates to outer space is now lower, and therefore so is the OLR.

Thus, the incoming shortwave radiation is greater than the OLR, causing an energy imbalance (b) that leads to a warming of the atmosphere.

This warming causes the temperature profile to adjust (the solid green line moves to the right in Figure 2.4c) until the temperature at the new emission height is high enough to put the OLR in balance with incoming shortwave radiation again.

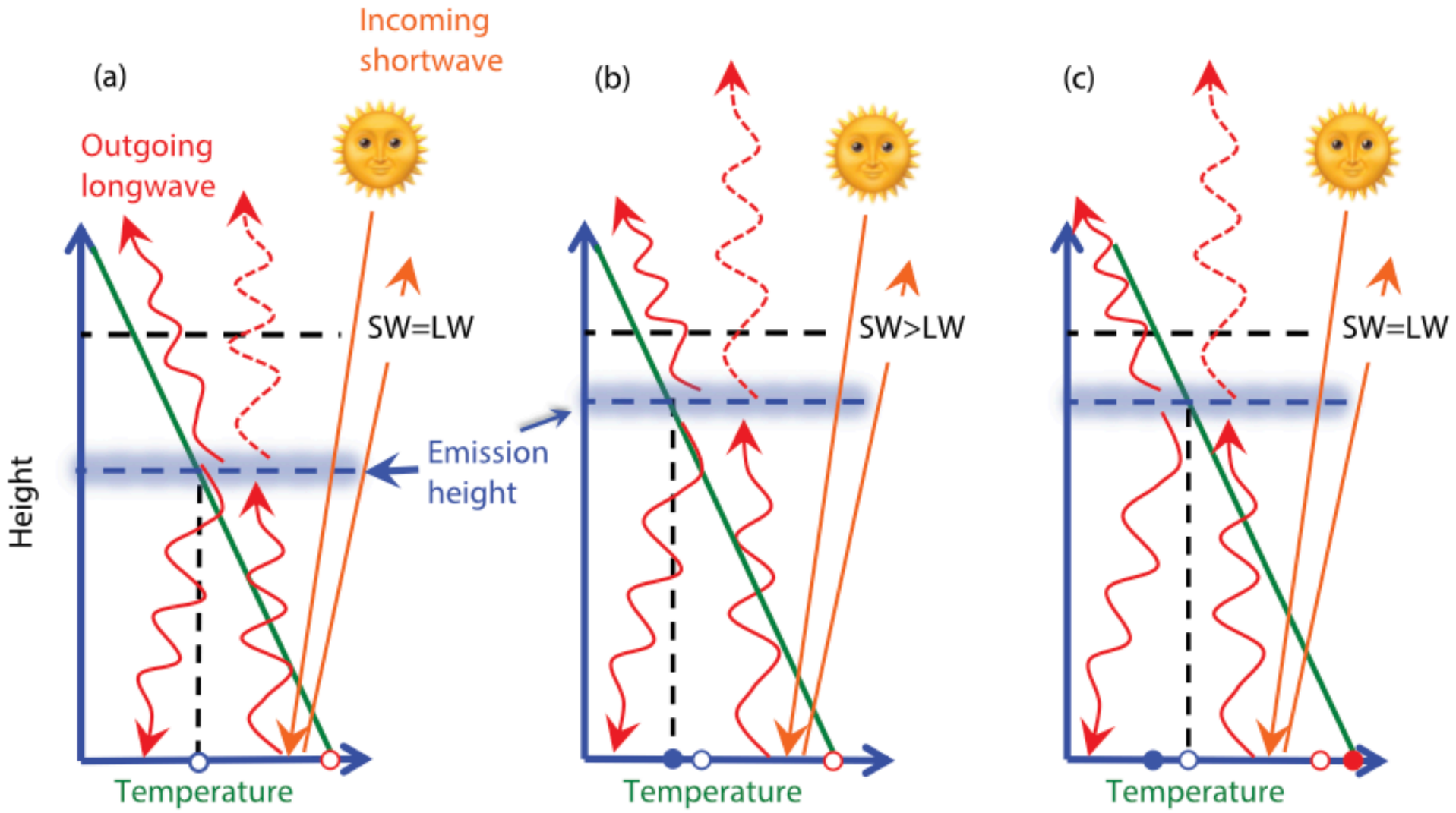


Figure 2.4: Role of the emission height and lapse rate in the greenhouse effect.

Emission height (dashed blue horizontal lines) and vertical temperature profiles (solid green) are shown (a) before a CO_2 increase, (b) immediately after a CO_2 increase but before the atmospheric temperature adjusted, and (c) after the atmospheric temperature adjusts. Red dots at the bottom of each panel denote the surface temperature. Blue dots denote the emission temperature. Before the warming and after the CO_2 increase (panel b), the emission temperature decreases (moving from the empty to the filled blue circle). As a result, the outgoing LW radiation decreases and is smaller than the incoming SW radiation. After the warming (panel c), the new emission temperature at the raised emission level is equal to the original emission temperature at the lower original emission level, denoted by the empty blue circle, thus balancing the incoming SW radiation again.

This increase in the emission height determines the response of the energy balance at the top of the atmosphere to higher CO_2, leading to atmospheric warming.

The surface is also warmed by a smaller, separate effect due to the increase in downward LW radiation at the bottom of the atmosphere, even before taking into account atmospheric warming: the increased atmospheric LW emissivity due to a higher CO_2 concentration implies that more of the downward LW at the surface arrives from lower in the atmosphere, where the temperature is higher; the higher emitting temperature means more LW radiation arriving at the surface and therefore a surface warming.

The key takeaway is that an increase in greenhouse gas concentration leads to an increase in the emission height and therefore to a warming at the surface.

The important factors leading to this warming are that the emission height depends on the CO_2 concentration and that the atmospheric temperature decreases with altitude within the troposphere.

2.2 GREENHOUSE GASES

2.2.1 Wavelength-dependent black-body radiation

While we have been treating the outgoing radiation so far as a single entity, it is made of different wavelengths of electromagnetic radiation.

The Earth emits from the surface approximately like a black body with a temperature of 288 K or so (blue curve in Figure 2.5a), while the Sun radiates as a black body of a much higher temperature (red curve), where both curves follow Planck's Law,

$$B(\lambda, T) = \frac{2hc^2}{\lambda^5} \frac{1}{e^{\frac{hc}{\lambda k_B T}} - 1}. \qquad (2.5)$$

Here, is the spectral radiance emitted by the black body at a wavelength λ(m), at thermal equilibrium at temperature T (K).

The radiance is the radiation in watts (W), per unit area (m^2) of the emitting surface, per unit solid angle (steradian), per unit wavelength (m): together, $W \cdot m^{-2}$ $ster^{-1}$ m^{-1} .

Also, $k_B = 1.38064852 \times 10^{-23}$ m^2 kg s^{-2} K^{-1} is the Boltzmann constant, $h = 6.62607004 \times 10^{-34}$ m^2 kg s^{-1} the Planck constant, and $c = 3 \times 10^8$ m s^{-1} the speed of light.

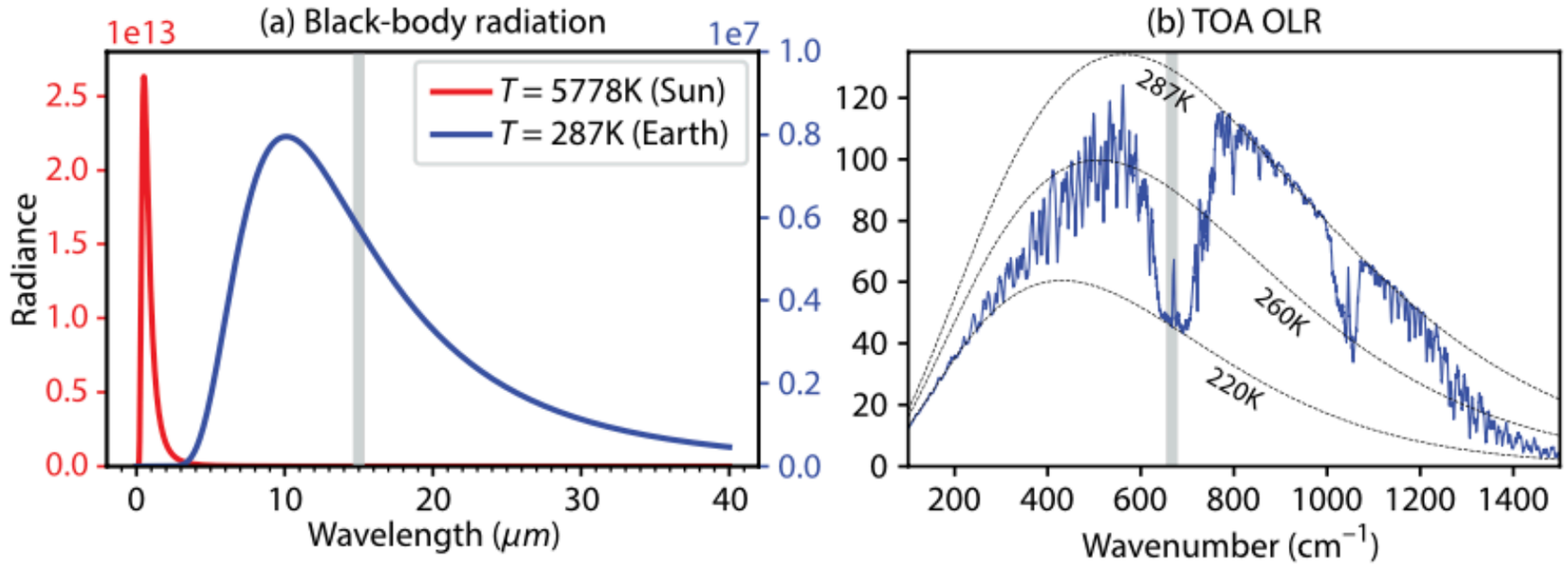
The integral of the spectral radiance over all wavelengths and angles yields the total black-body radiation, σT^4 (W m^{-2}), used previously, and σ can be expressed in terms of the constants appearing in equation (2.5).

2.2.2 Energy levels and absorption

A given molecule may be at different energy levels, depending on the energy levels occupied by its electrons and on which rotation and vibration modes of the molecular bonds are excited.

A photon of frequency ν and wavelength $\lambda = c/\nu$ has an energy $h\nu$.

Figure 2.5: Black-body radiation.



(a) Planck's black-body spectral radiance B(L,T) for the emission temperatures of the Earth (blue) and Sun (red) as a function of wavelength, in $Wm^{-2}ster^{-1}m^{-1}$. The Sun radiates much more strongly, of course and its peak emission is at a shorter wavelength (hence "shortwave" radiation), whereas Earth's peak emission comes at longer wavelengths ("longwave radiation"). (b) The blue curve shows an estimated outgoing longwave radiation at the top of the atmosphere as a function of wavenumber, with black-body radiation curves at different temperatures shown by dashed lines: $mWm^{-2}ster^{-1}(cm^{-1})^{-1}$. The deviations from the 287 K black body radiation curve indicate absorption bands due to CO_2, CH_4, H_2O, and O_3; the central CO_2 absorption line is shown on both panels as a vertical gray bar. The two panels use different units for the *x*-axis that are often used to display such spectra, wavelength (λ, μm) on the left and wavenumber ($1/\lambda$, cm^{-1}) on the right.

For specific frequencies, the photon may be able to deliver the exact amount of energy needed to cause the molecule to transition between its energy level and another level and thus be absorbed by the molecule.

Moving electrons to higher orbitals typically requires very energetic photons in the ultraviolet part of the SW range.

Exciting vibrations and rotations of greenhouse gas molecules in the atmosphere requires photons in the LW range, allowing these molecules to absorb LW radiation.

Photon frequencies or wavelengths corresponding to the absorption energies of greenhouse gases are referred to as absorption lines.

Figure 2.5b shows the estimated top of the atmosphere outgoing LW radiation; one can clearly identify wavelength bands where the OLR is significantly below the 287 K black body curve representing emission from Earth's surface.

These bands correspond to wavelengths in which different greenhouse gases absorb and re-emit outgoing radiation, eventually re-emitting it from a higher altitude and thus at a colder temperature to outer space, and therefore reducing the OLR at these wavelengths.

The absorption from 400 to 600 cm and beyond 1200 cm is due to water vapor, which is the dominant greenhouse gas, responsible for approximately 50% of the Earth longwave greenhouse effect.

The absorption from 1000 to 1100 cm is due to ozone, and the minimum in OLR around 600–800 cm marked by the gray bar (also shown by the gray bar at 15 μm in panel a) is due to absorption by CO_2.

2.2.3 Broadening

Molecules can absorb photons with energy in a broader range around the precise energies corresponding to transitions between pairs of molecular energy levels due to two main effects.

The first is *pressure broadening*, which may occur when two molecules collide while one of them is absorbing a photon.

In that case, as one molecule absorbs the photon, some excess energy can be transferred to/from the colliding molecule, allowing the absorbing molecule to retain exactly the energy needed for an energy-level transition.

As a result, the absorbed photon does not need to be at exactly the energy corresponding to the energy-level transition, showing up as a broadening of the lines of absorbed photon frequencies.

This broadening is enhanced at higher pressures, as the frequency of molecular collisions is enhanced then—hence the name of the effect.

The second (and significantly weaker effect in the Earth atmosphere) is Doppler broadening: the thermal motion of molecules leads to a shift in the frequency of emitted/absorbed photons as seen by the moving molecule due to the Doppler effect.

A photon with a slightly different frequency than that needed to induce a transition in a resting molecule can therefore seem by the moving molecule to have exactly the right frequency, allowing the moving molecule to absorb it.

Given these two effects, absorption is maximal at the absorption lines, at the frequencies/wavelengths corresponding to molecular energy transitions, and it gradually decays away from these lines.

Even when the atmosphere is saturated with respect to absorption at the CO_2 absorption lines, increasing CO_2 concentration leads to stronger absorption at wavelengths near the absorption lines that are not saturated.

The increased LW absorption near these lines again explains the greenhouse effect due to increasing CO_2 concentration, in addition to the increase in the emission height discussed earlier.

2.2.4 Radiative forcing, logarithmic dependence on CO_2

Consider radiation propagating through a pipe filled with a radiation-absorbing gas (say CO_2), and let $I(x)$ be the intensity (W/m^2) of the *transmitted* radiation as a function of distance x.

The transmitted radiation intensity is expected to decline along the pipe due to the absorption, leading to a negative rate of change, dI/dx, along the length of the pipe.

The absorption is typically proportional to the radiation intensity at any given location, and therefore the equation for the transmitted radiation, $I(x)$, may be written as $dI/dx=-\mu I$, where μ depends on the absorbing medium.

The solution for the radiation intensity along the pipe is therefore , so that the transmitted radiation decreases exponentially with distance and with the amount of CO_2 encountered by the radiation.

The *radiative forcing* due to a given atmospheric CO_2 increase is the radiation (in W/m^2) that is absorbed by the added gas and is prevented from being transmitted to outer space, over all wavelengths, assuming the tropospheric temperature, moisture, and clouds are kept at their values before the CO_2 increase.

Radiative forcing is calculated using radiation models rather than being observed directly, yet the uncertainty involved is small relative to many other factors affecting the climate response to anthropogenic greenhouse forcing.

Exact calculations show that the radiative forcing of CO_2 depends *logarithmically* rather than exponentially on CO_2 concentration.

That the response is not exponential is a result of the fact that, as discussed above, the atmosphere is saturated with respect to CO_2 absorption—that is, LW photons emitted from the surface at the wavelengths of the CO_2 absorption lines are already absorbed by the atmosphere at preindustrial CO_2 concentrations.

The shape of the absorption bands as a function of wavelength, and as a function of the distance from the absorption lines discussed above, leads to the logarithmic dependence of radiative forcing on the CO_2 concentration.

We show later that the warming and radiative forcing are approximately linearly related.

This implies that the temperature response also depends logarithmically on the CO_2 concentration.

We therefore write, schematically, $T = T_0 + A \log_2 (CO_2/280)$, where the use of log base 2 implies that A (in Kelvin) is the temperature increase if CO_2 concentration doubles from its preindustrial value of 280 ppm. This dependence means that the warmings due to one and two doublings of the CO_2 concentration are

$$\begin{aligned} T_{\times 2} - T_{\times 1} &= A \log_2 (2\mathrm{CO}_2/280) - A \log_2 (\mathrm{CO}_2/280) = A \log_2 2 = A, \\ T_{\times 4} - T_{\times 2} &= A \log_2 (4\mathrm{CO}_2/280) - A \log_2 (2\mathrm{CO}_2/280) = A \log_2 2 = A. \end{aligned}$$

That is, each doubling of CO_2 leads to the same increase in temperature!

2.2.5 Other greenhouse gases, global warming potential

Human activities led to the increase in concentration of different greenhouse gases that can absorb in the longwave range, including carbon dioxide (CO_2), methane (CH_4), nitrous oxide (N_2O), CFC12, CFC11, and more.

By 2020, the radiative forcing due to CO_2 relative to 1750 has been estimated at about 2 W/m^2, and when taking into account the other anthropogenic greenhouse gases mentioned above, the total radiative forcing is about 3 W/m^2.

For perspective, the RCP8.5 scenario, representing a business-as-usual worst-case scenario, corresponds to a radiative forcing increase of 8.5 W/m^2 by year 2100.

One way to present the effect of other greenhouse gases is via their CO_2-equivalent mixing ratio, which is the CO_2 concentration that would lead to the same radiative forcing as a given concentration of another given greenhouse gas.

While the concentration of CO_2 just passed 400 ppm by 2020, the combined CO_2-equivalent mixing ratio due to all greenhouse gases reached 500 ppm.

Some of these added anthropogenic gases have a larger radiative forcing than CO_2 per kilogram of gas, yet their lifetime in the atmosphere may be shorter.

For example, an added unit weight of methane (CH_4) leads to a significantly stronger increase in LW absorption than the same weight of added CO_2.

But methane has a shorter lifetime in the atmosphere (about a decade; it eventually reacts with the OH radical to form CO_2 and water) than CO_2 (thousands of years; see later discussion).

It is useful, therefore, to be able to compare the *long-term* radiative effects of different gases, not reflected by the CO_2-equivalent mixing ratio; this is done using the concept of global warming potential (GWP).

The GWP of a greenhouse gas is its forcing effect relative to that of CO_2, taking into account both its lifetime and its strength as an absorber, over a specified period referred to as a *time horizon* (TH).

The GWP of a gas *x* is

$$GWP(x) = \frac{\int_0^{TH} a_x \cdot [x(t)]dt}{\int_0^{TH} a_r \cdot [r(t)]dt},$$

where *x(t)* is the time-dependent decay in abundance of the gas and *r(t)* that of CO_2; a_x is the radiative forcing of *x* per 1 added kilogram of *x* (W/m2/kg), and a_r is that of CO_2.

Climate Background Box 1

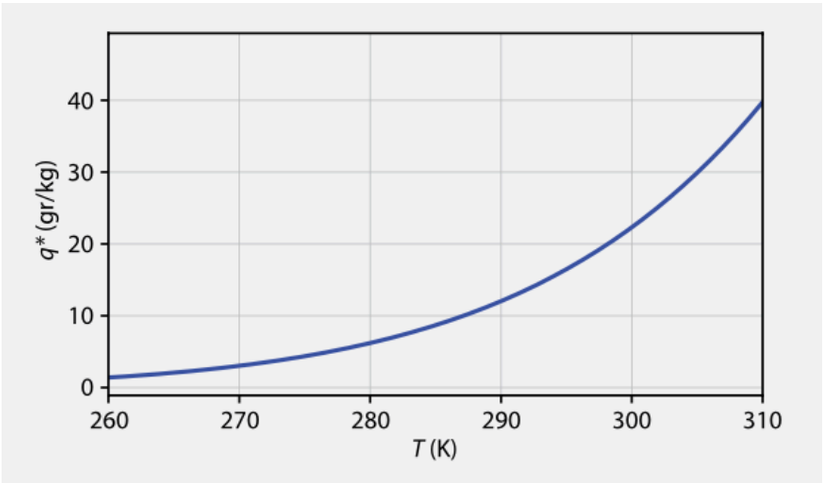
The Clausius-Clapeyron relation

Atmospheric water vapor plays a dominant role in the response to greenhouse gas increase, as well as in the many consequences of warming, from anticipated changes to droughts, forest fires, hurricanes, and extreme precipitation events.

The saturation water vapor pressure is the pressure of water vapor in equilibrium with water at a given temperature.

The relative humidity is defined as the ratio of the water vapor pressure and saturation water vapor pressure.

To a first approximation, the relative humidity may often be assumed not to change in response to warming, which means that the water vapor response is determined by the behavior of the saturation moisture as a function of temperature, given by the Clausius-Clapeyron relation.



The saturation specific humidity, denoted $q^*(T,p)$, is the mass (kg) of moisture in 1 kg of moist air, at saturation, at a temperature T in degrees Celsius and air pressure p in hPa, and may be calculated as follows.

Let $R_v = 461$ J K^{-1} kg^{-1} be the specific gas constant for water vapor and $R_d = 287$ J K^{-1} kg^{-1} the specific gas constant for dry air.

The saturation ($e^*(T)$ in hPa) is then given by the approximate Clausius-Clapeyron relation

$e^*(T) = 6.112 \exp(17.67\ T/(T+243.5))$.

The saturation mixing ratio (kg water vapor per kg dry air) is given by $r^* = (R_d/R_v)\ e^*/(p-e^*)$, while the saturation specific humidity (kg water vapor per kg moist air) is, finally, $q^* = r^*/(1+r^*)$.

Note that is exponential in the temperature as shown in the figure, a strong dependence that is behind the important role of atmospheric moisture in climate change.

The above Clausius-Clapeyron relation is an approximate solution to the Clausius-Clapeyron equation,

$$\frac{1}{e^*(T)}\frac{de^*(T)}{dT} = \frac{L}{RT^2} \equiv \alpha_{cc}(T),$$

where $L \approx 2250$ J K^{-1} kg^{-1} is the (temperature-dependent) latent heat of vaporization, and is the expected relative change to the saturation moisture per degree Kelvin warming.

This equation is also useful on its own as it shows that a $\Delta T = 1$ K warming leads to a 7% increase in moisture: $\Delta e^*/e^* = a_{cc}(T)\Delta T \approx 0.07$.

Therefore, a mid-range estimated surface warming of 3 K in response to a doubled CO_2 concentration implies an over 20% increase in water vapor pressure or in specific humidity.

The consequences of such a dramatic increase are reviewed throughout this book where the Clausius-Clapeyron relation is repeatedly used.

Finally, we note that for a typical surface pressure of $p_s = 1000$ hPa, the saturation specific humidity may be further approximated as a function of the temperature T in Kelvin, as
q^* (T,1000 hPa) = 1.68 x 10^6 exp (-5415/T) (kg moisture per kg moist air).

2.2.6 The water vapor feedback

Greenhouse gases such as CO_2 and methane also have an indirect radiative effect.

A warmer atmosphere due to an increased greenhouse gas concentration allows for more atmospheric water vapor content in accordance with the Clausius- Clapeyron relation (discussed earlier).

Since water vapor is also a potent greenhouse gas, this causes further warming, which causes the moisture content to further increase and the temperature to further increase.

This amounts to a positive feedback: the initial warming leads to more water vapor, which further amplifies the warming.

Model calculations and physical arguments show that the relative humidity may be assumed to remain roughly unchanged as climate warms, allowing us to calculate the approximate expected increase in specific humidity directly from the Clausius-Clapeyron relation.

Estimates are that a surface warming of 1 C (say, due to the increase of CO_2) leads to increased atmospheric humidity that accounts for an extra radiative forcing of 2 W/m^2.

For perspective, CO_2 doubling induces a radiative forcing of about 4 W/m^2.

Put differently, the water vapor feedback roughly doubles the climate sensitivity, defined as the equilibrium surface warming response to the doubling of CO_2 (more layer).

Water vapor changes are a response (feedback) to an increase in greenhouse gas concentration rather than a direct forcing, because water vapor concentration responds to temperature changes on a very short timescale via surface evaporation and precipitation.

Any climate warming involves a strong negative feedback due to the outgoing LW radiation (cooling) increasing as , and thus leads to stronger radiative cooling in response to the climate warming.

The positive water vapor feedback in the Earth atmosphere only weakens this negative feedback by weakening the increase in LW radiation as the temperature increases, such that the net feedback is still negative.

3 - TEMPERATURE

The warming of surface temperatures over the past decades, and the expected further warming are, of course, at the heart of anthropogenic climate change and are the reason for many of the other issues related to global warming that affect us directly and indirectly.

Figure 3.1 shows a time series of the observed globally averaged surface temperature anomaly (relative to the mean of 1961– 1990) since the year 500, and the projected temperature time series for three different warming scenarios in a climate model.

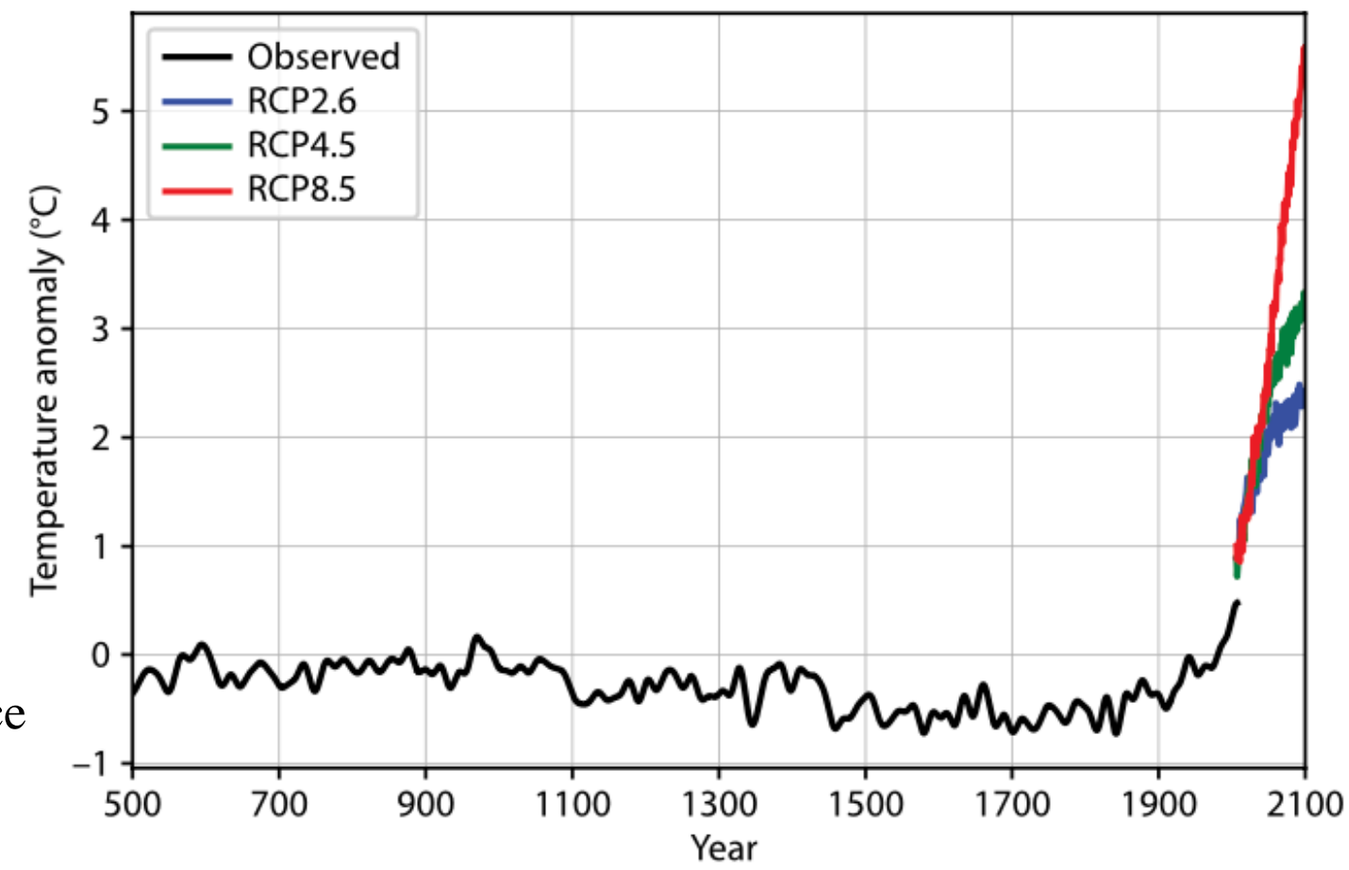


Figure 3.1: Observed and projected warming.

Observed globally averaged surface temperature anomaly (relative to the mean of 1961–1990) since year 500 (black) plus projections from three RCP scenarios in a climate model.

The observed record has been referred to as the *hockey-stick curve* due to its shape, showing a decline over most of the record and an increase in the past century or so.

Climate models that are run under a CO_2 doubling scenario (from a preindustrial value of 280 to 560 ppm) predict a globally averaged warming of 2–4 C as a result of the greenhouse forcing and all modeled feedbacks.

By 2020, CO_2 already increased from 280 to nearly 420 ppm (and the CO_2-equivalent mixing ratio, which includes the effects of other anthropogenic greenhouse gases, has already reached 500 ppm; see earlier discussion).

We have seen that the dependence of radiative forcing on CO_2 is logarithmic (section 2.2.4), which seems to suggest that we should have seen most of the warming that is expected for CO_2 doubling.

One therefore wonders if the warming that has so far been observed, of about 1.1 C, is consistent with the climate warming projected for a doubling of CO_2.

It turns out that the oceans play a dominant role in delaying the warming response to greenhouse gas increase; we will explore this issue later.

Figure 3.2 shows that the projected warming according to the RCP8.5 scenario is much larger in the Arctic than in mid-latitudes and that high latitudes in general are expected to warm more than low latitudes.

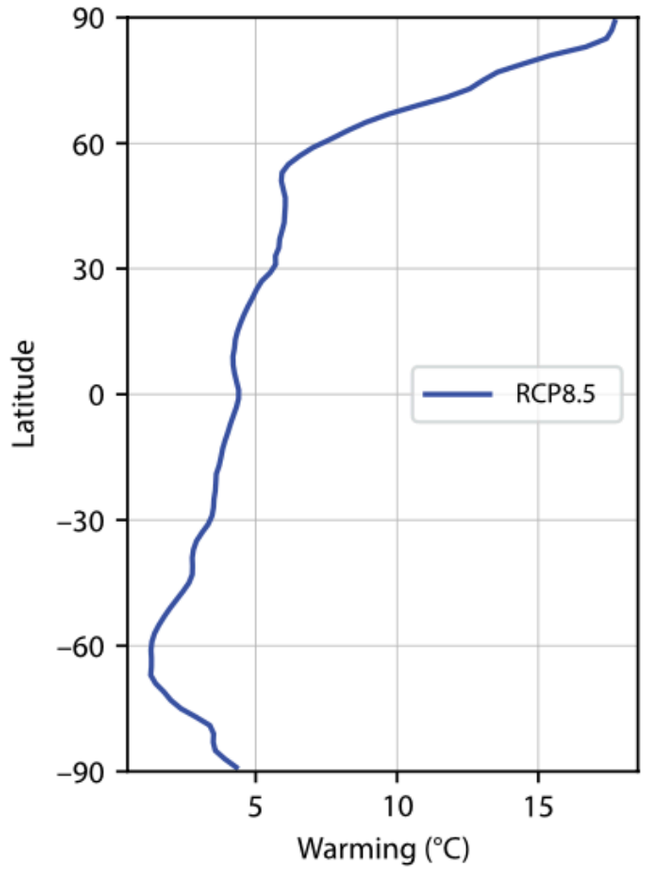


Figure 3.2: Polar amplification.

Warming over the 21st century as a function of latitude, based on the RCP8.5 scenario (Figure 2.1) in a climate model.

This is referred to as *polar amplification*, and three mechanisms for this robust observation are discussed later.

The warming over the past century is characterized by a couple of periods of weak warming or even weakly cooling trends, sometimes referred to as "*hiatuses*."

While the specific mechanisms for these periods of weaker warming trends are still hotly debated, they are an unavoidable result of combining natural climate variability with a warming trend, as discussed later.

Finally, a robust signature of CO_2-induced climate change is a cooling of the stratosphere (the atmospheric layer from 10 to 50 km above the surface), while the troposphere, the lower 10 km of the atmosphere, is warming.

In section 3.4 we extend the two-layer atmospheric energy balance discussed previously to a three-layer atmospheric energy balance model to explain this.

3.1 CLIMATE SENSITIVITY AND THE ROLE OF THE OCEAN

Suppose the expected long-term surface warming after the climate system equilibrated with a doubling of CO_2 (or more precisely, a doubling of the CO_2- equivalent mixing ratio; see earlier discussion), with all feedbacks taken into account, is 3 C, roughly in the middle of the IPCC range. Given that we expect warming to be logarithmic in CO_2 concentration, we write the expected surface warming as a function of the CO_2 concentration as $\Delta T = 3 \log_2(CO_2/280)$.

For a recently achieved equivalent CO_2 mixing ratio of 500 ppm, one therefore expects a warming of $\Delta T = 3 \log_2(500/280) = 2.5$ C, more than twice as much as has been observed.

This section discusses and explains this seeming discrepancy in terms of the concept of *transient* climate sensitivity, after discussing first the related concept of *equilibrium* climate sensitivity.

3.1.1 Equilibrium climate sensitivity

Consider first how we may estimate the eventual warming of the climate system, after it had equilibrated in response to a doubling of the CO_2 concentration, using present-day observations.

The increase in greenhouse gas concentrations implies an additional downward heat flux from the atmosphere toward the surface, which leads to a warming of the lower atmosphere and upper ocean (say upper 50 m), by an amount ΔT.

We term this additional heat flux *radiative forcing* (measured in W/m^2, defined earlier), denote it ΔF, and refer to the warming of the surface atmosphere and upper ocean as surface warming.

The heat capacity of an object is the amount of heat (J) required to raise its temperature by one K and its units are therefore J/K.

Note that the heat capacity of the atmosphere is of the same order of magnitude as the heat capacity of the upper 10 m of the ocean.

The ocean, which is able to (slowly) mix heat down to great depths of up to 4000 m, therefore dominates heat budget changes due to an increase in greenhouse gas concentration.

It is estimated that the oceans so far have absorbed about 90% of the excess heat due to increased greenhouse gas concentrations.

This excess heat is defined to be the radiative forcing integrated over time.

As we will see, this does not mean that the ocean *reduces* the eventual warming, but it can certainly *delay* it.

While the land covering a third of the Earth surface is massive, of course, heat penetrates into soil or rocks only by very slow diffusion.

Seasonal temperature variations can therefore penetrate about a meter into the land surface, while diurnal variations, having a much shorter timescale, penetrate only less than 10 cm.

As a result, only the uppermost part of the land surface contributes to the effective heat capacity of the Earth surface, and its contribution to the storage of excess heat due to radiative forcing is significantly less than that of the ocean.

Climate models estimate that the increase in outgoing top-of-the- atmosphere (TOA) radiative cooling due to a surface warming ΔT is, to a good approximation, linearly proportional to this warming and may therefore be written as λ_{LW} DT.

This linearity is further explained in equation (3.5).

The parameter $\lambda_{LW}(W/m^2K^{-1})$ is not well known, because different climate models produce different values.

In addition, part of the radiative forcing denoted $\Delta Q(W/m^2)$ is transported to the deeper ocean.

The statement of heat balance of the lower atmosphere and upper ocean at the current time (now) is therefore that the heat flux into the deeper ocean is equal to the radiative forcing minus the radiative cooling,

$$\Delta Q_{\text{now}} = \Delta F_{\text{now}} - \lambda_{LW} \Delta T_{\text{now}}. \qquad (3.1)$$

Consider a doubling (2x) of the CO_2 concentration, leading to a radiative forcing of ΔF_{2x}.

After the climate system has had time to warm and equilibrate to the new radiative forcing, the ocean absorbs no additional net heat (it is in a new, warmer, steady state and the net heat flux into the ocean must therefore vanish), and we have $0 = \Delta T_{2x} - \lambda_{LW} D T_{2x}$.

Our objective is to use observational constraints to calculate the anticipated final equilibrium warming for double CO_2, ΔT_{2x}.

From the above, we have

$$\Delta T_{2\times} = \frac{\Delta F_{2\times}}{\lambda_{LW}}$$

as well as

$$\lambda_{LW} = \frac{\Delta F_{\text{now}} - \Delta Q_{\text{now}}}{\Delta T_{\text{now}}},$$

which may be combined to give an equilibrium warming of

$$\Delta T_{2\times} = \frac{\Delta F_{2\times}\,\Delta T_{\text{now}}}{\Delta F_{\text{now}} - \Delta Q_{\text{now}}}. \qquad (3.2)$$

The present surface warming (ΔT_{now}) and current heat flux into the deep ocean (ΔQn_{ow}) appearing on the RHS can be estimated from present-day observations, while the radiative forcings at present and for a doubling of CO_2 (ΔF , ΔF_{2x}) can be estimated reliably from radiation models.

Thus, the RHS is known, and this can be used to estimate the desired equilibrium climate sensitivity on the LHS without relying on complex and somewhat uncertain climate models.

The result is, not surprisingly, in the same range predicted by these models.

3.1.2 Transient climate sensitivity

The deeper ocean absorbs some of the heat due to the excess radiative forcing, and its large heat capacity means that it warms up very slowly.

Because the relatively cool deep ocean continuously mixes with the more rapidly warming surface ocean, it slows down the surface warming.

This means that the currently observed surface warming, for example, reflects only part of the potential warming given the present greenhouse gas concentration; this section attempts to explain and quantify this effect.

Consider the response of the surface temperature (including the upper ocean) and the deeper ocean temperature to an abrupt doubling of the CO_2 concentration.

The following equations describe the temperature evolution governed by the heat budget (per unit area) of the lower atmosphere and upper ocean, whose temperature perturbation due to the increase in greenhouse gases is denoted $\Delta T_{surface}$, and of the deeper ocean, whose temperature perturbation is denoted ΔT_{deep},

$$
\begin{aligned}
C_{\text{surface}} \frac{d\Delta T_{\text{surface}}}{dt} &= \Delta F_{2\times} - \lambda_{LW} \Delta T_{\text{surface}} - \gamma(\Delta T_{\text{surface}} - \Delta T_{\text{deep}}) \\
C_{\text{deep}} \frac{d\Delta T_{\text{deep}}}{dt} &= \gamma(\Delta T_{\text{surface}} - \Delta T_{\text{deep}}). \qquad (3.3)
\end{aligned}
$$

The first equation is the perturbation heat budget per unit area for the atmosphere and upper ocean due to the increase in greenhouse gases; the second, for the deeper ocean.

$C_{surface}$ and C_{deep} are the heat capacities per unit area of the combined atmosphere and ocean surface and of the deep ocean, correspondingly.

The left-hand side of both equations is therefore the rate of change of the heat content of the atmosphere / surface ocean and of the deep ocean, per unit area.

The RHS represents heat fluxes responsible for this rate of change.

The term $\gamma(\Delta T_{surface} - \Delta T_{deep})$ represents the slow heat exchange between the deep ocean and the upper ocean, which was represented by ΔQ in equation (3.1).

The parameter γ is in units of heat capacity per unit area over a timescale, (J $K^{-1}m^{-2}$) s^{-1}, where the (long) timescale is that of the slow heat exchange between the deep and surface ocean.

Note that this term has opposite signs in the two budget equations, so that in a warming scenario it represents a source of energy for the deep ocean and a sink for the surface.

The first equation generalizes the equilibrium balance assumed previously in equation (3.1) with two differences: we now take into account the time rate of change of the surface temperature (on the LHS), and we consider the response to an *abrupt* CO_2 doubling, hence the use of in equation (3.3) rather than ΔF_{now}.

As for parameter values, for a CO_2 doubling, detailed calculations show that the radiative forcing is $\Delta F_{2x} = 4$ W/m^2, and if the equilibrium climate sensitivity is assumed to be 3 C, then $\lambda_{LW} = (4 \text{ W/m}^2)/(3 \text{ C})$.

$C_{surface}$, which represents the combined heat capacity of the atmosphere and upper 50 m of the ocean, may be approximated as $C_{surface} = \rho_\omega c_p 50\text{m}$; and the heat capacity of the deeper ocean is given by $Cd_{eep} = \rho_\omega c_p H$, where H is the ocean depth, $c_p = 4005$ J K^{-1}kg^{-1} is the specific heat capacity of seawater (heat required to raise the temperature of 1 kg of seawater by 1 K, in J K^{-1}kg^{-1}), and $\rho_\omega = 1024$ kg m^{-3} is the density of seawater.

The parameter γ is set to one, and the second equation in (3.3) implies that the timescale over which the temperature of the deep ocean is modified by the mixing term is γ/C_{deep}, or about 500 yr.

In a steady state, when the LHS of equation (3.3) vanishes, the second equation gives $\Delta T_{surface} = \Delta T_{deep}$.

This indicates that the deep ocean and surface ocean warmed to the same degree and are now in equilibrium.

The first equation then gives $\Delta T_{surface} = \Delta F_{2x}/\lambda_{LW}$, as in section 3.1.1.

Thus, the ocean depth and heat capacity do not enter the steady state solution and do not affect the equilibrium response to radiative forcing.

However, in the short term, the ocean heat capacity can make a big difference, as demonstrated by Figure 3.3.

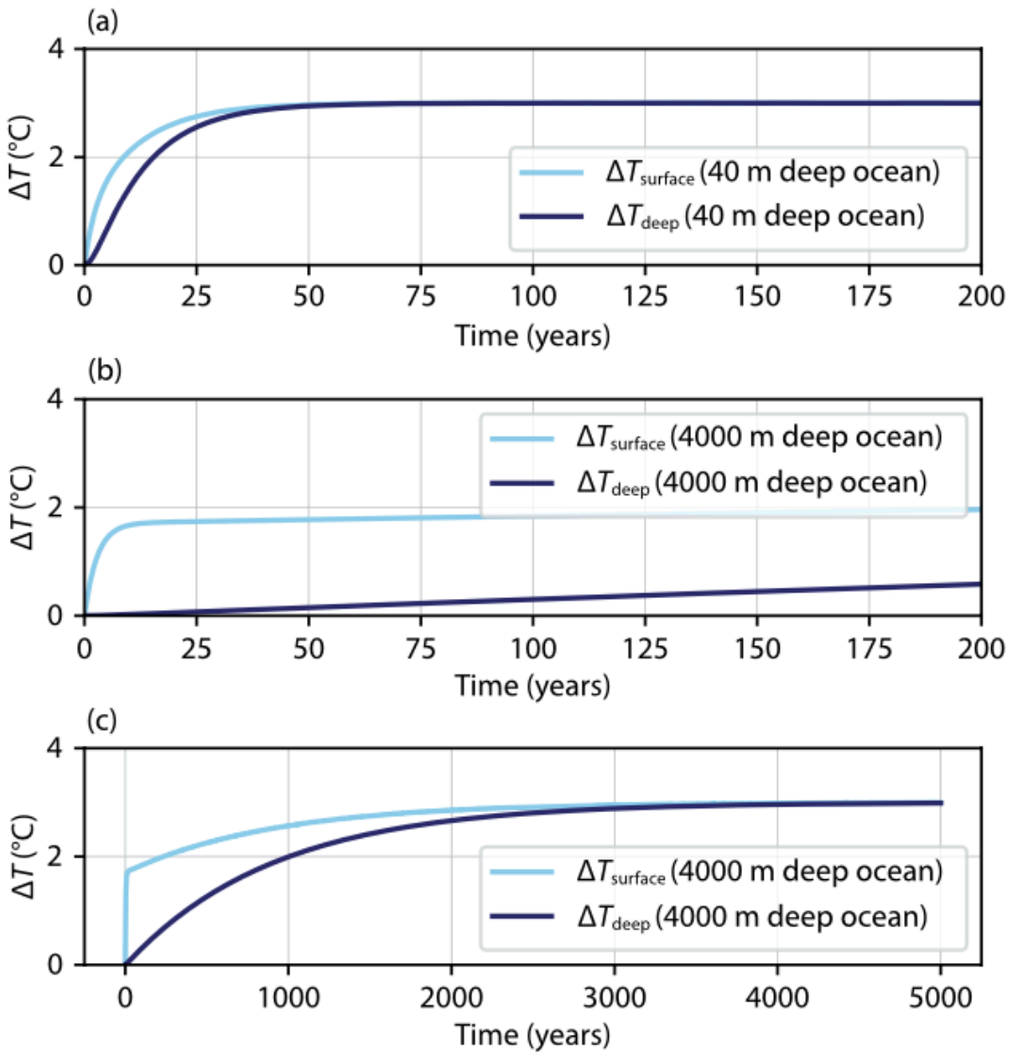


Figure 3.3: Transient climate sensitivity.

The temperature anomalies of the upper ocean and of the deep ocean, as a function of time in a scenario of instantaneous CO_2 doubling. (a) An artificial case assuming the subsurface ocean is only 40 m deep. (b) A more realistic scenario, assuming an ocean depth of 4000 m and showing only the first 200 yr of adjustment to an abrupt doubling of CO_2. (c) Same scenario as in (b), and showing the full period of adjustment.

The upper panel shows that in the case of a shallow ocean (where the deeper ocean is assumed to be only 40 m deep), both the ocean and atmosphere adjust to the new radiative forcing within about 50 yr.

For the case in which the deep ocean depth is assumed to be more realistic (4000 m), the middle panel shows the initial adjustment, showing that the atmosphere adjusts quickly but only partially and that the deep ocean is very slow to respond.

The lower panel shows the longer-term adjustment, demonstrating that the atmospheric adjustment is indeed delayed significantly and is only completed when the deep ocean has completed its warming after thousands of years.

The warming prior to full equilibration, which takes into account the delayed response due to the heat capacity of the deep ocean, is referred to as the *transient* climate sensitivity.

One may conclude that while we have seen a warming of about a degree so far, this is a result of the ocean's heat capacity slowing the atmospheric warming rather than being the result of a low climate sensitivity to CO_2 increase.

3.2 POLAR AMPLIFICATION

The significantly stronger warming seen in the Arctic in both the warming that has occurred so far and in that anticipated in future projections (Figure 3.2) is referred to as *polar amplification* or *Arctic amplification*.

This amplification means that the high latitudes are especially sensitive to global warming, with a range of potential implications from Greenland melting and sea level rise to Arctic wildlife to effects on remote mid-latitude weather and more.

The polar/Arctic amplification is due to three main factors, which are reviewed in this section.

The albedo feedback

The melting of high-latitude snow and ice in a warmer climate, over both land and ocean, leads to a decrease in albedo, an increase in absorbed solar radiation, and therefore further warming.

This is an important positive feedback leading to the warming of the high latitudes where snow and ice are currently widespread, and therefore to polar amplification.

But the reduced high-latitude albedo is not the only responsible mechanism nor necessarily the strongest one leading to polar amplification.

The Planck feedback

Consider a simple equation balancing incoming radiation flux F with an outgoing black-body LW radiation, $F = \epsilon\sigma T^4$, where the temperature T is in Kelvin. If a radiative forcing is now added to the LHS, we expect the temperature to increase by , leading to a new balance,

$$F + \Delta F = \epsilon\sigma(T + \Delta T)^4. \qquad (3.4)$$

A quick calculation using this equation shows, for example, that if T = 240K (-33°C), an increase in radiative forcing of ΔF = W/m^2 leads to about twice as large an increase ΔT than if T = 300K (27°C), letting the longwave emissivity be for simplicity.

Thus, the same radiative forcing leads to a larger temperature response if the base temperature is colder.

This *Planck feedback* contributes to Arctic amplification. Equation (3.4) may be simplified and linearized using the fact that (ΔT)/T << 1, so that higher powers of this ratio may be neglected.

Expanding dropping all terms with powers of (ΔT)/T higher than one, we find ΔF = 4 $\epsilon\sigma$T$^3\Delta$T, or

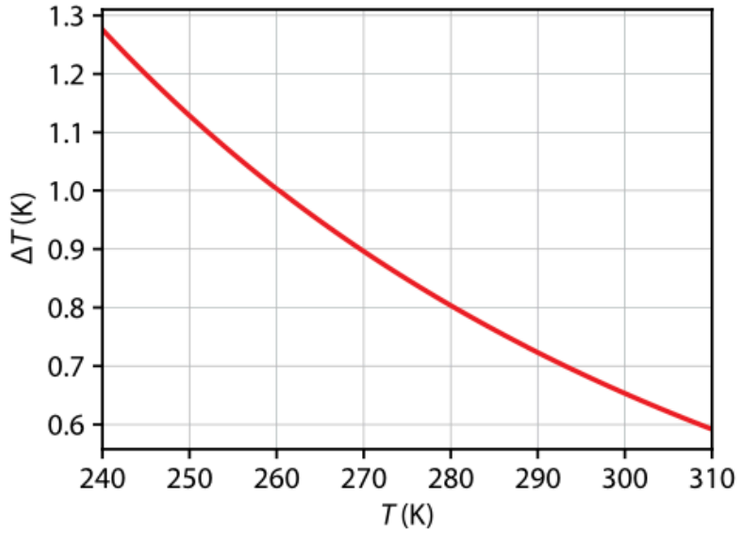
$$\Delta T \approx \Delta F / \left(4\epsilon\sigma T^3\right). \qquad (3.5)$$

This expression makes it even clearer that the temperature response ΔT to radiative forcing ΔF is stronger at low temperatures T, as shown in Figure 3.4.

The above discussion treated the radiative forcing as constant in latitude in order to isolate the effects of the Planck feedback on polar amplification, although in reality it may be latitude-dependent as well.

Figure 3.4: The Planck feedback.

The warming expected due to an increase in radiative forcing of 4 W/m² based on equation (3.5) with $\epsilon = 1$.



The tropical lapse rate feedback

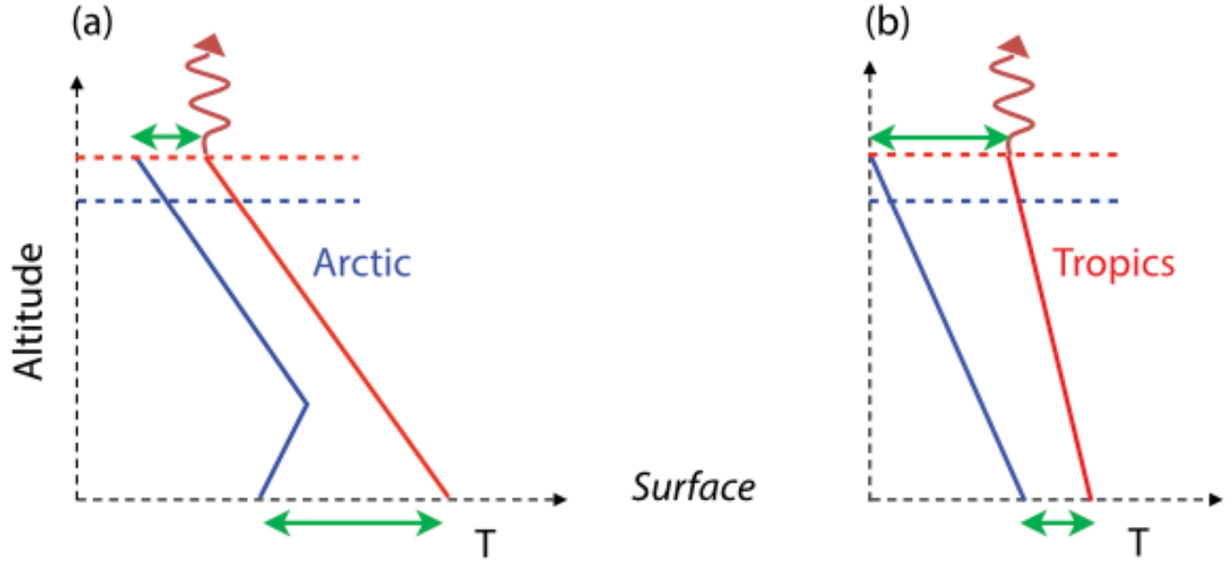
The atmospheric lapse rate, dT/dz, where *T* is the atmospheric temperature and *z* the altitude in meters, represents the rate of cooling with altitude and varies between about -5 and-9.8 K/km.

The lapse rate itself, as well as its change due to warming, depends on the surface temperature and on the atmospheric moisture, among other factors.

This dependence leads to a different change to the lapse rate as a result of surface warming in the tropics versus high latitudes, a difference that contributes to polar amplification.

Figure 3.5: Schematics of the Arctic (a) and tropical (b) lapse rate feedbacks.

Solid blue (red) lines show the temperature profiles before (after) warming. Dashed blue (red) lines show the emission level before (after) the warming (see earlier). The green double arrows show the warmings at the surface and at the emission level.



Starting with the tropics, consider a parcel of moist air adiabatically rising (that is, the air parcel rises without exchanging mass or heat with its environment) with some specified surface temperature and moisture content.

As the parcel rises, it encounters lower atmospheric pressure, expands and cools, and as a result experiences some condensation as well.

Its temperature profile with height is known as the moist adiabatic profile, and its rate of cooling with height is the moist adiabatic lapse rate.

A warming of the surface temperature (say, due to an increase in greenhouse gas concentration) changes the moist adiabatic temperature profile such that there is an even larger warming at high altitudes, as explained below.

This reduction in the difference between the high altitude and surface temperatures implies that the tropical lapse rate becomes less negative with a warming of the surface temperature.

Given that the tropical atmospheric temperature as a function of altitude tends to follow the moist adiabatic profile, we expect that in a warmer climate, as the surface warms in the tropics, the higher troposphere there will warm even more.

The stronger high-altitude warming leads to stronger radiative cooling from the upper troposphere to outer space via outgoing longwave radiation than would have been expected given the magnitude of the surface warming (see schematic Figure 3.5b).

This implies that the change in tropical lapse rate is a negative feedback: the surface warms, and the increased cooling to outer space more than compensates for this warming, thus reducing the tropical surface warming.

The weakening (in absolute value) of the tropical lapse rate with surface warming that is behind this negative feedback can be calculated using energy conservation for moist air parcels and can be qualitatively understood as follows.

Consider two air parcels at the surface, one initially warmer than the other, and assume for simplicity that both are at saturation with respect to water vapor (having relative humidity of 100%).

Due to the Clausius-Clapeyron relation (see earlier), the warmer parcel contains more moisture initially.

As the parcel rises and encounters lower pressures, it expands and therefore cools adiabatically.

The cooling leads to condensation and therefore to latent heat release.

The parcel that started out warmer and moister will, as a result, experience more latent heat release and therefore more heating as it rises.

This heating means a smaller rate of cooling as the parcel rises, and thus a less negative lapse rate. This implies that a warmer surface temperature leads to an even larger warming aloft.

Figure 3.6 shows two such temperature profiles and their difference, calculated by considering the energy conservation of the rising moist air parcels, as explained using the concept of moist static energy and by solving the corresponding equations.

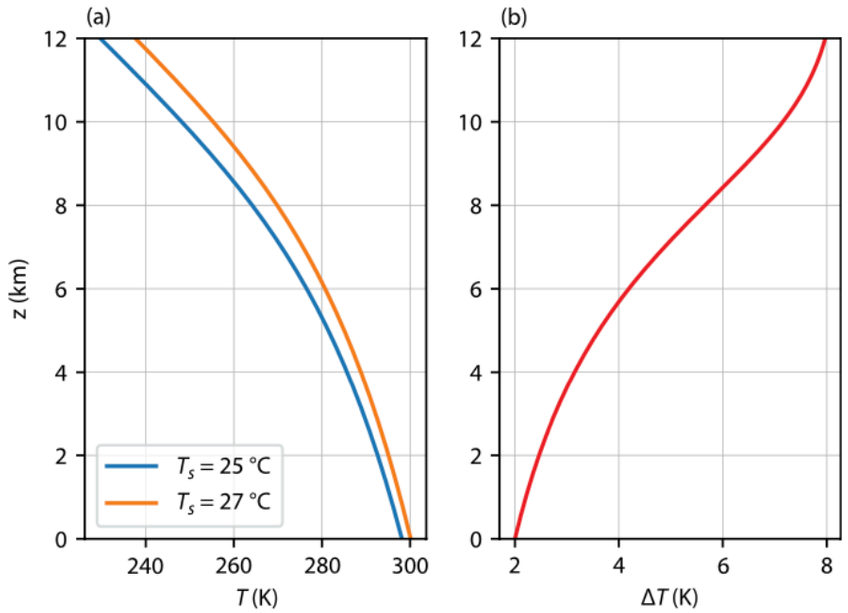


Figure 3.6: Tropical lapse rate feedback.

(a) Temperature profiles of two surface air parcels starting with a relative humidity of 100% and two different surface temperatures and rising adiabatically in the atmosphere. (b) The difference in temperature between the two profiles, showing the enhanced upper atmosphere warming of the parcel that starts with a slightly warmer surface temperature.

The high-latitude lapse rate response to warming

At high latitudes, the warming response to an increased greenhouse gas concentration tends to be surface-enhanced.

As a result, a given surface warming is accompanied by a smaller warming in the upper troposphere.

This leads to a smaller increase in outgoing LW radiative cooling to outer space than would have been expected from the surface warming, and this acts as a *positive* feedback on the surface warming.

This enhanced surface warming and thus the positive Arctic lapse rate feedback result to a large degree from the fact that the winter Arctic surface in the present climate, especially over sea ice and snow, is very cold, and the near-surface winter-time atmospheric temperature *increases* with height.

This is referred to as a surface *inversion* (blue line in Figure 3.5a).

A surface–enhanced warming, as shown by the red line in Figure 3.5a, reduces the near-surface inversion strength (or eliminates the inversion), resulting in a more significant warming at the surface than at high altitudes.

The weakening of the inversion in a warmer climate can occur for several reasons.

First, such an inversion typically occurs in the present climate over sea ice in the Arctic, so that a melting of the sea ice exposes the much warmer surface ocean and eliminates the inversion.

A second possible mechanism in a warmer climate involves the response of low clouds to warming.

As the ocean warms, so does the air over the ocean, which becomes moister as a result of the Clausius-Clapeyron relation.

The moister air travels from the ocean over the cold high-latitude continents (e.g., Canada, northern United States) during winter-time and cools as a result.

The moisture condenses and leads to the formation of clouds in the lower atmosphere.

Because the solar radiation during winter is weak at high latitudes, the cloud albedo effect is minimal.

At the same time, the low clouds absorb the LW radiation emitted by the surface and re-emit it downward and upward (see earlier), thus leading to a strong cloud greenhouse effect.

This leads to an enhanced warming of the lower atmosphere and thus to the weakening or elimination of the inversion and to the Arctic lapse rate feedback.

The combined response of the negative lapse rate feedback at low latitudes and positive lapse rate feedback at high latitudes leads to enhanced Arctic warming and reduced tropical warming, contributing to polar amplification.

3.3 “HIATUS” PERIODS

The globally averaged surface temperature record over the past century and a half shows a period of what seems to be temporary cooling (1940–1970 or so) and another period of seemingly paused warming (1998–2013).

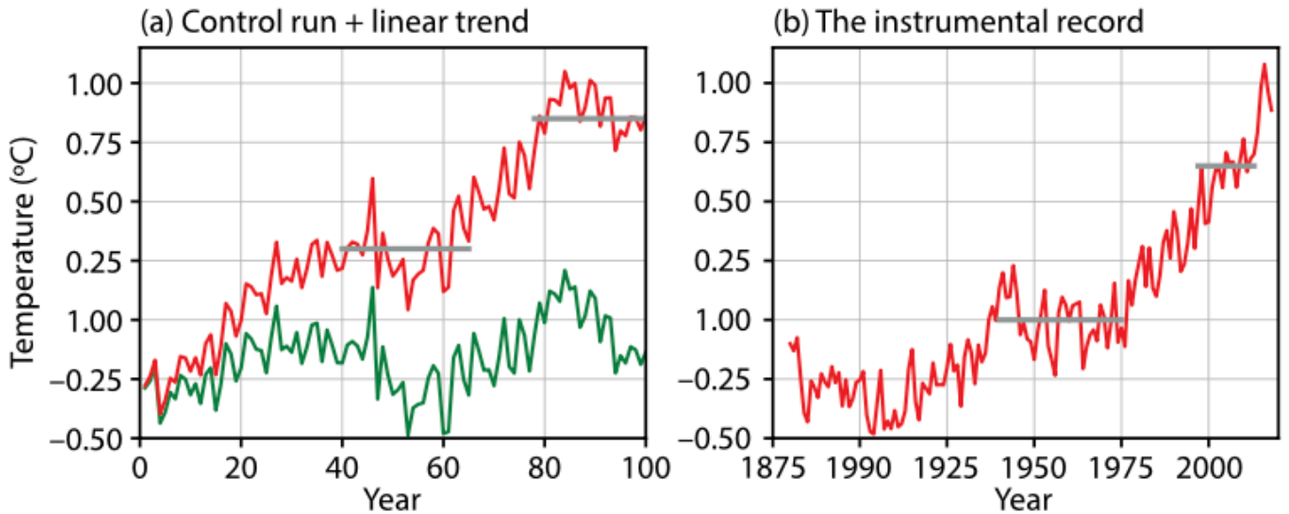
There is a significant body of research work looking at the mechanisms behind the temperature trends of these periods, especially the latter one.

Yet it is useful to remember that natural climate variability leads to oscillations in the globally averaged temperature, and when these oscillations are superimposed on a warming trend, the two combine to give what seem to be “hiatus” periods.

This is demonstrated in Figure 3.7, where the right panel shows the observed record for the surface temperature with the two “hiatus” periods marked by horizontal bars, and the left panel shows the globally averaged temperature in a control run (driven with a fixed preindustrial CO_2 concentration) of a climate model (green) and the same time series with an added linear trend (red).

Figure 3.7: “Hiatus” periods.

(a) Globally averaged surface temperature anomaly from a control (CO_2 fixed at preindustrial concentration) run of a climate model (green), and the same temperature record with a linear trend added (red). (b) The instrumental record of globally averaged surface temperature anomaly. Both panels show “hiatus” periods marked by gray horizontal bars.



The control run record plus a linear trend produces what seems to be “hiatus” periods, supporting the idea that such periods are to be expected in a global warming scenario.

Furthermore, observations show there was no “hiatus” in the warming of the oceans over 1998–2013.

Given that 90% of the excess heat trapped in the climate system due to the increase in greenhouse gas concentration ends up in the subsurface ocean, the “hiatus” in surface warming makes a little dent in the overall warming trend in the climate system.

3.4 STRATOSPHERIC COOLING

The distinguishing element of the stratosphere is its warming as a function of altitude, as opposed to the cooling as a function of altitude in the troposphere (Figure 3.8a).

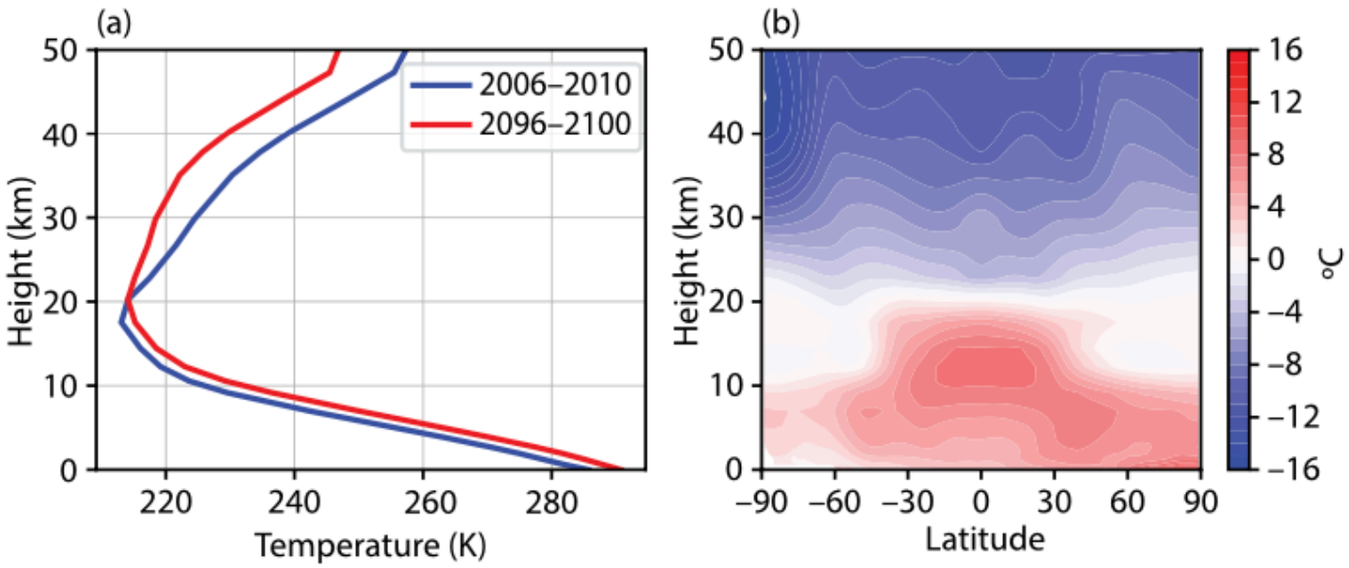


Figure 3.8: Stratospheric cooling.

(a) Mid-latitude (30 N - 50 N) zonally averaged temperature profiles for an RCP8.5during the 21st century to the RCP8.5 projection at the beginning and end of the 21st century. (b) The zonally averaged atmospheric temperature response during the 21st century to the RCP8.5 scenario, showing a tropospheric warming and a stratospheric

The distinguishing element of the stratosphere is its warming as a function of altitude, as opposed to the cooling as a function of altitude in the troposphere (Figure 3.8a).

This is a result of absorption of SW radiation by the ozone layer in the stratosphere.

Thus, unlike the troposphere, whose main source of radiative heating is LW radiation from the surface, the stratosphere is also heated by absorbed SW radiation.

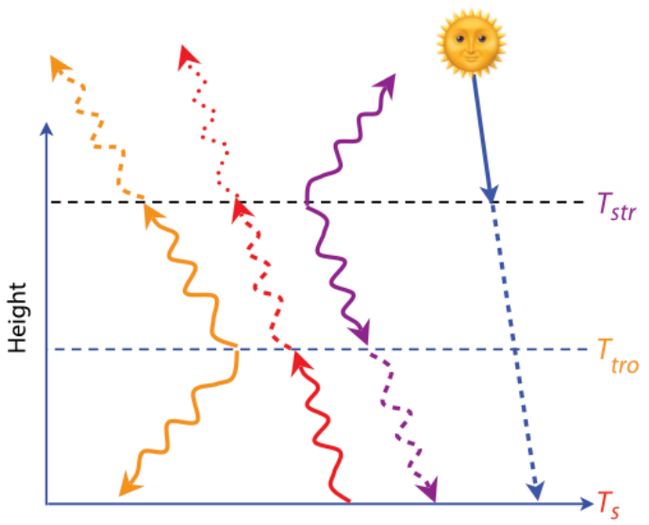
This leads to the very different stratospheric response to an increase in CO_2 concentration.

Figures 3.8a,b show the zonally averaged temperature change (C) in an RCP8.5 scenario during the 21st century as a function of height (a) and height and latitude (b), showing that while the troposphere warms, the stratosphere cools significantly.

To understand the response of the stratosphere, we add a third layer to our two-level energy balance model from earlier, such that T_s is the surface temperature, T_{tro} the tropospheric temperature, and T_{str} the newly added stratospheric temperature, as shown in the schematic Figure 3.9.

Figure 3.9: Understanding stratospheric cooling.

An energy balance model including a stratospheric layer, used to explain the stratospheric cooling in a global warming scenario.



We assume that a fraction β_{str} of the SW solar radiation is absorbed by the stratosphere. The energy balances of the surface, troposphere, and stratosphere can then be written with the energy sources on the LHS and sinks on the RHS as

$$
\begin{aligned}
(1-\beta_{str})\tfrac{1}{4}S_0 + \epsilon_{tro}\sigma T_{tro}^4 + (1-\epsilon_{tro})\epsilon_{str}\sigma T_{str}^4 &= \sigma T_s^4 \\
\epsilon_{tro}\sigma T_s^4 + \epsilon_{tro}\epsilon_{str}\sigma T_{str}^4 &= 2\epsilon_{tro}\sigma T_{tro}^4 \\
\beta_{str}\tfrac{1}{4}S_0 + \epsilon_{str}(1-\epsilon_{tro})\sigma T_s^4 + \epsilon_{str}\epsilon_{tro}\sigma T_{tro}^4 &= 2\epsilon_{str}\sigma T_{str}^4. \quad (3.6)
\end{aligned}
$$

In these equations, each term represents a heat flux in W/m^2, and all temperatures are in Kelvin.

The terms involving represent incoming ¼S_0 solar radiation, as explained earlier; ϵ_{tro} is the tropospheric emissivity and the stratospheric.

The different terms may be interpreted using the schematic arrows in Figure 3.9.

For example, the solid red arrow shows the LW radiation from the surface at a rate of σT^4_s.

When it encounters absorption in the troposphere, the red arrow changes from solid to dashed, and a heat flux $\epsilon_{tro}T^4_s$ is deposited there.

Then, the remainder surface emission flux, $(1 - \epsilon_{tro})\sigma T^4_s$, continues upward, as represented by the dashed red arrow.

The line changes again in the stratosphere from dashed to dotted, corresponding to the deposition of a heat flux $\epsilon_{str}(1 - \epsilon_{tro})\sigma T^4_s$ there as well, with the remainder continuing to outer space (dotted red arrow).

These equations can be written as an easily solved linear set of equations for the temperatures raised to the fourth power,

$$\begin{matrix} \sigma & -\epsilon_{tro}\sigma & -(1-\epsilon_{tro})\epsilon_{str}\sigma \\ -\epsilon_{tro}\sigma & 2\epsilon_{tro}\sigma & -\epsilon_{tro}\epsilon_{str}\sigma \\ -(1-\epsilon_{tro})\epsilon_{str}\sigma & -\epsilon_{str}\epsilon_{tro}\sigma & 2\epsilon_{str}\sigma \end{matrix} \begin{matrix} T_s^4 \\ T_{tro}^4 \\ T_{str}^4 \end{matrix} = \begin{matrix} (1-\beta_{str})\frac{1}{4}S_0 \\ 0 \\ \beta_{str}\frac{1}{4}S_0 \end{matrix} .$$

As a crude qualitative numerical example, consider $\beta_{str} = 0.05$ (the stratosphere, in the range of 1–100 hPa, absorbs about 5–15 W/m^2 depending on latitude and season), $\epsilon_{tro} = 0.55$, $\epsilon_{str} = 0.1$, and, for an increase in CO_2, let $\epsilon_{tro} = 0.6$, $\epsilon_{str} = 0.15$.

The solution before the increase in CO_2 is $T_s = 302K$, $T_{tro} = 257.4K$, and $T_{str} = 260.1K$.

The solution with the emissivity values representing the increased CO_2 concentration is $T_s = 306K$, $T_{tro} = 261.8K$, and $T_{str} = 254K$, showing the warming of the surface and the troposphere and the cooling of the stratosphere, as expected.

To understand the reason for this cooling, consider first the two-level energy balance model of equation (2.3) from earlier.

As CO_2 concentration increases, the tropospheric emissivity (denoted ϵ there) increases, increasing both the source and sink terms in the tropospheric heat budget (ϵ drops out of that equation, appearing on both sides).

For the surface equation, the increased atmospheric emissivity increases the heat source of downward LW and thus leads to surface warming.

The atmosphere then responds to the increase in emission from the warmer surface and warms up as well (while this is a helpful idealization in the current context, remember the refinements discussed earlier involving the lapse rate and emission height).

Adding a stratospheric layer as the third equation in (3.6) does not qualitatively change this response of the troposphere and surface (first two equations).

However, for the stratosphere, because of the presence of the SW absorption heating term (β_{str} ¼ S_0), the stratospheric emissivity does not drop out of the third equation in (3.6).

The LW cooling on the RHS is balanced by both LW absorption from the lower atmosphere and surface and by SW absorption.

When the stratospheric emissivity increases, say by *P%*, due to the increase in greenhouse gas concentration, the cooling term on the RHS increases by this percentage.

However, in the source terms on the LHS, only the two LW radiative heating terms increase by *P%*, while the SW heating does not increase with the increased stratospheric emissivity.

As a result, the sum of the heating terms on the LHS increases by a smaller fraction than the radiative cooling term $2\epsilon_{str}\sigma T^4_{str}$ on the RHS, leading to a net cooling of the stratosphere with increasing CO_2.

3.4.1 Detection and attribution

A successful *detection* of a signal due to anthropogenic climate change (say, in global mean surface temperature, stratospheric temperature, or Arctic sea ice area) requires showing that the observed change is inconsistent with internal/natural variability.

An *attribution* of observed events (say, of the strong Arctic sea ice melting in 2012 or the Siberian heat wave of 2020 or the California forest fires of 2020) to climate change due to greenhouse gas increase specifically requires further showing that the observed change is *only* consistent with the suggested forcing scenario (e.g., natural variability plus increase in greenhouse forcing) and is *not consistent* with alternative, plausible explanations of recent climate change due to either anthropogenic or natural sources (e.g., a change in ozone concentration, aerosols, or land use or a response to volcanoes).

A stratospheric cooling was predicted as early as 1967 to be an expected consequence of CO_2 increase, and combined with a warming of the troposphere, it is a clear signature of CO_2-induced climate change.

This is therefore an important example of the attribution of observed atmospheric temperature changes to the increase in greenhouse gas rather than to other anthropogenic effects (e.g., ozone depletion) or natural causes (e.g., volcanoes).

4 - SEA LEVEL

Global mean sea level (GMSL) has increased by about 35 cm over the past 150 yr (Figure 4.1)

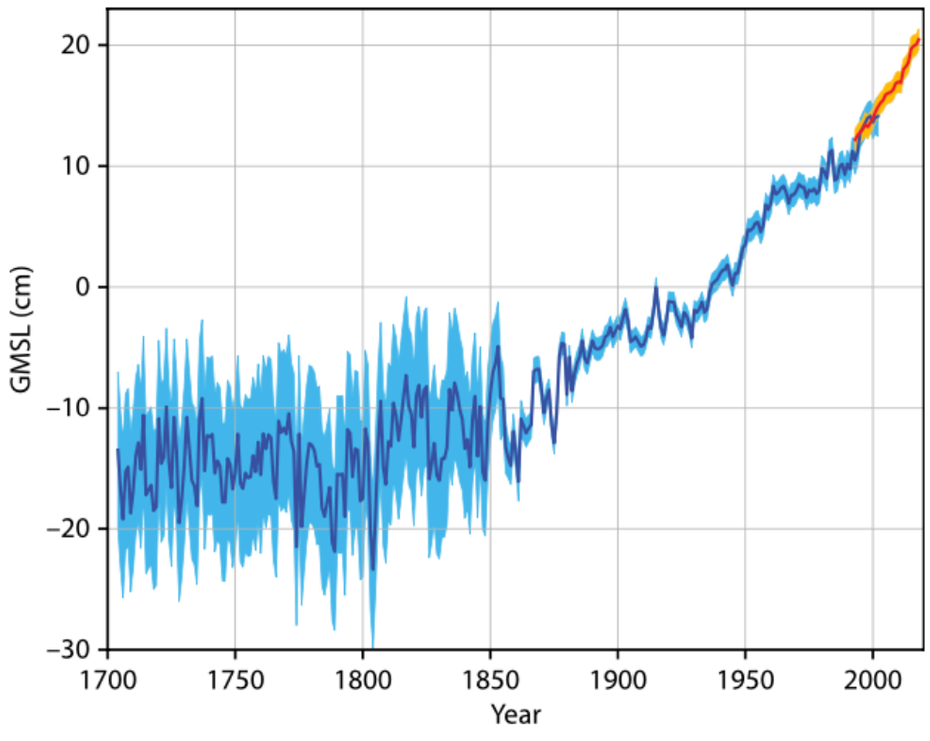


Figure 4.1: Observed global mean sea level rise.

Observed global mean sea level anomaly (defined as the deviation from a specified reference value) since 1700 (blue curve) and the estimated uncertainty (light-blue shading). The recent satellite record and its uncertainty range are similarly shown in red and orange.

While some of this increase is likely related to the natural warming due to the exit from the Little Ice Age (extending roughly from 1300 to about 1850), the signature of anthropogenic climate change on sea level is clearly seen.

Recent events, from the flooding of Manhattan during Hurricane Sandy in 2012 to the flooding of Venice, Italy, in 2019 that involved a 1.8 m local sea level rise, brought sea level to the forefront of the news.

We will see that both natural processes and anthropogenic climate change can contribute to such events.

Additionally, evidence from past warm climates suggests the potential for a significant sea level rise in response to fairly modest greenhouse warming (Box 4.1).

Sea level changes involve a wide range of regional and global processes, and we attempt here to understand and quantify the main ones.

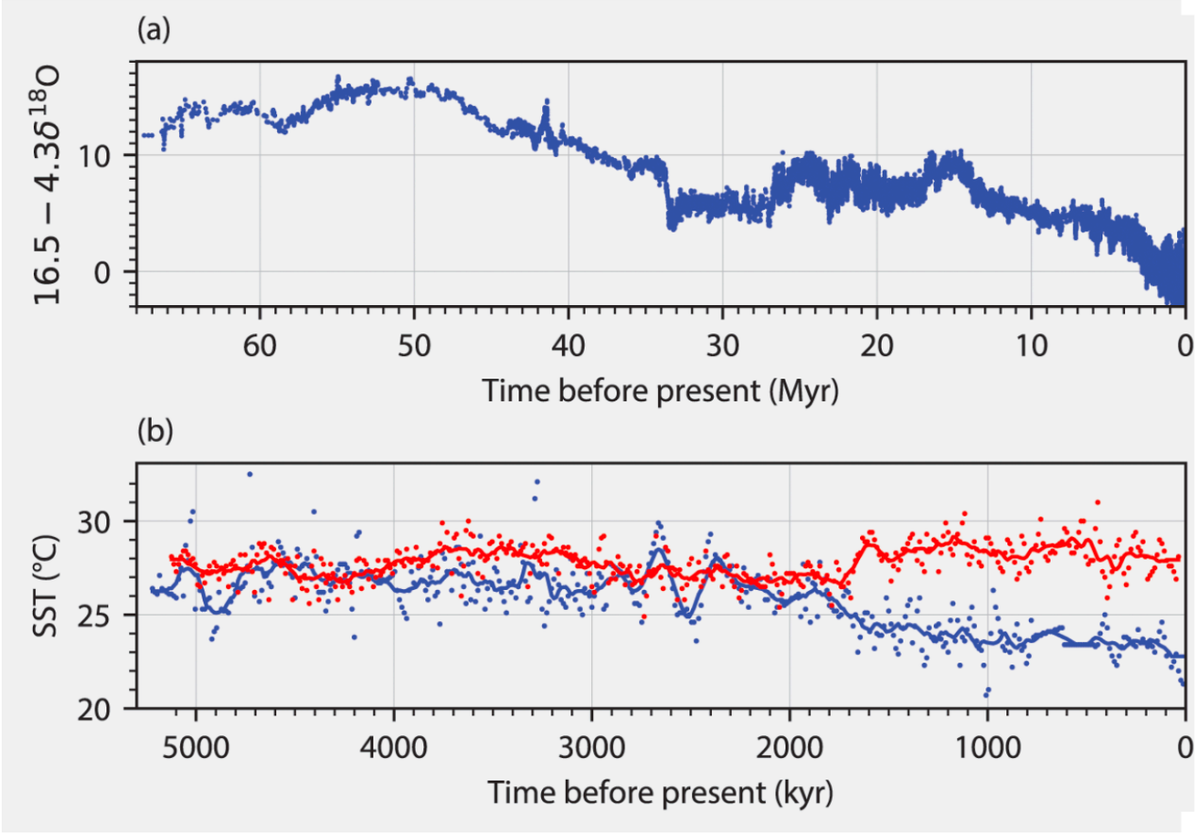
Climate Background Box 4.1

Past warm climates

An important perspective on possible future climate response to warming, including of sea level, can be obtained by examining past warm climates.

Panel a in the figure below shows a proxy of the deep ocean temperature over the past 60 Myr, showing the gradual cooling since the very warm *hot-house* climate of the Eocene (56–33.9 Myr).

Past ocean temperatures may be deduced from extracted deep sea sediment cores that contain the shells of plankton deposited at the ocean bottom for millions of years. The isotopic composition of these shells (e.g., $\delta^{18}O$, which is a function of the ratio between the concentrations of oxygen isotopes ^{16}O and ^{18}O) depends on the temperature in which the plankton grew, thus recording past temperatures.



The vertical axis in panel a is a function of the oxygen isotopic composition of plankton shells from deep ocean sediments, converted to a Celsius-like scale for convenience.

There is fossil evidence of frost-intolerant species of plants and animals existing during the Eocene in areas such as present-day northern North America, where temperatures currently often reach -40°C in winter.

This suggests that surface winter temperatures at that time did not drop below freezing, even in high-latitude continental areas away from the moderating effects of the ocean.

Similarly, deep ocean temperatures may have been as warm as 12°C, as opposed to around 0°C at present, indicating that the high-latitude areas where deep ocean water sinks during winter to fill the abyssal ocean (Box 6.1) were also warm.

During the Eocene, there was no significant ice over Antarctica nor Greenland, so that sea levels were more than 70 m higher than at present. Note the large scatter of dots in the past 2–3 Myr, an expression of the large variability due to the glacial cycles (Climate Box 4).

The last time atmospheric CO_2 concentration is believed to have been around 400 ppm, similar to that at present, was the Pliocene period, 5.3– 2.6 Myr, which is therefore considered the nearest past analog of a future global warming.

Global mean surface temperature was some 2–3°C warmer than at present, sea level has been estimated to have been up to 20 m above present levels, the Arctic was possibly 8°C or more warmer, and the equatorial Pacific may have been in a permanent El Niño state, as shown in panel b.

This panel shows reconstructed surface ocean temperature time series from the east and west Pacific, showing them to be separated by some 4°C all the way to about 2 Myr ago, and to be much closer prior to that. During present-day El Niño events (Climate Box 3), as the east Pacific warms, the east and west Pacific have a similar temperature, as seen in the Pliocene record during 2–5 Myr, hence the interpretation of a permanent El Niño.

Additionally, during the Pliocene some mid-latitude ocean upwelling sites (such as off the coast of California) were some 10°C warmer than today, suggesting that the upwelling was disrupted.

These upwelling sites are currently among the richest fisheries in the world ocean, because the upwelling brings nutrients to the surface, which supports the growth of plankton and attracts fish.

This Pliocene climate may represent an equilibrium response to present-day CO_2 levels, toward which our climate might be progressing assuming current CO_2 levels are sustained for a long time, as discussed in section 3.1.

These past warm climates are still not well understood and not well simulated by state-of-the-art climate models, indicating the need to improve our understanding and modeling of warm climate dynamics.

Sea level rises and falls on many timescales.

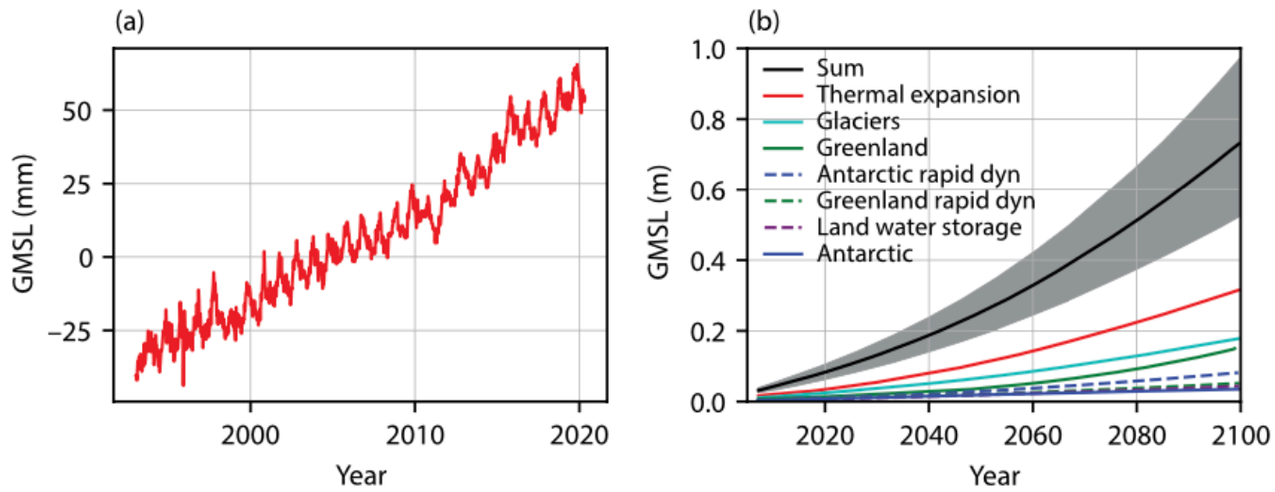
For example, GMSL rose 130 m since the last glacial maximum 21 kyr ago, due to the melting of ice sheets that covered the high-latitude continents during the last ice age (Climate Box 4).

More recently, global mean sea level has been rising at a rate of 3.5 mm/yr and seems to have accelerated in recent decades (Figures 4.1, 4.2a).

Future projections suggest that GMSL will continue rising, by up to 1 m by the end of the century under the Representative Concentration Pathway 8.5 scenario (Figure 4.2b).

Figure 4.2: Observed and projected global mean sea level change.

(a) Satellite record since 1992. (b) Future projections under RCP8.5, including the contribution of different processes, are discussed in this chapter. The black curve and gray shading show the projected total sea level rise and its uncertainty.



Starting with some definitions, the *global mean sea level* is a measure of the total water volume in the oceans.

Relative sea level (RSL) is the height of the sea surface above the solid ocean bottom at a particular location.

Steric sea level reflects the variation of the ocean volume due to ocean water density changes alone, induced by warming, for example.

In the following sections we first examine GMSL changes due to processes such as glacier and ice sheet melting, acceleration of ice flow from Greenland and Antarctica, expansion of warming ocean water, and more (see later discussion).

Local RSL due to wind or ocean current changes, local land uplift or sinking of the coast, coastal land erosion, and the gravitational effect of melting in Greenland and Antarctica will be examined later.

4.1 GLOBAL MEAN SEA LEVEL CHANGES

Processes that affect GMSL include, although are not limited to, (1) thermal expansion, (2) glacier and ice sheet mass balance, and (3) land water storage.

4.1.1 Thermal expansion

As discussed earlier, about 90% of the excess radiative forcing due to the increase in greenhouse gas concentration has so far been stored in the warming oceans.

This causes seawater to expand, contributing to GMSL rise.

Fresh water freezes at 0^{o}C, and its density as a function of temperature peaks at 4^{o}C.

However, for salty ocean water, the freezing temperature is -1.8^{o}C, and its density decreases with temperature above this freezing point.

The typical vertical ocean temperature profile is characterized by a few different regimes (Box 6.1).

The upper 25–50 m or so is typically well-mixed by wind and waves and is termed the *mixed layer*, where temperature is nearly uniform as a function of depth.

Below the mixed layer, the temperature decreases within the *main thermocline* down to about 1 km depth, below which the temperature is low and nearly uniform.

When the atmosphere warms, the mixed layer quickly warms within months, the above-thermocline ocean warms up over decades, while the deep water may take hundreds to thousands of years to warm.

Based on the observed warming of the oceans for the period 1971–2010 and their deduced thermal expansion, the rate of GMSL rise due to the expansion of ocean water from the surface to 700 m depth is estimated at about 0.6 mm/yr, and the rate becomes 0.8 mm/yr if the deeper ocean is included.

By the end of the century, warming according to the RCP8.5 scenario will lead to a sea level rise due to thermal expansion of over 30 cm, a significant fraction of the anticipated rise (Figure 4.2b).

A back-of-the-envelope calculation

Consider a warming of the upper ocean and the associated sea level rise due to the thermal expansion of seawater.

Initially, d_0 is the depth of the water column experiencing warming, its temperature is denoted T_0, and its initial density is $\rho(T_0)$.

After experiencing a warming ΔT, the water has a temperature $T = T_0 + \Delta T$, an expanded depth $d = d_0 + \Delta d$, and a reduced density $\rho(T_0 + \Delta T)$.

Given that the mass of ocean water is not changed by the warming, we can write an equation stating that the mass (per unit area) before warming is equal to that after the warming,

$$m = d \times \rho\,(T_0 + \Delta T) = d_0 \times \rho\,(T_0),$$

which implies that $d = d_0\,\rho(T_0)/\rho(T_0 + \Delta T)$, or

$$\begin{aligned} \Delta d &= d - d_0 = d_0 \left(\frac{\rho(T_0)}{\rho(T_0+\Delta T)} - 1 \right) \\ &= d_0 \frac{\rho(T_0) - \rho(T_0+\Delta T)}{\rho(T_0+\Delta T)}. \end{aligned}$$

To a good approximation, $\rho(T_0 + \Delta T)$in the denominator may be replaced by a constant reference (averaged) density ρ_0, giving

$$\Delta d = d_0 \frac{\Delta \rho}{\rho(T_0+\Delta T)} \approx d_0 \frac{\Delta \rho}{\rho_0}, \qquad (4.1)$$

so that sea level rise is proportional to d_0, the depth of penetration of the warming.

As an example, consider a uniform warming of $\Delta T = 3^oC$ corresponding to a mid-range IPCC surface warming due to CO_2 doubling.

Assume this warming penetrated to a depth of 500 m, and let the temperature before the warming be $T_0 = 16^oC$.

The ocean water density depends on temperature, as mentioned above, but also on the salt concentration (salinity, measured in parts per thousand, ppt, roughly representing kg salt per metric ton of seawater).

Let the density before warming be $\rho(16^oC,35\ ppt) = 1025.75\ kg/m^3$, and the density after warming be $\rho(16^oC,35ppt) = 1025.75\ kg/m^3$.

Note that these density values justify replacing $\rho(T_0 + \Delta T)$ with $\rho_0 = 1025\ kg/m^3$ in the denominator of equation (4.1) with a negligible effect on the accuracy of the calculated sea level rise.

Using the above expressions, we find a sea level rise of cm, similar to the IPCC estimate of sea level rise due to thermal expansion in the RCP8.5 scenario (red line in Figure 4.2b).

Calculating GMSL change from ocean warming

More generally, we need to take into account that ocean warming varies in both depth and the horizontal direction and that the expansion of seawater depends on its temperature.

Consider the *equation of state* relating ocean water density to temperature.

A good first approximation to this equation of state is that the density at a constant salinity S_0 changes linearly with temperature according to

$$\rho\left(T, S_0\right) = \rho_0\left(1 - \alpha_T\left(T - T_0\right)\right), \qquad \alpha_T = -\frac{1}{\rho}_0 \frac{\partial \rho}{\partial T}\Big|_{T=T_0},$$

where T_0 is the temperature prior to the warming, T the temperature after the warming, and α_T the *thermal expansion coefficient*, itself a function of temperature and evaluated here at T_0.

This is a linearization of the equation of state using Taylor expansion.

To calculate the change to GMSL, let $\Delta T(x,y,z) = T - T_0$ be the warming as a function of horizontal location and depth, divide the ocean into layers of infinitesimal thickness dz, and apply the above mass balance to each vertical layer in the ocean with dz replacing d_0 in equation (4.1).

Integrating over all layers from the ocean bottom at to the surface at z = 0 and over all horizontal locations (x,y), we find that the increase in ocean volume is

$$\Delta V = \int dx \int dy \int_{-H}^{0} \frac{1}{\rho_0} \Delta\rho\,(x, y, z) dz.$$

The change to GMSL is this change in volume divided by the ocean area A.

Using the fact that $\Delta\rho = \rho(\mathrm{T}) - \rho(\mathrm{T} + \Delta\mathrm{T}) \approx \rho_0 \alpha_\mathrm{T} \Delta\mathrm{T}(\mathrm{x,y,z})$, we find the change in GMSL to be

$$\begin{aligned} \Delta\,\mathrm{GMSL} &= \frac{1}{A} \int dx \int dy \int_{-H}^{0} \frac{1}{\rho_0} \Delta\rho\,(x, y, z) dz \\ &= \frac{1}{A} \int dx \int dy \int_{-H}^{0} \alpha_T \Delta T\,(x, y, z) dz. \end{aligned}$$

The dependence of the expansion coefficient on temperature is significant.

For example, assuming a salinity of 35 ppt, $\alpha_\mathrm{T}(2^\mathrm{o}\mathrm{C}) = 0.78 \times 10^{-4}\ ^\mathrm{o}\mathrm{C}^{-1}$, while $\alpha_\mathrm{T}(10^\mathrm{o}\mathrm{C}) = 1.6 \times 10^{-4}\ ^\mathrm{o}\mathrm{C}^{-1}$, reflecting the sensitivity of the density to the temperature: $\rho(2^\mathrm{o}\mathrm{C}) - \rho(3^\mathrm{o}\mathrm{C}) = 0.09\ \mathrm{kg/m^3}$, while $\rho(10^\mathrm{o}\mathrm{C}) - \rho(11^\mathrm{o}\mathrm{C}) = 0.18\ \mathrm{kg/m^3}$.

As a result, warming the already warm low-latitude surface ocean leads to a larger contribution to sea level rise than warming the colder high-latitude or deep waters.

4.1.2 Ice sheets and mountain glaciers

Greenland and Antarctica

Ice sheets gain mass via snow accumulation and lose it via *ablation*, which is the combination of all processes that lead to ice loss.

This includes the net surface mass balance (snow accumulation minus surface ice melting/sublimation), ice flow toward the ocean in ice streams, iceberg calving into the ocean, and more.

For the period of 1993 to 2010, the contribution to global mean sea level rise from Greenland was 0.4 mm/yr, and the contribution from Antarctica was 0.27 mm/yr, although it is currently difficult to estimate how much of this signal is due to natural variability and how much is due to anthropogenic climate change (as a reminder, GMSL is currently increasing by about 3.5 mm/yr).

Note that the RCP8.5 projections in Figure 4.2b include a contribution due to a possible acceleration of ice flow from Greenland and Antarctica.

This contribution is referred to as *rapid dynamics*, and is distinct from changes to surface accumulation or melting.

A warmer climate is expected to lead to more snow accumulation over Antarctica (reducing the rate of sea level rise) due to the increase in atmospheric water vapor with warming (Clausius-Clapeyron, Box 2.1).

At the same time, (uncertain) future projections anticipate increased ice flow, which is projected to dominate the expected increased snow accumulation by the end of the century.

The expected total contribution of Antarctica to GMSL is therefore positive, although less than the expected contribution of rapid dynamics alone.

For much more on the contribution of Greenland and Antarctica, and on the processes and uncertainty involved.

Mountain glaciers

Mountain glaciers have been retreating for about 150 yr, contributing to GMSL rise (for more on mountain glaciers, see chapter 11).

Some of the melting occurred before any significant anthropogenic warming and likely represents delayed climate response to exiting from the Little Ice Age.

There has been an acceleration in mountain glacier melting in recent decades reflecting the effect of anthropogenic warming, and the US National Snow and Ice Data Center has reported, for example, that the mass balance of 41 monitored glaciers has become more negative from 1980 to 2012.

The contribution to GMSL from glaciers excluding Greenland and Antarctica was 0.76 mm/yr during 1993 to 2010.

4.1.3 Land water storage

Mountain glaciers have been retreating for about 150 yr, contributing to GMSL rise.

Some of the melting occurred before any significant anthropogenic warming and likely represents delayed climate response to exiting from the Little Ice Age.

4.1.4 Detection of anthropogenic climate change in GMSL

Figure 4.1 leaves little doubt that there is a strong trend over the past century that is very distinct from the natural variability prior to 1850 or so.

Thus, one might conclude that the signs of climate change are clearly detected in the GMSL record.

While we will later examine a few quantitative statistical approaches to this detection issue, here we instead use a visualization approach, as formal statistics do not seem necessary in this case.

Figure 4.3 shows the sea level trend as a function of year and averaging period.

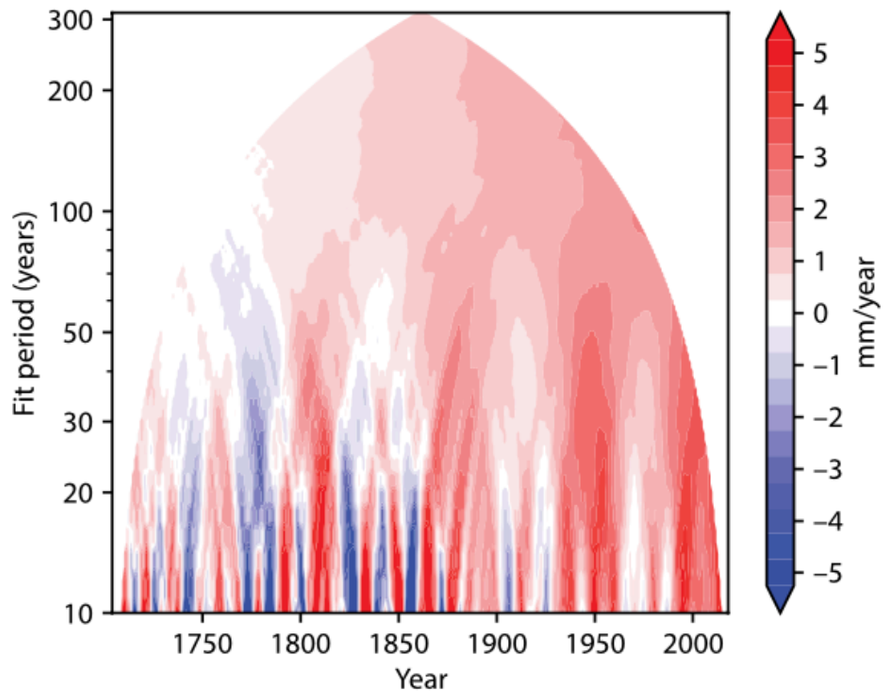


Figure 4.3: Detecting climate change in sea level rise.

Shown are the rates of sea level rise in mm/year as a function of year (horizontal axis) and averaging (fit) period (vertical).

Consider first the bottom of the plot, for averaging periods of 10 to 15 yr or so.

At the beginning of the record, the trends can be either positive or negative, alternating rapidly and representing natural variability and perhaps also noise due to larger errors in the less well observed earlier record.

But later, especially after 1990, the trends are uniformly positive, where the trend clearly masks any variability signal.

At averaging periods longer than about 30 yr, the right side of the plot is uniformly red, again showing that rise in GMSL dominates the variability.

4.2 REGIONAL SEA LEVEL CHANGES

Regional sea level changes can be much faster and larger than GMSL changes, and the associated local sea level drop or rise can be very dramatic.

The main factors that influence regional patterns of relative sea level change include atmosphere-ocean interaction (wind stress, atmospheric pressure loading, and ocean current changes), spatial differences in the rate of ocean warming, land changes due to coastal erosion, isostatic rebound of a continent in response to the melting of past land ice sheets, and the gravitational effect due to current melting of ice sheets.

4.2.1 Atmosphere-ocean interaction

Atmospheric pressure loading

Regional sea level adjusts to atmospheric pressure changes.

Atmospheric sea level pressure (SLP) exerts a vertical force on the ocean surface; as a result, a high SLP leads to a low sea surface height (SSH), and a low SLP to a high SSH.

For example, a low sea level pressure in the center of a hurricane—say, 50 hPa below normal SLP—can lead to a sea level rise of about 50 cm (roughly 1 cm per 1 hPa).

This effect is distinct from the wind-driven storm surge caused by storms, as discussed below.

In a warmer climate, the large-scale sea level atmospheric pressure is expected to change due to changes in atmospheric circulation, increase in the subtropics and mid-latitudes, and decrease at higher latitudes, especially over the Arctic.

In the RCP8.5 scenario this leads to a fairly modest, although consistent among different climate models, sea level rise of about 2.5 cm in the Arctic and in the Southern Ocean near Antarctica and a smaller sea level decrease spread over lower latitudes.

Smaller-scale changes in SLP involved in local weather conditions can be significantly larger.

To calculate the sea level response to SLP changes, consider an initial SLP that does not vary horizontally, consistent with a flat SSH and a motionless ocean (Figure 4.4a).

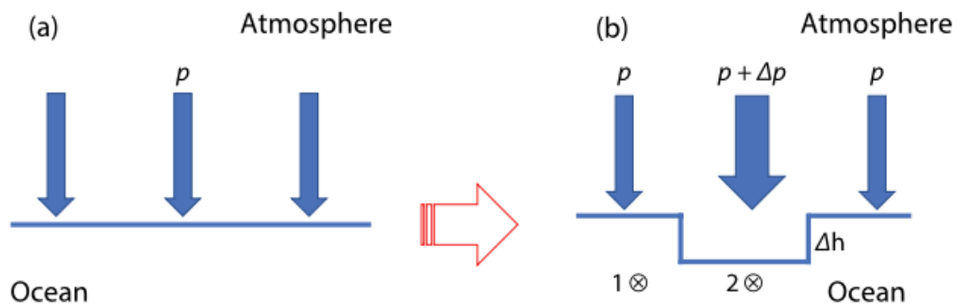


Figure 4.4: Atmospheric pressure loading.

How a change to sea level atmosphere pressure (SLP) from spatially uniform on the left to spatially variable on the right affects regional sea level.

A positive SLP perturbation , applied only regionally (Figure 4.4b), pushes water away, causing a sea level drop under the higher pressure and sea level rise elsewhere.

Horizontal pressure differences in the ocean lead to water movement, and for the new ocean state to be at equilibrium, the water pressure within the ocean should remain horizontally uniform.

The pressure below the ocean surface is determined by both the sea level atmospheric pressure and the sea surface height, which reflects the weight of ocean water above a given point.

For the ocean to be at rest, the changes to ocean pressure due to the sea level difference should balance the SLP perturbation.

The pressure difference between two horizontal locations within the ocean (e.g., those marked 1 ⊗ and 2 ⊗ in Figure 4.4b) due to an SSH difference Δh between these locations is the weight of water (per unit area) due to this difference, equal to $\varrho g \Delta h$, where ϱ is the ocean water density and g is gravity.

The condition that the total subsurface pressure within the ocean under the SLP perturbation is equal to the total subsurface ocean pressure away from there is therefore $\Delta p + \varrho g \Delta h = 0$, allowing us to calculate the SSH response Δh.

Writing $\Delta h = -\Delta p/(\varrho g)$, with $\varrho \approx 1025$ kg m^{-3} and g = 9.8 ms^{-2}, we find that an SLP signal of 50 hPa (5000 Pa) leads to a sea level change of 0.5 m, as mentioned above. This simple calculation does not take into account the Coriolis force due to the Earth rotation, discussed shortly.

Wind stress

Wind changes due to weather or climate variability, or due to climate change, can lead to a significant regional sea level response.

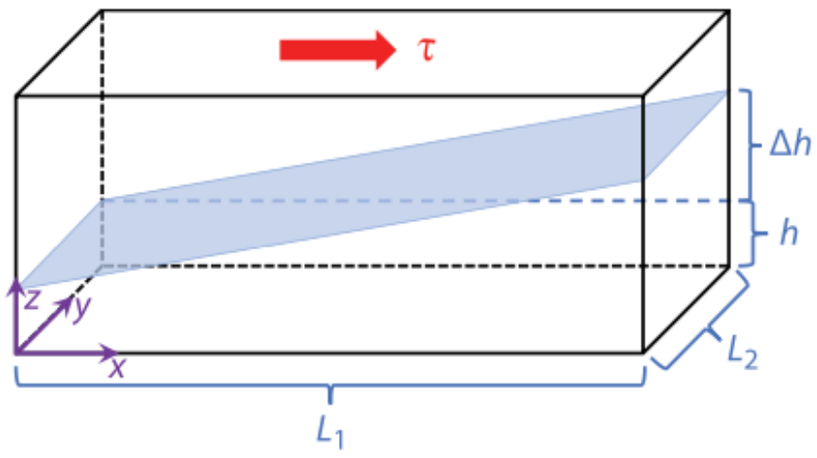
A near-surface wind of velocity U (m/s), typically measured at 10 m above the ocean surface, results in a horizontal force per unit area (*stress*, in units of N/m^2) on the ocean surface.

Consider an ocean channel, crudely representing the Adriatic Sea, for example, of length L_1 and width L_2.

Figure 4.5 shows the wind stress τ pushing the water in a channel to the right, creating a sea level height difference Δh across the length of the channel.

Figure 4.5: Wind-driven sea level change.

Wind stress drives a sea surface height slope and affects sea level height.



The higher sea level on the right implies that the ocean pressure is higher there as well, exerting a push on the water toward the left.

When the two forces due to the wind and pressure exactly balance, the water level is in a new equilibrium, allowing us to calculate the equilibrium SSH response to the wind.

The wind force over the ocean is given by the *bulk formula* $\tau = \rho_{air} C_D |U| U$, where the force is quadratic in the wind speed, and the use of absolute value of the wind speed times the wind speed itself guarantees that the sign (direction) of the force is the same as that of the wind.

Here, surface air density is $\rho_{air} = 1.2$ kg/m^3, and C_D is a nondimensional *drag coefficient*, with a typical value of 0.0013.

The total wind force over the channel surface is $F_{wind} = \tau L_1 L_2$.

The pressure at a depth z in the ocean may be calculated as the weight per unit area of water over that level, $p(z) = \rho g(h - z)$, where h is the local sea surface height, and z is the vertical coordinate, increasing upward and set to zero at the bottom, so that $h - z$ is the height of the water column above a depth z.

Assuming that the pressure at a given depth is the result of the water weight above that point is referred to as the *hydrostatic balance*.

Integrating over the cross-section area perpendicular to the wind direction, in the (y,z) plane, the horizontal pressure force on this section is $L_2 \int_0^h \rho g(h-z)dz = \frac{1}{2} L_2 \rho g h^2$.

The net horizontal pressure force on the channel in Figure 4.5 is the pressure force from the left (where the ocean depth is h) minus that from the right (where it is $h + \Delta H$), or

$$F_{\text{pressure}} = \left[\frac{1}{2}\rho g h^2 - \frac{1}{2}\rho g (h+\Delta h)^2\right] L_2 \approx -\rho g h \Delta h L_2. \qquad (4.2)$$

In the last equation we have neglected terms proportional to the square of the height difference, $(\Delta h)^2$, because they are smaller.

Using the horizontal force balance $F_{wind} + F_{pressure} = 0$, we have $\tau L_1 L_2 = \rho g \Delta h L_2$, so that the sea level change across the channel is estimated as

$$\Delta h = \tau L_1 / \left(g \rho \bar{h}\right),$$

where $\tilde{h}$ is the averaged ocean depth.

Thus, a large SSH response can result from a small mean depth $\tilde{h}$, a large wind stress, or a long distance L_1 over which the wind blows.

This expression shows that a persistent wind due to a weather event over the Adriatic Sea could lead to a large sea level rise, as observed in Venice, Italy, in 2019 and in 1966.

It is very difficult to attribute such individual strong wind events to ACC; what can typically be done is to attempt to estimate to what degree the probability of a given wind event was enhanced due to ACC, as we demonstrate below in other contexts.

Therefore, in general, the detection of and attribution of regional sea level events is more challenging than for the GMSL record (discussed earlier).

Storm and tidal surges

A strong storm, such as a tropical cyclone, can affect regional sea level via both the pressure loading and wind stress effects discussed above, as well as via the forcing of large-amplitude ocean waves.

When combined with local high tides that happen to occur at the same time, these can lead to a very dramatic local sea level change, coastal flooding, and significant damage.

Local infrastructure meant to address storm and tidal surges can be overwhelmed when these effects are combined with longer-term changes due to GMSL rise, longer-term regional sea level rise associated with climate change, and changing storm strength due to climate change.

Ocean currents and the Coriolis force

On a large enough scale of tens of kilometers and above, the Coriolis force enters the ocean dynamics by which pressure loading and wind stress affect sea level response.

The Coriolis force acts on every object moving relative to the rotating Earth, to the right of the motion in the Northern Hemisphere and to the left in the Southern Hemisphere.

Due to the Coriolis force, ocean currents such as the Gulf Stream are accompanied by an SSH difference across the current (there is about a meter SSH difference between the two sides of the Gulf Stream!).

A change to ocean currents will change the height difference across the current and can therefore also lead to a regional SSH change.

This may be quantified by analyzing the force balance dominating ocean currents.

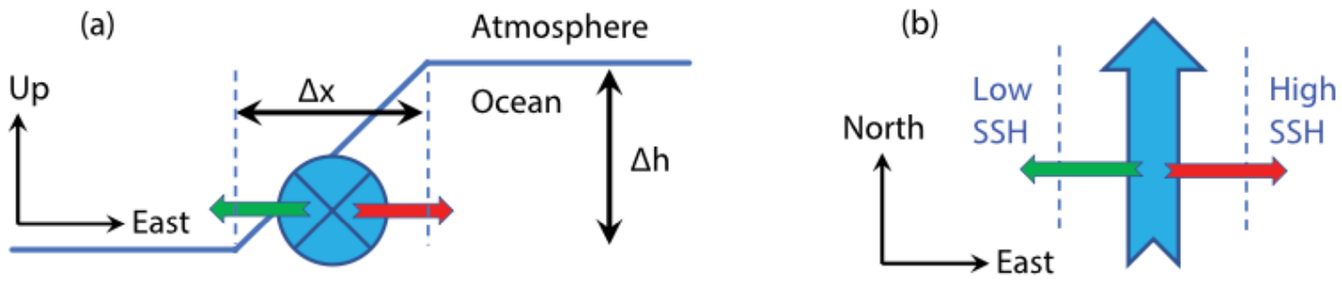
The dominant force balance for practically all major ocean currents (and large-scale atmospheric winds and weather systems) is between the Coriolis force and the pressure force.

These forces generally balance each other, such that the net force is zero; as a result the current is at a steady state and does not accelerate.

Consider a schematic east-west section across the northward flowing Gulf Stream (drawn as flowing into the page and denoted by the blue ⊗ symbol in the vertical section of Figure 4.6a and by the blue arrow in the plane view of Figure 4.6b).

Figure 4.6: The Coriolis force and sea level.

Sea surface height variations across the northward flowing Gulf Stream, showing the pressure force from high to low sea level (green arrows) and Coriolis force to the right of the current (red arrows). (a) A vertical east-west section (current is denoted by the blue symbol ⊗ of an arrow pointing into the page). (b) A horizontal schematic with the Gulf Stream shown by the blue arrow.



The SSH to the east of the Gulf Stream is about 1 m higher than that west of the stream.

This leads to a westward pressure force from high sea level to low sea level (green arrow).

This force is balanced by the Coriolis force: The Gulf Stream flows northward, and in the Northern Hemisphere this leads to a Coriolis force to the right of the flow—that is, eastward (red arrow)—that exactly balances the pressure force.

To quantify this discussion, consider an ocean current flowing northward (in the y direction) associated with an east-west sea level difference Δh (in the x direction).

The horizontal pressure force due to a height difference ΔH across the current, per unit distance along the current, is derived above (eqn 4.2) as

$$F_{\text{pressure}} \approx -\rho g h \Delta h. \qquad (4.3)$$

The acceleration tendency experienced due to the Coriolis force by an ocean current moving northward at a velocity v (m/s) is given by fv, where $f = 2\Omega \sin\theta$ is the Coriolis parameter (units of s^{-1}), the rotation rate of the Earth is given by $\Omega = 2\pi/\text{day} = 0.0000727 (s^{-1})$ and θ is the latitude.

The Coriolis force is this acceleration times the mass of the relevant ocean section (per unit distance along the current), $\Delta x h \rho$, where Δx is the width of the current and h the ocean depth.

That the Coriolis force and the pressure force sum to zero in an equilibrium implies that $0 = (\Delta x h \rho) fv - \rho g h$, or $-fv = -g\, \Delta h/\Delta x$.

Writing $\Delta H/\Delta x = \partial h/\partial x$, we find that the force balance of a northward flowing ocean current such as the Gulf Stream is

$$-fv = -g \frac{\partial h}{\partial x}.$$

The left-hand side represents the eastward acceleration driven by the Coriolis force on the northward flowing Gulf Stream, while the right-hand side represents the westward acceleration due to the east-west SSH gradient leading to a pressure force across the stream.

A similar balance between the pressure and Coriolis forces holds for other ocean currents and is referred to as *geostrophy*.

If a current velocity changes, so would the corresponding sea surface height gradient and as a result also the local sea level near the current, possibly affecting sea level at nearby coastal locations.

Ocean currents may respond to climate change, for example, due to a wind change driven by an atmospheric response to a warmer ocean temperature.

Substituting order-of-magnitude values for the Gulf Stream, of a velocity v = 1 m/s, width Δx = 100 km, Coriolis parameter $f = 2\Omega \sin(30^o) = 7.27 \times 10^{-5}$ s^{-1}, and g = 9.8 m/s^2, we find the height difference across the Gulf Stream to be about $\Delta h = fv\Delta x/g$ = 74 cm. current weakens by 10% due to surface wind changes associated with global warming, this will reduce the height difference across the current by a corresponding fraction, possibly leading to a sea level rise along the coast of North America.

For the sake of completeness, we note that if a current is flowing in the east-west *x* direction at a velocity *u*, its geostrophic balance can be similarly derived to be $fu = -g\,\partial h/\partial y$.

These expressions for geostrophy do not take into account the important effect of change in ocean water density with depth, which is beyond our scope here.

For the sake of simplicity, our discussion of the effects of wind stress on sea level height did not take into account the Coriolis force, although in reality the Coriolis force does play a role in how sea surface height responds to wind changes.

Finally, we note again that geostrophy is also the force balance of major wind patterns in the atmosphere, from weather systems to atmospheric jet streams.

4.2.2 Land changes

Coastal erosion and land subsidence

Coastal erosion is the loss or displacement of the upper part of the land surface adjacent to the coastline due to the action of waves, currents, tides, wind-driven surges, waterborne ice, precipitation runoff, and other impacts of storms.

This may lead to the retreat of the coastline landward and to an increase in relative sea level over the eroded coast.

The connection of specific erosion events to global warming, due for example to an expected strengthening of tropical storms, may be indirect and therefore uncertain.

Dramatic recent land erosion events occurred in Kutubdia, an offshore island in Bangladesh in 2012, due to tidal waves, high winds, and strong rainfall, in New Jersey and Long Island in response to Hurricane Sandy, and elsewhere.

In addition, land subsidence (lowering) in or adjacent to an ocean-covered area similarly leads to the increase in relative sea level rise.

Subsidence can be caused by the over-pumping of groundwater, by the compaction of the ground due to centuries of building, and by long-term changes due to plate tectonics.

Venice, Italy, is estimated to have subsided by about 120 mm in the 20th century, contributing about half of its observed sea level rise.

Of course, natural processes such as sediment transport by rivers or by waves and tides can also lead to the *buildup* of coastal areas and to the decline of relative sea level.

Such natural processes may be disrupted by coastal building and by the elimination of natural vegetation, adding to the effects of climate change.

Post-glacial rebound and isostatic adjustment

The lingering rising of the solid Earth after the melting of a land ice sheet or glacier can last thousands of years; therefore, the influence of the melting of the large ice sheets of the last glacial maximum (21 kyr; see Box X.2) still affects sea level changes observed today.

When a large ice mass forms on a continent, its weight presses the continental surface down, leading to a subsurface flow within the upper mantle away from the glacier until a new equilibrium is reached.

Given that ice density (900 kg/m) is about a third of that of the solid Earth, a third of the height of the ice is submerged below the Earth surface at equilibrium (Archimedes' law).

The response of the continent to the presence or melting of ice sheets is referred to as isostatic adjustment.

When the ice melts away, the continent rebounds, and there is a return subsurface flow within the upper mantle.

As a result, the continent rebounds near the location of the melting ice sheet and sinks away from there.

This leads to a downward trend in the relative sea level close to the location of the ice sheet and to an opposite trend away from it.

These sinking land trends can lead to a *regional* sea level increase of about a millimeter per year along parts of the Atlantic coast of the United States, and need to be taken into account when interpreting local sea level records for the signal of global warming.

This effect is distinct from the more straightforward effect on the *global* sea level of the melting of the ice sheet and the flow of its melt water to the ocean.

4.2.3 Gravitational effects: sea level fingerprints of melting

A sea level gravitational fingerprint of melting is a spatial pattern of the response of sea level to ice sheet melting due to gravitational effects.

Consider a melting of the Greenland ice sheet.

Because of the addition of water mass to the ocean, GMSL will rise, *but the rise is not uniform* as a function of horizontal location.

Currently, the gravitational attraction by the mass of ice over Greenland means that sea level is increasing toward Greenland.

The reduction of ice mass due to melting results in less gravitational attraction of ocean water toward Greenland, which would tend to *reduce* sea level near Greenland. Since sea level near Greenland is reduced, sea level elsewhere is expected to rise more than the GMSL rise expected due to the melting.

This gravitational effect leading to local reduction of sea level in response to Greenland melting extends some 2000 km away from Greenland and therefore influences the sea level rise near parts of Northern Europe.

Figure 4.7 shows the fingerprints of melting in Greenland and west Antarctica.

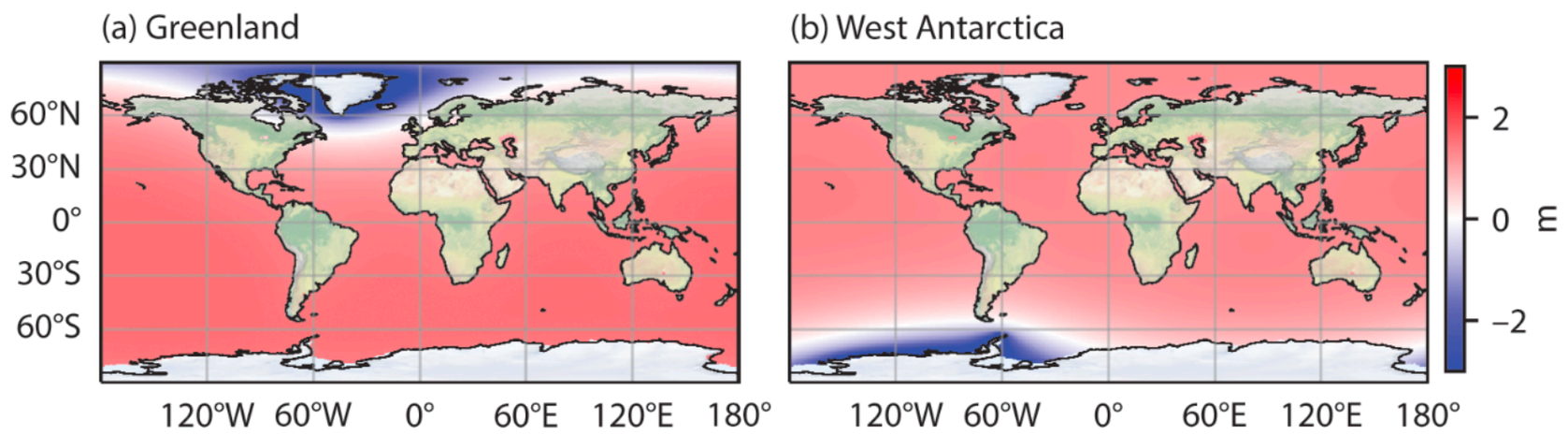


Figure 4.7: Gravitational effects.

Sea level fingerprint of melting in (a) Greenland, (b) West Antarctica, showing distribution of sea level rise in meters, where the GMSL rise is 1 m in both cases.

In parallel, the reduction of ice weight will make the underlying continent adjust upward elastically over a short period and further adjust slowly due to isostatic adjustment over a longer timescale.

The loss of ocean water near Greenland due to the weakening of the gravitational force will make the sea floor adjust upward as well.

On the other hand, the addition of water away from Greenland will push the sea floor there downward. The resulting spatial pattern of sea level change is the fingerprint of Greenland melting and is different from the fingerprint of Antarctic melting.

Due to these processes, a rate of Greenland melting that corresponds to a 1 mm/yr rise in GMSL is expected to lead to sea level drop within 2000 km of Greenland and at the shore of Greenland itself can lead to a *drop* of up to 10 mm/yr, 10 times larger than the global rate of opposite sign!

5- OCEAN ACIDIFICATION

About a quarter of the CO_2 emitted by human activity has been absorbed by the world oceans, and another quarter by soils and the terrestrial biosphere, with the rest accumulating in the atmosphere.

While this ocean absorption delays atmospheric warming, it leads to an acidification of the world oceans, referred to as "the other CO_2 problem."

That the already observed non-negligible acidification of the world ocean will be followed by a significant further acidification is a robust prediction for a higher CO_2 world.

The consequences for ocean biology are more difficult to predict, but it is difficult to imagine how the anticipated significant changes to ocean acidity would not lead to a disruption to ocean ecosystems.

The acidity of the ocean is measured in terms of its pH, which is defined as the negative of the of the *molar concentration* (moles per liter) of H ions in seawater.

Figure 5.1 shows the observed surface ocean pH since 1850, as well as that projected under the RCP8.5 scenario.

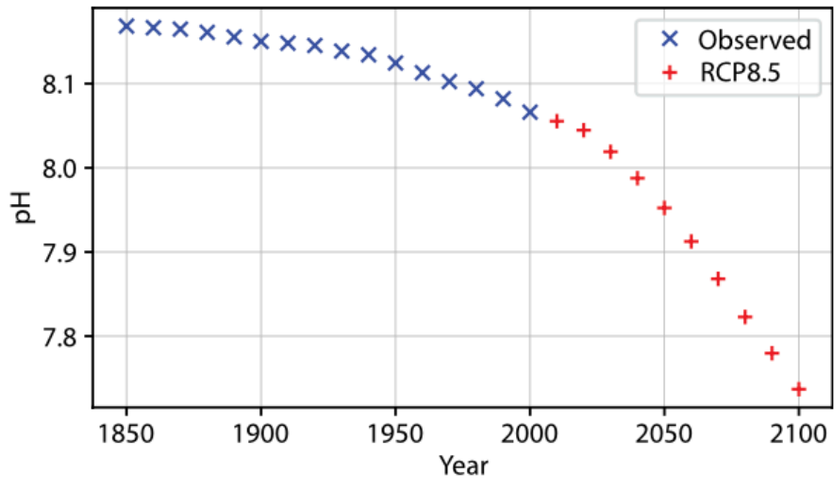


Figure 5.1: Ocean pH changes.

Observed globally averaged surface ocean pH from 1850 to 2000 (blue), and the projected pH to the end of the 21st century based on the RCP8.5 scenario (red).

The observed reduction in pH is from 8.16 to 8.06, which implies an increase of $100x(10^{-8.06}-10^{-8.16}/10^{-8.16} = 25\%$ in the concentration of H+ ions.

When CO_2 dissolves in seawater, it reacts with water to form carbonic acid, H_2CO_3, making the water more acidic (lower pH).

This, as we will see, increases the concentration of bicarbonate (HCO_3^-) ions and decreases that of carbonate (CO_3^{2-}) ions.

Multiple marine organisms deposit calcium carbonate ($CaCO_3$) structures as one of two minerals (crystal forms) known as aragonite and calcite, with the first being more soluble.

This deposition becomes more difficult as the concentration of the carbonate ion decreases with decreasing pH.

Among the marine organisms affected are single-celled phytoplankton such as coccolithophores and the common single-celled foraminifera, as well as pteropods, tiny (< 1 cm long) marine snails that are at the base of the oceanic food chain and whose shell is expected to begin to dissolve at acidity levels anticipated within this century, according to some high-emission scenarios.

Furthermore, the increased acidification expected by year 2100 is suggested to lead to reduced olfactory functioning of fish and possibly disrupt their ability to detect prey and to avoid predators.

Similarly, some studies suggest an effect of acidification on fish retinal function and vision.

The solubility of calcium carbonate depends on pH but also on temperature and pressure.

Due to these dependencies, the deep ocean is *undersaturated* to the forms of calcium carbonate deposited by marine organisms, which means that it is already sufficiently acidic to dissolve these calcium carbonate structures.

The upper ocean is *supersaturated* with respect to these forms; they therefore do not dissolve there.

The transition zone between the dissolving deep ocean and the upper ocean has risen by 50–200 m in the past 150 yr due to the increase in atmospheric CO_2 concentration.

Because cold water tends to be more undersaturated, the effects of ocean acidification are expected to be felt first in the colder higher latitudes and in the deep ocean.

As the ocean warms, its solubility to CO_2 decreases, reducing its ability to absorb anthropogenic CO_2 emissions and thus acting as a positive feedback on anthropogenic warming.

Understanding and predicting this effect is important to our ability to predict warming later this century.

Assuming anthropogenic emissions cease at some point, the emitted CO_2 will eventually be eliminated and the natural state recovered, although this can be expected to occur on a timescale of thousands to tens of thousands of years.

In the following, we attempt to understand calcium carbonate ($CaCO_3$) dissolution, describe the ocean carbonate system that controls this dissolution, understand this system by simplifying it to a set of reactions that can be solved explicitly, and then use this understanding to explore the response of pH and calcium carbonate dissolution to increasing atmospheric CO_2 concentrations.

The mechanisms and consequences of decreasing CO_2 solubility in a warming ocean and the restoring of the natural level of CO_2 over tens of thousands of years after emissions ceased are discussed later.

5.1 CALCIUM CARBONATE ($CACO_3$) DISSOLUTION

Many marine organisms produce calcium carbonate shells or skeletons in the form of calcite or aragonite and may be affected by a lower ocean pH.

The equilibrium reaction between solid calcium carbonate and the ions it is composed of is written as

$$\mathrm{CaCO_3} \rightleftharpoons \mathrm{Ca^{2+}} + \mathrm{CO_3^{2-}},$$

and the equilibrium concentrations of the ions are related via the solubility product,

$$K_{sp} = \left[\mathrm{Ca^{2+}}\right]\left[\mathrm{CO_3^{2-}}\right], \qquad (5.1)$$

where the brackets denote the molar concentration of a given ion.

The solubility product constant itself is the equilibrium constant for a solid substance dissolving in an aqueous solution and represents the concentration levels at which a solute dissolves.

The more soluble a substance is, the higher the value is.

This constant for calcium carbonate depends fairly strongly on temperature and pressure (depth in the ocean), as shown in Figure 5.2.

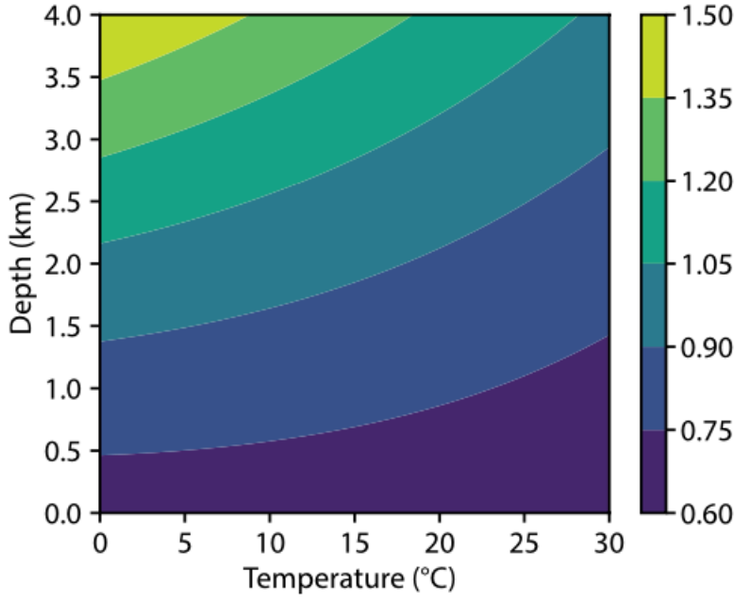


Figure 5.2: Solubility constant.

The calcium carbonate (specifically, aragonite) solubility product constant, K_{sp} (in $(mol/l)^2 \times 10^6$) as a function of temperature and depth (reflecting ocean pressure).

Given the solubility product, the potential for dissolving $CaCO_3$ is measured by the *saturation state* omega (Ω) defined as

$$\Omega = \frac{[Ca^{2+}][CO_3^{2-}]}{K_{sp}}. \qquad (5.2)$$

When $\Omega < 1$, $CaCO_3$ tends to dissolve.

As ocean pH decreases due to the increase in atmospheric CO_2, the carbonate ion concentration, $[\,CO_3^{2-}]$, decreases as well, making Ω smaller and leading toward more dissolution of calcium carbonate.

In the following, we attempt to understand the link between atmospheric CO_2, pH, and CO_3^{2-} by formulating and solving the equations for the ocean carbonate system.

Climate Background Box 5.1

The Carbon Cycle

Carbon is exchanged between different reservoirs in the climate system, including in the deep ocean (which contains about 38,000 gigatons carbon, GtC), surface ocean (1000 GtC), atmosphere (750 GtC), biosphere (600 GtC), and soils (1600 GtC).

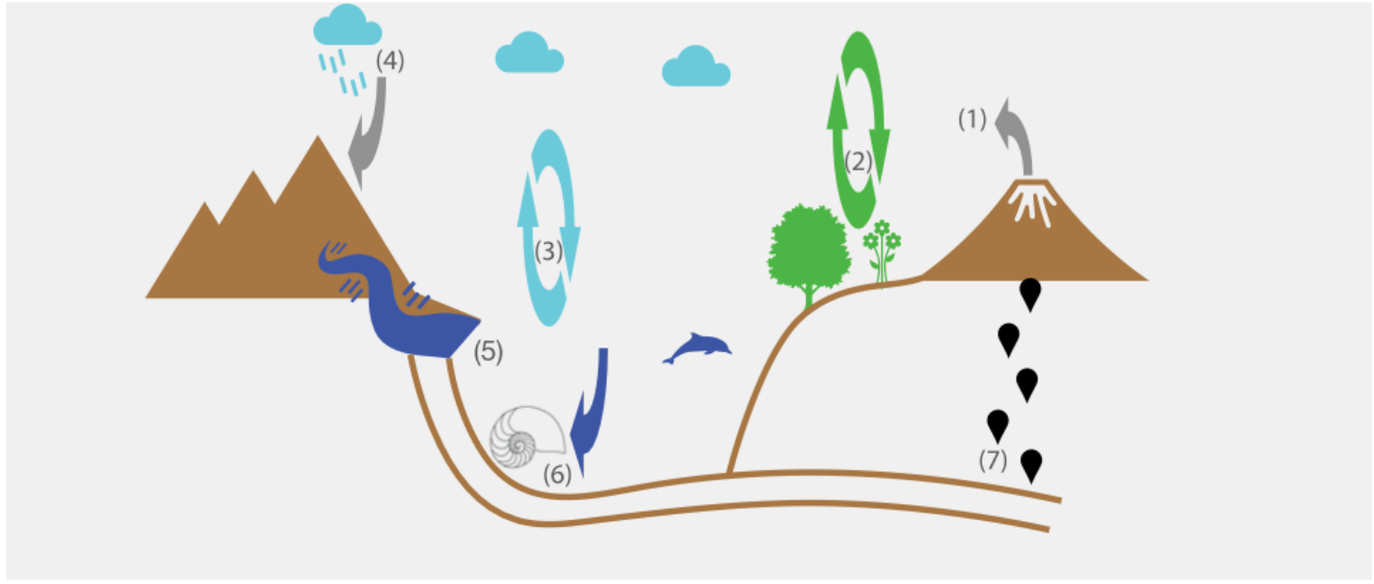
The fluxes between these components involve a fast exchange of large fluxes between the land biosphere/soils and the atmosphere, due to photosynthesis and respiration, of about 120 GtC/yr (2 in the accompanying figure) and between the ocean biology and the atmosphere (90 GtC/yr (3)).

In addition, the source of carbon in the climate system, and of atmospheric CO_2 in particular, is a slow release by volcanic activity, at a rate of 0.07 GtC/yr (1).

There is also a slow sink due to the dissolution of CO_2 in rain (4), an interaction of the resulting solution with rocks and the formation of calcium carbonate that is dissolved in rivers (5), flows to the ocean, and is deposited in sediments there (6).

On geological timescales, the ocean floor with the calcium carbonate sediments subducts under continental margins due to plate tectonics, and the carbon is recycled into the CO_2 emitted by volcanoes (7).

Human emissions are currently about 10 GtC/yr, far outweighing the natural source due to volcanoes. With available fossil fuel reservoirs estimated at 5000 GtC, there is a potential for a significantly larger future disruption of the carbon cycle.



5.2 THE CARBONATE SYSTEM

Our objective is to find the relation between atmospheric CO_2, dissolved CO_2, CO_3^{2-}, ocean pH, and other related variables.

We start by introducing the relevant chemical reactions that are part of the ocean carbonate system.

Carbon dioxide is soluble in water, and its dissolution occurs in two steps.

First, Henry's law states that atmospheric CO_2(g) concentration is in equilibrium with dissolved CO_2(aq),

$$CO_2\,(g) \rightleftharpoons CO_2\,(aq),$$

and then its reaction with water is given by

$$CO_2\,(aq) + H_2O \rightleftharpoons H_2CO_3\,(carbonicacid).$$

Because it is difficult to distinguish between $CO_2(aq)$ and H_2CO_3, they are treated together as a single variable defined as

$$H_2CO_3^* \left(\equiv CO_2^*\right) \equiv CO_2\,(aq) + H_2CO_3. \quad (5.3)$$

In terms of this variable, Henry's law is

$$CO_2\,(g) \rightleftharpoons H_2CO_3^*. \quad (5.4)$$

Now, carbonic acid is a weak diprotic acid (diprotic acids are able to release two protons),

$$H_2CO_3^* \rightleftharpoons H^+ + HCO_3^- \quad (5.5)$$
$$HCO_3^- \rightleftharpoons H^+ + {CO_3}^{2-}. \quad (5.6)$$

Finally, water dissociation is given by

$$H_2O \rightleftharpoons H^+ + OH^-. \quad (5.7)$$

In order to solve for the six unknown concentrations of $CO_2(g)$, $H_2CO_3^*$, HCO_3^-, CO_3^{2-}, OH^-, H^+, we so far have only four reactions that will later be written as explicit equations: (5.4), (5.5), (5.6), and (5.7).

We therefore need to specify two more constraints.

One is mass conservation for the total number of carbon atoms, expressed via a quantity known as total dissolved inorganic carbon (DIC), also referred to as total CO_2 and denoted ΣCO_2 or C_T, which is conserved in the above reactions.

The other constraint is the conservation of electric charge, which again must be satisfied by the above reactions and is expressed via a parameter called *alkalinity* (*Alk*).

C_T and *Alk* are both measurable quantities for which one can write conservation equations affected by various sources and sinks, as well as by the movement and mixing of water masses.

Once DIC and alkalinity are specified at a given location in the ocean, the carbonate system is completely determined (that is, there are the same number of unknowns and equations) and the concentrations of the different ions can be calculated.

Consider these two constraints in some detail now.

Total CO_2

The number of moles carbon atoms per liter (MC/l) is given by the sum of the different species of the carbonate system,

$$\begin{aligned} \mathrm{DIC} \equiv \Sigma \mathrm{CO}_2 \equiv C_T &= \left[\mathrm{H_2CO_3^*}\right] + \left[\mathrm{CO_3^{2-}}\right] + \left[\mathrm{HCO_3^-}\right] \\ &\approx \left[\mathrm{CO_3^{2-}}\right] + \left[\mathrm{HCO_3^-}\right], \end{aligned}$$

where the partition in the ocean of the species appearing on the RHS of the first line is 1%, 10%, 90%, correspondingly (this partition is a strong function of pH), hence the approximation in the second line.

Alkalinity

The concept of alkalinity arises when one considers the charge balance of seawater.

The net charge needs to be zero, which, when we take into account the major ions in the ocean, means that the number of positive charges minus the number of negative charges should vanish,

$$\begin{aligned} 0 &= \left(\left[\mathrm{H^+}\right] + \left[\mathrm{Na^+}\right] + \left[\mathrm{K^+}\right] + 2\left[\mathrm{Mg^{2+}}\right] + 2\left[\mathrm{Ca^{2+}}\right]\right) \\ &\quad - \left(\left[\mathrm{HCO_3^-}\right] + 2\left[\mathrm{CO_3^{2-}}\right] + \left[\mathrm{OH^-}\right] + \left[\mathrm{Cl^-}\right]\right. \\ &\quad \left. + 2\left[SO_4^{2-}\right] + \left[\mathrm{NO_3^-}\right] + \left[\mathrm{HBO_3^-}\right]\right). \end{aligned} \tag{5.8}$$

For this purpose, it is useful to differentiate between strong bases and acids whose concentration does not change with pH and weaker ones that do change.

For example, when NaCl dissolves in seawater, it separates completely into Na^+ and Cl^- regardless of the pH.

However, in the dissociation of the weak acid HCO_3^-, $HCO_3^- \rightleftharpoons H^+ + CO_3^{2-}$, the concentrations of the ions on the RHS and LHS of this equilibrium vary with pH.

Alkalinity is a measure of the charge balance due to these weak acids and bases.

It is defined as the sum of negative ions that belong to weak acids or bases that change their dissociation with the ocean pH, minus the sum of positive ions that originate from such weak acids/bases.

Separating the charge balance (eqn 5.8) into the parts due to the weak acids and bases (first line) and strong ones (second and third lines), we have

$$\begin{aligned} 0 &= \left(\left[H^+\right] - \left[OH^-\right] - \left[HCO_3^-\right] - 2\left[CO_3^{2-}\right] - \left[HBO_3^-\right]\right) \\ &\quad + \left(\left[Na^+\right] + \left[K^+\right] + 2\left[Mg^{2+}\right] + 2\left[Ca^{2+}\right] - \left[Cl^-\right]\right. \\ &\quad \left. - 2\left[SO_4^{2-}\right] - \left[NO_3^-\right]\right). \end{aligned}$$

Defining the alkalinity to be the negative of the first line, we have

$$\begin{aligned} Alk &\equiv \left[\mathrm{HCO_3^-}\right] + 2\left[\mathrm{CO_3^{2-}}\right] + \left[\mathrm{HBO_3^-}\right] - \left[\mathrm{H^+}\right] \\ &= \left[\mathrm{Na^+}\right] + \left[\mathrm{K^+}\right] + 2\left[\mathrm{Mg^{2+}}\right] + 2\left[\mathrm{Ca^{2+}}\right] - \left[\mathrm{Cl^-}\right] \\ &\quad - 2\left[\mathrm{SO_4^{2-}}\right] - \left[\mathrm{NO_3^-}\right]. \end{aligned} \tag{5.9}$$

Alkalinity is measured in units of *Equiv*/l, which is the sum of ion concentrations (e.g., in micromoles per liter), each multiplied by their charge.

We will see below that the concentrations of H^+ and OH^- are very small, and we approximate the alkalinity by what is known as the *carbonate alkalinity*, as defined in the first line below, and then further approximate it as shown in the second line,

$$\begin{aligned} Alk_C &\equiv \left[\mathrm{HCO_3^-}\right] + 2\left[\mathrm{{CO_3}^{2-}}\right] + \left[\mathrm{OH^-}\right] - \left[\mathrm{H^+}\right] \\ &\approx \left[\mathrm{HCO_3^-}\right] + 2\left[\mathrm{CO_3^{2-}}\right]. \end{aligned} \tag{5.10}$$

5.2.1 Carbonate system equations

To perform calculations, we write the above carbonate system reactions as the following set of equations using the equilibrium constants.

The six unknowns are

$$\left[\mathrm{CO_2\,(g)}\right], \left[\mathrm{H_2CO_3^*}\right], \left[\mathrm{OH^-}\right], \left[\mathrm{H^+}\right], \left[\mathrm{HCO_3^-}\right], \left[\mathrm{CO_3^{2-}}\right], \tag{5.11}$$

where we remember that

$$\mathrm{H_2CO_3^*}\left(\equiv\mathrm{CO_2^*}\right)\equiv\mathrm{CO_2\,(aq)}+\mathrm{H_2CO_3}, \quad (5.12)$$

and the first four equations are

$$K_H\equiv K_0\,(T,S,p) = \frac{\left[\mathrm{H_2CO_3^*}\right]}{[\mathrm{CO_2(g)}]} \quad (5.13)$$

$$K_1\,(T,S,p) = \frac{[\mathrm{H^+}]\left[\mathrm{HCO_3^-}\right]}{\left[\mathrm{H_2CO_3^*}\right]} \quad (5.14)$$

$$K_2\,(T,S,p) = \frac{[\mathrm{H^+}]\left[\mathrm{CO_3^{2-}}\right]}{\left[\mathrm{HCO_3^-}\right]} \quad (5.15)$$

$$K_w\,(T,S,p) = \left[\mathrm{H^+}\right]\left[\mathrm{OH^-}\right]. \quad (5.16)$$

For typical values of these reaction coefficients on the LHS, which depend on temperature *T*, salinity *S*, and pressure *p*.

To close the system, we need two more equations, the definitions of carbonate alkalinity and total CO_2,

$$Alk_C = \left[\mathrm{HCO_3^-}\right]+2\left[\mathrm{{CO_3}^{2-}}\right]+\left[\mathrm{OH^-}\right]-\left[\mathrm{H^+}\right] \quad (5.17)$$

$$C_T = \left[\mathrm{HCO_3^-}\right]+\left[\mathrm{CO_3^{2-}}\right]+\left[\mathrm{H_2CO_3^*}\right]. \quad (5.18)$$

Given the values of Alk_C, C_T, K_H, K_1, K_2, and K_w, we can use the four carbonate system equations (5.13, 5.14, 5.15, 5.16) and the definitions of alkalinity and total CO_2(5.17, 5.18) to solve for the six unknowns (5.11).

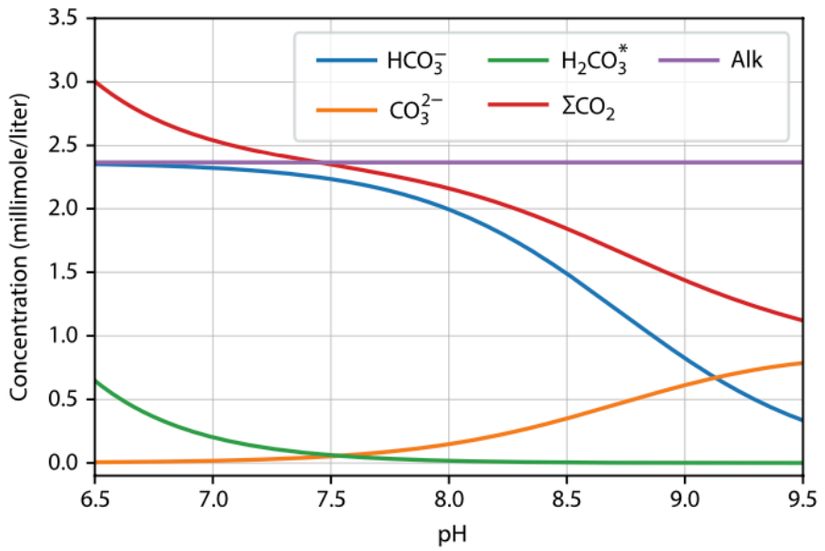
This is a nonlinear system of equations, and it may be solved numerically.

Note that instead of specifying Alk_C and C_T, we could have specified any two of the carbonate system variables, such as CO_2(g) or Alk_C and C_H and pH.

Once two are specified, the rest are calculated using the above equations.

The solution of the carbonate system as a function of pH for a constant alkalinity is shown in Figure 5.3.

Figure 5.3: The solution of the carbonate system, showing the concentration of carbonate species as a function of pH for a fixed alkalinity.



This figure was obtained by varying the DIC(C_T), specifying a fixed alkalinity, and solving for all variables for each value of C_T.

The calculated pH is then used as the *x*-axis coordinate, as is customary for such plots.

As a reminder, the current average surface ocean pH is about 8.1.

Note that the carbonate ion (orange line), whose concentration controls the dissolution of calcium carbonate via the saturation state (eqn 5.2), decreases significantly for lower pH values, while the concentration of the bicarbonate ion (blue) increases.

This solution provides the information regarding the response of the carbonate system as a function of pH required for us to calculate the ocean response to increased CO_2.

The solution used to plot Figure 5.3 also includes the value of the atmospheric CO_2(g) concentration that would have been in equilibrium with a volume of water with each particular value of the prescribed DIC.

We use this to plot pH and the carbonate ion concentration as a function of CO_2(g) in Figure 5.4, showing a significant reduction of the carbonate ion concentration as the atmospheric CO_2(g) increases.

An even better understanding of the system can be developed by considering an approximate set of carbonate system equations that can be solved directly, as we do next.

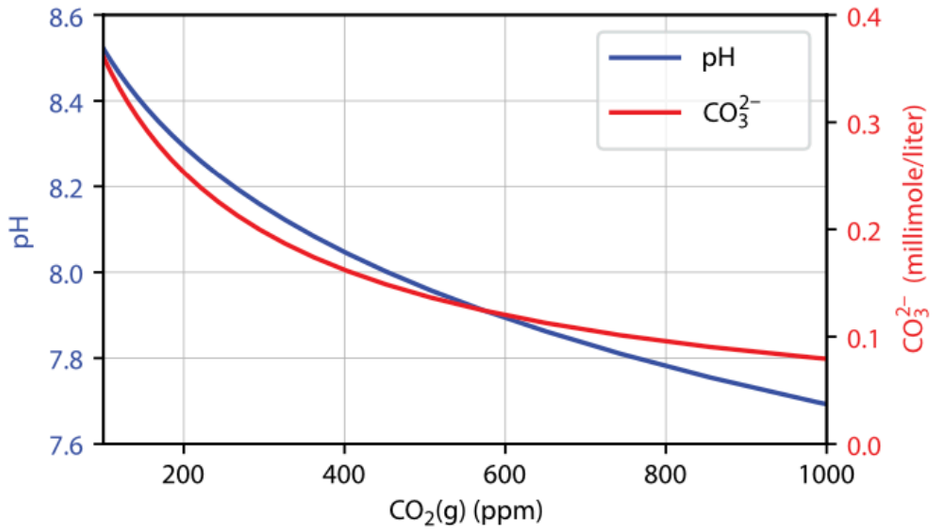


Figure 5.4: The response of pH and carbonate ion to CO_2 increase. The solution of the carbonate system for a fixed alkalinity as in Figure 5.3, showing the ocean pH (blue) and the carbonate ion CO_3^{2-} concentration (red) as a function of atmospheric CO_2.

5.2.2 Approximate solution of the carbonate system

Consider an approximate solution to the carbonate system that allows us to better understand the response of the system to increased atmospheric CO_2 in a future global warming scenario.

The approximation used here is valid only at pH values around 8, and for small perturbations to the observed level of ocean DIC, and is consistent with the ocean pH values at present or those that are anticipated in the coming decades.

For this pH range we may assume

$$\left[HCO_3^-\right], \left[CO_3^{2-}\right] \gg \left[H^+\right], \left[OH^-\right], \left[H_2CO_3^*\right].$$

The ions on the left are measured in hundreds to thousands of micromoles per liter (Figure 5.3).

The smallness of $[\,H_2CO_3^*\,]$ may be deduced from Figure 5.3 (see green line near pH = 8).

That is small is clear from the pH level, which implies that $[H^+] \sim 10^{-8}$ mole/l.

That is small is similarly deduced from the water dissociation equation (5.16) and the fact that $K_w \sim 2 \times 10^{-14}$ $(mol/l)^2$.

With this approximation, let the carbonate system unknowns be the values of the five concentrations [CO2(g)], $[\,H_2CO_3^*\,]$, $[\,HCO_3^-\,]$, $[\,CO_3^{2-}]$ and $[H^+]$.

With a total of five unknowns ($[OH^-]$ is not calculated, nor needed now), we need five equations (the one for water dissociation is not needed),

$$
\begin{aligned}
K_H &= \frac{\left[H_2CO_3^*\right]}{[CO_2(g)]}, \\
K_1 &= \frac{\left[HCO_3^-\right][H^+]}{\left[H_2CO_3^*\right]}, \\
K_2 &= \frac{\left[CO_3^{2-}\right][H^+]}{\left[HCO_3^-\right]}, \\
Alk_C &= \left[HCO_3^-\right] + 2\left[CO_3^{2-}\right], \\
C_T &= \left[HCO_3^-\right] + \left[CO_3^{2-}\right]. \qquad (5.19)
\end{aligned}
$$

The last two equations give

$$\left[\mathrm{HCO_3^-}\right] = 2\mathrm{C}_T - \mathrm{Alk_C}, \quad (5.20)$$
$$\left[\mathrm{CO_3^{2-}}\right] = \mathrm{Alk_C} - \mathrm{C_T}. \quad (5.21)$$

Using the K_2 equation,

$$[\mathrm{H^+}] = K_2 \frac{2C_T - Alk_C}{Alk_C - C_T}. \quad (5.22)$$

Next, using the K_1 equation,

$$\left[\mathrm{H2CO_3^*}\right] = \frac{K_2}{K_1} \frac{(2C_T - Alk_C)^2}{Alk_C - C_T},$$

which, using Henry's law, gives

$$[\mathrm{CO_2(g)}] = \frac{K_2}{K_1 K_H} \frac{(2C_T - Alk_C)^2}{Alk_C - C_T}. \quad (5.23)$$

We have now solved for all unknowns in terms of the equilibrium constants and the specified total CO_2 (DIC) and alkalinity.

Given their definitions, total CO_2 (DIC) is necessarily smaller than alkalinity in this approximation (see last two lines in eqn 5.19), $Alk_C > C_T$.

At the same time, the typical values given below indicate that $Alk_C < 2C_T$.

The last equation (5.23) therefore makes it clear, for example, that if the DIC increases, the atmospheric CO_2 increases as well; we explore more such responses of the carbonate system to various perturbations in the next section.

To calculate the numerical values of the above solution, one may use the typical values $Alk = 2350\ \mu$mol/l, $C_T = 2075\ \mu$mol/l, and the constants that are derived for a temperature and salinity of $T = 15^oC$ and $S = 35$ ppt at a depth of 0 m: $K_H = 0.0375$ mol/l/ ppt, $K_1 = 1.15 \times 10^{-6}$ mol/l, $K_2 = 7.43 \times 10^{-10}$ mol/l, $K_w = 2.37 \times 10^{-14}$ (mol/l)^2; the calcite and aragonite solubility constants are $K_{sp,c} = 4.32 \times 10\text{-}7$ (mol/l)2 and $K_{sp,a} = 6.72 \times 10\text{-}7$ (mol/l)2 , correspondingly

5.3 RESPONSE TO PERTURBATIONS

We now consider the response of the ocean carbonate system to various perturbations that may be expected as part of anthropogenic climate change, including increased atmospheric CO_2 concentration, warming, and the long- term recovery from increased CO_2.

5.3.1 Response to increased atmospheric CO_2 concentration

If the atmospheric CO_2 increases, so would the ocean reservoir of total CO_2 due to Henry's law.

Consider therefore that the DIC increases due to anthropogenic emissions by one unit.

The alkalinity does not change in this scenario, as no ions related to weak acids are added to the ocean.

The approximate solution from earlier then allows us to calculate the response of the pH and carbonate ion concentration,

$$\begin{aligned} \Delta C_T &= \mathbf{1}\uparrow, \quad \Delta Alk_C = \mathbf{0} \\ [\mathrm{H^+}] &= K_2 \frac{2C_T - Alk_C + \mathbf{2}}{Alk_C - C_T - \mathbf{1}} \uparrow \Rightarrow \mathrm{pH}\downarrow \\ \left[\mathrm{CO_3^{2-}}\right] &= \mathrm{Alk_C} - \mathrm{C}_T - \mathbf{1}\downarrow. \end{aligned}$$

The solution thus indicates that the pH should drop, and the ocean should become more acidic, as expected.

The decrease in the concentration of the carbonate ion $\mathrm{CO_3^{2-}}$ means that the reaction of dissolution of calcium carbonate,

$$CaCO_3 \rightleftharpoons Ca^{2+} + CO_3^{2-}, \qquad (5.24)$$

will be driven toward its right-hand side and therefore lead to more dissolution (or less deposition) of calcium carbonate, moving upward the transition zone between the dissolving deep ocean and the upper ocean.

This is also seen by writing this reaction as Ksp = [Ca^{2+}] [CO_3^{2-}], where reduction in the carbonate ion would lead to calcium carbonate dissolution, which increases the concentration of both ions and restores the product to its equilibrium value between the solid phase of calcium carbonate and its ions.

The fact that much of the dissolved CO_2 is converted into bicarbonate and carbonate ions via the carbonate system reactions means that for a given change to the atmospheric CO_2, the ocean absorbs much more CO_2 than it would have without these reactions, an effect referred to as the *buffer capacity* of the ocean carbonate system.

Henry's law (eqn 5.4) controls the concentration of [HCO_3^-], and without the carbonate reactions, the increase in dissolved CO_2 would correspond to the increase in [HCO_3^-] alone.

However, as noted above, the distribution of carbon among [$H_2CO_3^*$], [CO_3^{2-}], and [HCO_3^-] is approximately 1%, 10%, and 90%, correspondingly (Figure 5.3)

.A small increase in $H_2CO_3^*$ due to the increase in atmospheric CO_2 and Henry's law is therefore accompanied by a larger increase in bicarbonate ions, and thus in a larger increase in dissolved CO_2 than would have been deduced by the change in $H_2CO_3^*$ alone.

The bottom line is that most of the newly dissolved atmospheric CO_2 is converted into bicarbonate ions. The actual increase in carbon in the ocean carbonate system is therefore much larger than it would have been due to Henry's law and without these reactions, allowing the ocean to absorb more of the emitted anthropogenic CO_2.

5.3.2 Response to warming

So far the ocean has been absorbing a significant fraction of the anthropogenic CO_2 emission, leading to the observed acidification yet also to a reduction in the greenhouse effect and warming that would have been experienced otherwise.

However, a warming of the ocean would lead to changes in the solubility of CO_2 in seawater.

As the warming intensifies, less emitted CO_2 will be dissolved, and some dissolved CO_2 may be released to the atmosphere, further amplifying the greenhouse warming.

During the last glacial maximum 21,000 yr ago, for example (Box X.2), the ocean temperature was colder by a few degrees, and the CO_2 concentration was 180 ppm, about 100 ppm less than its preindustrial value of 280 ppm.

About a third of this drop in CO_2 can be attributed to the cooler glacial ocean temperatures, an effect which is referred to as the *solubility pump*.

The effect of the dependence of solubility on temperature is demonstrated in Figure 5.5, where the pH, atmospheric CO_2, and reaction constants are shown as a function of temperature, in a calculation specifying constant DIC and alkalinity.

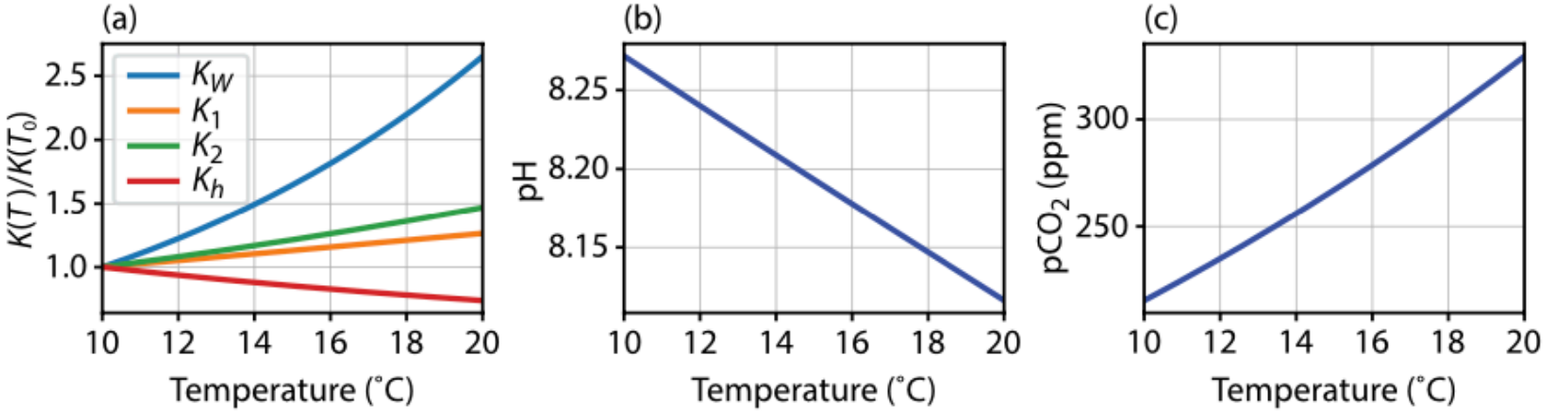


Figure 5.5: Response of the carbonate system to warming, showing quantities as a function of the ocean temperature, when the DIC and alkalinity are assumed fixed. (a) Reaction constants normalized by their values at 10 C. (b) pH. (c) Atmospheric partial pressure of CO_2, denoted pCO_2.

The dependence of the reaction constants on temperature leads to the changes seen to ocean pH and to the atmospheric CO_2 with temperature, showing increasing atmospheric CO_2 with warming, as expected.

The figure suggests that a few degrees of warming can lead to a significant atmospheric CO_2 increase of a few tens of parts per million, even without the direct contribution of anthropogenic CO_2 emissions.

Equation (5.22) suggests that the pH changes as a function of temperature mostly due to the variation of K_2 with temperature, while equation (5.23) indicates that K_1, K_2, and K_H all play a role in leading to the change in atmospheric CO_2 with temperature, rather than, say, only Henry's law constant being responsible for the direct dissolution of CO_2 in seawater, as one might think naively.

Note that in the fraction appearing in the solution for the atmospheric CO_2 concentration, K_2/K_1K_H, Henry's constant K_H decreases with temperature, while the other two increase.

The ratio overall increases, leading to the increase in atmospheric CO_2 with warming.

5.3.3 Long-term decline of anthropogenic CO_2

Human release of carbon into the atmosphere so far amounts to an order of 700 GtC (the atomic weight of carbon is 12 atomic mass units, while that of carbon dioxide is 44; implying a 1 ton carbon in 44/12 = 3.67 tons of CO_2).

The current anthropogenic emission rate is about 10 GtC/yr, and business-as- usual scenarios anticipate over 1000 GtC total further emission by year 2100.

These emissions far exceed the emission of CO_2 by the main natural source due to volcanoes (Box 5.1).

The lifetime of emitted CO_2 affects long-term warming projections and is clearly of interest.

The first stage in the adjustment to the added anthropogenic carbon is a relatively fast (years) absorption by the terrestrial biosphere and an equilibration with the upper ocean where some of the CO_2 dissolves, leading to the lowering of ocean pH, as we have seen above.

The deeper ocean takes up and responds to the dissolved CO_2 more slowly (hundreds of years) due to the slow mixing between the upper and deeper ocean, until a final partitioning of CO_2 is achieved between the atmosphere and ocean.

The lowering of the ocean pH leads to a dissolution of calcium carbonate in ocean bottom sediments (or reduced burial of calcium carbonate), as discussed earlier.

This dissolution restores the ocean pH, which may be understood using our simplified carbonate system as follows.

The dissolution of one unit of $\mathrm{CaCO_3}$ into $\mathrm{Ca^{2+}}$ and $\mathrm{CO_3^{2-}}$ implies the addition of one unit of total CO_2 (ΔC_T, in the form of the carbonate ion) and two units of carbonate alkalinity (ΔAlk_C , due to the double negative charge of the carbonate ion).

The resulting pH and atmospheric CO_2, based on our simplified solution, are then

$$
\begin{aligned}
\Delta C_T &= \mathbf{1}\uparrow, \quad \Delta Alk_C = \mathbf{2}\uparrow \\
[\mathrm{H}^+] &= K_2 \frac{2C_T - Alk_C + \mathbf{0}}{Alk_C - C_T + \mathbf{1}} \downarrow \Rightarrow \mathrm{pH}\uparrow \\
[\mathrm{CO_2(g)}] &= \frac{K_2}{K_1 K_H} \frac{(2C_T - Alk_C + \mathbf{0})^2}{Alk_C - C_T + \mathbf{1}} \downarrow .
\end{aligned}
$$

That is, pH increases (because decreases) in response to calcium carbonate dissolution, and the atmospheric CO_2 (eqn 5.23) decreases.

The timescale for this process, which involves interaction with ocean sediments, though, is thousands of years, and more detailed calculations show that a significant fraction of emitted anthropogenic CO_2 may therefore remain in the atmosphere for tens of thousands of years.

Eventually, on an even longer timescale, the natural sink of CO_2, involving its dissolution in rain water, formation of a weak acid that interacts with silicate rocks, leads to the formation of calcium carbonate whose dissolved components are washed into the ocean, leading to increased $\mathrm{CaCO_3}$ burial there and restoring the atmospheric CO_2 to its natural level.

This natural level has very slowly varied over millions of years due, for example, to changes in emission rates by volcanoes or other tectonic processes, leading to higher CO_2 during past warm climate periods such as the Pliocene (5.3–2.6 Myr) and Eocene (56– 33.9 Myr), discussed in Box 4.1.

Final word

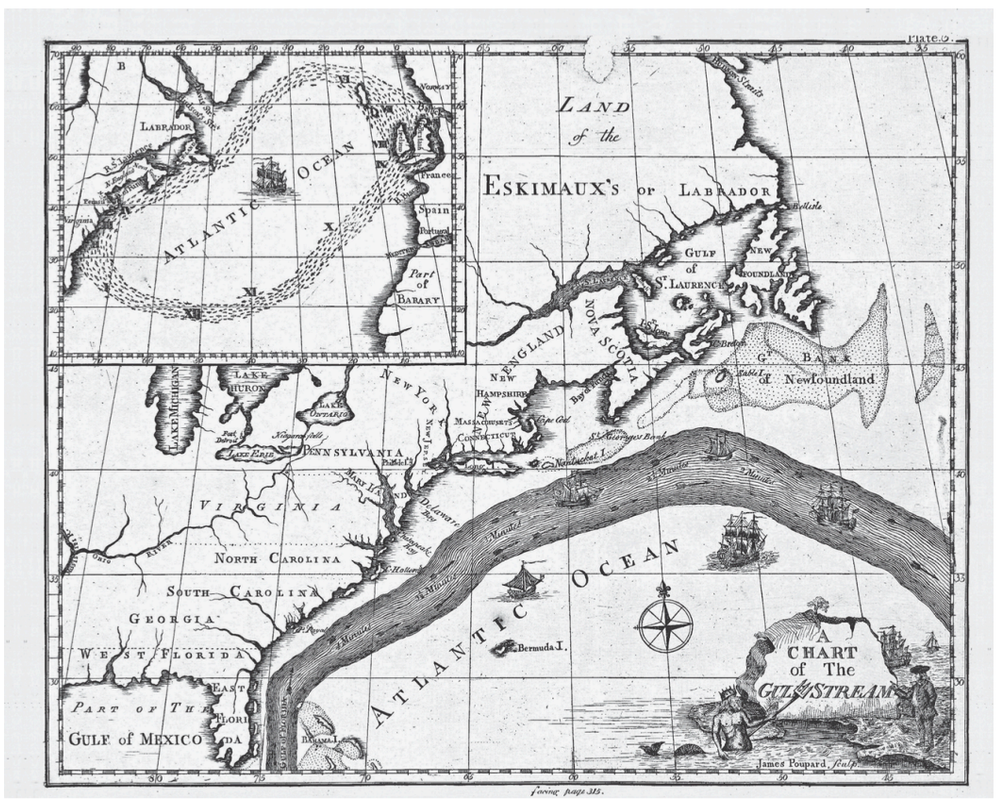
In the case of ocean acidification, unlike many other issues discussed in this book, the uncertainty in the response to anthropogenic forcing is small: given an atmospheric CO_2 increase, the pH will drop.

The signal of ACC is already clearly seen as the reduction in the observed ocean surface pH (blue markers in Figure 5.1); given that it is not difficult to calculate the expected acidity change due to the increase in atmospheric CO_2, the attribution to ACC (specifically to the increase in atmospheric CO_2) in this case is straightforward.

As mentioned, the response of the ocean biology to the change in ocean acidity is a more complex issue, yet if CO_2 increases significantly following some high-end emission scenario, the ocean pH drop will be large, and one may anticipate a non-negligible disruption to ocean biology.

6 - OCEAN CIRCULATION

The Atlantic meridional overturning circulation (AMOC), also referred to as the *thermohaline circulation* or the ocean conveyor belt, is the zonally integrated northward flow of water in the upper 1 km of the North Atlantic Ocean, which is cooled in the northern North Atlantic, becomes denser and sinks, and returns south between 2 and 3 km depth as the North Atlantic Deep Water (NADW; see Box 6.1).



The circulation magnitude is about 20 Sverdrups (Sv), where 1 Sv is cubic meters of seawater per second.

For perspective, all the world rivers combined transport about 1 Sv, and the Gulf Stream transports up to about 100 Sv.

Because the AMOC involves the flow of warm water northward and a return of cold water southward, it carries a significant amount of heat poleward, of the order of 1 petawatt, or 10^{15}W.

This heat is released to the atmosphere in the northern North Atlantic, where the water cools and sinks.

AMOC is therefore often credited for contributing to the warmth of Northern Europe relative to areas of eastern North America that are at the same latitude or even farther south (London is at 51.5 ^{o}N with average maximum/minimum January temperatures of 9 ^{o}C/4 ^{o}C, while in Boston, at 42 ^{o}N, these temperatures are 3 ^{o}C/ -5 ^{o}C).

Climate projections suggest that this circulation might weaken dramatically by 2100, mostly due to the increase in fresh water input into the northern North Atlantic from increased precipitation and ice melting, which would make the surface ocean water fresher and therefore lighter and would not allow it to sink, weakening the northward ocean heat transport.

The reason Northern Europe is warmer than eastern North America turns out to involve additional factors, including the atmospheric circulation and the location of the atmospheric jet stream, but as we will see, a collapse of AMOC may still lead to local cooling of sea surface temperatures in the northern North Atlantic Ocean.

Furthermore, the ocean meridional circulation may be an example of the climate system encountering a tipping point—a *gradual* change in CO_2 concentration past a certain threshold may lead to an *abrupt* AMOC collapse.

Such an abrupt response may be accompanied by irreversibility of the change: if the CO_2 concentration stops increasing and is gradually *reduced* past the threshold where the abrupt change occurred, AMOC may not necessarily recover its original value.

While it is far from obvious that AMOC will experience such abrupt changes and tipping point dynamics, we use it below as an opportunity to understand how climate tipping points are analyzed, why they occur, and what exactly happens when they do.

The sinking of deep water in the North Atlantic is supplemented by similar sinking to even greater depths near Antarctica (Box 6.1), and the Atlantic meridional overturning circulation is part of a global meridional overturning ocean circulation (MOC).

Some of the deep return flow from the North Atlantic upwells in the Southern Ocean, driven by the winds there, and some proceeds to the Pacific and Indian Oceans, gradually warms and slowly upwells over large areas on its way there, and returns as an upper ocean warm flow to the North Atlantic Ocean.

This overturning circulation co-exists with the wind-driven ocean circulation in the great upper ocean (top ~1 km) gyres. In the western North Atlantic, for example, the narrow (~100 km width) Gulf Stream transports up to about 100 Sv northward, compensated by a slow southward return flow over the upper North Atlantic Ocean east of the Gulf Stream.

Similar wind-driven upper ocean gyres exist in all ocean basins, with intense narrow flows on the west and a slow broad return flow in the basin interior.

Had we followed a specific water parcel in the ocean, we might have seen it circling a few times over a few decades in these horizontal wind-driven gyres, then escaping northward and sinking to the deep ocean, eventually returning hundreds of years later to the surface in a different ocean basin.

We now discuss present observations and future warm-climate projections of AMOC.

Next, we analyze a highly idealized yet surprisingly insightful model of the AMOC, allowing us to understand how tipping points occur in this case and more generally in the climate system and discuss the role of simple climate models.

We conclude with a brief discussion of the consequences of an AMOC collapse in full-complexity climate models and with a survey of the roles the ocean plays in anthropogenic climate change.

Climate Background Box ——— Ocean temperature, salinity, and water masses

The zonally averaged temperature and salinity across the Atlantic Ocean are contoured in the accompanying figure as a function of latitude and depth in panels a and c, while the profiles of temperature and salinity versus depth in panels b and d are averaged within the Atlantic over latitudes 30 ^{o}S–0^{o}.

The deep ocean is seen to be much colder than the surface, being filled via sinking at the cold high latitudes.

The temperature away from the high latitudes is typically vertically uniform in the upper 25–50 m *mixed layer* that is well mixed by wind and waves.

It then decreases rapidly in the upper kilometer, in a depth range known as the *main thermocline*.

Below the thermocline, the temperature is nearly uniform, gradually decreasing toward the bottom.

Salinity (measured in parts per thousands; say a kg of salt in a metric ton of seawater) variations are a result of surface evaporation, which increases the salinity (and the freezing of sea ice that results in the rejection of brine), and precipitation and river runoff, which decrease it.

Water bodies with characteristic temperature and salinity are referred to as *water masses*.

A tongue of salty water spreading south from the North Atlantic is seen at a depth of 2–3 km, known as the North Atlantic Deep Water (NADW).

Two fresher water masses are seen to spread northward from the south, the Antarctic Intermediate Water (AAIW), at a depth of about 1 km and above the NADW, and the Antarctic Bottom Water (AABW), near the bottom and under the NADW.

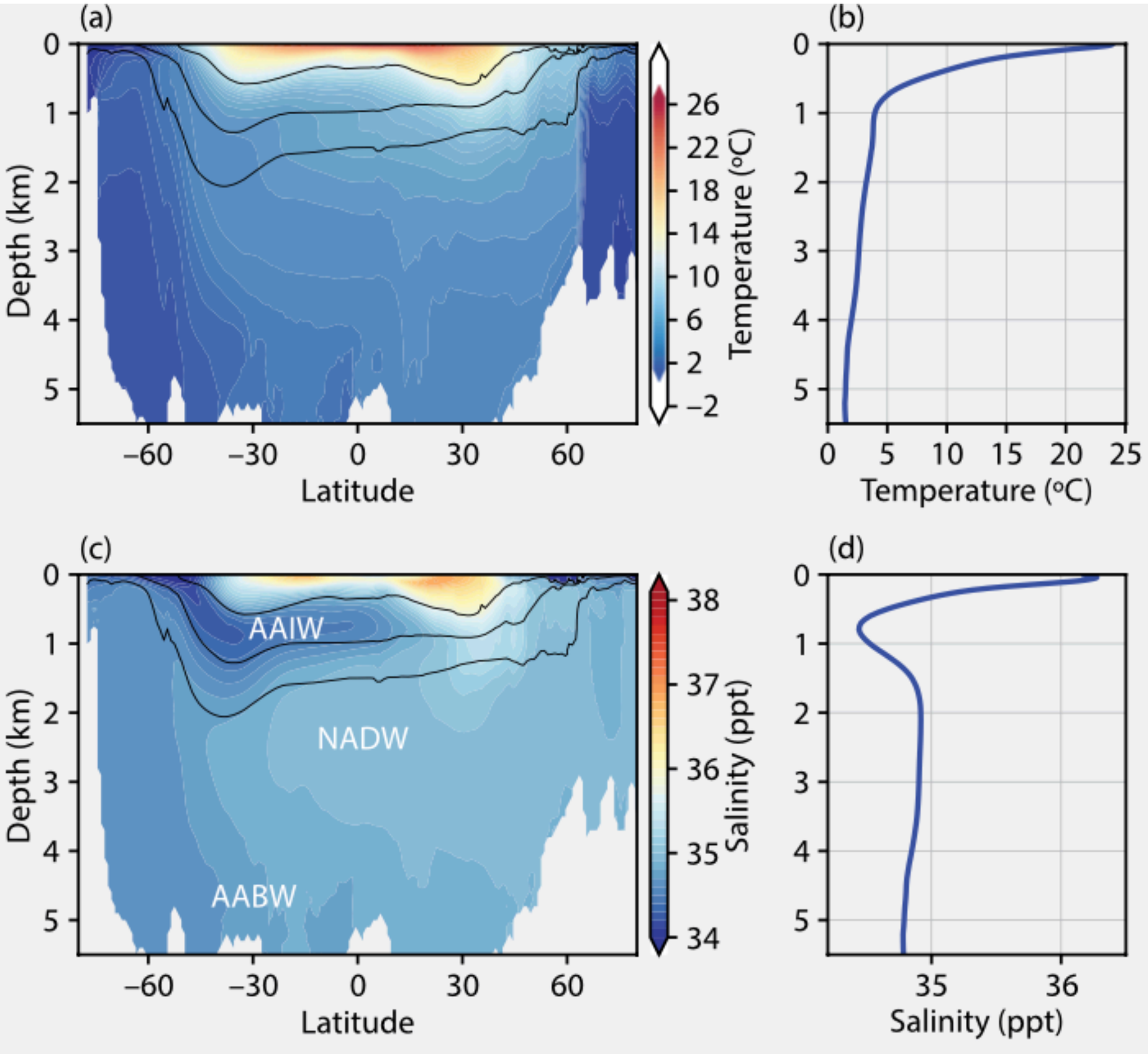
Note that the vertical dimension is stretched significantly, as the actual aspect ratio between the length and depth of the Atlantic Ocean is about 1000:1.

These water masses are therefore essentially thin sheets of ocean water spreading horizontally with very little mixing due to small-scale turbulence across density surfaces.

The spreading largely occurs over surfaces of constant density, such as the three marked by the black contours.

The Stommel box model discussed later idealizes the formation and spreading of the NADW.

A shutdown of the Atlantic meridional overturning circulation may disrupt this very distinct large-scale water mass structure.


(a)
(b)
(c)
(d)
Depth (km)
Latitude
Temperature (°C)
Salinity (ppt)
AAIW
NADW
AABW
0
1
2
3
4
5
–60
–30
0
30
60
26
22
18
14
10
6
2
–2
38
37
36
35
34
0
5
10
15
20
25
35
36

6.1 OBSERVATIONS AND PROJECTIONS

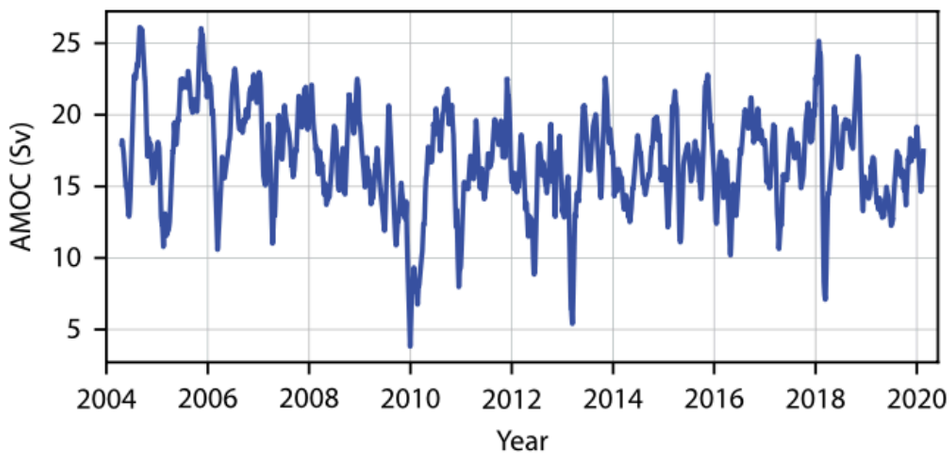
The AMOC may be defined as the total northward water volume transport in the upper 1 km of the North Atlantic Ocean, often calculated at one specific latitude.

Since 2004, AMOC has been estimated continuously in the RAPID program using observations across 26.5 ºN in the North Atlantic Ocean, including regular ship surveys that measure temperature, salinity, and density across the Atlantic Ocean and observations of the wind that drives upper ocean transports, of the Gulf Stream, and more.

The observed AMOC time series shows a strong, yet noisy and irregular, seasonal cycle (Figure 6.1).

Figure 6.1: The observed AMOC.

A time series from the RAPID campaign.



Historical ship observations have been used to estimate the transport across 25 ºN over the past few decades—for 1957, 1981, 1992, 1998, and 2004—suggesting a detection of a significant AMOC slowdown from 23 Sv to 15 Sv over this period.

However, this seeming weakening is likely a result of the large-amplitude seasonal cycle we are now aware of thanks to the RAPID campaign, and for now it is probably fair to say that we are unable to detect an AMOC slowdown with confidence, even if future projections of a slowdown seem robust across climate models.

While the observed record is still short, AMOC oscillations seen in climate models have periods from a couple of decades to centuries.

These oscillations affect the large-scale sea surface temperature in the North Atlantic in particular.

A variability mode known as the *Atlantic multidecadal oscillation* (AMO) is a pattern of alternating warming and cooling with an amplitude of some 0.2 ^{o}C that is inferred from sea surface reconstructions over the past century or so.

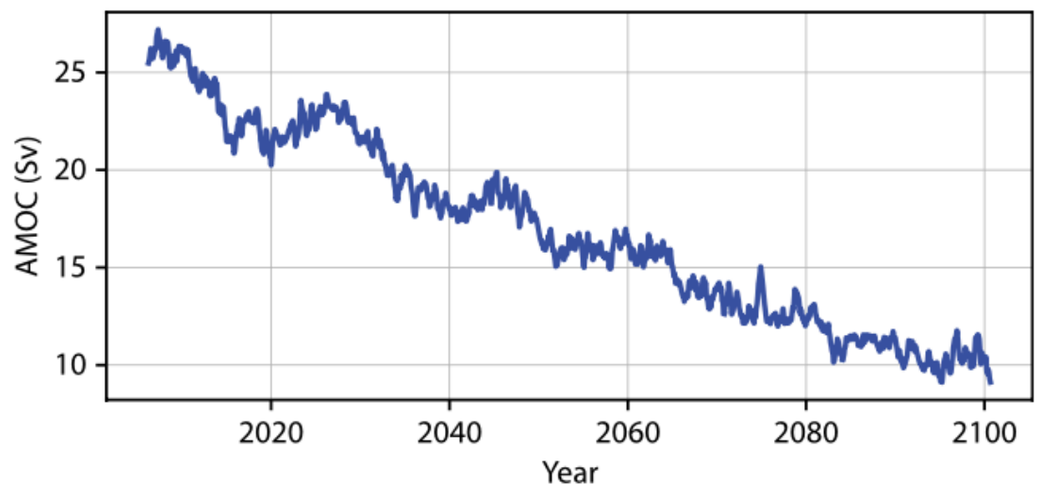
This mode was suggested to be related to AMOC variability, although its very existence and driving mechanisms are being debated.

While speculative, the AMO was suggested to affect the occurrence of droughts and hurricanes, among other influences.

Future RCP scenarios (e.g., RCP8.5) suggest a dramatic future slowdown of the AMOC (Figure 6.2), mostly due to a freshening of the northern North Atlantic by increased precipitation and ice melting.

Figure 6.2: An AMOC future projection.

A projected AMOC transport time series in a climate model run under the RCP8.5 scenario.

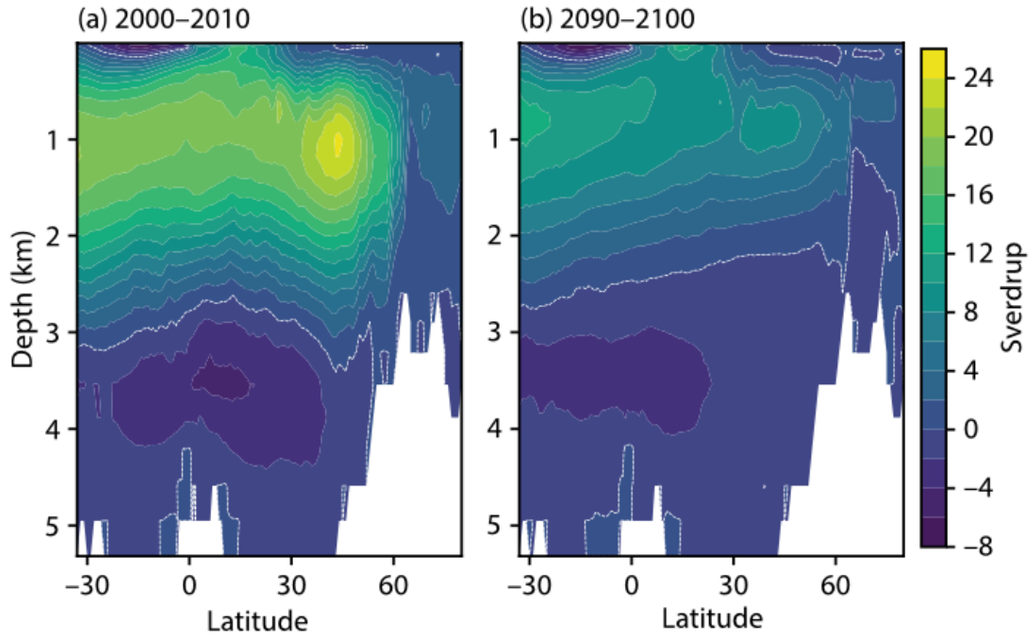


The freshening reduces the density of the surface water in the northern North Atlantic, and this leads to a lowering of the north-south density difference and makes the high-latitude surface water lighter, preventing its sinking.

AMOC is often represented in model simulations and projections via a *streamfunction*, which is depicted by a contour plot as a function of latitude and depth across the North Atlantic Ocean (Figure 6.3).

Figure 6.3: The AMOC streamfunction.

Shown in units of Sv from a climate model run under the RCP8.5 scenario, during the first and last decades of the 21st century.



The zonally integrated AMOC transport occurs along contours of constant streamfunction values, and the difference between contour values (say at 24 ^{o}N) at two depths is equal to the total meridional mass transport in this depth range.

The AMOC streamfunction is shown in Figure 6.3 for the first and last decades of the 21st century, for an RCP8.5 projection, demonstrating the projected weakening.

AMOC is driven by a complex set of factors, including north-south ocean water density differences due to temperature and salinity differences between the low and high latitudes in the North Atlantic Ocean (Box 6.1), winds in the Southern Ocean, and small-scale ocean mixing due to turbulence on a scale of meters or less within the ocean interior.

Such ocean mixing is therefore another sub-grid scale process that has global climate implications and needs to be parameterized in ocean models.

As an introduction to the analysis in the next section, we note that a somewhat naive, yet perhaps not useless, view of the AMOC is that it is dominantly driven by a north-south ocean water density difference, which leads to a pressure gradient that can drive a northward flow.

In this view, the cold polar water versus warm tropical oceans leads to a meridional density gradient that *drives* AMOC.

On the other hand, the salty tropical ocean versus the relatively fresh high-latitude ocean has the opposite effect on the meridional density gradient, *braking* the AMOC.

The presence of the Earth rotation and the resulting Coriolis force seem to suggest that this view is too naive (a north-south pressure gradient would drive a zonal ocean current; see section 4.2.1).

Yet there are reasons to believe that the north-south density gradient still plays a dominant role, and relying on it still leads to very robust insights.

6.2 THE STOMMEL MODEL

Our objective is to understand what might happen to the AMOC in a warmer climate, when a freshening of the northern North Atlantic is expected due to ice melting and increased precipitation.

In particular, we are interested in the prospect of an abrupt collapse of AMOC as CO_2 gradually increases beyond a certain threshold, a scenario often referred to as the crossing of a tipping point.

Divide the North Atlantic Ocean very crudely into two boxes (Figure 6.4), one representing high altitudes (box 1) and the other mid-latitudes/subtropics (box 2).

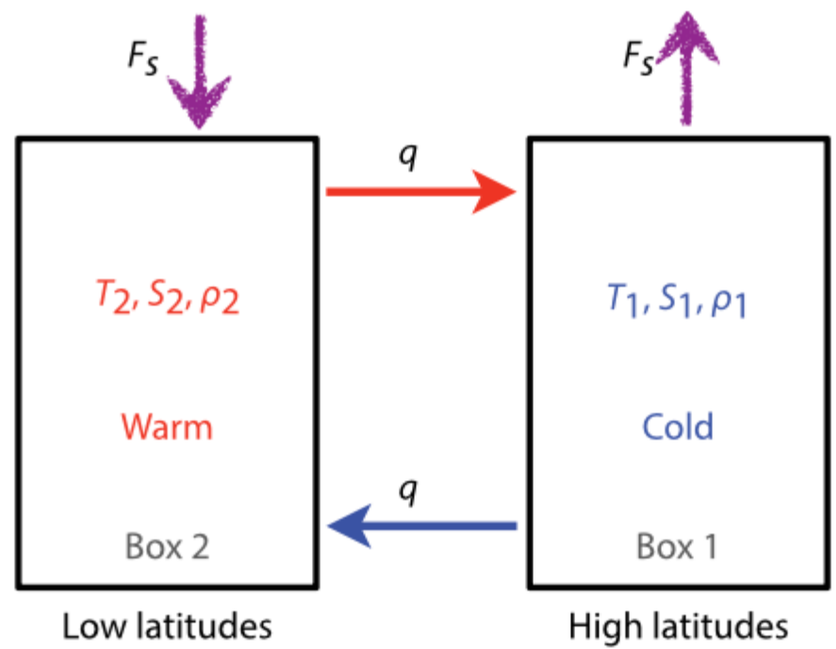


Figure 6.4: Stommel box model schematic.

The boxes are assumed to extend from the ocean surface to the ocean bottom and be well mixed (clearly a crude approximation, as we know the ocean temperature, salinity, and density to vary significantly with depth; see Box 6.1).

The two boxes are connected via a circulation representing AMOC, denoted by the red and blue arrows, which transports a volume flux q between the boxes.

The AMOC volume transport (m^3/s) is assumed proportional to the density difference between the boxes, $q = K(\rho_1-\rho_2)$.

The density of seawater is approximated to be a linear function of temperature and salinity, $\rho(T,S) = \rho_0 - \alpha_T(T-T_0)+\beta_S(S-S_0)$, increasing with salinity and decreasing with temperature.

Surface forcing: salinity and evaporation/precipitation

To see how evaporation and precipitation affect salinity, consider a bucket filled with a volume V of seawater and undergoing evaporation.

First, the volume budget of the water in the bucket is

$$\frac{dV}{dt} = -EA,$$

where E is the net evaporation rate per unit area and A is the surface area of bucket.

Letting the mass of salt per unit mass of seawater be denoted by the salinity S, the total salt in the bucket, not affected by evaporation, is equal to ϱSV, where ϱ is the water density.

Assuming the density is nearly constant, the salt conservation statement is, therefore,

$$\frac{d}{dt}(SV) = 0,$$

which we can expand using the above mass conservation equation into

$$V\frac{dS}{dt} = -S\frac{dV}{dt} = SEA \approx S_0 EA.$$

Salinity in the ocean varies mostly between 34 and 36 ppt (1 ppt roughly representing a kg salt per metric ton of seawater), and $S_0 = 35$ ppt denotes an averaged reference salinity.

We term S_0E the *virtual salt flux* per unit area.

Because evaporation rates are not large, the changes to the ocean salinity are typically relatively small and the above approximation of using S_0 instead of S is a very good one, accurately describing ocean salinity changes due to evaporation and precipitation.

A typical order of magnitude of evaporation in the ocean is 1 m/yr, and given the ocean depth of 4000 m, evaporation does not significantly affect ocean volume.

This also means that the water volume exchanges involved in evaporation and precipitation may be neglected, and we may represent their effect by the virtual salt flux, which is denoted in Figure 6.4 and in the derivation below, also referred to as the *fresh water forcing*.

Salt budget equations for the Stommel model

We assume that the ocean temperature is set by the interaction with the atmosphere, which is cold in polar areas and warmer in subtropical areas.

We therefore let the ocean box temperatures T_1 and T_2 be fixed to a first approximation, so that circulation changes occur only due to ocean salinity changes.

The only unknowns, therefore, are the salinities of the two boxes and the circulation q.

Write the salt budget equation for box 1, first assuming that $q > 0$, representing a salt transport qS_2 into box 1 and a transport qS_1 from box 1 to box 2.

A freshening effect induced by excess precipitation in the high-latitude box 1 is represented by a $(-F_s)$ virtual salt flux contribution to the budget,

$$V_1 \frac{dS_1}{dt} = qS_2 - qS_1 - F_s.$$

If q < 0, we have salt transport of (-q)S_2 into box 1 and a transport (-q)S_1 out of box 1 to box 2,

$$V_1 \frac{dS_1}{dt} = (-q)S_2 - (-q)S_1 - F_s.$$

Together, this may be combined into a single form, which we write for both boxes as

$$\begin{aligned} V_1 \frac{dS_1}{dt} &= |q|\,(S_2 - S_1) - F_s \\ V_2 \frac{dS_2}{dt} &= |q|\,(S_1 - S_2) + F_s. \end{aligned} \quad (6.1)$$

Note the absolute value of the transport appearing here, combining the above two cases of positive and negative AMOC volume transport q.

The fresh water forcing is assumed positive, $F_s > 0$, corresponding to a freshening of the polar box and to making the low-latitude box saltier.

Solution

Take the difference of the two salt budget equations (6.1), define $\Delta T = T_1 - T_2 < 0$ and $\Delta S = S_1 - S_2$, and let the box volumes be the same, $V_1 = V_2 = V$, for simplicity.

The transport may then be written as

$$q = K\,(\rho_1 - \rho_2) = K\,(-\alpha_T \Delta T + \beta_S \Delta S),$$

and the equation for the salinity difference becomes

$$-V\frac{d\Delta S}{dt} \quad = 2|q|\Delta S + 2F_s$$

or

$$-V\frac{d\Delta S}{dt} = 2K\Big|\,(-\alpha_T \Delta T + \beta_S \Delta S)\Big|\Delta S + 2F_s. \qquad (6.2)$$

Define a rescaled known temperature difference variable, $X = \alpha_T \Delta T < 0$, and an unknown rescaled salinity difference, $Y = \beta_S \Delta S$.

In a steady state, $d\Delta S/dt = 0$, so that equation (6.2) leads to a simple quadratic equation for $Y = \beta_S \Delta S$,

$$\Big|Y - X\Big|Y = -\frac{\beta_S F_s}{K}.$$

Because of the absolute value, there are two cases to consider q > 0 and q < 0, equivalent to Y > X and Y < X.

Noting again that X < 0, we first consider the case Y < X,

$$Y^2 - XY - \frac{\beta_S F_s}{K} = 0$$

$$Y = \frac{X}{2} - \frac{1}{2}\left(X^2 + 4\frac{\beta_S F_s}{K}\right)^{1/2}. \quad (6.3)$$

Next, consider Y > X,

$$Y^2 - XY + \frac{\beta_S F_s}{K} = 0$$

$$Y = \frac{X}{2} \pm \frac{1}{2}\left(X^2 - 4\frac{\beta_S F_s}{K}\right)^{1/2}. \quad (6.4)$$

Note that the plus solution for *Y* in the first case is positive and is therefore not consistent with the assumption Y < X used to obtain that solution, so there are no more than three solutions for a given value of the fresh water forcing F_s.

6.3 MULTIPLE EQUILIBRIA, TIPPING POINTS, HYSTERESIS

Multiple equilibria

Given the solution for $Y = \beta_S \Delta S$ as a function of the fresh water forcing F_s (eqns 6.3 and 6.4, plotted in Figure 6.5a), we can calculate the salinity difference ΔS, and from that the circulation q as a function of the fresh water forcing F_s, as in Figure 6.5b.

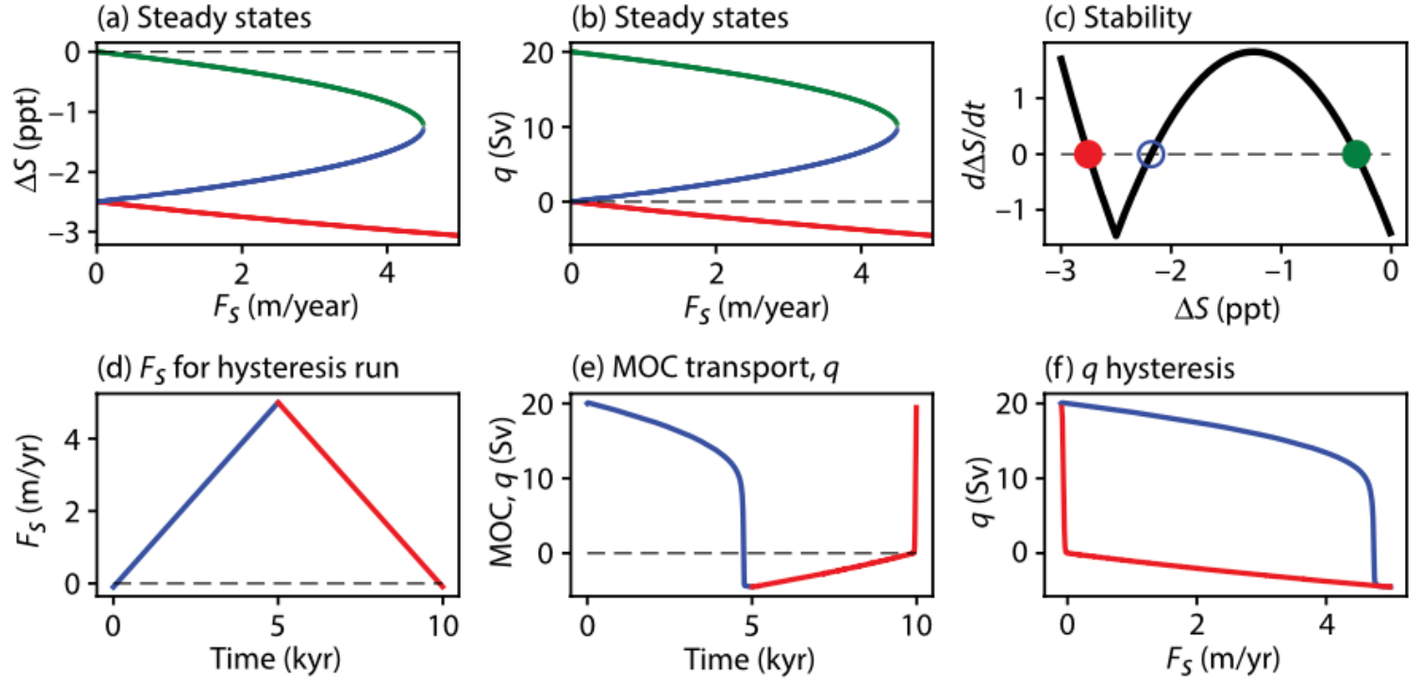


Figure 6.5: Solution of the two-box model.

(a, b) Steady states of salinity difference and MOC as a function of fresh water forcing F_s. (c) Stability analysis: $d\Delta S/dt$ as a function of ΔS for F_s corresponding to 2 m/s (to convert this to a virtual salt flux, multiply by the box area and by the reference salinity S_0), from eqn (6.2). (d) Time- dependent fresh water forcing used for the hysteresis run. (e, f) Results of the hysteresis run.

The number of real-valued solutions varies from one to three, depending on the value of the fresh water forcing.

Note that the solution shown by the green line in Figure 6.5b represents a northward upper ocean flow, while that shown in red is a weak, reversed flow that is the only possible solution for large fresh water forcing.

The results are already remarkable at this point: a given fresh water forcing can lead to three different solutions for the overturning circulation.

As we will see, the existence of such *multiple equilibria* is the first ingredient needed for tipping points to occur.

Stability

The intermediate solution (blue in Figure 6.5a,b,c) is unstable.

That is, a small perturbation from that steady state would result in the circulation transitioning to one of the other two solutions that exist for the same fresh water forcing.

To see this, consider $d\Delta S/dt$ plotted as a function of ΔS in Figure 6.5c using equation (6.2).

If the solution is exactly at the steady state denoted by the empty blue circle, $d\Delta S/dt = 0$ and the steady state is maintained.

However, suppose that the solution deviates a bit to the right (the salinity difference between the two boxes, ΔS, increases due to some random weather event affecting precipitation and evaporation).

At that point, Figure 6.5c shows that $d\Delta S/dt > 0$, and the salinity difference will therefore keep increasing and getting away from this steady state.

Similarly, if the solution deviates to the left of the steady state, marked by the empty blue circle in panel c, $d\Delta S/dt$ is negative there, and therefore the salinity difference will keep decreasing away from the steady state.

The solution marked by the blue circle in Figure 6.5c and by the blue lines in panels a and b is therefore referred to as *unstable*. If we start near one of the stable solutions (filled red and green circles in Figure 6.5c), a small increase in ΔS leads to $d\Delta S/dt < 0$, while a small decrease in ΔS leads to $d\Delta S/dt > 0$.

Thus, in both cases the deviation decreases back toward these two steady states, and they are therefore referred to as *stable* steady states.

Tipping points and hysteresis

The existence of multiple equilibria for a given fresh water forcing value leads to both the possibility of abrupt changes as the forcing changes gradually and to the possibility of irreversible changes to the circulation in a greenhouse scenario.

To see this, suppose the circulation is at a steady state corresponding to a point on the upper (green) solution for q in Figure 6.5b and that we gradually and very slowly increase the fresh water forcing such that the solution for the overturning circulation is always at equilibrium with the fresh water forcing.

The circulation solution then moves to the right along the green line and thus weakens until the fresh water forcing amplitude is at the critical value of $F_s \approx 4.2$ m/s, at which the system switches from three solutions to only one.

That one remaining solution is the very weak (reversed) circulation on the red curve in Figure 6.5b, so the circulation must abruptly switch to that solution—that is, collapse!

The lower three panels of Figure 6.5 and the schematic in Figure 6.6 demonstrate this abrupt collapse scenario.

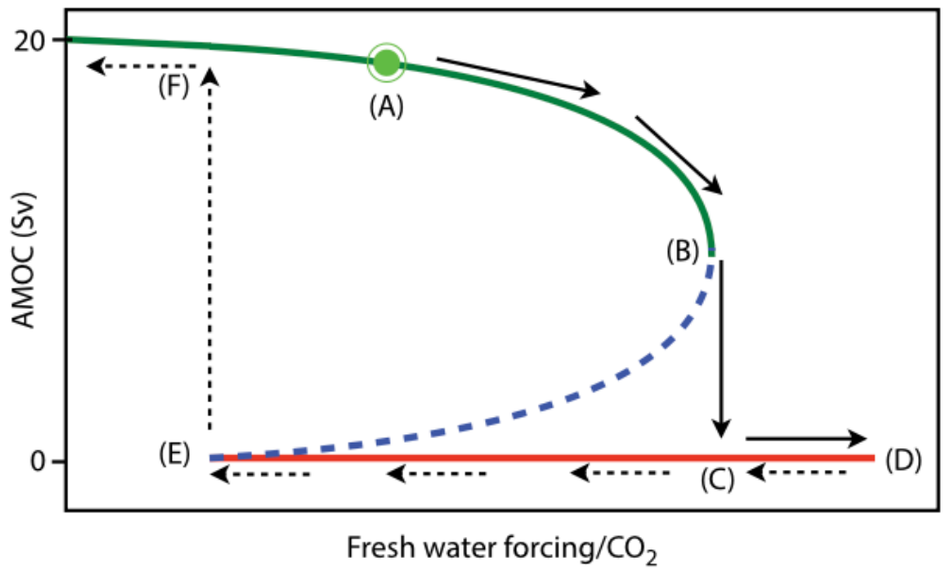


Figure 6.6: A schematic of the hysteresis and tipping point behavior in the two-box Stommel model.

Suppose that in a global warming scenario, a gradual increase of CO_2 leads to a gradual increase in precipitation/melting and thus in the fresh water forcing F_s as a function of time (blue curve in Figure 6.5d).

Figure 6.5e shows the transport as a function of time for this forcing, showing (blue line) a gradual decrease and then a collapse just before year 5000.

This abrupt collapse, corresponding to the switch between the two equilibria solutions, denotes the occurrence of a tipping point.

The schematic Figure 6.6 complements the picture.

Starting with the bright green point (A) representing the present-day climate, the fresh water forcing increases slowly due to a CO_2 increase, and the AMOC solution responds by moving to the right (black solid arrows).

As the solution reaches point (B), the stable green solution ceases to exist and the AMOC has to transition to the only stable solution available for larger fresh water values, which is a collapsed circulation shown by point (C) on the red line, denoting the other stable solution.

If the fresh water forcing is now made gradually weaker in time (red curve in Figure 6.5d), say, because CO_2 values are finally gradually decreasing, the circulation strengthens again (red curve in Figure 6.5e, dashed arrows in Figure 6.6) gradually at first.

Even as the fresh water forcing is reduced below the critical value F_c (point (C) in Figure 6.6), the circulation does not recover (does not jump back to the green line).

Recovery happens only when (point (E) in Figure 6.6).

The different evolution of the solution for increasing and decreasing forcing, expressed as the loop showing the AMOC transport as a function of the fresh water forcing in Figures 6.5f and 6.6, is termed *hysteresis*.

The existence of multiple stable and unstable solutions and the resulting hysteresis are all a result of the nonlinear nature of equation (6.2) for the salinity difference between the two boxes (that is, the right-hand side is quadratic in ΔS).

In general, tipping points, hysteresis, and multiple equilibria are due to nonlinearities, of which there are numerous in the climate system.

6.4 KEEPING IT SIMPLE

Remarkably, some full-complexity ocean models, of the same complexity class used in climate studies, show abrupt changes to AMOC and a corresponding hysteresis loop when an appropriate scenario of increasing and then decreasing fresh water forcing is applied.

The main difference from our above Stommel box model analysis is that AMOC seems to weaken and possibly vanishes at high fresh water forcing values in more realistic climate models rather than reverse as in the above box model.

Furthermore, it was suggested that the present-day circulation may be close to the threshold that leads to such an abrupt irreversible collapse.

On the other hand, the decline of AMOC in the full-complexity coupled ocean-atmosphere climate model, shown in Figure 6.2, is gradual rather than an abrupt collapse.

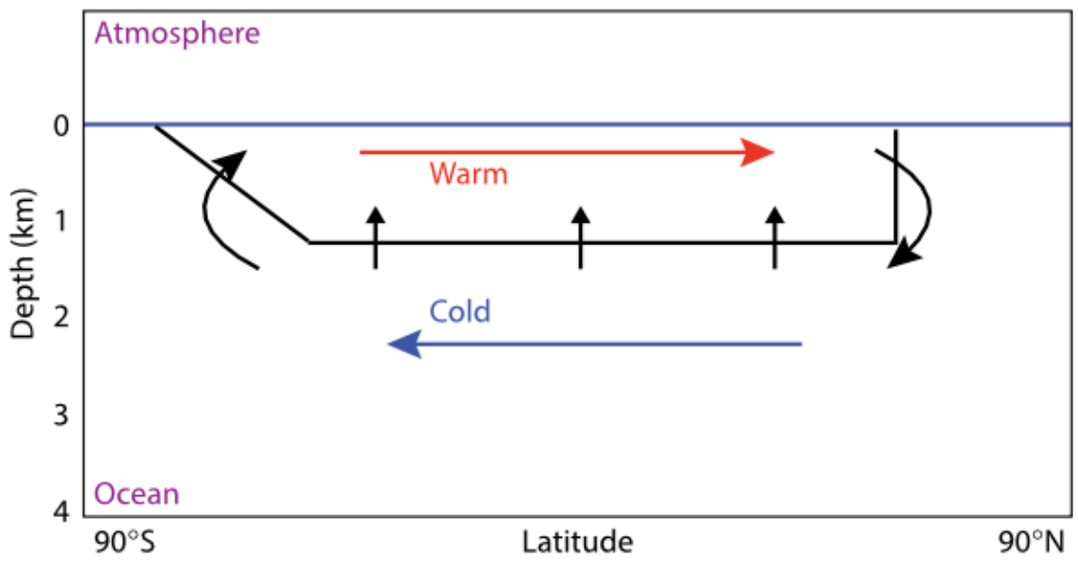
It is possible that the CO_2 change in this scenario was too fast to allow the tipping point to be clearly expressed or that this model does not have a tipping point for AMOC, demonstrating the uncertainty in this prediction.

The Stommel model in Figure 6.4 is by no means a reasonable representation of the actual ocean thermal structure in panel a of the figure in Box 6.1.

A more appropriate idealization is of a layer of warm water overlying a colder deep ocean, as in the schematic Figure 6.7; many other idealized models have been proposed for this problem as well.

Figure 6.7: An alternative AMOC idealization.

A schematic of an idealized AMOC model that is more appropriate as an idealization of the ocean temperature structure than the Stommel model, showing a warm pool of water in the upper kilometer or so, overlying a deep cold ocean that outcrops in the high latitudes of both hemispheres. Meridional heat transport by the Atlantic Ocean is the result of the upper layer's northward flow of warm water and the deeper return flow of colder water (crudely representing the NADW). Some of the cold water returns to the surface in the Southern Ocean and some as upwelling along the way.



This brings up an important point about climate science: Observations and realistic climate model simulations are critical tools for us to be able to study and predict the present and future climates.

But a deeper understanding is often achieved using idealized (toy) models that can be analyzed in depth.

Of course, "everything should be made as simple as possible but not simpler" (attributed to Einstein), so is the Stommel model too simple?

When an analysis similar to the above is applied to alternative idealized models, an AMOC decline at higher fresh water forcing (higher CO_2) is often seen, but it is not necessarily abrupt and there isn't always a hysteresis.

As mentioned, it is also not clear that full-complexity coupled ocean-atmosphere climate models robustly show such a hysteresis of AMOC when forced with slowly increasing and decreasing CO_2, leaving the robustness of this prediction of an AMOC tipping point uncertain.

The existence of tipping points for increasing greenhouse forcing has been proposed in the context of other climate components as well, from Arctic sea ice to Arctic clouds to subtropical clouds to ice sheets, underlying the possibility that a gradual CO_2 change may lead to abrupt changes and thus surprises whose precise timing is difficult to predict.

In nearly all cases the mathematical structure of the system allowing a tipping point behavior is similar to the schematic Figure 6.6, with the number of solutions changing as a function of the CO_2 from one to three (including an unstable one) to one.

Whether a robust climate tipping point behavior is in fact expected in any specific climate component as CO_2 increases, with the accompanying irreversibility when CO_2 eventually decreases, is still an open issue.

6.5 CONSEQUENCES OF AMOC COLLAPSE

The sea surface temperature warming over the 21st century under the RCP8.5 scenario is shown in Figure 6.8.

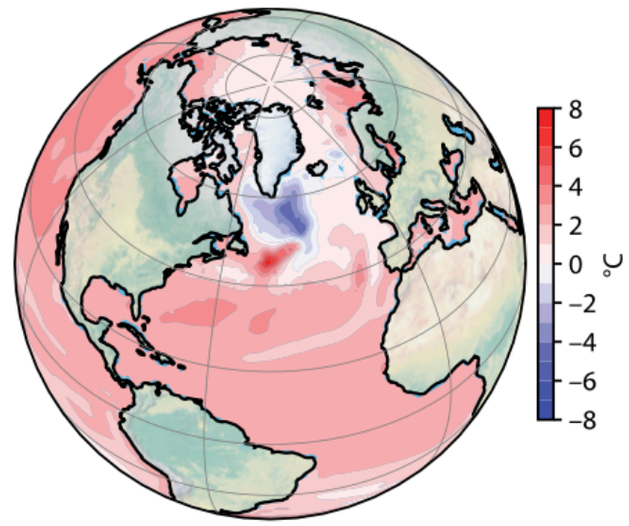


Figure 6.8: Consequences of AMOC collapse.

The SST at year 2100 minus that at 2006 in an RCP8.5 scenario.

While there is a strong overall warming, there is also a large-scale patch in the North Atlantic south of Greenland where the SST has cooled.

This local cooling is probably related to the ceased northward heat transport by the collapsed AMOC.

(Importantly, the 2004 movie *The Day After Tomorrow* predicted that such a cooling due to AMOC collapse would lead to an ice age over North America within a few weeks. See Box X.2 for a more reality-based picture of ice ages.)

While the above analysis of the Stommel box model assumed the temperature to be fixed, one does expect the SST to be affected by the collapse, as demonstrated by this image.

The change to the temperature is not sufficiently large to invalidate our above analysis, yet it has a significant impact on the nearby ocean regions and continents.

Other consequences involve the reduced ventilation of the deep ocean by the North Atlantic Deep Water currently sinking in the North Atlantic (Box 6.1), with possible effects on the large-scale deep ocean distributions of temperature, salinity, oxygen, and more.

A similar-looking SST signal to the above projected one that has been observed in the North Atlantic in recent years is referred to as the North Atlantic warming hole.

Whether it can be attributed to an already occurring AMOC weakening and/or to anthropogenic climate change is being debated.

6.6 THE OCEANS AND GLOBAL WARMING

It may be helpful to briefly survey some roles the world oceans play in the context of anthropogenic climate change, many of which have been discussed.

We saw in section earlier that the deep ocean with its immense heat capacity can delay the atmospheric warming response to greenhouse forcing by taking a longer time to warm.

Similarly, when atmospheric CO_2 will finally be decreasing, the already occurring ocean warming may delay the expected climate cooling.

Ocean warming, of course, leads to the expansion of seawater and thus to sea level rise (chapter 4).

By absorbing about a quarter of the anthropogenically emitted CO_2, the oceans reduce the greenhouse effect and therefore also delay warming, yet at the same time they experience ocean acidification, which causes its own set of problems (earlier discussion).

Sea surface temperature plays a dominant role in several contexts, starting with hurricanes, whose maximum possible magnitude may be controlled by the temperature and heat capacity of the upper ocean, which fuel hurricanes via evaporative cooling and subsequent condensation and latent heat release in the atmosphere.

Droughts are often forced by the deviation of SST, in areas that can be thousands of kilometers away from the drought location, from their long-term averaged values.

Such deviation patterns of SST that lead to droughts are caused by natural climate variability modes, such as La Niña cooling in the eastern equatorial Pacific (Box X.1), which forces droughts in California and elsewhere.

Similarly, droughts and therefore forest fires in Australia can be forced by deviations of the sea surface temperature from its long-term mean in both the equatorial Pacific and Indian Oceans.

If these natural SST variability modes changed in a warmer future climate, this would have a significant effect on regional precipitation, droughts, and fires.

Sea ice in the Arctic Ocean has been dramatically melting in past decades, possibly forced by anthropogenic climate change and putting in motion its own set of feedbacks that lead to local and remote effects on weather and climate.

Sea ice around Antarctica has not responded similarly so far, demonstrating the complexity of the climate system and the uncertainty in attributing changes to anthropogenic climate change.

The warming of ocean water near both Greenland and Antarctica can very effectively melt the massive (~1 km thick) floating ice shelves there.

This can weaken the force that these ice shelves exert on the ice flows that feed them from continental ice sheets.

The weakening of this force, in turn, can lead to acceleration of ice flows from Antarctica and Greenland and to a further speedup of sea level rise.

Apart from this long yet still incomplete list of ocean effects, processes, and feedbacks involved in anthropogenic climate change, the oceans also add to the uncertainty involved in projecting future changes.

The long timescales of the ocean response to climate forcing make it more difficult to model the eventual response of the climate system.

Ocean eddies, the continuously changing ocean motions that are the equivalent of atmospheric weather systems, are characterized by much smaller length-scales (ocean eddies have scales of tens to hundreds of km, while weather systems extend over thousands of km).

Even smaller-scale ocean motions also play a critical role, meaning that ocean/climate models must cover the global oceans and at the same time have a high resolution that can properly represent all relevant small-scale motions.

This leads to the need to perform very computationally expensive and in fact infeasible simulations, forcing us to compromise on the representation of critical ocean motions and leading to less reliable projections.

Our ability to model variability modes such as El Niño is getting better, yet we are still not in a position to reliably project future changes to these modes.

This means that our ability to project future changes to droughts, regional precipitation, and fires is necessarily still limited.

It is important to note that the reliability of future climate change projections is still mostly limited by our incomplete understanding of many issues rather than by computational resources.

Ocean science therefore has many important, challenging, and exciting open problems requiring further progress.

Climate Background Box X1

El Niño, La Niña

These are the warm (panel a of the accompanying figure) and cold (b) phases of the El Niño–Southern Oscillation (ENSO) climate variability mode in the equatorial Pacific, which occurs due to a large-scale interaction between the ocean and atmosphere.

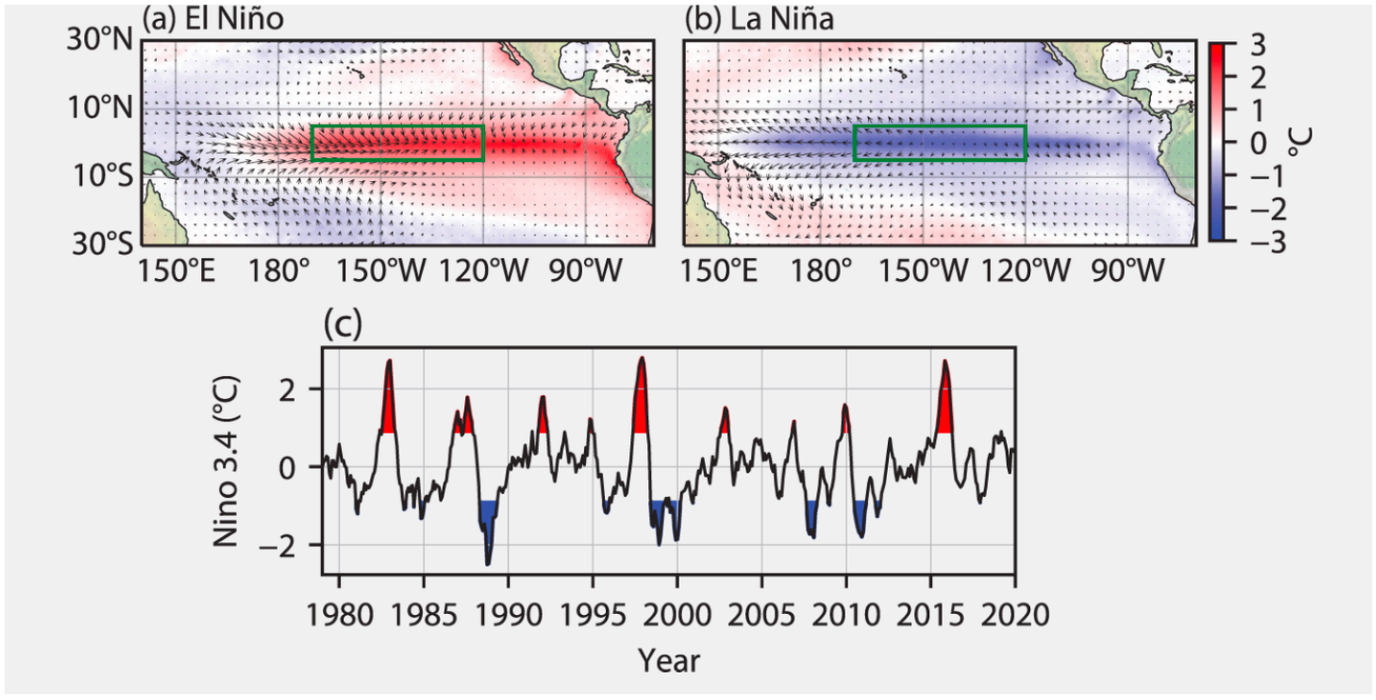
Southern Oscillation refers to a variability of the sea level atmospheric pressure difference between Tahiti (in the Pacific) and Darwin, Australia, which co-varies with the sea surface temperature in the eastern equatorial Pacific.

The time series (c) shows the SST anomaly (deviation from monthly mean) averaged over what's known as the NINO3.4 index region, 120 ^{o}W–170 ^{o}W, 5 ^{o}S–5 ^{o}N, shown by the green box in panels a and b.

Panel a (b) shows the El Niño (La Niña) SST composite—that is, the SST anomaly averaged over all months during which the index is above 1 std (below minus 1 std).

The superimposed vectors show the corresponding wind anomaly composites.

The time series shows warm El Niño events in red and cold La Niña events in blue, with warm events occurring irregularly every 2–7 years.



The two phases of ENSO have significant and different effects on weather, affecting precipitation, temperature, and winds over large areas of the globe.

In spite of their irregular occurrence, these events can now be predicted a few months in advance, providing advance warning to farmers, fisherman, and other interested parties around the equatorial Pacific and elsewhere, that are affected by this variability.

The simulation of ENSO by climate models is getting better, although predictions of how it might change in a future climate still seem very uncertain due to sensitivity to model formulation and differences between the projections of different models.

Climate Background Box X2

Ice ages

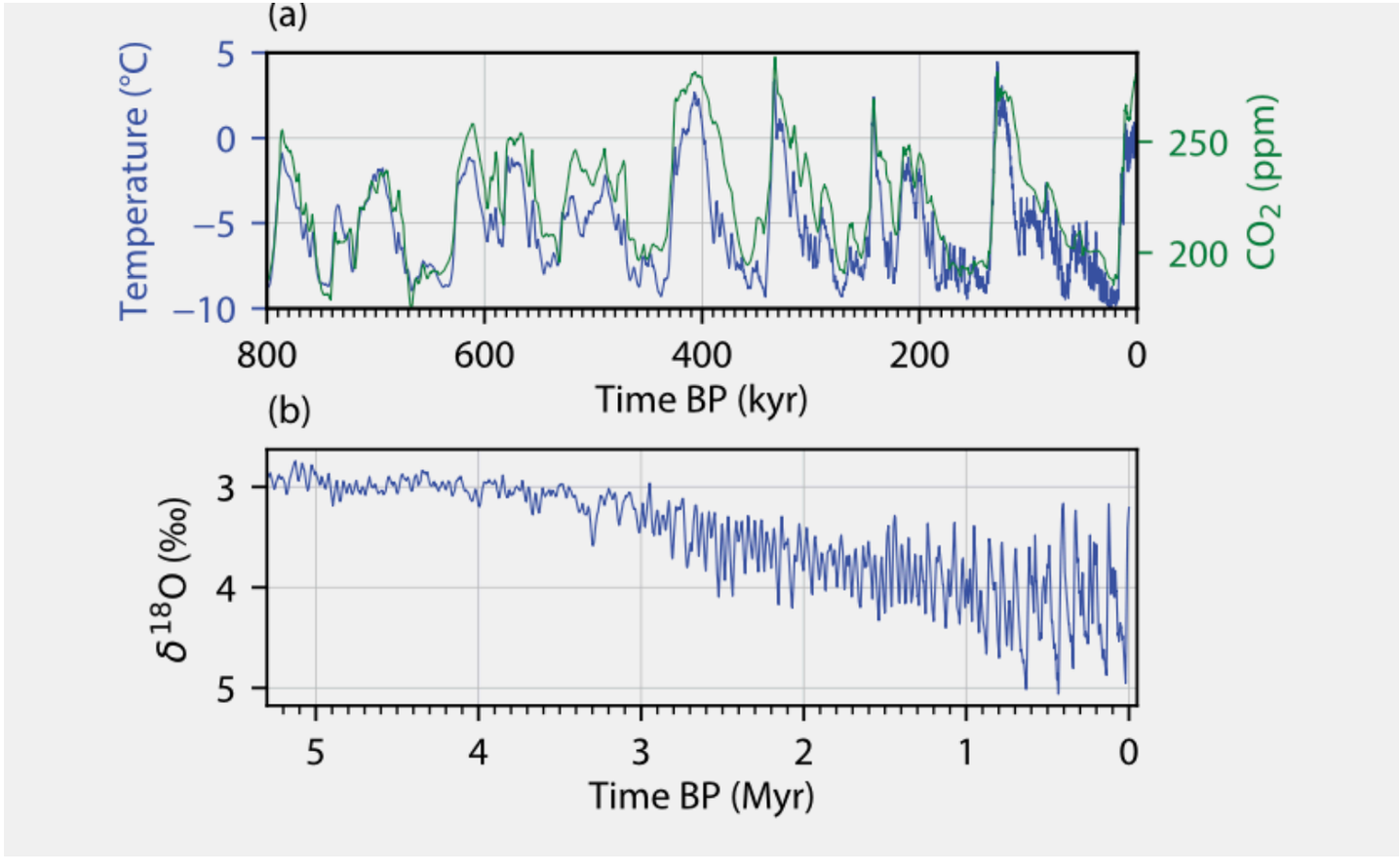
Over the past 800,000 yr, Earth has gone through some eight glacial cycles, with ice ages separated by interglacial periods similar to the one that has existed over the past ~12 kyr (the Holocene); each glacial cycle has lasted about 80–120 kyr.

During the last glacial maximum (LGM, 21 kyr before present [BP]), sea level dropped by about 130 m due to ice buildup over high-latitude continents, global average temperature was some 4 ^{o}C lower than at present, and atmospheric CO_2 dropped to 180 ppm, a significant reduction relative to the 280 ppm concentration during interglacial periods (green line in panel a in the accompanying figure).

The temperature (e.g., the Antarctic temperature estimated from ice core data shown by the blue line in panel a) and land ice volume records have a characteristic saw-tooth shape representing gradual cooling and ice buildup over 90 kyr followed by a faster warming and melting over 10 kyr.

During the LGM, northern North America was covered by an ice sheet of thickness of up to 3–4 km that extended all the way to present-day New York City, and Northern Europe was similarly ice covered.

Panel b shows a deep ocean sediment isotopic record ($\delta^{18}O$, reflecting the fraction of different oxygen isotopes in buried plankton shells), which is a measure of both ice volume over land and the deep ocean temperature.



It shows that ice ages began 2.7 Myr ago following a gradual cooling since the previous warm climates (Box 4.1) as shorter-period 41 kyr cycles, changing to a larger amplitude and a longer period of about 100 kyr around 0.8 Myr.

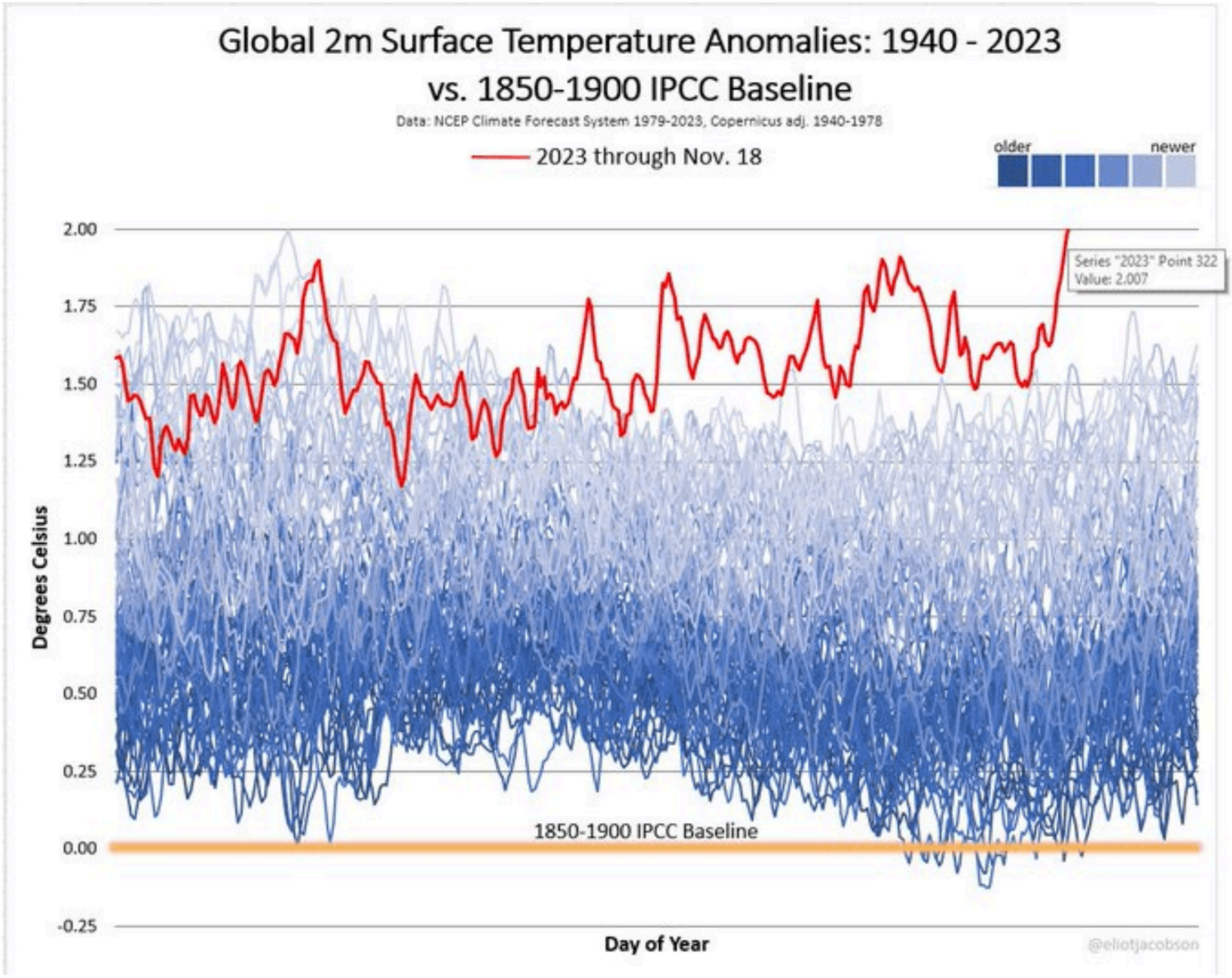
Figuring out the causes of glacial cycles is a test of our understanding of climate dynamics.

In spite of significant progress, decades of studies of ice ages—which are the largest climate variability signal over the past million years—have still not yielded a consensus on the mechanism of the cycles, the reason for their timescale or their change of period at 0.8 Myr, or the reason for CO_2 changes.

Successfully modeling the detailed climate of the LGM would increase our confidence in climate models that are used for future projections.

A precise knowledge of the state of the climate system during ice ages also has consequences for present-day sea level studies, because the Earth crust is still rebounding from the weight of the LGM ice sheets, resulting in current regional sea level changes (as reviewed in section 4.2.2).

We were not expected to breach 2.0 Celsius until 2060, but we did Friday. Yay for us!

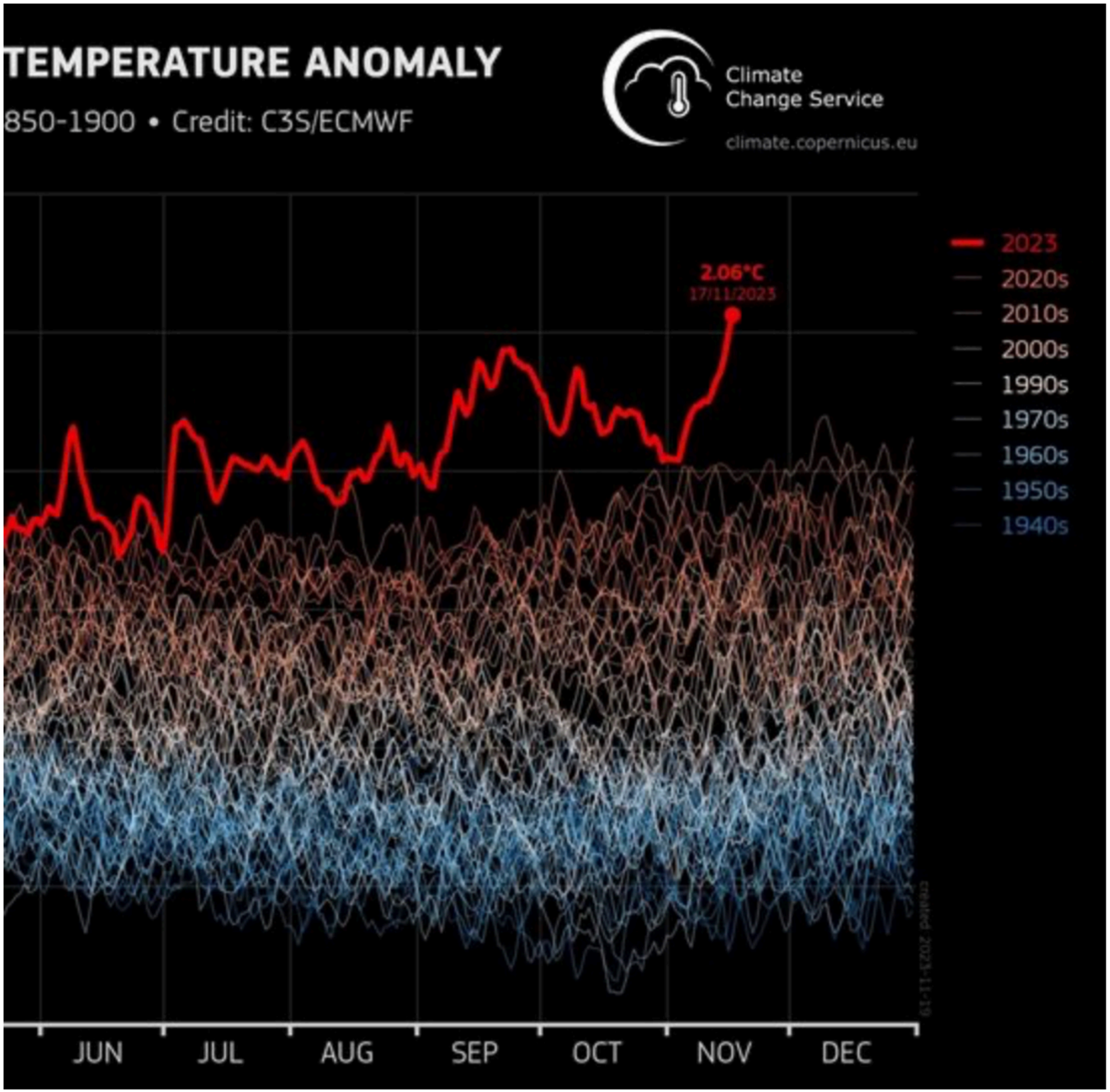


On November 18, 2023, the global mean temperature briefly exceeded 2.0 degrees Celsius (2.06) for the first time in recorded history, exceeding the IPCC 1850-1990 baseline (considered pre-industrial).

The terrifying rise is temporary, according to climate scientists I follow. The long-term average is below 1.5 degrees Celsius, 1.3 C; however, with **record-high climate-warming gases** pouring into the atmosphere, even that temperature will rise.

Two data sets, one by Doctor Elliott Jacobson and one by Leon Simons, confirmed the breach.

This is what unchartered territory looks like. Let’s hope it is not too late to prevent the worst impacts.



Attribution: Copernicus EU

‘No one wants to be right about this’: climate scientists’ horror and exasperation as global predictions play out.

We knew by the mid-1990s that lurking in the tails of our climate model projections were monsters: monstrous heatwaves, catastrophic extreme rainfall and floods, subcontinental-scale wildfires, rapid ice sheet collapse raising sea level metres within a century. We knew – just like we know gravity – that **Australia’s Great Barrier Reef** could be one of the earliest victims of uncontained global warming.

But as today’s monstrous, **deadly heatwaves overtake large parts of Asia, Europe and North America** with temperatures the likes of which we have never experienced, we find even 1.2C of global warming isn’t safe.

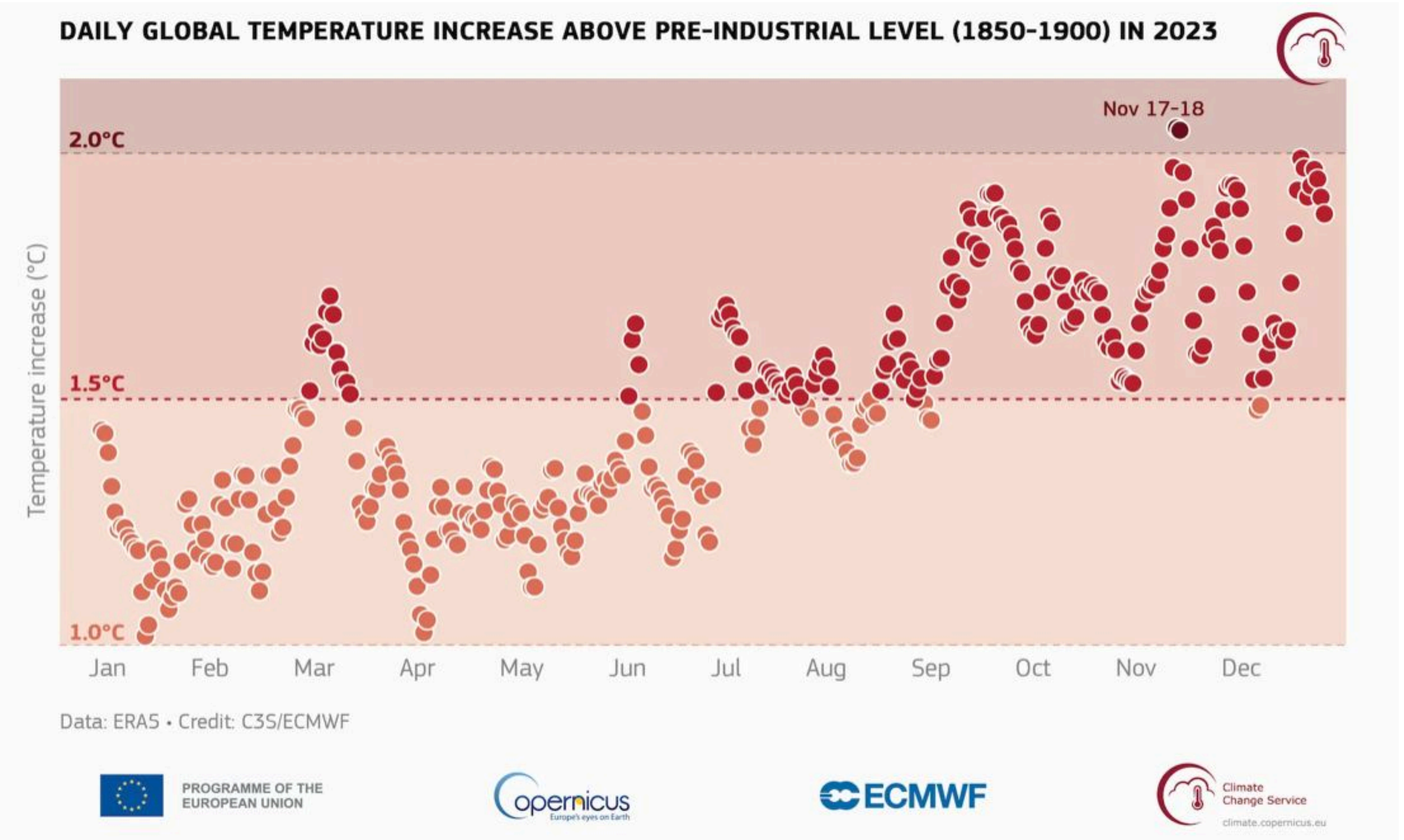
Driving all this is the fossil fuel industry. Enabling it are political leaders unwilling to bring this industry under control and who promote policies such as offsetting and **massive gas expansion** that simply enable this industry to continue.

- Bill Hare, physicist and climate scientist and chief executive of Climate Analytics

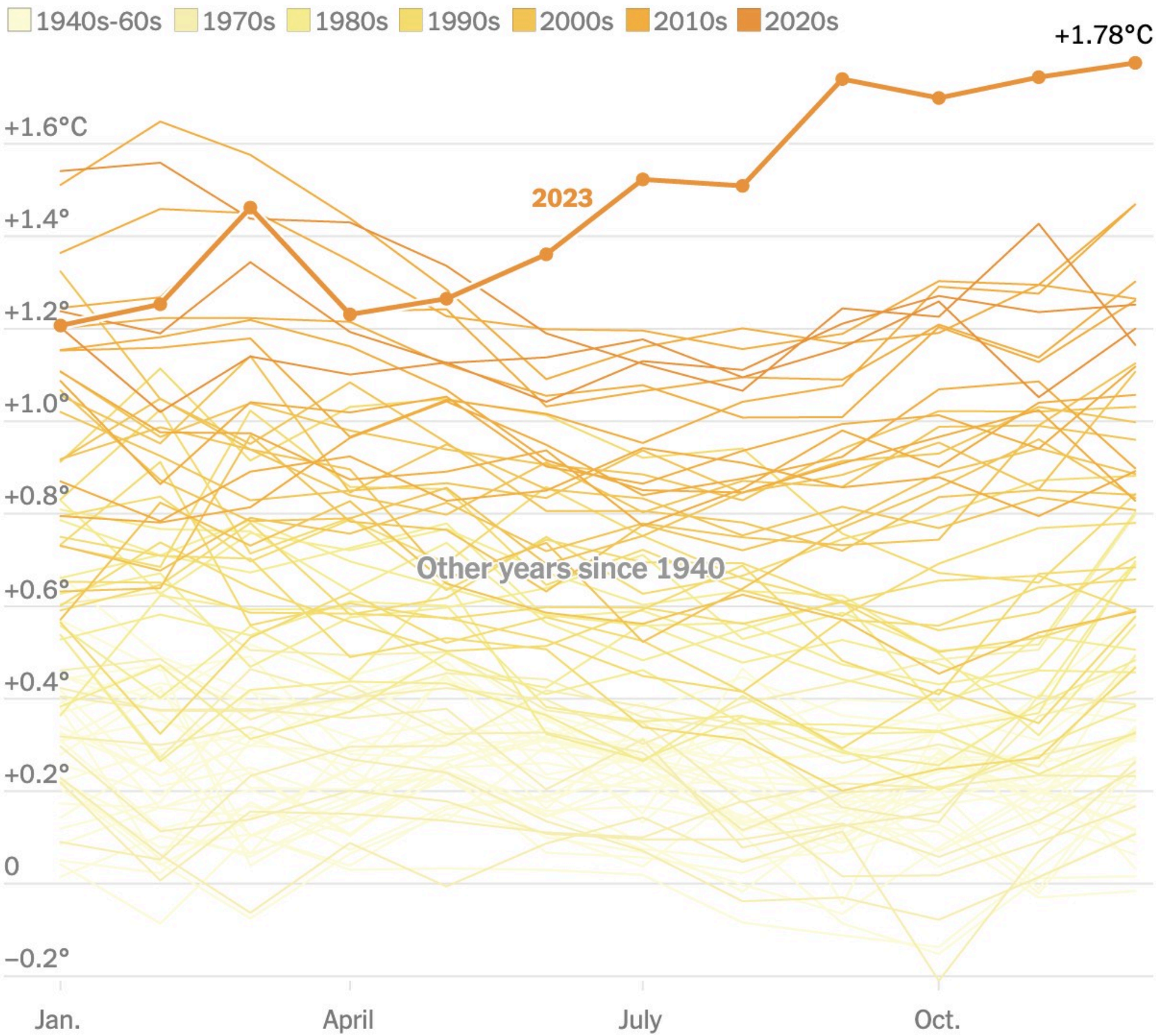
‘No one wants to be right about this’: climate scientists’ horror and exasperation as global predictions play out.

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>
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>
> - Bill Hare, physicist and climate scientist and chief executive of Climate Analytics

More late stuff….



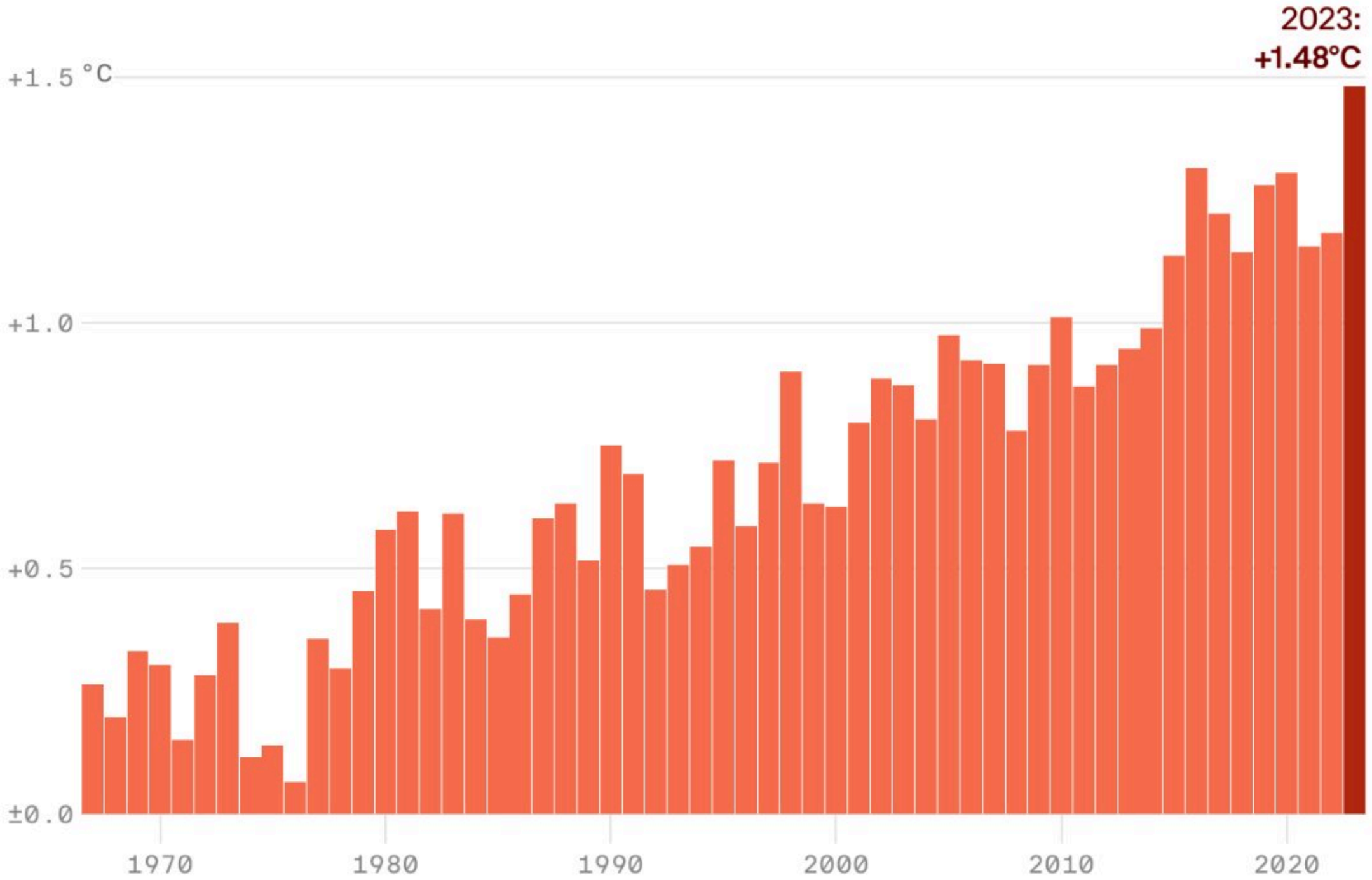
Monthly global temperature compared with pre-industrial levels



Source: Copernicus/ECMWF

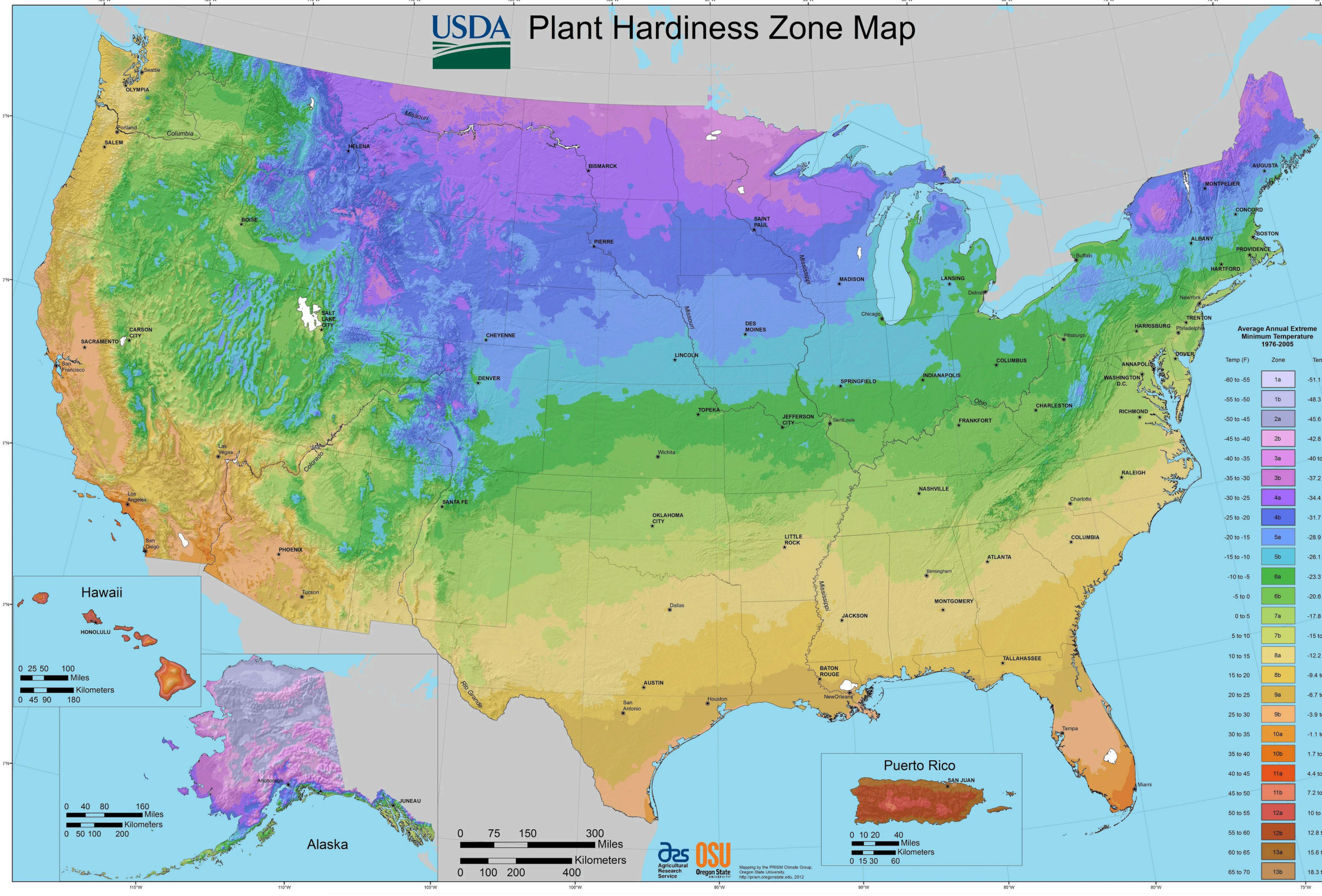
Annual average global surface temperature anomalies

Relative to pre-industrial levels (1850–1900); Annually; 1967–2023



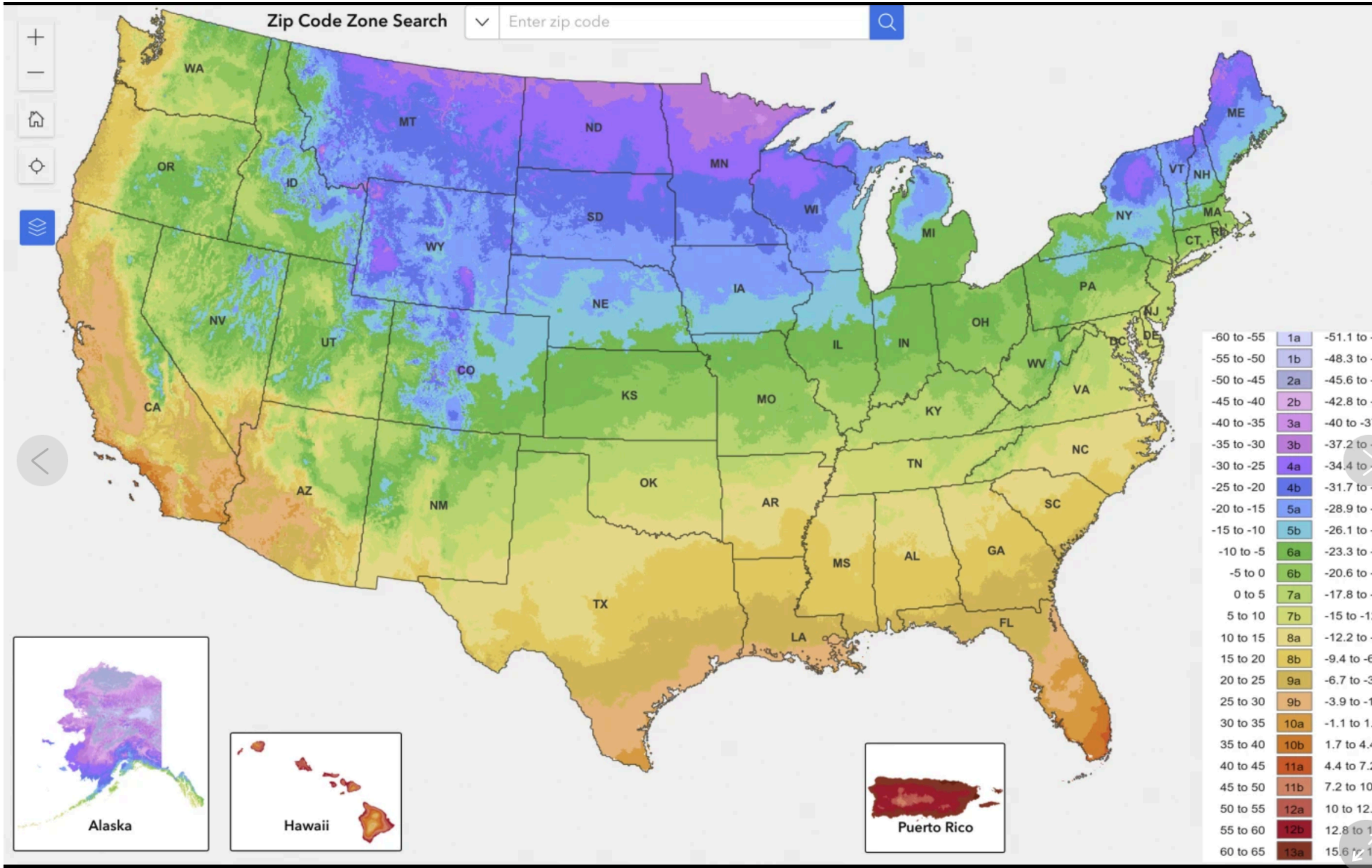
Data: ERA5, Copernicus Climate Change Service; Chart: Alice Feng/Axios

USDA Plant Hardiness Zone Map



Average Annual Extreme Minimum Temperature 1976-2005

Temp (F)	Zone	Temp (C)
-60 to -55	1a	-51.1 to -48
-55 to -50	1b	-48.3 to -45
-50 to -45	2a	-45.6 to -42
-45 to -40	2b	-42.8 to -40
-40 to -35	3a	-40 to -37.2
-35 to -30	3b	-37.2 to -34
-30 to -25	4a	-34.4 to -31
-25 to -20	4b	-31.7 to -28
-20 to -15	5a	-28.9 to -26
-15 to -10	5b	-26.1 to -23
-10 to -5	6a	-23.3 to -20
-5 to 0	6b	-20.6 to -17
0 to 5	7a	-17.8 to -15
5 to 10	7b	-15 to -12.2
10 to 15	8a	-12.2 to -9.4
15 to 20	8b	-9.4 to -6.7
20 to 25	9a	-6.7 to -3.9
25 to 30	9b	-3.9 to -1.1
30 to 35	10a	-1.1 to 1.7
35 to 40	10b	1.7 to 4.4
40 to 45	11a	4.4 to 7.2
45 to 50	11b	7.2 to 10
50 to 55	12a	10 to 12.8
55 to 60	12b	12.8 to 15.6
60 to 65	13a	15.6 to 18.3
65 to 70	13b	18.3 to 21.1



1 of 2 | This image from the U.S. Department of Agriculture shows the agency's new plant hardiness zone map updated on Wednesday, Nov. 15, 2023. The map was updated for the f
time in a decade, and it shows the impact that climate change will have on gardens and yards across the country. (USDA via AP)

Circuits of Change **MODELING THE FUTURE CLIMATE - Simulations**

In 1957, a study from the Scripps Institution of Oceanography described the intersection of fossil fuels and global climate change in words that proved far more resonant than one might have imagined at the time.

The landmark study included this oft-quoted passage:

> ***Human beings are now carrying out a large-scale geophysical experiment of a kind that could not have happened in the past nor be reproduced in the future. Within a few centuries we are returning to the atmosphere and oceans the concentrated organic carbon stored in sedimentary rocks over hundreds of millions of years.***

Like any good scientists, we must be curious about the outcome:

"This experiment, if adequately documented, may yield a far-reaching insight into the processes determining weather and climate."

It may seem callous to describe our current situation—glaciers melting, cities baking, species disappearing—as an interesting experiment.

It's also quite accurate.

Each year our emissions push the chemistry of the atmosphere a little further into territory never before explored in human history.

If there were a million Earths in our solar system, each with a different concentration of greenhouse gases, we'd have a better idea of just what to expect in the future.

Unfortunately, we can't replicate the experiment that we're now conducting on planet Earth.

The next best thing is to simulate it, and the only place that's feasible is inside a computer.

Reproducing the atmosphere with **computer models** has been a key activity in computer science ever since the field was born, just after World War II.

At first the emphasis was on weather prediction. By the 1960s, forecasters were using 3-D models of atmospheric flow, or **general circulation models**, to gaze into the future of weather patterns several days out.

Since then, breakthroughs in weather and computer science, together with Moore's Law (the maxim that computer processor speed doubles about every 18 months), have allowed for steady improvements in weather forecasting models and thus in forecast skill. At the beginning, one might have been happy to have a solid three-day forecast; now, 8- to 10-day outlooks with some measure of skill are commonplace.

The most obvious weather changes emerge from the interplay among a fairly limited set of ingredients in the atmosphere: pressure, temperature, moisture, wind, and clouds.

Weather is also shaped by other factors in the surrounding environment, such as ocean temperature, vegetation, and sea ice.

But these latter factors don't change much over the course of a few days, so a weather model can focus on tracking the atmosphere while keeping the rest of the environment constant.

Climate models don't have that luxury.

As the weeks roll into months, vegetation thrives and decays.

Sea ice comes and goes. Greenhouse gases accumulate.

The ocean gradually absorbs heat from our warming atmosphere, sometimes releasing it in giant pulses during El Niño events.

In the very long term, even the topography of Earth changes.

All of these variations feed into the atmosphere and influence weather—often so subtly that the effect isn't obvious until it's been playing out for years.

For all of these reasons, it's extremely difficult to go beyond a weather model's focus on the atmosphere and to depict Earth's whole environment accurately in a **global climate model.**

Difficult doesn't mean impossible, though.

Climate modeling has undergone a rapid transformation in the last 30-plus years.

As recently as the 1980s, the most sophisticated climate models portrayed the atmosphere and the land surface—and that was about it.

Gradually, scientists have forged models that depict other parts of the Earth system.

Many of these are incorporated in ever-growing global climate models.

The best of the bunch now include land, ocean, sea ice, vegetation, and the atmosphere, all interacting in a fairly realistic way.

Global models are far from ideal; as contrarians often point out, there are still major uncertainties about climate processes that no model can resolve.

However, the skill of recent global models in replicating twentieth-century climate, as we'll see below, is one sign that the models are doing many things right.

Climate, Weather, and chaos

One of the most common gripes about climate change projections is the idea that a computer model can tell us anything worthwhile about climate in the distant future.

It's often phrased this way: "If they can't get the forecast right for next week, how can they predict the climate a hundred years from now?"

Even the late Michael Crichton employed this tactic often.

A weather forecast and a climate projection are two different beasts.

Weather models can't see clearly beyond a few days because of inherent limits to the predictability of small-scale weather (i.e., what will happen in your neighborhood, as opposed to the globe as a whole).

Weather observing stations are situated a few kilometers or miles apart at best, and in much of the world you can go 100 km (60 mi) between stations, not to mention the largely unsampled oceans.

Small disturbances missed in the data can influence larger weather events over time-the "butterfly effect," as discovered in the 1960s by Edward Lorenz, the father of chaos theory.

We'll never have enough weather stations to catch every one of these tiny weather makers.

That's why we can't give a specific temperature forecast in February for the first day in August, although we can say with great confidence for most midlatitude cities that the month of August will be warmer than February.

Climate models aren't interested in the vagaries of individual weather events so much as the influence of long-term climate shapers such as greenhouse gases and polluting aerosols.

Chaotic variations in weather normally average out over years and decades, so they don't corrupt a long-term climate projection the way they might a short-term weather forecast.

The downside is that global models can't provide the precision in time and space that many policymakers and the public want to see, although progress is being made through techniques such as "downscaling" from global output to regional depictions.

How computer modeling got started

The best analogy to a climate model may be a state-of-the-art computer game.

Instead of cars tearing down a racecourse, or warlords fighting over territory, this game follows the heating, cooling, moistening, and drying of the atmosphere.

This isn't quite as sexy, but it's far more important to the future of the planet.

The action unfolds in a virtual space that's typically divided into a three-dimensional grid of rectangles.

Imagine the steel latticework in a building under construction, but extended across the entire surface of the planet.

At the center of each rectangle formed by the intersection of the imaginary beams is a **grid point**, where atmospheric conditions are tracked as the model's climate unfolds.

In the atmospheric part of a typical global model, each grid point accounts for a horizontal area of roughly 1.25° longitude × 1.25° latitude.

At midlatitudes (40°N), that represents an area of around 140 km × 100 km (87 mi × 62 mi)—about the size of Connecticut or roughly half the size of Sicily.

Some climate models now have a global atmospheric resolution as fine as 0.5° latitude and longitude, which corresponds to a midlatitude area of about 56 km × 43 km (35 mi × 27 mi), roughly the size of a good-sized city and its suburbs.

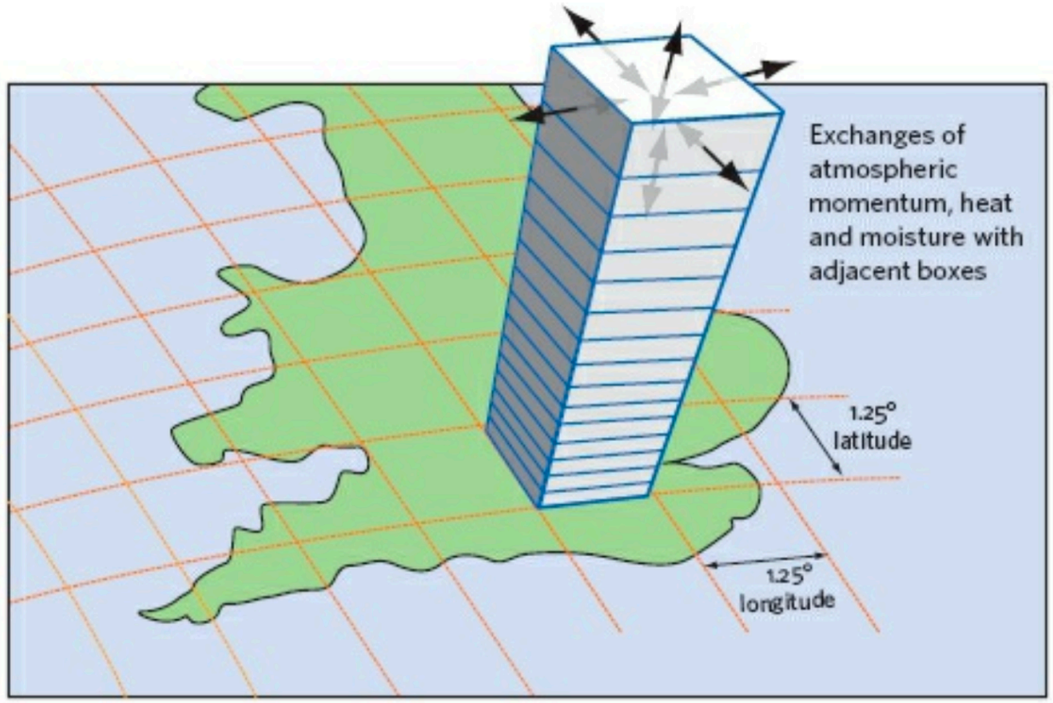
That's still a much larger area than represented by the grid points in a high-end weather model.

Toward the poles, the width of each rectangle shrinks toward zero, which calls for special computational treatment.

Some experimental weather and climate models are now replacing the 3-D global latticework with a hexagonal structure, resembling a soccer ball with thousands of tiny faces, that makes for smoother and more efficient computing.

The typical global model also includes perhaps 25 or so vertical layers.

Each layer spans anywhere from 100 to 500 m (330–1650 ft) close to Earth's surface, with the vertical distance between grid points increasing as you go up.



A climate model divides the atmosphere into 3-D chunks and measures the change in each one as weather unfolds over days, years, and decades of model time.

A climate model operates in a parallel universe of sorts, one in which time runs far more quickly than in real life.

Each model advances in **time steps** that move the simulated climate forward by anywhere from five minutes to a half hour, depending on the model's complexity.

At each time step, the model calls on formulas derived from the laws of physics to compute how all the atmospheric and environmental variables have changed since the last time step—the amount of sunlight hitting each spot on Earth, the convergence of winds, the formation and dissipation of clouds, and so forth.

Each step only takes a few seconds to process, but it can still take weeks of dedicated time on a supercomputer to depict a century of climate in detail.

The output from all this number crunching is saved in various formats along the way.

After the processing is done, scientists can go back, sift through the results, calculate yearly averages and other statistics, and produce graphics.

In order to depict long-term climate change, scientists have to provide the model with input on how the environment is changing.

One critical variable is fossil fuel emissions.

Modelers rely on two major types of simulations to show how an increase in greenhouse gases affects climate.

In both cases, carbon dioxide is traditionally used as a stand-in for all human-produced greenhouse gases in order to save on computing time and expense, though in recent years the most detailed models have also incorporated methane and other key gases.

- **Equilibrium** runs typically start off with an immediate, massive injection of carbon dioxide—for instance, enough to bring the airborne amount to twice or even four times the preindustrial level. The model's climate gradually responds to this extra CO_2 and eventually attains equilibrium at some new (and presumably warmer) state. The idea isn't to see how long the climate takes to respond, since we'd never add that much CO_2 instantaneously in the real world, but to see where the climate ends up.

- **Transient** runs more closely resemble reality. In these simulations, the CO_2 concentration in the model atmosphere is typically boosted in smaller increments that correspond to the actual amounts added by human activity—typically 1% per year, compounded year over year. In this case, scientists are interested as much in the process as the outcome. For instance, does the model climate warm smoothly, or does the temperature increase in fits and starts? The various global models have tended to show closer agreement for transient runs than for equilibrium runs—an encouraging sign, as the transient scenario is closer to how greenhouse gases increase in real life and is thus more relevant to climate policy over the twenty-first century.

Of course, you can't simply add CO_2 to a global model and leave everything else alone.

As discussed earlier, a large part of that CO_2 gets absorbed by vegetation and the oceans.

There are also other processes that kick in as carbon dioxide warms Earth.

Consider the sea ice that melts as the atmosphere warms up.

As we saw earlier, the loss of that ice means less reflection of sunlight and an amplifying, positive-feedback loop that leads to even more warming.

The positive feedback from water vapor also has to be portrayed accurately: the warming produced by the extra CO_2 helps evaporate more moisture from the sea, and that water vapor raises global temperature further.

These environmental processes are quite different from the physics that drives day-to-day weather, and they can't be added to a climate model in an easy or simple way.

The general circulation models of the 1960s and 1970s couldn't hope to account for these processes in a realistic fashion, although more simple models of energy balance confirmed the big-picture conclusion that greenhouse gases would lead to global warming.

As computing power has improved, global climate models have grown steadily in sophistication and detail.

They have incorporated a growing number of submodels, dealing with each of the major climate-influencing elements.

When an atmospheric model is yoked to one or more of these submodels, with information shared among each, the new hybrid is called a **coupled model** (even when it has more than a couple of components).

If such a model also includes the flow of carbon into and out of the atmosphere, it's dubbed an **Earth system model.**

In a simple coupling, a submodel runs side by side with a global model and passes it information—a one-way dialogue, as it were.

For instance, a global model might be fed information on airborne nitrogen levels from a chemistry model, but the resulting changes in climate wouldn't feed back into nitrogen production, as they would in real life.

More realistic are **integrated** couplings, which allow two-way exchanges of information between model and submodel.

However, it can take years to develop such linkages and verify their accuracy.

Climate modelers have typically avoided incorporating current weather conditions when projecting the climate decades into the future, since there's so much day-to-day variability in weather.

However, bringing current ocean conditions into the model can help nail down how global temperature is likely to unfold over the next decade.

The Met Office Hadley Centre began using this strategy in the late 2000s.

By running a global model in two ways—either including or omitting the patchwork of present-day ocean temperatures—the group has shown that incorporating current ocean conditions, and the "memory" they bring to the atmosphere, can add measurable benefit to a climate outlook.

The Met Office began issuing 5- and 10-year forecasts of global average temperature on a routine basis starting in 2009.

Such forecasts are now showing at least a modest level of skill.

Ocean conditions were also fed into 10- and 30-year model runs carried out in support of the 2013 IPCC report.

Some of these runs were launched with the state of the ocean in the years 1960 or 1980, in order to see how well the modeled climate matched the observed conditions.

Other 10- and 30-year runs peered into the future; these began with ocean conditions in 2005.

With these simulations and related work to draw on, the IPCC's 2013 assessment was the first to provide near-term climate projections.

The panel foresees a likely global temperature rise of 0.3°–0.7°C (0.5°–1.2°F) for the period 2016–2035 compared to 1986–2005, barring a major volcanic eruption or a sustained change in solar output.

Especially when the focus is on a weather or climate feature with restricted geographic extent, such as hurricanes, it makes sense to limit the model's extra-sharp precision (and the related computing burden) to the areas where it's most needed.

This can be done using a **nested grid**, whereby the globe as a whole is simulated at a lower resolution and the areas of interest are covered at higher resolutions.

Similarly, when studying how climate might evolve in a fixed area, such as parts of a continent, global models are often coupled to **regional climate models** that have extra detail on topography.

The output from nested global models and regional models is especially useful in creating the climate assessments that are sorely needed by policymakers.

However, bumping up a model's resolution isn't a simple matter: it involves a careful look at the physics of the model to ensure that processes that don't matter so much in a low-resolution simulation, such as showers and thunderstorms, are realistically depicted at the new, sharper resolution.

Otherwise, improving a model's resolution can actually degrade its performance.

Some of these issues are avoided in the variable-grid structure described above.

By simulating the last century of climate, scientists can see if a model accurately depicts the average global temperature as well as regional and seasonal patterns, ocean circulations, and a host of other important climate features.

Once it's passed this test, a model can go on to project the climate of the future—or replicate the distant past, the goal of paleoclimate modeling.

Another safety check is that the world's leading models are regularly checked against each other through such activities as the Program for Climate Model Diagnosis and Intercomparison, conducted at the U.S. Lawrence Livermore National Laboratory.

When the models agree, that bolsters confidence in the results; if one model deviates markedly from the others, it's a flag that more work may be needed.

One of the program's main activities is running a mammoth Climate Model Intercomparison Project (CMIP), which is designed to structure and archive the modeling carried out at various centers around the world in support of each IPCC assessment.

The 2013 IPCC report draws on the suite of modeling runs known as CMIP5.

Researchers embarked in the late 2010s on a new round of modeling, dubbed CMIP6, that will feed into the next major IPCC assessment, slated for the early 2020s.

The limitations

In both weather and climate models, it's not easy to calculate every iota of atmospheric change.

Clouds, for example, are still a tremendous challenge.

If they change in a systematic way, clouds can have major implications for how much radiation gets reflected and absorbed from both the sun and Earth.

But clouds are far more varied and changeable in real life than any model can accurately represent, so modelers have to simplify them in their simulations.

There's currently no gold standard for how to do this, and new techniques are being explored all the time.

Typically, each grid point in a model will be assigned an amount or percentage of cloud cover on the basis of how much water vapor is present and whether the air is rising or sinking (rising air is conducive to cloud formation).

The type of cloud and its reflective value are estimated from the height and temperature of the grid point.

To a greater or lesser degree, climate models also include **aerosols**—pollutants, dust, and other particles that can shield sunlight and serve as nuclei for cloud droplets and ice crystals.

This allows scientists to see how changes in pollution might affect cloud formation and temperature, as in the case of the global dimming observed in the late twentieth century.

There's been substantial progress in modeling the direct effects of aerosols on sunlight and other radiation.

All told, this direct influence appears to be a net cooling, even though some aerosols, such as soot, tend to absorb more sunlight than they reflect.

Aerosols can also influence climate indirectly by shaping the formation of clouds.

Although the cloud–aerosol interaction has been persistently tough to observe and model, the best estimates to date point toward a slight overall cooling effect.

Aerosols can also fall onto snow and ice and darken their surface, which hastens melting and slightly counterbalances the cooling effects noted above.

In order to test whether their cloud schemes are working well, scientists run their models for present-day climate and then compare the overall distributions of cloud cover to the real-world situation observed by satellite.

These datasets have their own limitations, though, since most satellites can't yet delineate multiple cloud layers very well.

The results are also compared to special cloud-resolving models that are too expensive to run routinely but that provide a valuable cross-check.

Oceans are also tricky.

Real-life oceans feature a **mixed layer** on top, where the turbulence of waves and close contact with the atmosphere help keep the temperature fairly constant.

The mixed layer may be as shallow as 10 m (33 ft) or as deep as 200 m (660 ft), depending on season, location, time of day, and weather conditions.

Underneath the mixed layer is a zone of deeper, colder water that interacts with the mixed layer much more slowly.

The earliest climate models depicted the mixed layer as a uniform slab but lacked a deep layer, a problem for long- term simulations.

Fixed oceans were a frequent source of **climate drift**, which occurs when a model-depicted climate gradually veers into unrealistically warm or cold territory.

Today's global models are coupled with much more realistic ocean submodels that allow for the two-way exchange of heat, moisture, chemistry, and momentum among the surface, the mixed layer, and deeper levels, with 25 or more layers now depicted in some models.

Still, the complicated circulations of the real ocean remain hard to portray.

For instance, the North Atlantic's flow is often characterized as a simple loop, with warm water flowing north off the U.S. East Coast toward Europe, cold water streaming back south toward Africa, and warmer waters flowing west with the trade winds toward the Caribbean.

In reality, however, the northward flow branches into several components as it approaches Greenland and the British Isles, and the water descends far undersea into a cold return flow that moves southward in several distinct layers.

Where the contrasts in temperature and salinity are strong, small but powerful eddies can spin off.

All this complexity is hard enough to measure, much less portray in a model.

The tropical Pacific remains one of the most difficult areas for climate models to depict in detail, not only because of ocean circulation but also because of the gigantic areas of rainfall spawned by the warm waters, especially in and near Indonesia.

This warm pool periodically expands eastward or contracts westward during El Niño and La Niña.

Most global models can now depict El Niño and La Niña, and some are able to represent the frequency of their comings and goings fairly well, but their precise geography and long-term behavior are still hard to pin down.

Even in neutral conditions, it's a challenge for models to realistically capture the amount of heat and moisture pumped into the upper atmosphere by tropical showers and thunderstorms.

A sensitive topic: Predicting the warm-up

The question of what will happen in a greenhouse-warmed world is so essential to climate modeling that it's spawned an index that's widely used to assess model performance.

This is the **sensitivity** of a global model, most often defined as the amount of temperature change produced in the model when carbon dioxide is doubled in an equilibrium-style run, as discussed above.

The climate eventually settles down at a new global temperature, and the difference between that temperature and the old one is the sensitivity.

It provides a convenient way to compare models and how readily they respond to greenhouse gas.

This definition isn't cast in concrete: one might consider a tripling or quadrupling of CO_2 instead, and various starting or ending points can be used, such as the peak of the last ice age.

However, the doubling of CO_2 since preindustrial times (generally taken as the mid-1800s) serves as a useful and time-tested convention familiar to all climate modelers.

At the current rate of increase, greenhouse gases will have roughly doubled in the atmosphere by the mid- to late twenty-first century (though we might be able to postpone the doubling, or perhaps even forestall it entirely, through a concerted effort to reduce greenhouse emissions).

The doubling of CO_2 used in model sensitivity gives us a good idea of how much warming to expect once the atmosphere and oceans have adjusted to the extra greenhouse gas, which could take centuries longer than the doubling itself.

Should there be big changes along the way—such as the partial or total loss of major ice sheets—then it could take many thousands of years more for global temperature to settle down.

This potential longer-term rise is usually analyzed separately from the more conventional definition of sensitivity used below.

Even though sensitivity is expressed as a single number, it's actually the outgrowth of all the processes that unfold in a model.

That's why equally sophisticated climate models can have different sensitivities.

That said, what's striking about equilibrium climate sensitivity is how little the average value has changed as models have improved.

Some of the earliest global climate models showed a doubled CO_2 sensitivity of around 2.5°C (4.5°F).

As more models of varying complexity came online, the range of possibilities was estimated to be **1.5°–4.5°C (2.7°–8.1°F.**

This range—something of a best guess at the time—was first cited in a landmark 1979 report on climate change by the U.S. National Academy of Sciences.

That value was reassessed and reaffirmed by the IPCC in its first three major reports and was narrowed only slightly in the 2007 report (to 2.0°–4.5°C or 3.6°–8.1°F).

Most of the major coupled models now show sensitivities clustered within a degree of **3.0°C (5.4°F**, which is the best estimate pegged in the latest IPCC report.

The fact that this value hasn't changed much after more than 40 years of modeling is, arguably, testimony to how well modelers have captured the most important processes at work.

A skeptic might complain that the models have simply been tuned—their physics adjusted— to produce the result that people expect.

Indeed, modelers have to guard against making so many adjustments to a model that it gets a "right" answer for the wrong reasons.

But there's also a physical rationale behind the 3°C number.

It's fairly easy to show that, if the atmosphere had no clouds or water vapor, the warming from doubled CO_2 would be somewhere around 1.0°–1.2°C (1.8°–2.2°F).

It's believed that water vapor at least doubles the impact from the extra CO_2.

Together with smaller effects from other greenhouse gases, this brings the total change into the 3°C ballpark.

In the end, the IPCC report—while considering some of the new work on either side of the spectrum—concluded that the long-accepted interval of 1.5°–4.5°C remained the most likely range of equilibrium climate sensitivity in its 2013 report.

The panel noted, "Even though this assessed range is similar to previous reports . . . confidence today is much higher as a result of high quality and longer observational records with a clearer anthropogenic signal, better process understanding, more and better understood evidence from paleoclimate reconstructions, and better climate models with higher resolution that capture many more processes more realistically."

Clearly, there's still disagreement about just how sensitive climate might be, but recent work reiterates that the most likely outcome of a doubling of carbon dioxide is a global temperature increase close to the one scientists have been projecting for decades: somewhere not too far from **3°C (5.4°F.**

One recent piece of evidence is a 2018 analysis led by the University of Exeter.

This study examined natural variability as depicted in 22 global climate models, looking to see how well each one captured the range of short-term ups and downs in global temperature that occurred from the 1940s to the 2010s.

The best-performing models pointed to an equilibrium climate sensitivity centered at 2.8°C (5.0°F), with a likely range notably tighter than the IPCC's: from 2.2° to 3.4°C (4.0°–6.1°F).

None of this rules out a much more or much less sensitive climate.

It simply adds to the weight of evidence for an amount of warming that's certainly nothing to be cavalier about.

A 3°C rise, after all, would bring Earth to its warmest point in millions of years.

Plus, if greenhouse gas concentrations do reach the benchmark of doubled CO_2, they could easily go further still.

Listening to the models

It's up to policymakers and the public to decide how society ought to respond to the onslaught of data that climate models give us.

Probabilities like the ones above can help immensely.

It may seem confusing to think of a temperature range and a probability at the same time, but that's the direction modelers have gone, and it's the type of output that's most useful in making decisions.

Once upon a time, weather forecasts would describe "possible showers" or "a chance of snow" without ever specifying a number.

That made it hard to tell exactly how likely it was that your picnic would get rained out.

It wasn't until computing became widespread in the 1960s that a new set of statistical weather models enabled forecasters to look at a wide range of outcomes and assign likelihoods.

By the 1970s, "probabilistic" forecasts were the norm and people had grown accustomed to phrasings such as "a 30% chance of rain."

Some modern-day snow forecasts include the odds that total accumulation will end up higher or lower than certain thresholds—a more complex but far more useful prediction.

Climate modeling is now affordable enough that some high-end models can be run a number of times.

Each run might use a different rate of CO_2 increase, for example, to show the many ways the climate might unfold depending on how serious we get about reducing emissions.

Alternately, the model might simulate the same CO_2 increase a number of times, but with starting conditions that vary slightly, in order to see how natural variations in climate affect the results.

These large sets of simulations are called **ensembles**, and they've become increasingly important in climate modeling.

As in the examples cited above, some ensembles include more than 1000 simulations, which was unheard of as recently as the turn of the century.

The sheer size of these ensembles allows scientists to calculate statistics and probabilities that a single model run can't provide.

Apart from creating ensembles, scientists are hard at work making other improvements to their models. Among the current areas of keen interest are weaving in more interactive chemistry, adding more detailed land cover and ecosystems, and capturing the various effects of aerosols more precisely.

One might ask why we ought to believe anything that global climate models tell us.

Some folks who deny significant human-produced climate change pose this question often, and it's healthy for everyone to recognize what climate models can and can't tell us.

Along these lines, the 2013 IPCC report examined results for the period 2046–2065 as well as 2081– 2100, with a separate chapter focusing on near-term projections (2016–2035).

The next major IPCC assessment, scheduled for release in 2021-2022, will include a range of projections extending 10 to 30 years out(see below).

In 2007 *a study from the* University of Exeter urged their fellow scientists to focus their research energies on the time frame from about 2040 to 2060.

For instance, models will always struggle with uncertainty about the starting-point state of the atmosphere, which affects shorter-term weather forecasts.

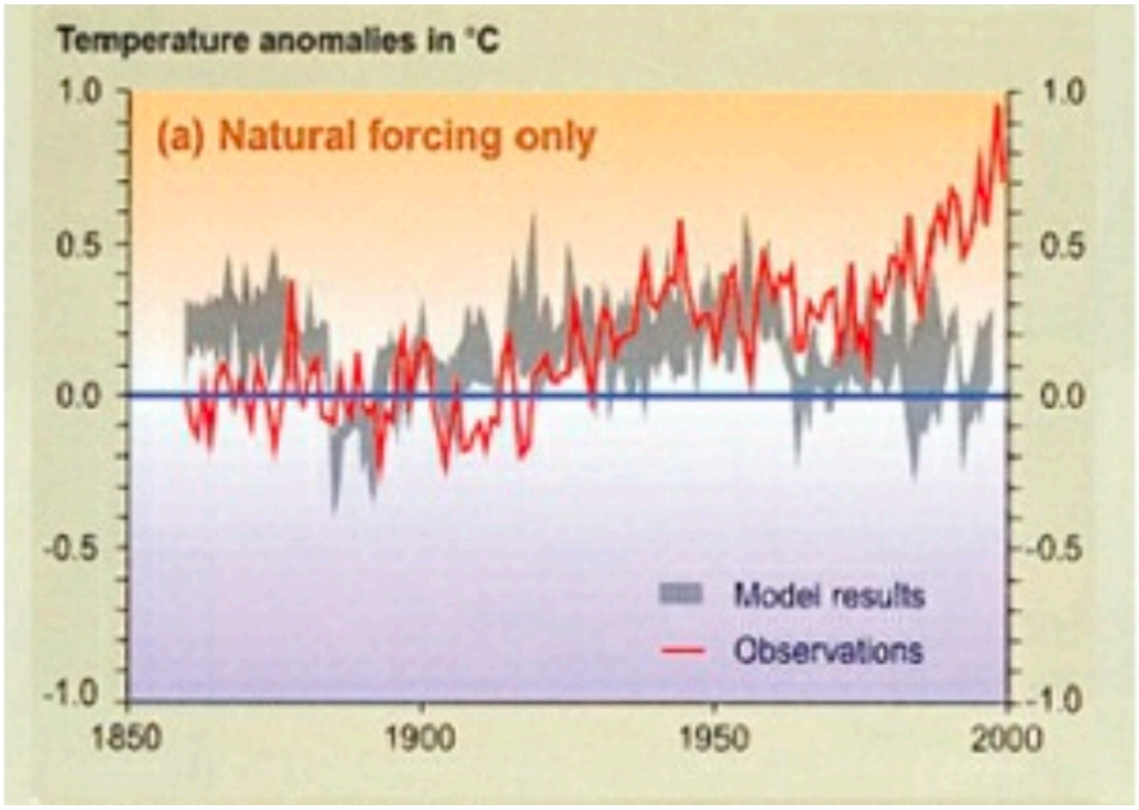
At the same time, our lack of foresight about greenhouse emissions becomes enormous when looking a century out.

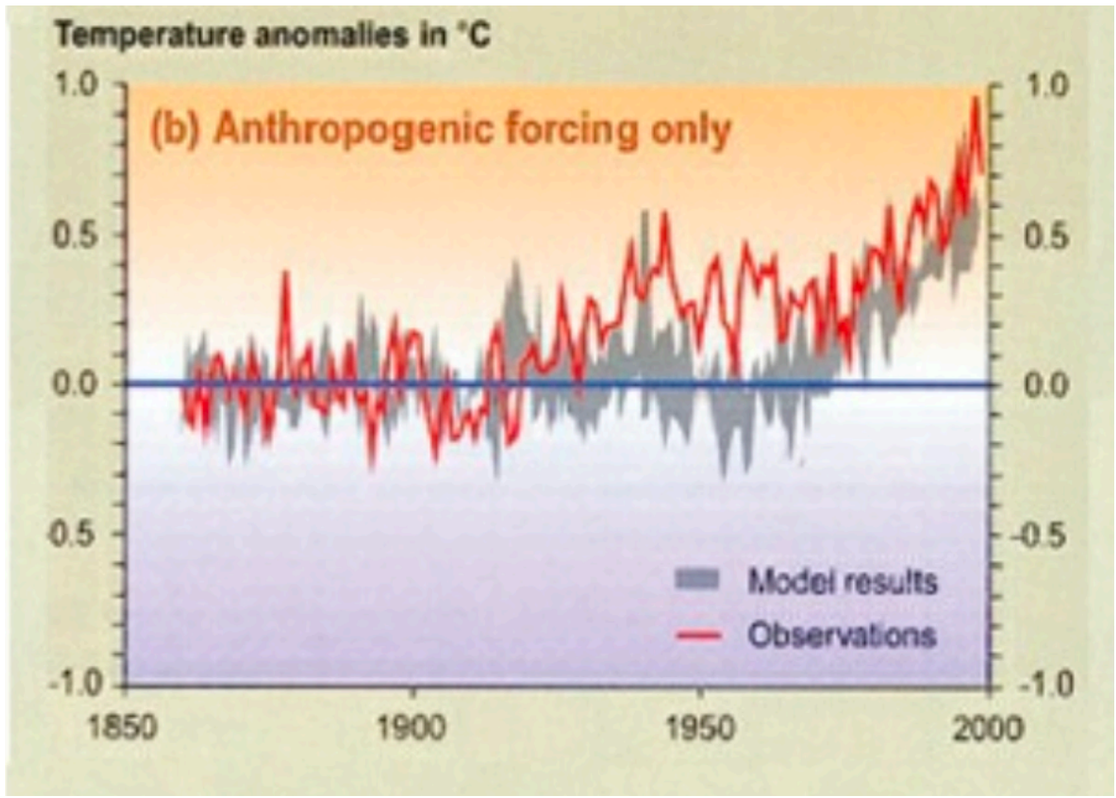
This leads to the interest in “sweet spots”—time frames close enough to be of interest for long-term business and government planning but far enough out so that models can provide insight that other techniques cannot.

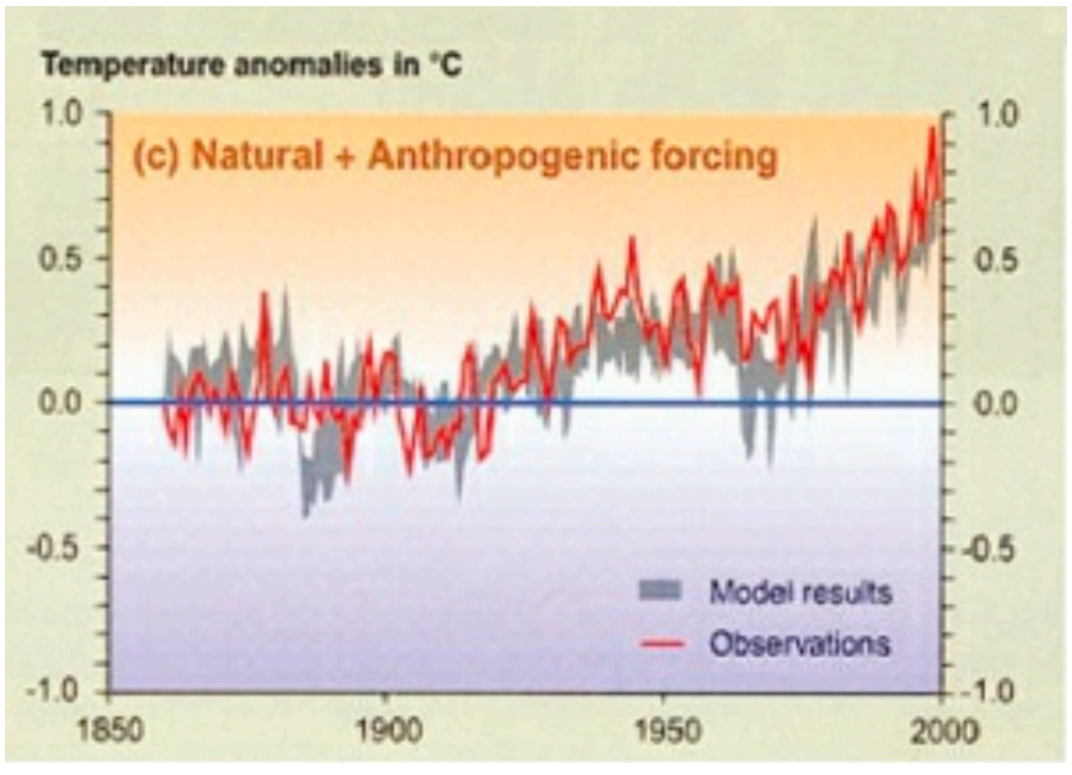
One good reason to be confident in the general quality of today’s climate models is their ability to simulate the past—especially the twentieth century, for which we have by far the most complete climate record.

The three graphs shown next, first prepared for the 2001 IPCC report, have stood the test of time.

They show how one climate model captures the ups and downs in global temperature over the last century.







The model doesn’t reproduce every year-to-year wiggle, but that’s not the idea, just as a weather model doesn’t aim to tell you the exact temperature in your backyard at 8:27 p.m.

What this model does show accurately is the character of twentieth- century climate.

It does so by including the gradual increase in greenhouse gases and sun- shielding pollutants, as well as the effects of natural features like volcanoes and the sun.

When such a model goes on to tell us what might happen as this century unfolds, we’d be foolish to ignore it.

When models look only at natural climate forces(top) or human-produced ones, such as greenhouse gases and sulfate pollution(middle), they fail to accurately reproduce the last century's climate (the red line in each graph).

A model that incorporates both kinds of forces (bot- tom) does a much better job.

The gray shading shows the range of each model ensemble. (IPCC)

Let us begin our effort to understand details……

That will enable us to build simple climate models.

Model #1

First - Some details about Radiation and Energy Balance

Temperature and energy

To a physicist,

energy = capacity to do work

- lift weight, turn wheel, or compress spring.

Unit of energy = joule(J).

1 J = energy required to lift 100 g about 1 m - or to lift apple about 3 ft.

Energy moves from one place to another.

Rate at which energy flowing = power(watts W).

1 watt = 1 joule/sec so 60-W light bulb consumes 60 J of energy every second.

Analogy illuminates difference between power and energy.

Gallon is quantity <-> joule, = quantity of energy.

Rate water flows through pipe measured in gallons/min.

Rate at which energy flows = power,

measured in watts (joules/sec).

Climate is all about energy flows, so climate calculations focus on power.

How much power is needed to run human body?

Human consumes about 2,000 food calories/day.

Calories = unit of energy,

1 food calorie = 4,184 J;

2,000 food calories = 8,368,000 J = 8.37 MJ.

1 day has 24 hours x 60 minutes x 60 seconds = 86,400 seconds

-> dividing 8,368,000 J by 86,400 s yields 97 J /s = 97 W.

-> typical human requires roughly 100 W to power body

- same as power required to run old-style light bulb.

One horsepower = 740 W

-> takes 1/7 of horsepower to run your body.

Internal energy of object -> how fast atoms and molecules in object are moving.

In cup of water, if water molecules moving slowly,

then water has less internal energy than another cup in which molecules moving rapidly.

In solid, movements of atoms approximately fixed in space by intermolecular forces -> why it is solid.

In solids, atoms can still move small distances around fixed positions.

Faster atoms move about fixed positions -> more internal energy.

Temperature = measure of internal energy of object.

As object's internal energy increases and molecules speed up, temperature also increases

for 2 cups of water, one hot and other cold

-> water molecules in hot cup moving faster than water molecules in cold cup, on average.

Along with Celsius temperature scale, is another temperature favored by physicists = Kelvin scale.

Temperature in Kelvin = temperature in degrees Celsius plus 273.15 ($K = C + 273.15$).

Freezing temperature, $0^\circ C = 273.15$ K, boiling temperature, $100^\circ C = 373.15$ K.

"Room temperature" $22^\circ C = 295$ K.

Temperatures in Earth's atmosphere between 200 K and 300 K,

average surface temperature of Earth (today) = 288 K.

Physicists prefer Kelvin scale because temperature expressed in Kelvin is proportional to internal energy.

If temperature doubles from 200 K to 400 K, internal energy also doubles.

If internal energy increases by 10%, then temperature in Kelvin increases by 10%.

0 K is absolute zero -> temperature molecules have 0 internal energy and cease moving;

—> coldest possible temperature.

Physics equations require energy in Kelvin.

Electromagnetic radiation

Climate warmth provided by Sun.

Sun 150 million km from Earth; vacuum of space in between.

How does energy from Sun reach Earth?

Energy transported from Sun to Earth by electromagnetic radiation.

Electromagnetic radiation includes visible light, X-rays, microwaves, and radio-frequency waves.

Electromagnetic radiation = stream of photons = small discrete packages of energy.

As photons travel from Point A to Point B - Sun to Earth -

each carries small amount of energy -> how energy transported.

Photons have characteristic size = wavelength

-> how photons interact with world.

Photons with wavelengths of between 0.3 and 0.8 micron

(a micron, abbreviated as μm = millionth of a meter;

human hair is 100 μm in diameter)

if can be seen with human eye -> visible photons.

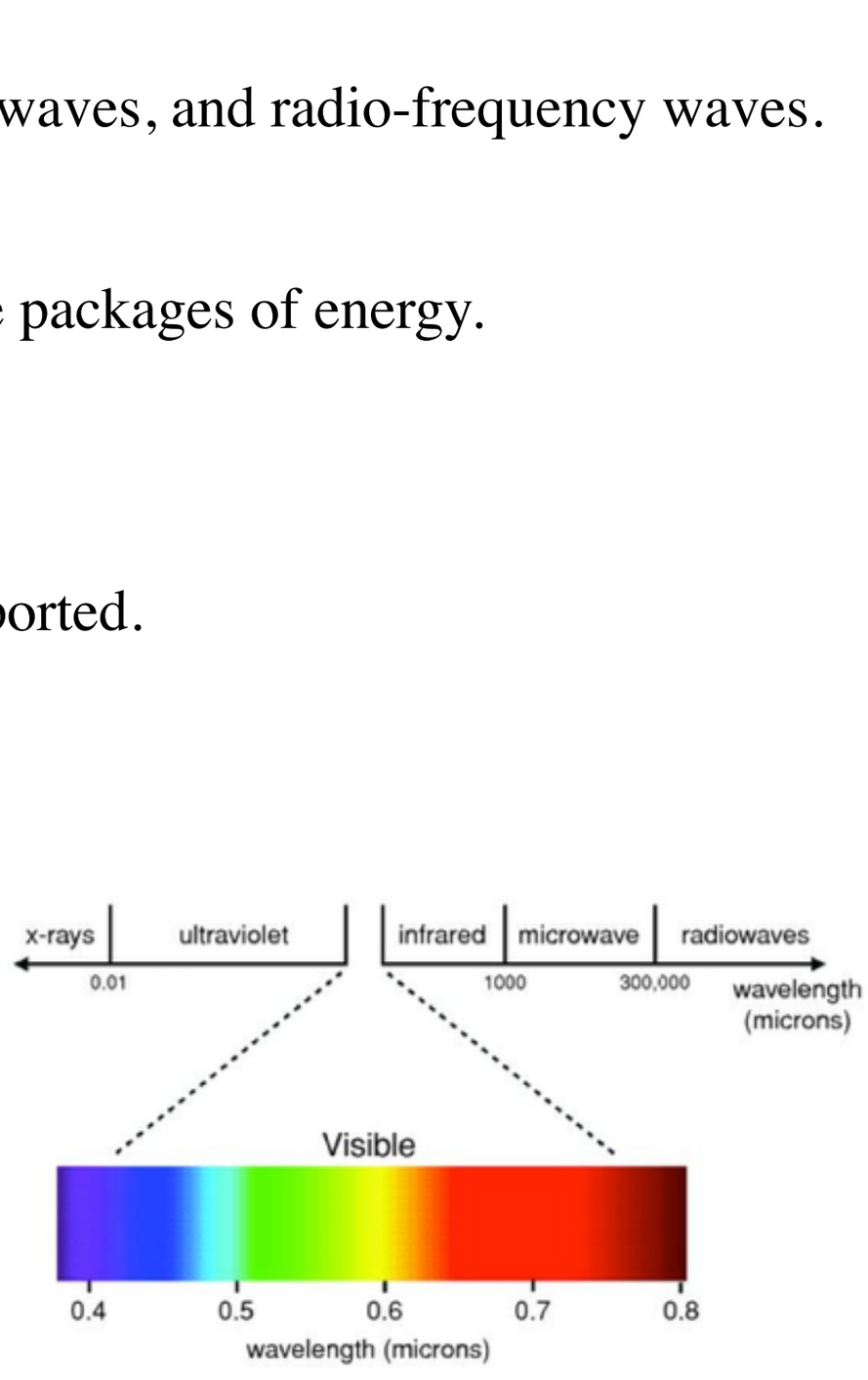


Figure 3.1: The electromagnetic spectrum. Note that the visible part makes up only a minor part of this spectrum.

Within visible range, different wavelengths

appear to human eye as different colors (Figure 3.1).

Humans see photons with wavelengths near 0.4 μm as blue, 0.6 μm as yellow, and 0.8 μm as red.

Photons with longer wavelengths, 0.8 μm to 1,000 μm = infrared

- they are beyond red end of visible spectrum.

Despite being invisible to humans, these photons

play an important role in both Earth's climate and in everyday lives.

Photons with wavelengths just below human detection limit of 0.3 μm

called ultraviolet because wavelength beyond violet end of visible spectrum.

Photons with wavelengths between 1,000 μm (1 mm) and 0.3 m = microwaves,

and photons in this wavelength range used from cooking to radar.

Wavelengths longer than about 0.3 m are radio frequency waves.

Wavelength determines photon's physical properties - how it interacts with matter,

i.e., visible and infrared photons cannot pass through walls,

but radio-frequency photons can.

Human eye can detect visible photons but not infrared or microwave photons.

Full body scan at airport security checkpoint, machine using microwaves

- goes through clothes but stopped by denser materials such as flesh, bomb, or gun.

Finally, atmosphere transparent to visible photons but not infrared photons;

-> enormous implications for our climate.

Blackbody radiation

For both Sun and lamp emitting photons - can see visible photons emitted.

Not, however, only things around you emitting photons.

In fact, everything around you is emitting photons.

You are emitting photons, walls of room, your table, etc.

Everything!

If everything emitting photons,

then why does everything not glow like light bulb?

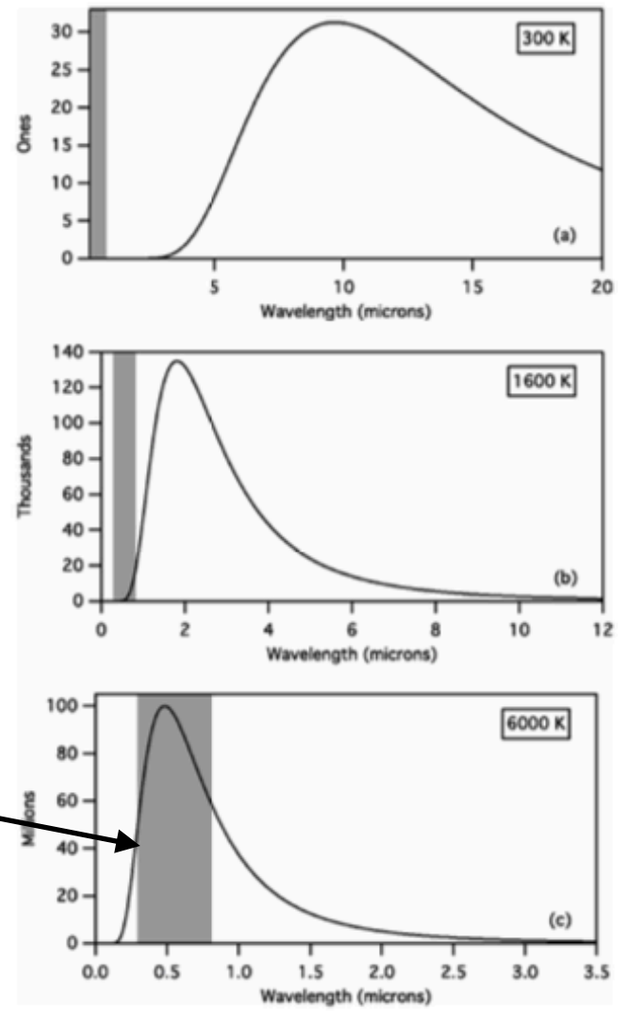
Turns out wavelength emitted

determined by object's temperature.

Figure 3.2 plots emissions spectra

for idealized objects called "blackbodies" at 3 temperatures.

note very different vertical scales



visible

Figure 3.2: Power emitted at different wavelengths from objects (with surface area of 1 m^2) at three temperatures: (a) 300 K, (b) 1,600 K, and (c) 6,000 K The vertical axes are in units of 1 W/μm, 1,000 W/μm, and 1 MW/μm of wavelength range, respectively. Gray bars show the wavelength range visible to human eyes.

Emissions spectrum shows power

carried away from object by photons at each wavelength.

Figure 3.2a shows distribution of photons emitted by 300-K blackbody(room temperature).

Photons emitted by this object mostly wavelengths > 4 μm.

These wavelengths outside range visible to humans (gray shading in figure).

Thus, all room-temperature objects emitting photons, but cannot see photons

because they fall outside visible range -> origin of term “blackbody”.

At room temperature, object appears black

because photons emitted by these objects are invisible to humans

(also the case that blackbodies absorb all such photons that fall on them

- they do not reflect any of them).

Blackbodies are idealized constructs, but most objects nevertheless behave like one, at least approximately, so we assume they are blackbodies..

Figure 3.2a also shows that peak of the emissions spectrum for 300-K blackbody occurs near 10 μm.

Turns out that there is simple relation between temperature and peak of object’s emission spectrum

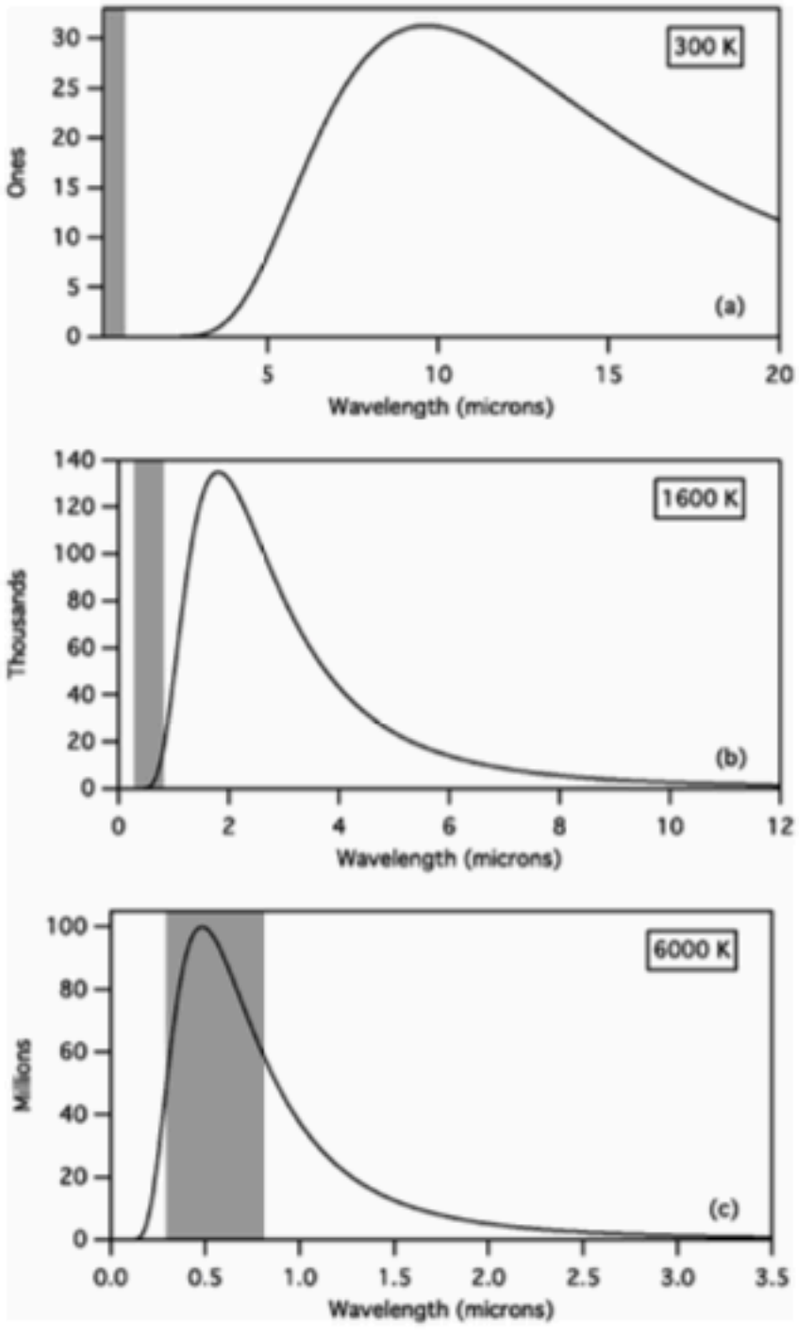


Figure 3.2: Power emitted at different wavelengths from objects (with surface area of 1 m^2) at three temperatures: (a) 300 K, (b) 1,600 K, and (c) 6,000 K The vertical axes are in units of 1 W/μm, 1,000 W/μm, and 1 MW/μm of wavelength range, respectively. Gray bars show the wavelength range visible to human eyes.

-> Wien's displacement law:

$$\lambda_{max} = \frac{2897}{T}$$

T is temperature of blackbody in Kelvin and λ_{max} is wavelength of peak of emission spectrum in microns.

If we put 300 K into Equation get 9.7 μm,

which is in good agreement with Figure 3.2.

Wien's displacement law also tells us that,

as object heats up,

λmax decreases,

shifting peak of emission spectrum

to shorter wavelengths.

Thus, 1,600-K object has λmax = 1.9 μm

and 6,000-K object has λmax = 0.5 μm,

values consistent with the emissions spectra

in Figures 3.2b and 3.2c.

Note, varying y-axis ranges

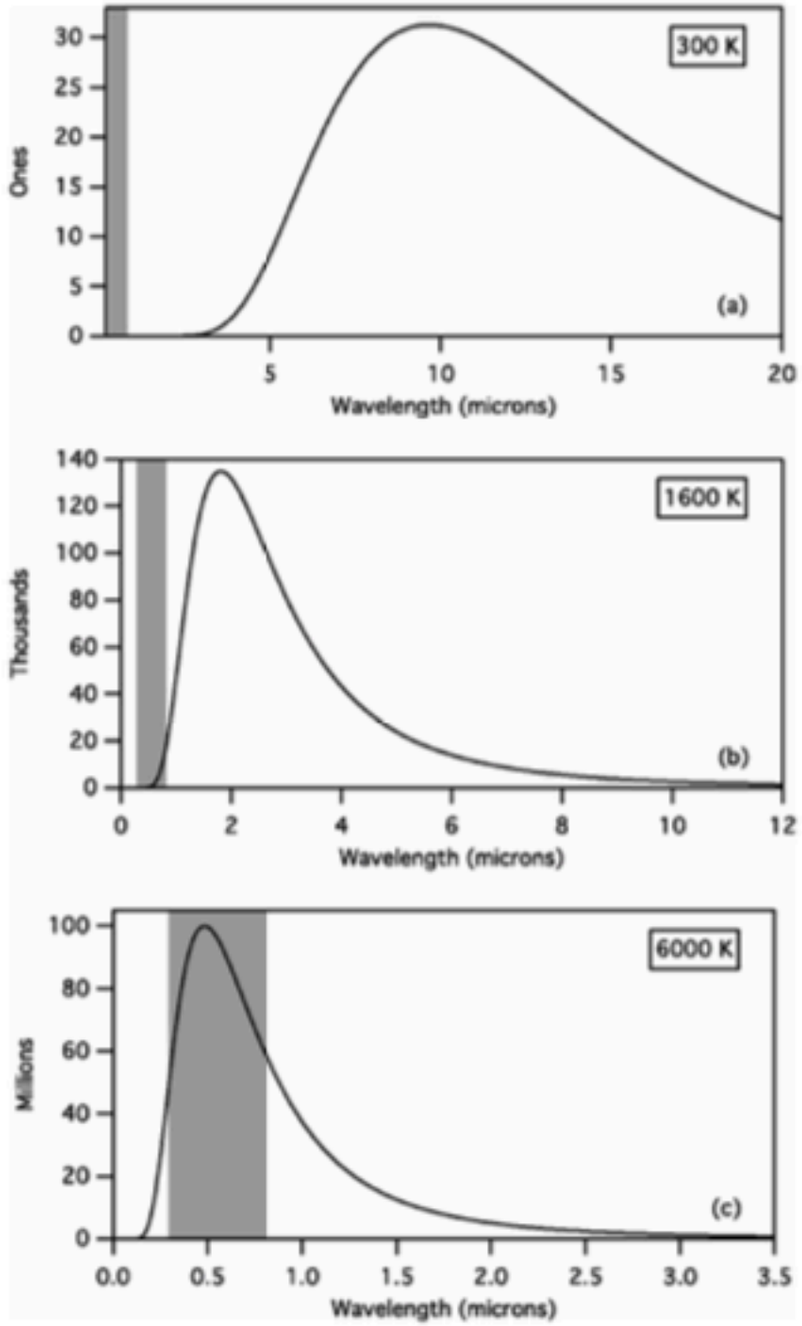


Figure 3.2: Power emitted at different wavelengths from objects (with surface area of 1 m^2) at three temperatures: (a) 300 K, (b) 1,600 K, and (c) 6,000 K The vertical axes are in units of 1 W/μm, 1,000 W/μm, and 1 MW/μm of wavelength range, respectively. Gray bars show the wavelength range visible to human eyes.

Objects do not just emit photons at λ_{max}; emit them over range of wavelengths around λ_{max}.

So, while λ_{max} = 1.9 pm for 1,600-K object, object emits photons over range of wavelengths from 0.7 to 10 μm.

Because fraction of photons emitted by this object have wavelengths between 0.7 and 0.8 μm,

—> at red end of visible spectrum, humans perceive 1,600-K object as having slight reddish glow

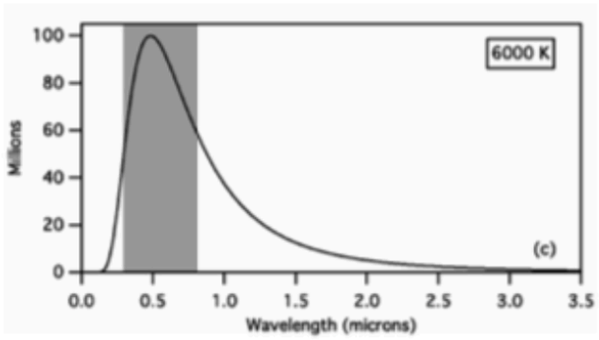
-> object glowing “red hot”.

Blacksmiths use this to determine when piece of metal has reached an appropriate temperature,

and necessity of seeing a faint glow from an object is one reason

that blacksmiths often work in dim, low-light conditions.

For 6,000-K object, most of photons emitted fall within visible range.



Sun is, to good approximation, 6,000-K blackbody,

and distribution of photons from Sun closely approximated by this blackbody spectra.

Because being able to see confers strong advantage in surviving (evolution),

it is no surprise that the eyes of humans and other animals

have evolved to see in this range of wavelengths.

In fact, human eye is maximally sensitive to light with wavelength near 0.5 μm,

which is λ_{max} for 6,000-K blackbody.

The chlorophyll molecule, which is a key component of photosynthesis,

strongly absorbs photons in visible range,

showing that plants also evolved to take advantage of photons emitted by Sun!

Finally, if photons emitted by room-temperature objects are not visible to our eyes,

how can we see room-temperature objects?

What you see when you look at a room-temperature object are visible photons

(emitted by Sun, light bulb or other hot object) that have bounced(scattered) off object.

Everyday object that uses the concepts discussed
is an incandescent light bulb.

Incandescent light bulb consists of glass envelope
containing small filament made of metal,
such as tungsten.

When light bulb turned on,
electricity flows through filament,
heating to 3,000 K (Figure 3.3).)

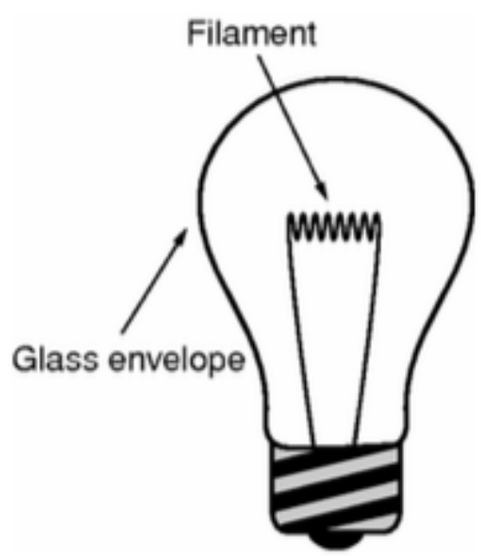


Figure 3.3: A schematlc of a typical lncandescent light bulb.

Figure 3.4 shows wavelength distribution of
photons emitted by 3,000-K blackbody.

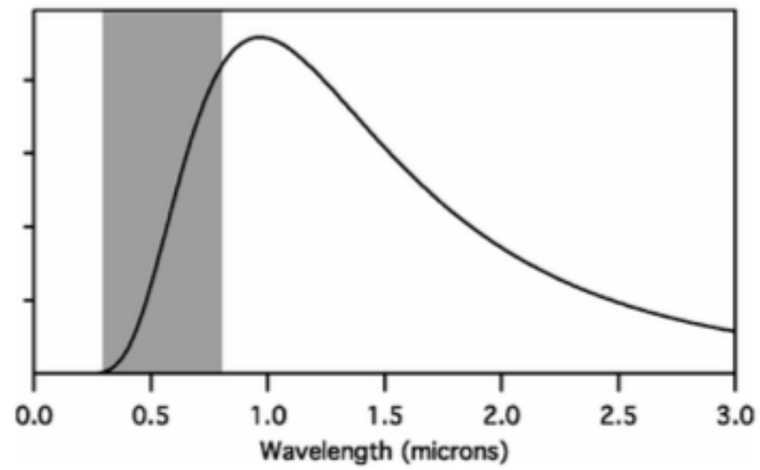


Figure 3.4: Emissions spectrum for a 3,000-K blackbody, a typical filament temperature for an incandescent light bulb. The numbers on the y-axis are omitted.

As figure shows, filament is hot enough -> some photons emitted visible
-> will see light bulb glowing.

However, 85% of photons emitted

-> wavelengths in infrared

- humans cannot detect

-> provide no lighting for humans and so energy to produce them wasted

-> incandescent bulbs **inefficient** as light sources.

Conventional incandescent bulbs are run at about as high a temperature as they can be.

Because of chemical reactions between any halogen gas and filament,

filament in halogen light bulbs can survive

at temperatures several hundred degrees hotter than in a regular incandescent bulb

-> halogen light bulbs put out more visible photons

-> more efficient than regular incandescent bulbs.

Obtain high efficiency lighting <-> change technology.

Compact fluorescent light bulbs and light emitting diode light bulbs

use different technologies to emit most of bulb's photons in visible wavelength range

-> five times more efficient.

12-W CFL -> same amount of visible light

as a 60-W incandescent by reducing

amount of infrared light emitted.

Not only does wavelength of emission

change with temperature,

but total power emitted also increases with temperature.

Mentioned earlier to note the varying y-axis ranges in

Figure 3.2, but is explicitly shown by Figure 3.5a

-> 4 different blackbody-emission curves on single plot.

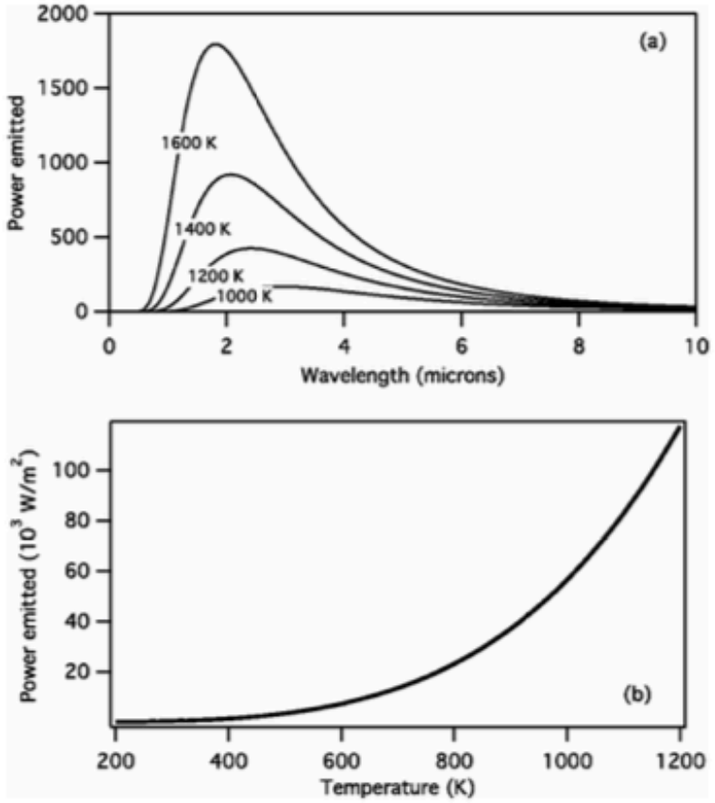


Figure 3.5: Plots of (a) the distribution of power emitted by a blackbody at four different temperatures (1,600, 1400, 1,200. and 1,000 K). in (W/m^2)/μm. and (b) total power emitted by a blackbody as a function of temperature. in W/m^2.

Power emitted is area under the curve

Plot shows that, at every wavelength, warmer objects emit more power than cooler objects.

Figure 3. 5b plots total power emitted by blackbody as function of temperature

-> as temperature of object increases, so does power emitted

-> Stefan-Boltzmann equation between total power radiated by blackbody and temperature:

$$\frac{P}{a} = \sigma T^4$$

P/a = power emitted by blackbody per unit surface area - watts per square meter;

σ is Stefan-Boltzmann constant $\sigma = 5.67 \times 10^{-8} [W/m^2)/K^4]$;

and T = temperature of object in Kelvin.

If multiply P/a by surface area a of object (square meters)

-> total power emitted by blackbody, in watts.

Stefan-Boltzmann equation has wide applications.

By measuring amount of power emitted by an object, astronomers

-> measure temperature of distant stars and planets.

U.S. military uses equation to build sensors to identify and lock onto hot jet engines against cold sky

in guidance systems of heat-seeking missiles.

Ear thermometer uses same physics to convert infrared emission from eardrum
and surrounding tissue into estimate of body temperature.

Figure 3.6 shows image of a dog, in infrared.

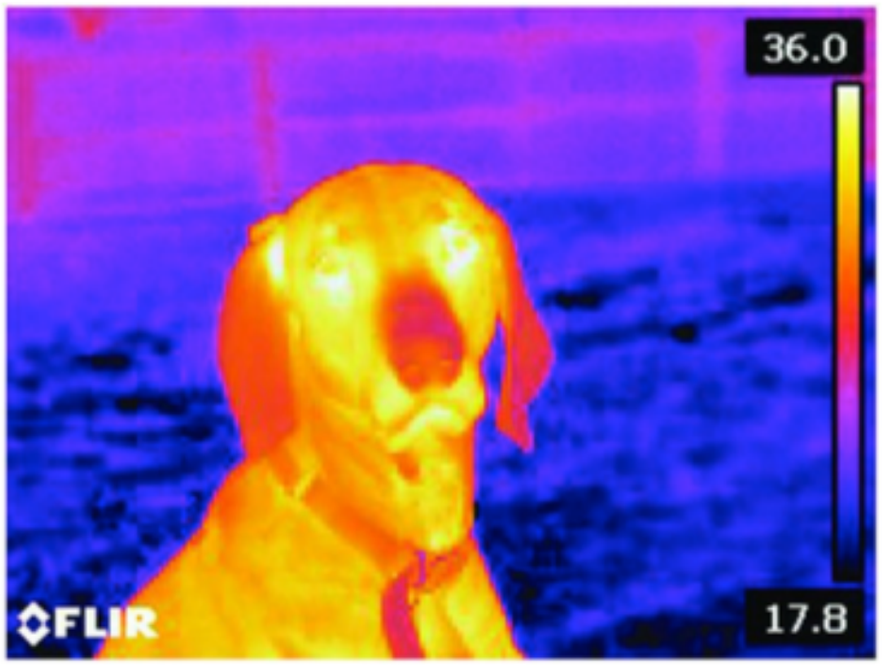


Figure 3.6: Photo of dog in the infrared, with colors assigned to different temperatures.

To construct image,
temperature determined by measuring
infrared emission and
converting to temperature.

Bright colors indicate warm temperatures
and dark colors indicate cool temperatures.

Like humans, dogs are mammals and body temperature is around 38°C.

Fur is insulator -> fur-covered regions of dog tend cooler.

Areas not fur covered - eyes -> close to dog's internal temperature.

Note dog's cold nose.

Now that we understand the radiation side of the equations needed for our model development, we turn to the Energy Balance side of the equations.

One of cornerstones of modern physics is 1st law of thermodynamics

-> energy is conserved.

If some object loses some energy, then some other object must gain that same amount of energy.

Furthermore, because photons are just little packets of energy,

1st law -> when object emits photon,

object's internal energy must decrease - since temperature is measure of internal energy,

emission of photon therefore causes object to cool.

Similarly, if photon hits an object and is absorbed,

then energy of photon transferred to objects internal energy and object warms.

If energy flowing into object (energy in) exceeds energy flowing out (energy out),

then internal energy (and temperature) of object **increases**.

Written mathematically:

Change in temperature $\propto$ Change in internal energy = energy in – energy out

If energy in and energy out are equal, internal energy and temperature do not change

-> equilibrium state.

Example - Consider a home oven.

How does an oven cook?

NOT directly because hot inside.

Physics is subtler than you realize.

Ovens do not cook because air in oven hot

- air is terrible conductor of heat.

Ovens cook by infrared radiation.

Electric oven turned on,

heating element heats up,

reaching temperatures high enough

-> glows dark orange -> element radiating enormous amount of power,

typically several thousand watts.

Photons emitted by heating element absorbed by walls of oven, heating them.

When walls reach predetermined temperature, 350-450°F (450-500 K),

oven “preheated” and put turkey in oven.

Assume turkey came out of refrigerator at temperature of 3°C or 276 K

-> turkey radiating 330 W/m^2.

If turkey has surface area of 0.1 m^2

-> total power radiated by turkey is 33 W.

Turkey also absorbing photons from oven's hot walls.

Oven walls, at 375°F (465 K), radiating 2,650 W/m^2.

Total surface area of oven's six walls approximately 1.3 m^2;

total power radiated by oven's wall roughly 3,500 W.

Most of energy radiated by oven's walls misses turkey in middle and hits other walls.

Turkey absorbs photons emitted by area of walls equal to surface area of turkey, 0.1 m^2.

Given that walls emit 2,650 W/m^2 -> turkey absorbing 265 W of power.

Because turkey emitting 33 W but absorbing 265 W, internal energy of turkey increasing -> warming.

Eventually, turkey reaches temperature -> "done"; cook removes from oven.

That is how conventional oven cooks.

Microwave oven also cooks food by bombarding food with photons.

However, instead of bombarding food with infrared photons,

microwave oven bombards food with microwave photons, which have longer wavelengths.

Microwave ovens cook faster because deliver higher rates of power to food than conventional oven.

By using microwaves, oven able to deliver five to ten times amount of conventional oven.

Net result - food is heated more rapidly in microwave oven than conventional oven.

We now have enough new ideas to get started with our model building!

A Simple Climate Model

The source of energy for our climate system

1st step to understanding climate is an **energy budget calculation**,

-> calculate energy **in** and energy **out** for Earth.

Ultimate source of energy is Sun -> amazing 3.8×10^{26} W [380 trillion trillion W] of power!!

Sun emits photons in all directions, so only small fraction of this energy falls on Earth.

1st step in calculating energy -> determine intensity of sunlight at Earth's orbit.

To estimate, imagine sphere surrounding Sun

radius = Sun-Earth distance, 150 million km (Figure 4.1).

Because sphere completely encloses Sun,

all of sunlight emitted by Sun

must fall on interior of the sphere.

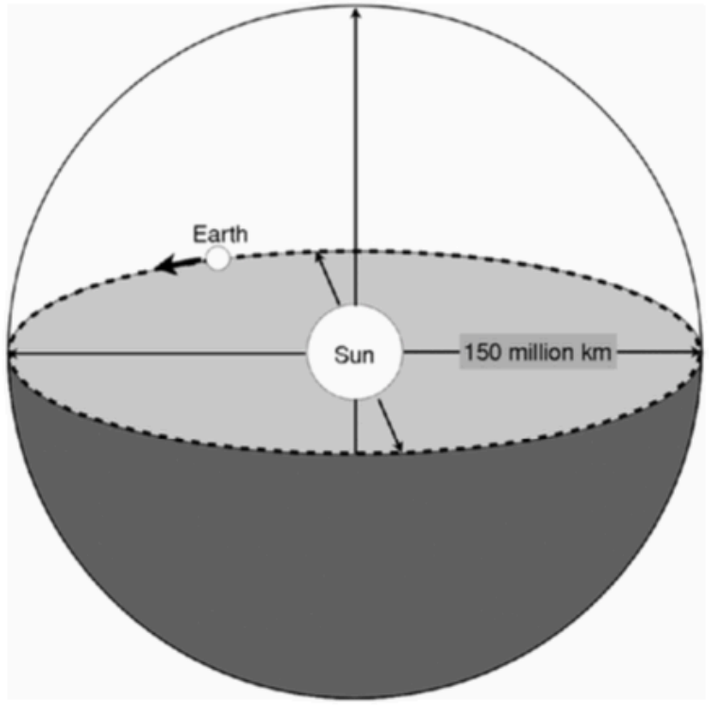


Figure 4.1: Solar constant calculation: A sphere (gray) surrounds the Sun with a radius equal to the Earth's orbit (dashed line); all radiation emitted by the Sun [black arrows) falls on this sphere.

Surface area of sphere is

$$4\pi r^2 = 4\pi(150 \text{ million km})^2 = 2.8 \times 10^{17} km^2 = 2.8 \times 10^{23} m^2$$

Dividing total energy emitted by Sun by area of sphere

produces estimate of intensity of solar radiation at Earth's orbit:

$$\frac{3.8 \times 10^{26} W}{2.8 \times 10^{23} m^2} = 1360 W/m^2$$ = solar constant for Earth = S.

Solar constant is function of how far planet is from Sun.

As planet gets closer to Sun, solar constant increases; gets further away, solar constant decreases.

Venus is located 107 million km from Sun.

Thus, solar constant for Venus is calculated as 3.8×10^{26} W divided by

surface area of sphere with radius of 107 million km, $1.43 \times 10^{23} m^2$ -> 2600 $W/m^2 = S_{Venus}$

Can then use Earth's solar constant

.

to determine total solar energy falling on Earth.

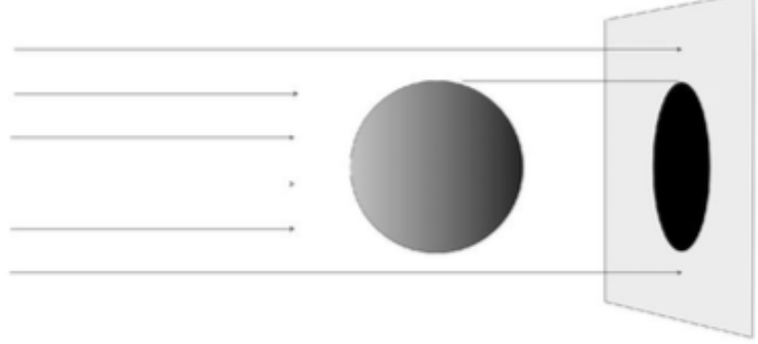


Figure 4.2: The Earth is casting a shadow on a screen placed right behind it because it blocks sunlight. The total amount of solar energy falling on the Earth is the same as what would have fallen into the shadow area.

Easiest way to calculate is to realize that,

if set up screen behind Earth,

Earth would cast circular shadow on screen,

with radius equal to radius of Earth (Figure 4.2).

Amount of sunlight falling on Earth

= amount that would have fallen into shadow area if Earth were not there.

Shadow area is πR^2 where R is radius of Earth

-> total solar energy that would have fallen into that area is πR^2 times solar constant S.

Given that radius of Earth is approximately 6400 km = 6.4×10^6 m, and S = 1360 W/m^2,

solar energy is falling on Earth at rate of 1.8×10^{17}W or 180000 TW

(1 TW, called a terawatt, is 10^{12} or a trillion watts).

An immense amount of power!!

Human society today consumes about 16 TW,

That is why we CAN convert to Solar Power!!!!!!

So, a simple calculation shows why solar energy is Holy Grail of renewable energy.

If we could capture just 0.01 percent of solar energy falling on Earth,

we could satisfy all of world's current energy needs!!!!!!!

Not all of photons from Sun that fall on Earth are absorbed by it.

Some of photons are reflected back to space by clouds, ice, and other reflective elements of Earth system.

Reflectivity of a planet is called albedo, from Latin word for “whiteness”.

Symbol α = fraction of incident photons reflected back to space:

for Earth, α is 0.3 -> 1 – α is fraction of photons that are absorbed by Earth.

Taking all ideas into account, energy in (E_{in}) for Earth (from Sun) is

$$E_{in} = S(1-\alpha)\pi R^2$$

Evaluating equation yields estimate of E_{in} for Earth of 120,000 TW.

More useful to do calculation per square meter of Earth’s surface area,

so divide Equation by surface area of Earth, $4\pi R^2$:

$$\frac{E_{in}}{area} = \frac{S(1-\alpha)\pi R^2}{4\pi R^2} = \frac{S(1-\alpha)}{4}$$

Note that πR^2 terms cancel, so net amount of solar energy absorbed per square meter is not function of Earth’s size.

Plugging values of S= 1360 W/m^2 and $\alpha = 0.3$ into equation,

obtain value of 238 W/m^2 for Earth's E_{in}.

So Earth absorbs average of 238 W/m^2 from Sun;

This does not mean that every square meter of Earth absorbs this amount, however.

In fact, amount of solar energy absorbed varies widely across planet.

1st, nighttime half of Earth receives no energy from Sun at all.

2nd, amount falling on square meter of daytime half

determined by orientation of square meter with respect to incoming beams of sunlight.

Amount of sunlight received is maximum if surface oriented perpendicular to incoming beam (Figure 4.3a).

As surface rotates away from perpendicular, amount of solar energy intercepting surface decreases (Figure 4.3b), eventually reaching zero for surface parallel to incoming beam (Figure 4.3c).

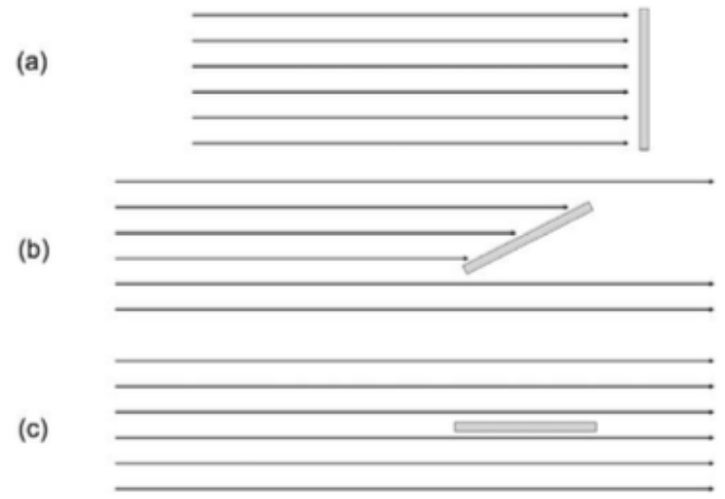


Figure 4.3: Schematic showing how the amount of energy falling on a surface is dependent on the angle between the surface and the incoming beams of light: (a) perpendicular, (b) rotated away from perpendicular, and (c) parallel.

Figure 4.4 shows how this leads to variations in amount of solar energy falling on Earth's surface with latitude.

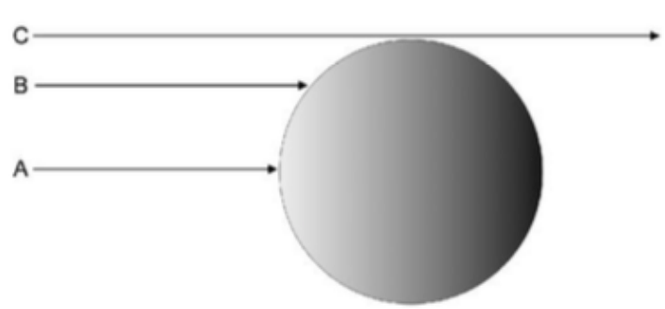


Figure 4.4: Schematic showing how the amount of solar energy falling on a square meter of the Earth's surface is determined by the latitude.

In tropics (Arrow A), surface of Earth is perpendicular to incoming solar light beams, corresponding to situation in Figure 4.3a.

Surface in mid-latitudes (Arrow B) is at moderate angle to incoming solar light beams, corresponding to situation in Figure 4.3b.

This means that mid-latitudes receive less solar radiation per square meter than tropics.

Finally, polar regions (Arrow C) correspond to situation in Figure 4.3c, so this region receives even less solar energy.

In addition to variations in incoming sunlight with latitude, albedo of planet also varies widely.

Tropics are mainly open ocean, which is dark and therefore has low albedo.

Combined with large amount of solar energy per square meter,

tropics therefore experience far more solar heating than anywhere else on planet.

High latitudes, however, are generally covered by snow and ice, giving them high albedo.

Combined with small amount of solar energy received per square meter,
this means that polar regions absorb least amount of solar energy.

This provides us with a **simple but fundamentally correct** explanation of
why tropics tend to be warmest place on planet and polar regions coldest.

Energy loss to space

In early 19th century, Joseph Fourier, one of history's great mathematicians, asked a deceptively simple question: Because energy is always falling on Earth from Sun, why does Earth not heat up until it is same temperature as Sun?

The **answer** he determined is that Earth is losing energy at rate equal to rate at which it is receiving energy from Sun.

On the basis of what we learned earlier, we may rightly guess that Earth loses energy back to space by means of **blackbody** radiation (Figure 4.5).

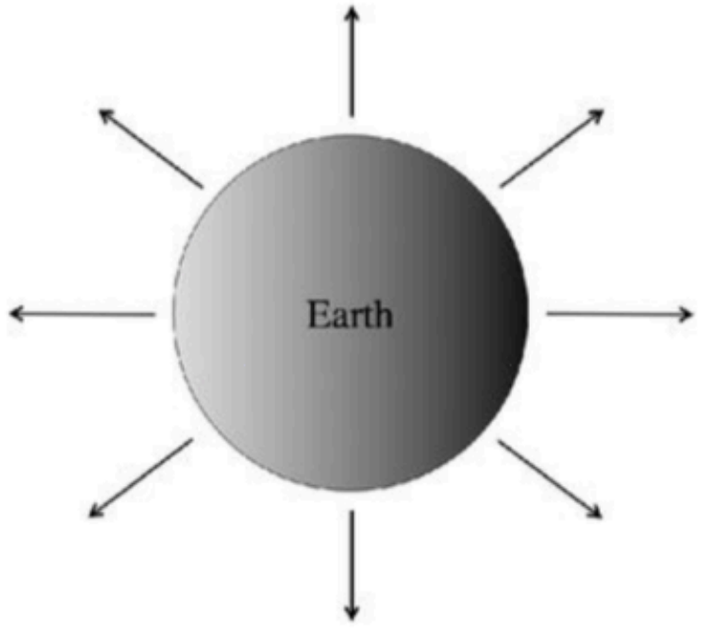


Figure 4.5: Diagram of the Earth, showing the emission of infrared blackbody radiation in all directions.

For a blackbody, P /a is σT^4,

where P /a is power emitted per square meter,

T is temperature of planet,

and σ = Stefan-Boltzmann constant, $5.67\times 10^{-8}(W/m^2)/K^4$.

Setting E_{in} equal to P/a, rate of energy out, we get our 1st equation:

$$\frac{S(1-\alpha)}{4} = \sigma T^4$$

Solving for T, we get the temperature equation.

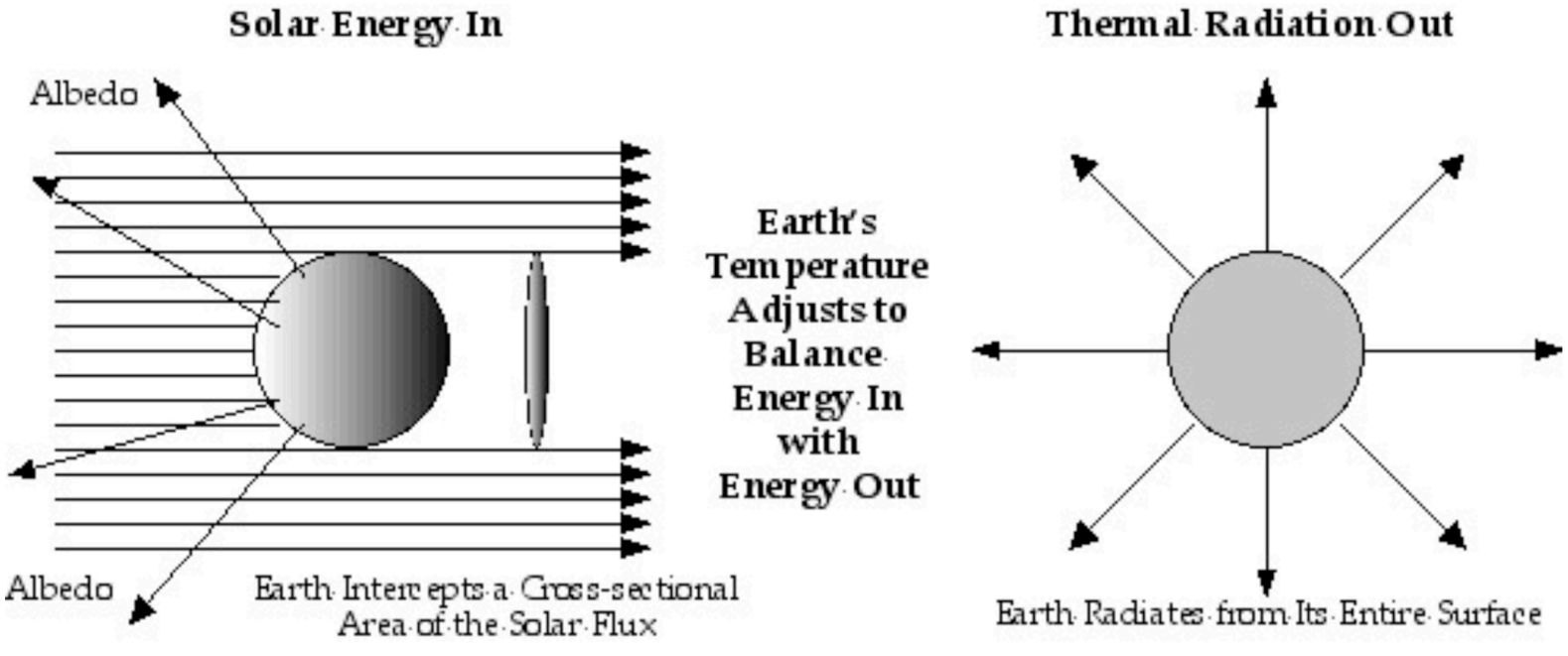
$$T = \left(\frac{S(1-\alpha)}{4\sigma}\right)^{1/4}$$

Plugging S = 1,360 W/m^2 and $\alpha = 0.3$ yields temperature T = 255 K (-18°C).

Actual average temperature of Earth closer to 288 K (15°C), so our **first** estimate of Earth's temperature **too cold** by 33°C.

Summarizing

Earth's Effective Temperature



Power In $= S(1-A)\pi R_E^2$

σT_E^4 , Stefan–Boltzmann law

$\sigma = 5.67 \times 10^{-8}$ J/sm^{2}K^4

S = 1366 J/sm^2 dS = 0.07%

A = 0.30 (28 - 30)

Power Out $= 4\pi R_E^2 \sigma T_E^4$

$$T_E = [S(1-A)/4\sigma]^{1/4}$$

$T_E = 255^{\circ}K = -18^{\circ}C$ That's too Cold!!

Where did our calculation go wrong?

Turns out that what we have **neglected** is heating of planet by a heated Earth's atmosphere

-—> **greenhouse effect.**

The greenhouse effect

One-layer model

To understand impact of atmosphere on planet's temperature,

we make following assumptions (reasonably accurate).

1. The Earth's atmosphere is transparent to visible photons emitted by Sun (wavelengths from 0.3-0.8 μm), so these photons pass through atmosphere and are absorbed by surface.

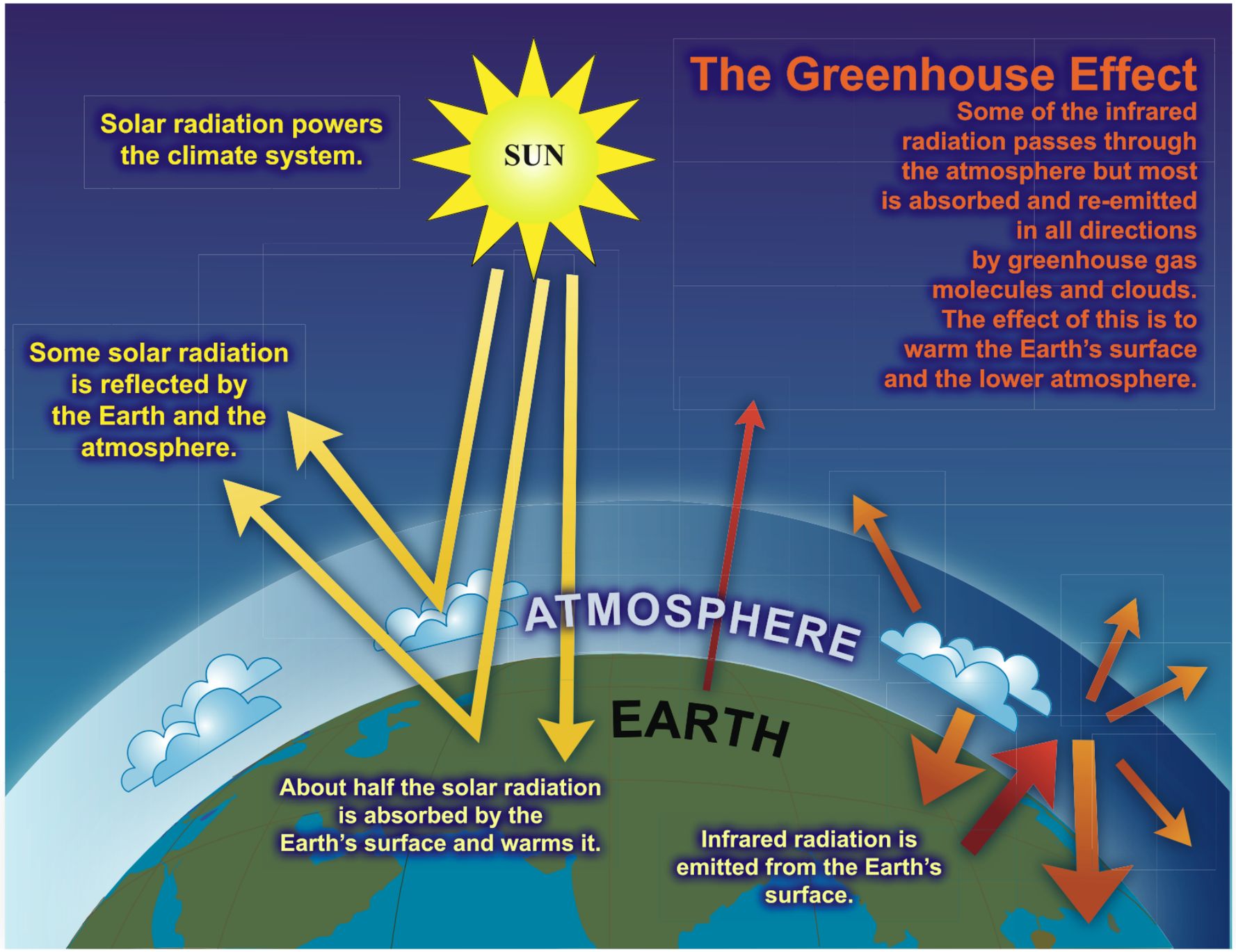
2. Atmosphere is opaque to infrared photons emitted by surface (wavelengths > 4 μm), -> these photons are absorbed by the atmosphere.

3. Atmosphere also behaves like blackbody, so emits photons based on its temperature. It emits photons **equally** both upward and downward.

4. Photons emitted by atmosphere in upward direction **escape** to space and carry energy away front Earth. Photons emitted downward are **absorbed** by surface.

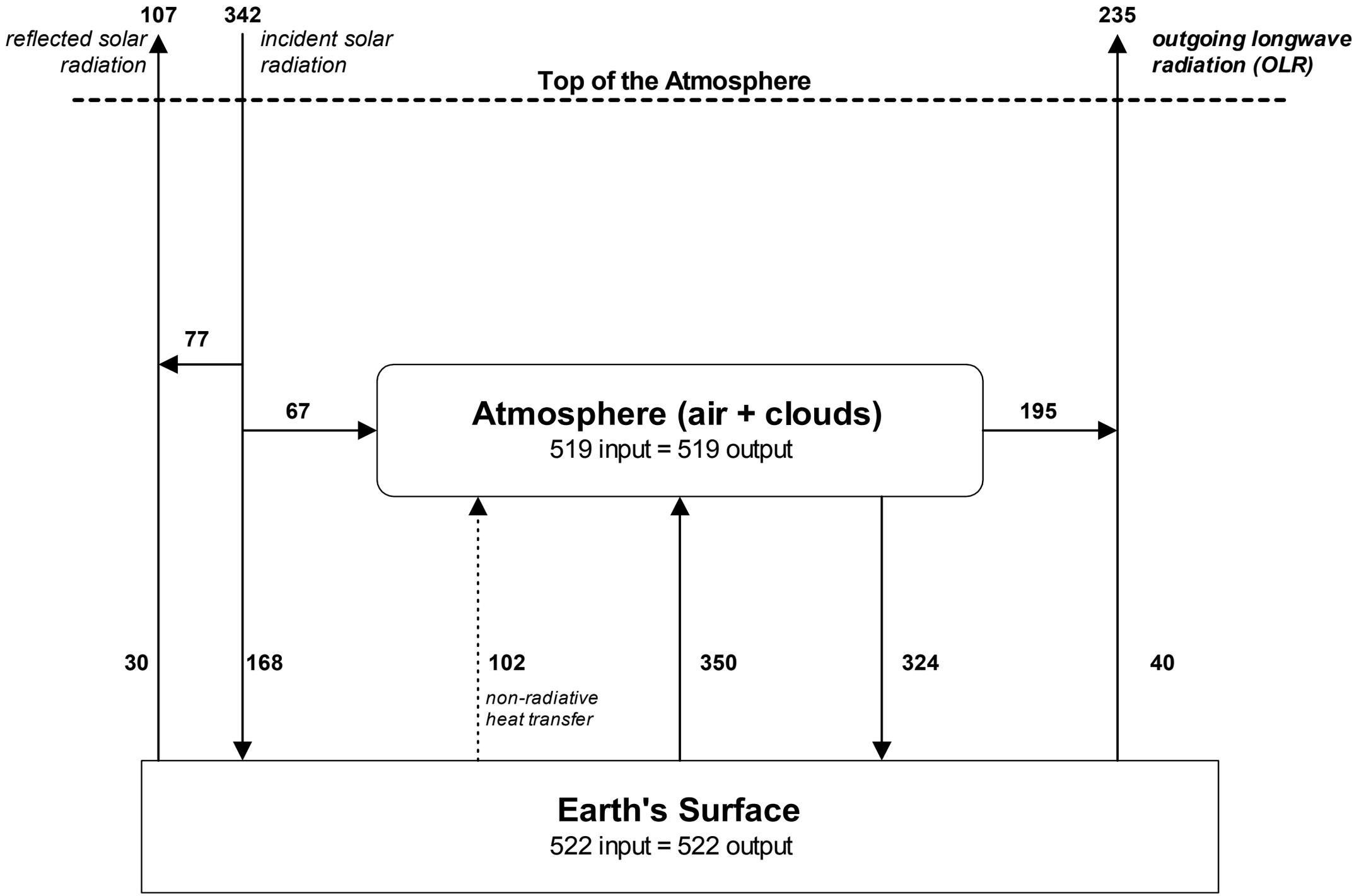
The Greenhouse Effect - overall picture

– Without greenhouse effect, global average surface temperature of Earth would be about -19 ºC instead of 15 ºC.



Now to build a detailed model

Summary - Earth's Energy Balance: Detailed Balance



Last two slides give very broad picture with lots of unexplained numbers.

Now for the details……

One-layer model is diagrammed in Figure 4.6.

For conceptual simplicity, diagram

shows effects of atmosphere concentrated

in single thin layer, which is why this model

is frequently called a "one-layer" model.

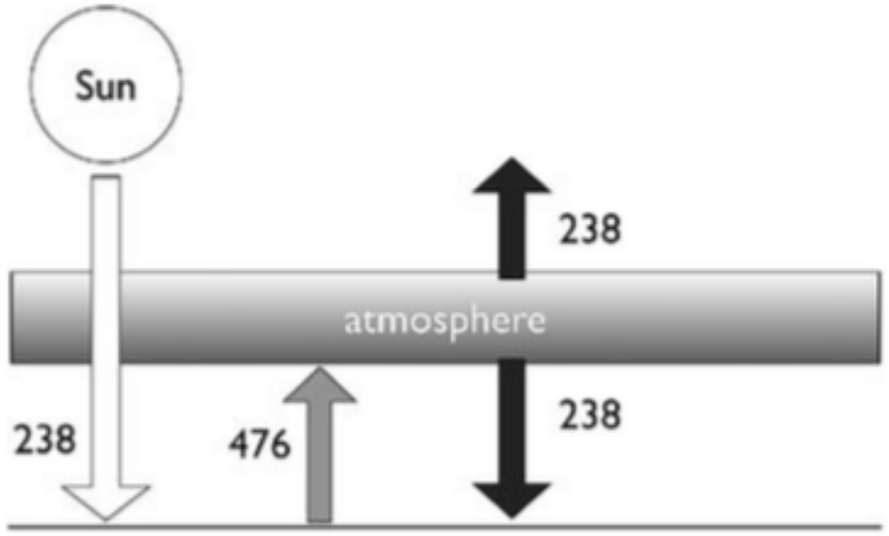


Figure 4.6: Schematic of energy flow on a planet with a one-layer atmosphere. The atmosphere is represented by a single layer that is transparent to visible photons but absorbs all infrared photons that fall on it. The arrows show global average energy flows with values in W/m^2.

To calculate surface temperature in model, **assume** that planet as a whole, as well as surface and atmosphere individually, must **all** be in energy balance (where E_{in} equals E_{out}).

1st, consider energy balance for planet as a **whole**.

Energy into planet coming entirely from Sun.

Energy out to space coming entirely from atmosphere

(remember: any photons emitted by surface are absorbed by atmosphere).

Using Earth's values of solar constant/albedo, energy in from Sun is 238 W/m^2

-> atmosphere must be radiating 238 W/m^2 upward to space **so** planet as a whole is in energy balance.

Because atmosphere radiates **equally** upward and downward,

atmosphere also radiating 238 W/m^2 back toward Earth's surface.

Consider energy balance for surface.

Energy in for surface is 238 W/m^2 from Sun and 238 W/m^2 from atmosphere

- total of 476 W/m^2.

Means surface emitting 476 W/m^2 upward to achieve energy balance.

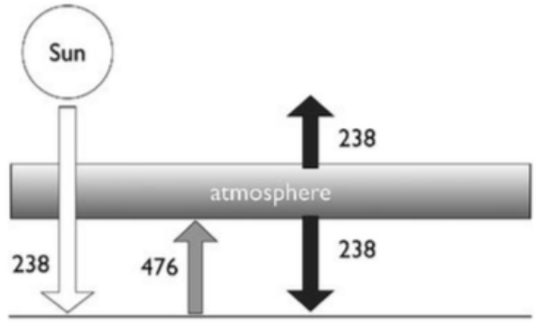


Figure 4.6: Schematic of energy flow on a planet with a one-layer atmosphere. The atmosphere is represented by a single layer that is transparent to visible photons but absorbs all infrared photons that fall on it. The arrows show global average energy flows with values in W/m^2.

What is temperature of surface?

If know that surface emitting 476 W/m^2,

can determine temperature by using Stefan-Boltzmann equation $E_{out} = P/a = \sigma T^4$.

Solving $\sigma T^4 = 476 W/m^2$ for T

yields a surface temperature of 303 K (30°C),

which is 48°C **warmer** than for planet **without** atmosphere (earlier result).

An incredibly important result!

The addition of atmosphere opaque to infrared radiation significantly warmed planet's surface.

Conceptually, this occurs because surface of planet with an atmosphere heated

not just by Sun but also by atmosphere.

Of course, if walk outside, cannot see atmosphere heating Earth's surface

because photons it emits are not visible, but they still carry energy.

When scientists talk about greenhouse effect

-> heating of surface by atmosphere.

Greenhouse effect

-> atmosphere warms surface by making it harder for surface to lose energy to space.

Without atmosphere,

all photons emitted by surface escape to space;

surface has to emit only 238 W/m^2 for planet to be in energy balance.

With one-layer atmosphere, though,

only half of photons emitted by surface end up escaping to space

- other half returned to surface

-> surface must emit twice as much, 476 W/m^2,

in order for 238 W/m^2 to escape to space.

Higher rate of emission requires **warmer** surface.

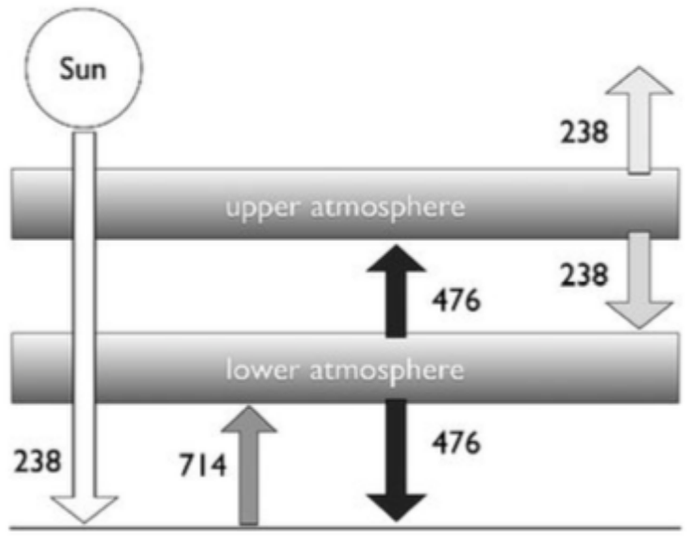


Figure 4.7: Schematic of energy flow on a planet with a two-ÃÎlayer atmosphere, with values in W/m^2.

Two-layer model

Consider planet with “two-layer” atmosphere (Figure 4.7).

Once again, atmosphere is transparent to visible radiation but opaque to infrared

-> that photons from Sun pass through atmosphere and absorbed by surface.

Photons emitted by surface absorbed in lower atmosphere.

Photons emitted by lower atmosphere in upward direction absorbed by upper atmosphere;

photons emitted in downward direction absorbed by surface.

Photons emitted by upper atmosphere in upward direction escape to space;

photons emitted in downward direction absorbed by lower atmosphere.

Key to determining surface temperature is

to enforce energy balance for planet as a whole,

surface, and both atmospheric layers.

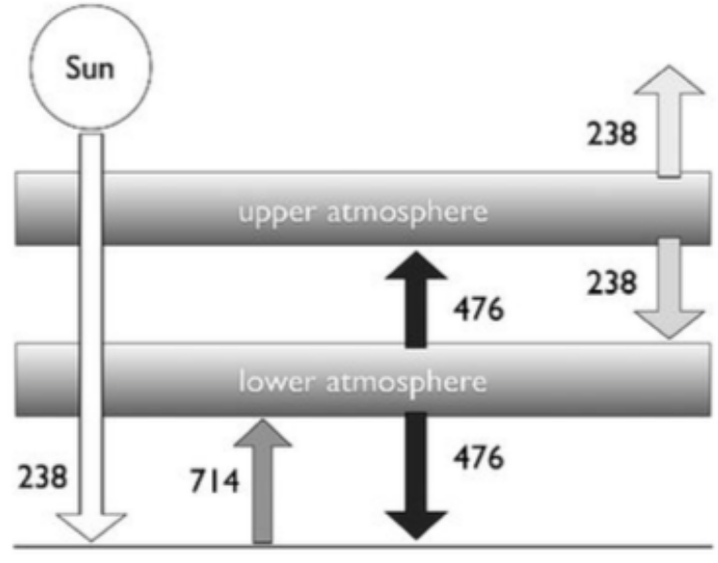


Figure 4.7: Schematic of energy flow on a planet with a two-ĂĬlayer atmosphere, with values in W/m^2.

Easiest way to do this is to start with planetary energy balance

and then work downward from topmost layer to surface.

Planetary energy balance requires energy out for planet to balance energy in from Sun.

Because energy out comes entirely front upper layer,

it must be emitting 238 W/m^2 to space in order to balance 238 W/m^2

that Sun is providing planet (assuming terrestrial values for S and E_{out}).

That, in turn, means that upper layer is also emitting 238 W/m^2 downward.

Totaling emissions in both directions, upper layer is emitting 476 W/m^2.

Because energy out must equal energy in for layer,

this layer must be receiving 476 W/m^2 from lower atmospheric layer.

Thus, know that lower layer is emitting 476 W/m^2 upward - and therefore downward, too.

For lower layer to achieve energy balance,

lower layer must also be receiving 476+476 = 952 W/m^2.

We already calculated that 238 W/m^2 are coming from upper layer,

so that means that 714 W/m^2 must be coming from the surface to lower layer.

Can verify our result by examining energy balance for surface.

Surface receives 476W/m^2 from lower atmosphere and 238 W/m^2 from Sun,

for total E_{in} of 714 W/m^2.

This corroborates what we calculated must be Eout for surface

based on energy-balance for lower atmosphere.

Finally, for surface to be emitting 714 W/m^2,

its temperature must be 335 K (62°C).

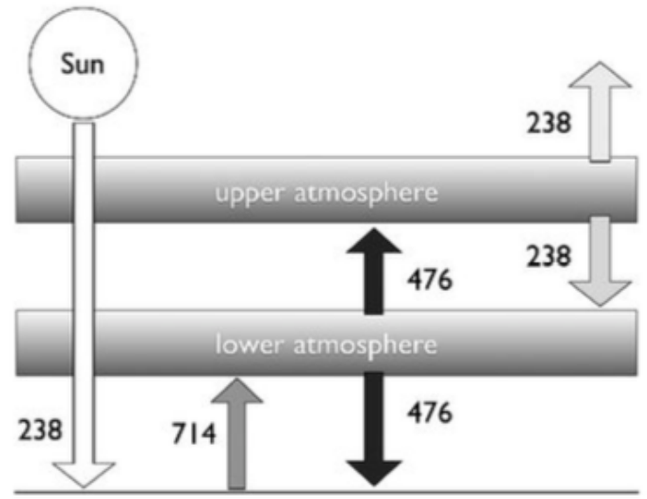


Figure 4.7: Schematic of energy flow on a planet with a two-ÃĬlayer atmosphere, with values in W/m^2.

This is 32°C warmer than surface of planet with one-layer atmosphere

and 80°C warmer than a planet with no atmosphere.

Thus, adding second layer to atmosphere **further increases** planet's surface temperature.

n-layer model

Now let us derive surface temperature for planet with n layer atmosphere(Figure 4.8).

For some variety,

assume that planet has solar constant

$S = 2{,}000\ W/m^2$ and an albedo $\alpha = 0.7$.

Thus, energy in for this planet is

$$\frac{S(1-\alpha)}{4} = 150\,W/m^2$$

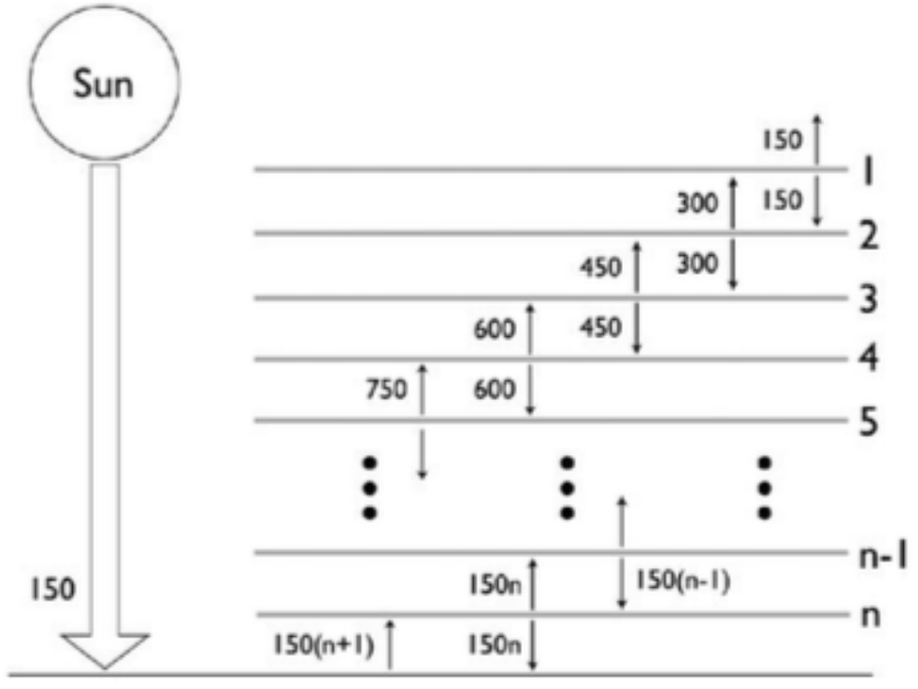


Figure 4.8: Schematic of energy flow on a planet with an n-layer atmosphere; layers are numbered from 1 to n (topmost to bottommost layers), with values in W/m^2.

Means upward emissions from topmost layer of atmosphere (Layer 1) must = 150 W/m^2.

And because upward and downward emissions must be same,

this layer is also emitting 150 W/m^2 downward,

so that total energy out for this layer is 300 W/m^2.

This in turn means that energy in for Layer 1 must also be 300 W/m^2.

Energy in for Layer 1 comes entirely from energy emitted by Layer 2.

Layer 2 must therefore be emitting 300 W/m^2 upward.

This means that it is also emitting 300 W/m^2 downward,

for a total energy out of 600 W/m^2.

Energy in for Layer 2 comes from downward emissions of Layer 1

and upward emissions of Layer 3

and must total 600 W/m^2 in order to balance energy out.

Downward emissions from Layer 1 are 150 W/m^2,

which means that upward emissions from Layer 3 must be 450 W/

Layer 3 must be emitting 450 W/m^2 both upward and downward,

for a total energy out of 900 W/m^2.

Energy in from downward emissions from Layer 2 is 300 W/m^2,

meaning that upward emissions from Layer 4 must be 600 W/m^2.

Figure 4.8: Schematic of energy flow on a planet with an n-layer atmosphere; layers are numbered from 1 to n (topmost to bottommost layers), with values in W/m^2.

By this time, a pattern has emerged

and we can extrapolate to the bottom-most layer, layer n.

Layer n is emitting 150n in both upward and downward directions.

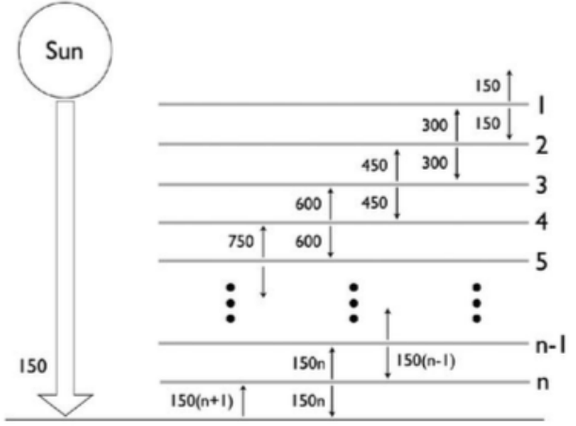


Figure 4.8: Schematic of energy flow on a planet with an n-layer atmosphere; layers are numbered from 1 to n (topmost to bottommost layers), with values in W/m^2.

This in turn means that the surface is receiving 150n emitted from the bottommost layer

and 150 W/m^2 from the Sun.

For energy to balance, the surface must be emitting 150(n + 1) W/m^2 upward.

Setting $150(n + 1) = \sigma T^4$,

we can solve for the surface temperature T of this planet:

$$T = \left(\frac{150(n+1)}{\sigma} \right)^{1/4}$$

Figure 4.9 shows surface temperature

as a function of n,

calculated by using above equation.

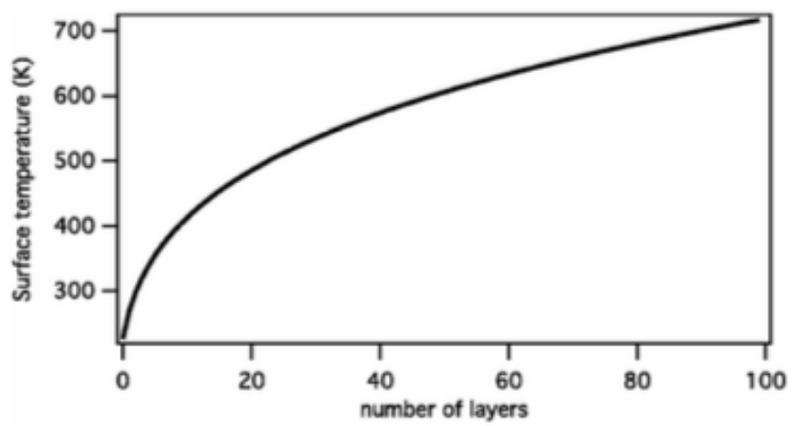


Figure 4.9: Surface temperature (K) for the n-layer planet as a function of the number of layers. Here, $S(1-\alpha)/4 = 150\,W/m^2$.

As you increase number of layers, surface gets hotter and hotter.

However, warming is not linear - each additional layer produces less warming than previous layer.

Can also write general solution for surface temperature of an n-layer planet:

$$T = \left(\frac{(n+1)S(1-\alpha)}{4\sigma} \right)^{1/4}$$

Equation says that surface temperature of planet basically

determined by three parameters:

number of layers in atmosphere (n),

solar constant (S),

and the albedo (α).

1. We can now answer question,

"Why is Earth's temperature what it is?

Temperature of our climate system is set by requirement that energy in and energy out balance.

And if n, S. or α change, then temperature will adjust as required to reestablish this energy balance.

2. As you add more greenhouse gases to planet (i.e., increase n), temperature of planet will increase.

Testing our theory with other planets

Important to emphasize that n-layer model makes many simplifying assumptions.

For example, not all energy transport on planet is by radiation

- some is transported by atmospheric motions, such as thunderstorms.

Another simplification is that model assumes an infinitely fast horizontal energy transport,

allowing us to use a single temperature for planet.

In reality, though, transport of energy is slow enough

that large temperature differences can develop between regions

(eg, tropics and polar regions or between night and day).

This means we **should not expect** model to produce quantitatively accurate surface temperatures.

Nonetheless, model captures **essential** physics of our climate system.

And model is successful in making qualitative predictions

of relative surface temperatures of Earth and its nearby neighbors, Mercury, Venus, and Mars.

Table 4.1 lists important characteristics of planets, and reveals some puzzles.

Mercury is planet closest to Sun,

yet Venus, twice as far from Sun as Mercury,

has surface temperature

approximately 300 K warmer.

Planet	Solar constant (W/m^2)	Albedo	Observed surface temperature (K)	Inferred n
Mercury	10,000	0.1	452	0.052
Venus	2,650	0.7	735	82
Earth	1,360	0.3	289	0.65
Mars	580	0.15	227	0.22

Table 4.1: Data on the four inner planets in our solar system

Result becomes even more puzzling when realize that,

because of its high albedo, energy in for Venus.

$S(1-\alpha)/4 = 200 W/m^2$,

is more than a factor of ten smaller than that for Mercury (2,250 W/m^2).

It is even less than energy in for Earth (258 W/m^2) - yet Venus is 450 K hotter than Earth.

Given surface temperature, albedo, and solar constant,

can solve equation for n, which is number of layers required to satisfy energy balance.

This “inferred n” is also listed in Table 4.1.

Inferred n for Mercury is near zero, suggesting it has almost no greenhouse effect.

This is correct, because Mercury has essentially no atmosphere.

For Mars, inferred n = 0.22.

This again makes some sense

- Mars has a thin atmosphere mostly containing carbon dioxide,

so it does have some greenhouse effect.

However, Martian atmosphere has fewer greenhouse gases than Earth’s atmosphere,

so greenhouse effect on Mars is expected to be weaker than that on Earth.

Our calculations confirm that.

Finally, our calculations suggest that Venus,

with inferred n = 82,

has a massive greenhouse effect.

This is again correct.

The surface pressure on Venus is ninety times that of Earth

(1300 psi, or pounds per square inch, compared with 14.5 psi here on Earth),

and atmosphere is mainly composed of carbon dioxide.

Result of this massive, greenhouse-gas-rich atmosphere

is a planet hotter than inside of your oven on broil

- hot enough even to melt lead.

Thus, see that equation successfully explains relative climates of innermost planets of our solar system.

Simple climate model #2

Overview

This model uses a simple climate model that shows how climate models can be tested and used to make future temperature change projections.

The climate model can be used to teach a variety of topics, including the utility of models, understanding of uncertainty, and understanding of how past temperature changes can help estimate future warming.

One of our focusses will be to experience the process of testing climate models, making projections of 21st century warming, and comparing projections to other model projections.

Each model can generate a 21st century temperature projection based on the model which is first tuned to 20th century warming.

The end product is a graph of projected 21st century warming based on the model.

This will show you what can go on in making climate projections and show how climate modelers can get uncertainty in projected temperatures of the future, even when they use the same historical observations to constrain the model parameters.

Questions to keep in mind during the discussion

- What is a model? Why do we even need a global climate model?
- What is one way to test a model's performance? (How can 20th century temperature records be used to test a model's behavior?)
- Why do the wiggles exist in the temperature records and what determines how wiggly they are?
- Where do uncertainties in climate projections come from?
- Why has the temperature been rising in the last hundred years?
- What is climate sensitivity and what aspect of the Earth's mean temperature does it control?

Background Information

Understanding the energy budget at the Earth's surface can take us a long way to understand recent changes in global mean temperatures.

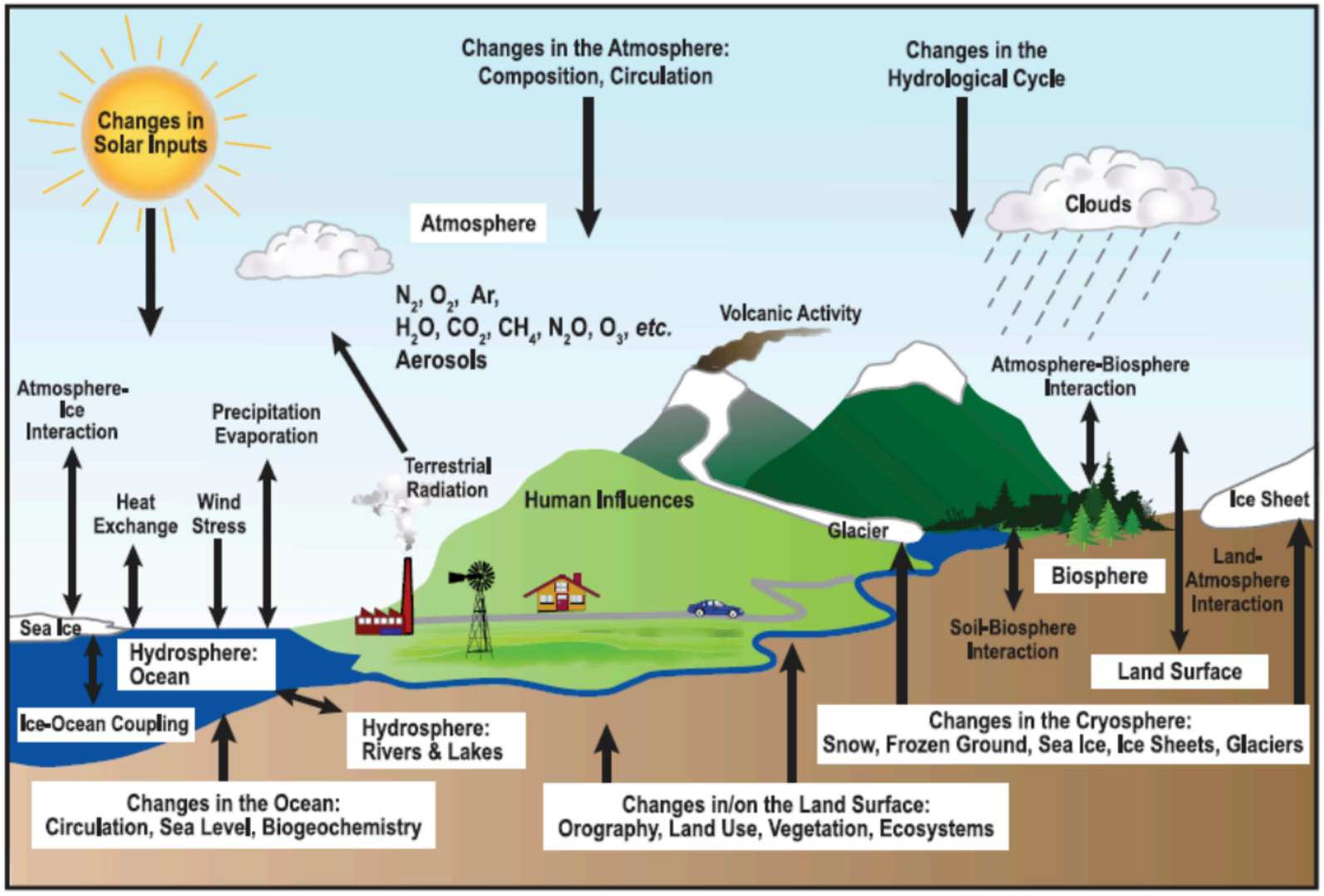
This climate model is essentially an energy balance model that tries to account for the major components of the climate system (feedbacks, natural variability, and an ocean) to help estimate changes in global mean temperature.

Before working with the simple climate model, some background information about models is necessary.

So why do we even need climate models?

Unlike many physics experiments that we can perform inside a laboratory, doing experiments on the Earth's climate is not as easy.

Therefore, scientists must use computer model representations of the climate that incorporate many of the processes and attributes that are essential to simulate the Earth's climate, some of which are represented in the picture.



However, we must test whether our models are accurately representing the Earth's climate.

Therefore, many of the climate models compare their climate models with past and present observations, such as global temperatures or global precipitation fields.

Now if we want our climate models to capture changes in the climate, such as what kind of temperature changes we would expect in the 21st or the 22nd century, we would need to see whether our models can even capture the warming over the 20th century.

Generally speaking, many of the current state-of-the-art models are tested in this way to see what areas of their model need to be improved.

We can easily use these climate models to make projections of 21st century warming, given a few possible emission scenarios.

One initially tunes the models so that they are better at estimating the warming that occurred over the 20th century.

Then one uses the same model parameters to make temperature projections for the future.

I now present a description of the climate model.

The climate model is an energy balance model.

When a system (like the Earth) is in energy balance, where the energy entering the system equals the energy leaving the system, the temperature of the system will stay the same.

If the system undergoes a positive forcing, like the increased intensity of the sun or increased greenhouse gases, the system will warm up until the excess energy entering the system is balanced by the energy leaving the system by longwave cooling.

Among many things, the amount that the surface warms will depend on the storage capacity of the surface, which we will assume is an ocean with some depth.

We can write this balance in terms of an equation

$$\text{Heat storage rate} = \text{longwave cooling} + \text{radiative forcing} \tag{1}$$

If we want to include effects from year to year differences in the weather, then we add weather noise such that

$$\text{Heat storage rate} = \text{longwave cooling} + \text{radiative forcing} + \text{weather noise} \tag{2}$$

When the system is in equilibrium, the heat storage rate will be on average zero, which means that longwave cooling + radiative forcing + weather noise will be zero.

Since weather noise is zero on average, the system will be in equilibrium if longwave cooling + radiative forcing is zero.

Now, Eq. 2 can be written in the following form

$$\rho c_p H \frac{dT(t)}{dt} = -bT(t) + R(t) + W(t) \tag{3}$$

where

$$\rho c_p H \frac{dT(t)}{dt} = -bT(t) + R(t) + W(t)$$

$\boldsymbol{\rho}$ is the density of sea water (1025 kg m^{-3}),

c_p is the specific heat of sea water (3985 Joules kg^{-1} $^oC^{-1}$)

H is the depth of the upper ocean where heat is absorbed (m)

T is the temperature change from preindustrial conditions (^{o}C)

t is time (seconds)

b is the climate sensitivity (Watts m^{-2} $^oC^{-1}$)

R is the radiative forcing from natural and anthropogenic effects (Watts m^{-2})

W is the year to year random weather forcing (Watts m^{-2})

Note that $bT(t)$ in Eq. 3 does not just represent longwave cooling.

Although we have talked about one feedback (the Planck's feedback, which is always negative) so far, we haven't included any of the other feedbacks in our previous discussion.

What we can do, however, is to combine the Planck's feedback with other feedbacks (water vapor, lapse rate, cloud, etc.) under the *b*, which is the climate sensitivity.

We can now track changes in temperature by taking the finite-difference form of Eq. 3 to look at the changes in temperature from year

$$\rho c_p H \frac{T^{n+1} - T^n}{\Delta t} = -bT^n + R^{n+1} + W^{n+1} \tag{4}$$

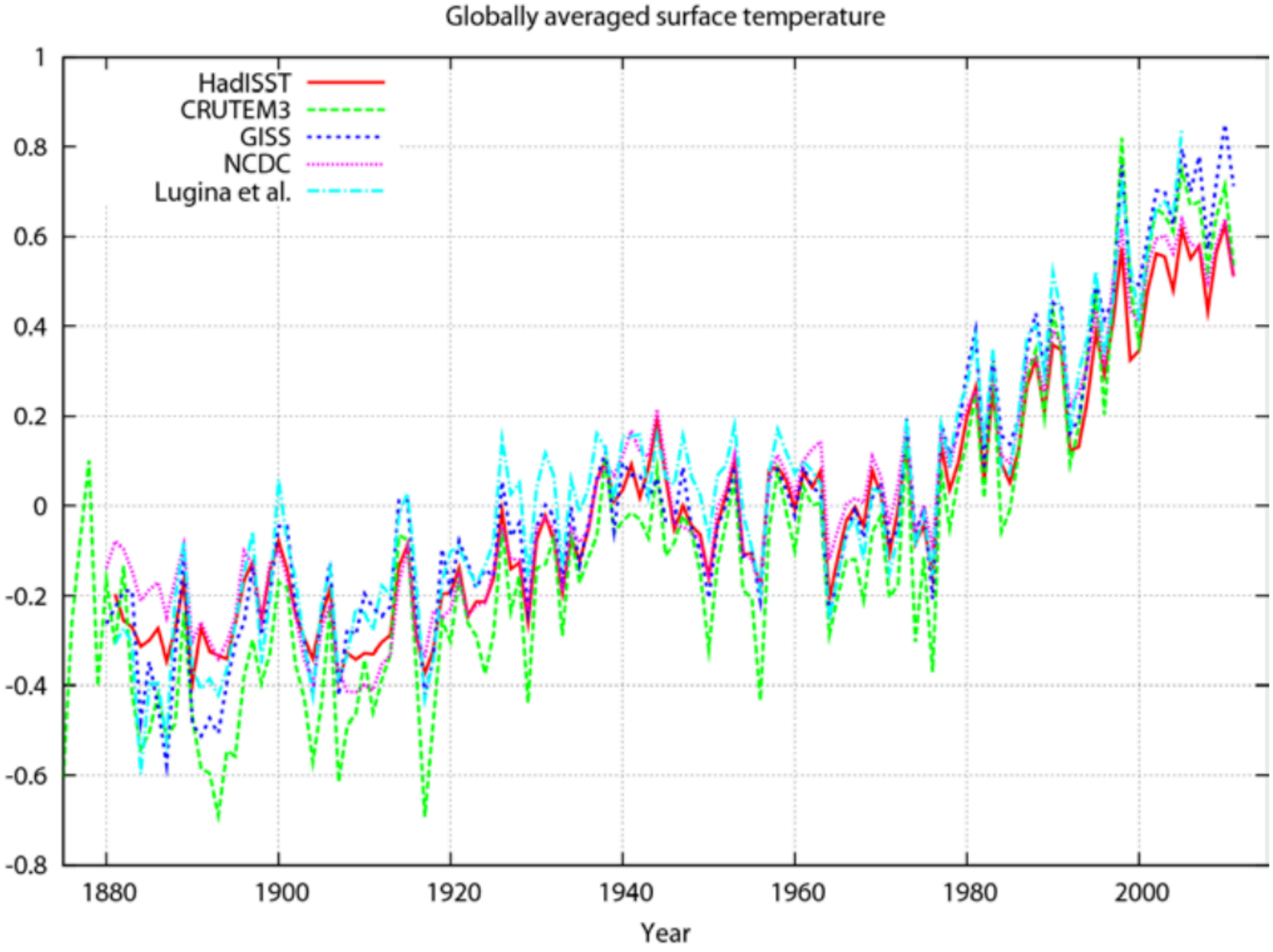
Where the n superscript is the value at time n where n goes from 1880 through 2006 and we wrote the hyearly change approximation to the derivative as

$$\frac{dT(t)}{dt} = \frac{T^{n+1} - T^n}{\Delta t} \tag{5}$$

We can rearrange Eq. 4 to get

$$T^{n+1} = T^n + \frac{-bT^n + R^{n+1} + W^{n+1}}{\rho c_p H} \Delta t \tag{6}$$

This is the equation governing the model.



Now if we look at the 20th century global mean temperature record in this context the temperature has been increasing, which means that the storage rate is increasing.

This also means that the radiative forcing has been larger than the longwave cooling (including effects from other feedbacks).

The rate of the temperature increase then is set by both the ocean heat storage capacity and the climate sensitivity.

The wiggles in the temperature can be either understood as either the effect of radiative forcing or weather noise.

We want to look at the 20th century temperature series and understand what controls the general trend, the rate of the trend, and the wiggles in the temperature.

After that we make temperature projections based on different scenarios, called Representative Concentration Pathways (RCPs).

The RCPs are meant to serve as input for climate models that capture different scenarios of changing emissions, concentrations, and land-cover change projections.

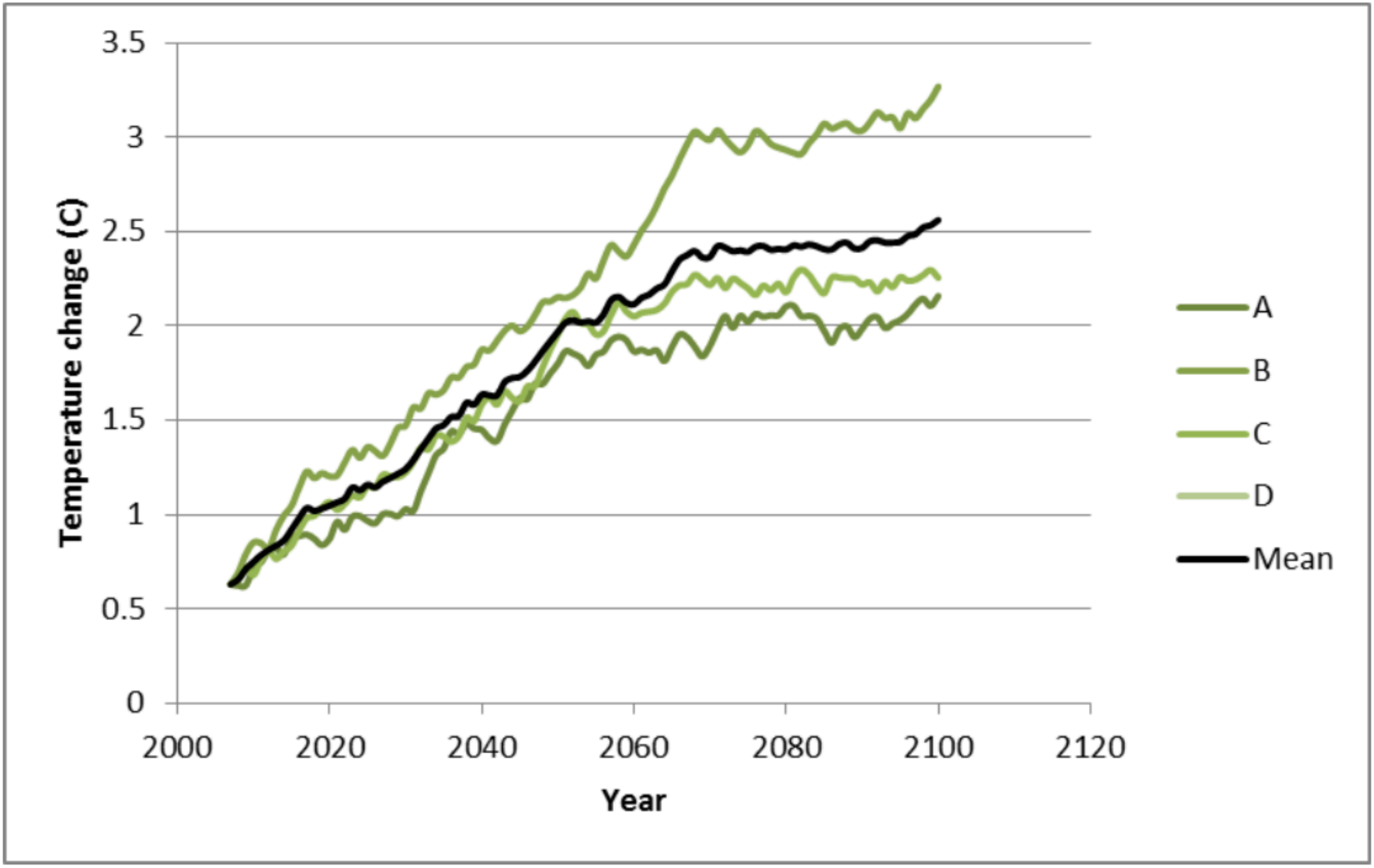
These scenarios provide plausible descriptions of how the future might unfold in several key areas—socioeconomic, technological and environmental conditions, emissions of greenhouse gases and aerosols, and climate.

When applied in climate change research, scenarios help to evaluate uncertainty about human contributions to climate change, the response of the Earth system to human activities, the impacts of a range of future climates, and the implications of different approaches to mitigation (measures to reduce net emissions) and adaptation (actions that facilitate response to new climate conditions).”(Moss et al., 2010)

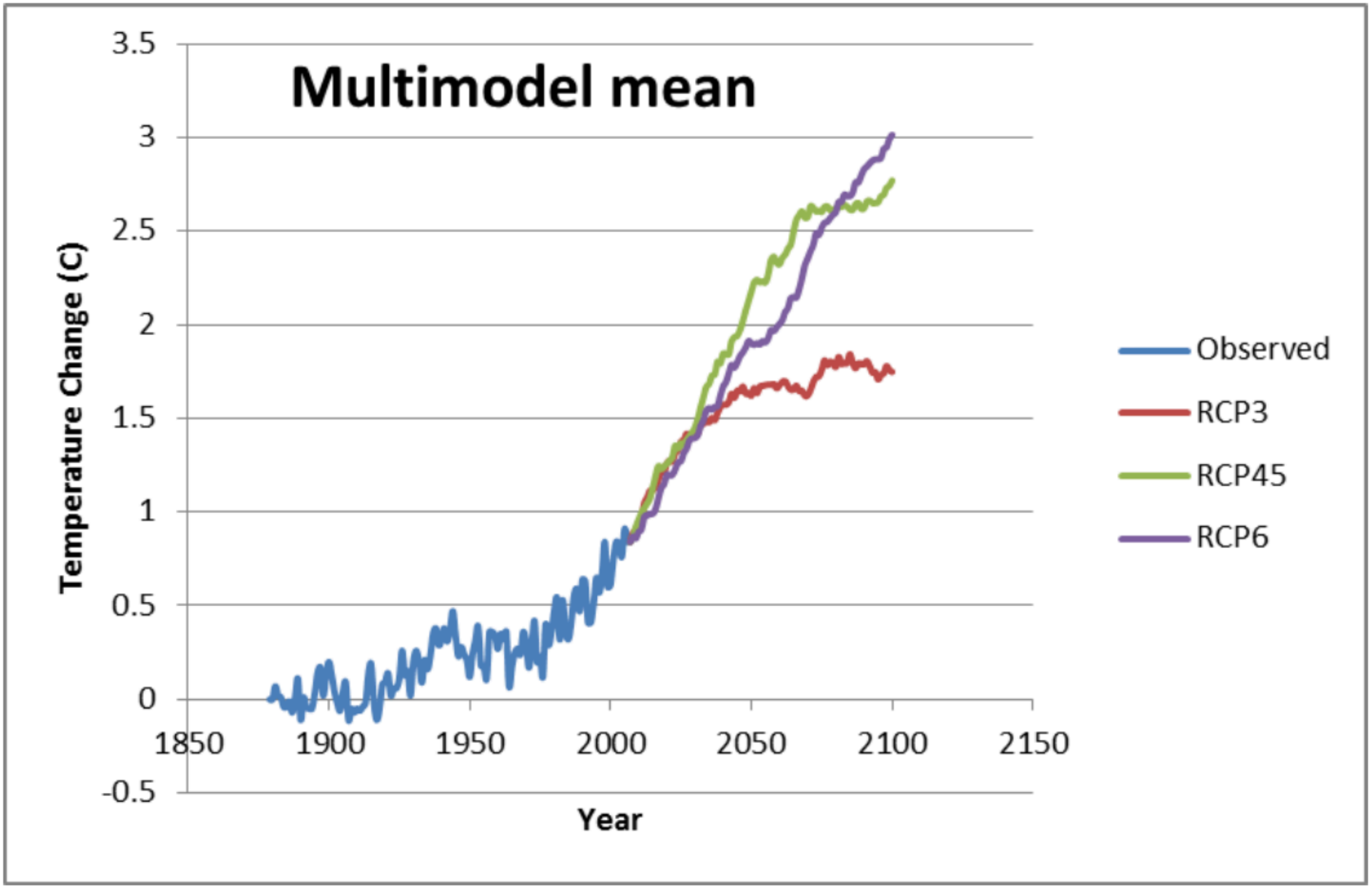
For our model, the RCPs will represent forcing inputs for the simple climate model.

The model generates a temperature projection from a scenario.

An example figure with three temperature estimates (A, B, & C) for RCP 45 is shown below.


3.5
3
2.5
2
1.5
1
0.5
0
Temperature change (C)
2000
2020
2040
2060
2080
2100
2120
Year
A
B
C
D
Mean

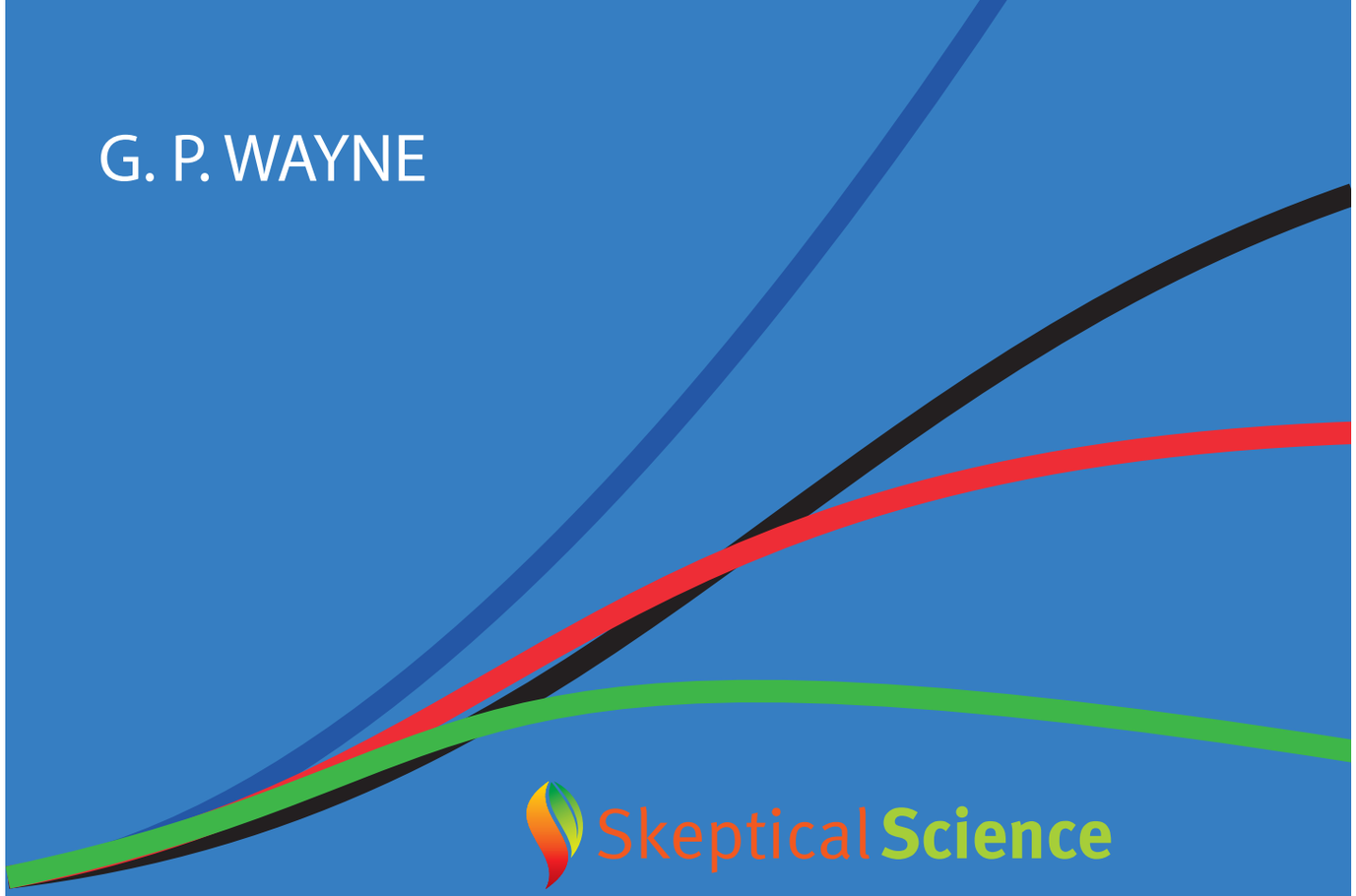
The multimodel mean plot, as shown below, compares different RCPs.



The Beginner's Guide to

Representative Concentration Pathways

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SkepticalScience

Discussion of RCP 4.5 as an example.

RCP4.5: A Pathway for Stabilization of Radiative Forcing by 2100

Allison M. Thomson*, Katherine V. Calvin, Steven J. Smith, G. Page Kyle, April Volke, Pralit Patel, Sabrina Delgado-Arias, Ben Bond-Lamberty, Marshall A. Wise, Leon E. Clarke and James A. Edmonds

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ABSTRACT

Representative Concentration Pathway (RCP) 4.5 is a scenario that stabilizes radiative forcing at 4.5 Watts per meter squared in the year 2100 without ever exceeding that value. Simulated with the Global Change Assessment Model (GCAM), RCP4.5 includes long-term, global emissions of greenhouse gases, short-lived species, and land-use-land-cover in a global economic framework. RCP4.5 was updated from earlier GCAM scenarios to incorporate historical emissions and land cover information common to the RCP process and follows a cost-minimizing pathway to reach the target radiative forcing. The imperative to limit emissions in order to reach this target drives changes in the energy system, including shifts to electricity, to lower emissions energy technologies and to the deployment of carbon capture and geologic storage technology. In addition, the RCP4.5 emissions price also applies to land use emissions; as a result, forest lands expand from their present day extent. The simulated future emissions and land use were downscaled from the regional simulation to a grid to facilitate transfer to climate models. While there are many alternative pathways to achieve a radiative forcing level of 4.5 W m^{-2}, the application of the RCP4.5 provides a common platform for climate models to explore the climate system response to stabilizing the anthropogenic components of radiative forcing.

Keywords: climate change mitigation; Integrated Assessment; Global Change Assessment Model; Representative Concentration Pathway; scenario.

Two simulations

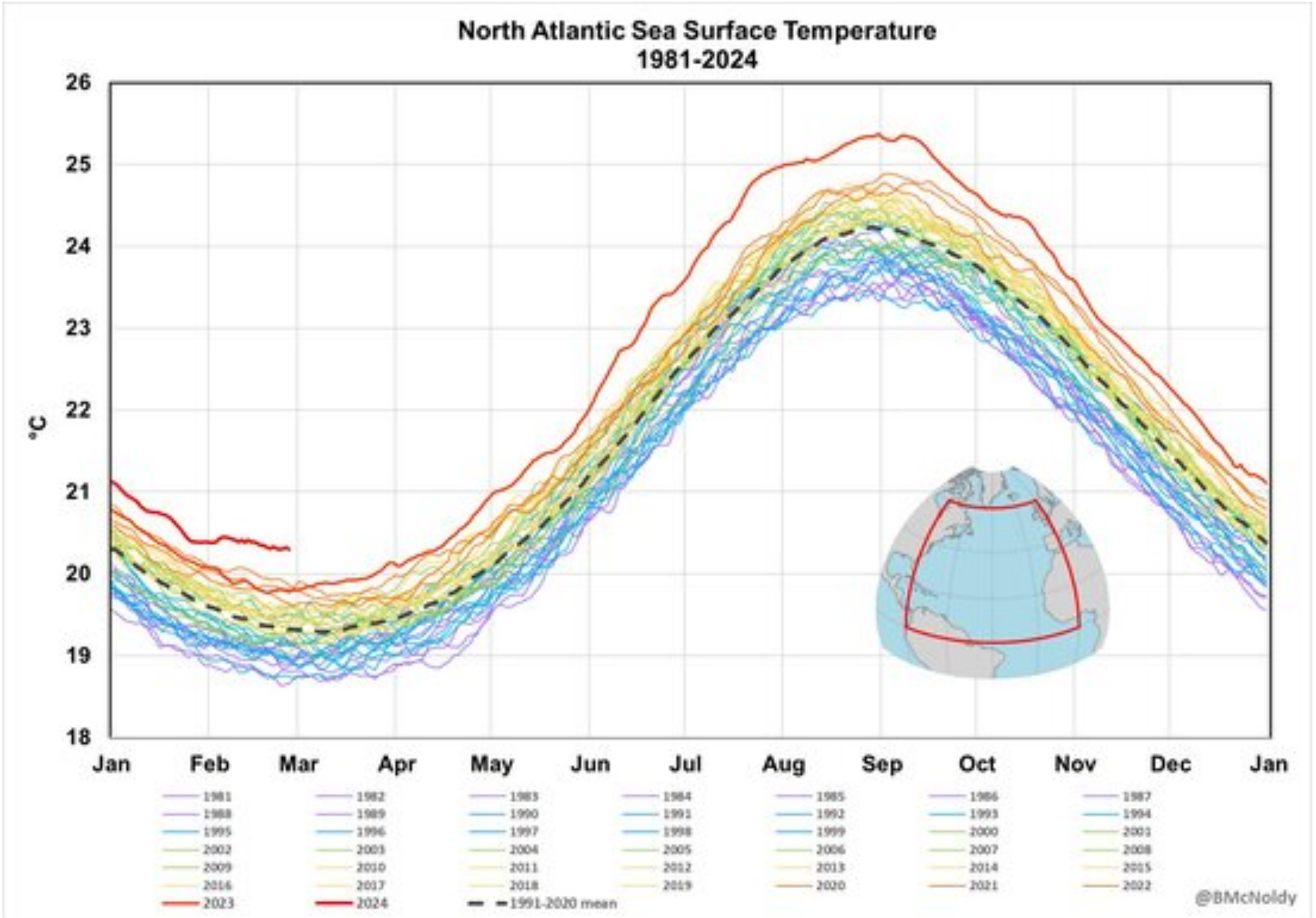
https://en-roads.climateinteractive.org/scenario.html?v=24.3.0

https://c-roads.climateinteractive.org/scenario.html?v=24.3.0

Developments since I wrote the lectures

One: **N. Atlantic Temperatures are "just astonishing. Like it doesn't seem real".**

This chart of ocean temperatures should really scare you



The thick orangey-red line that runs the length of the chart and hovers above nearly all the others is from 2023.

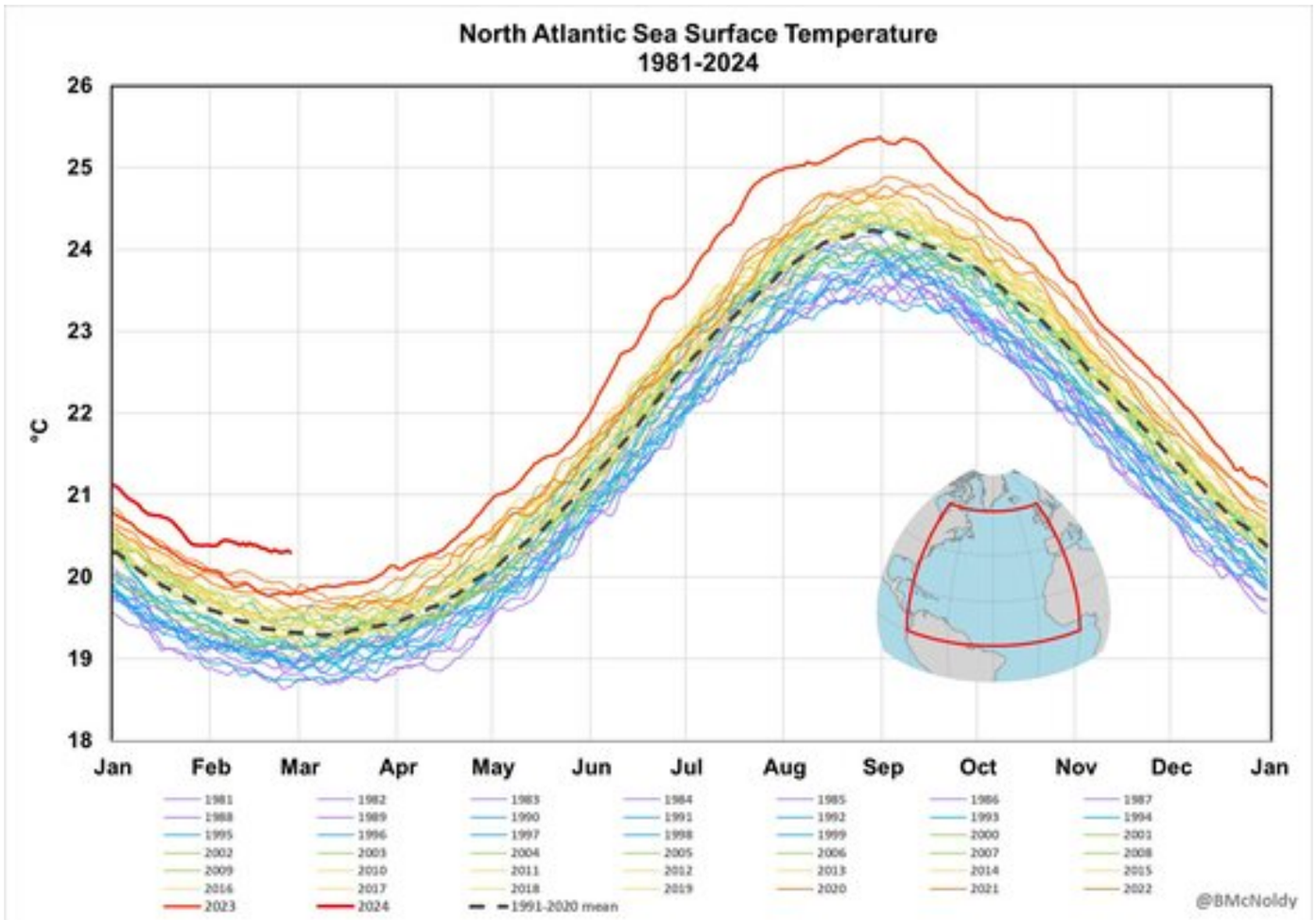
The North Atlantic started breaking heat temperature records in March of last year.

Even more alarming is the departure that the new, shorter line from 2024 represents.

It’s far above the rest, indicating this extreme, anomalous increase has continued into this year.

All this heat is not only a problem for marine ecosystems today, scientists say.

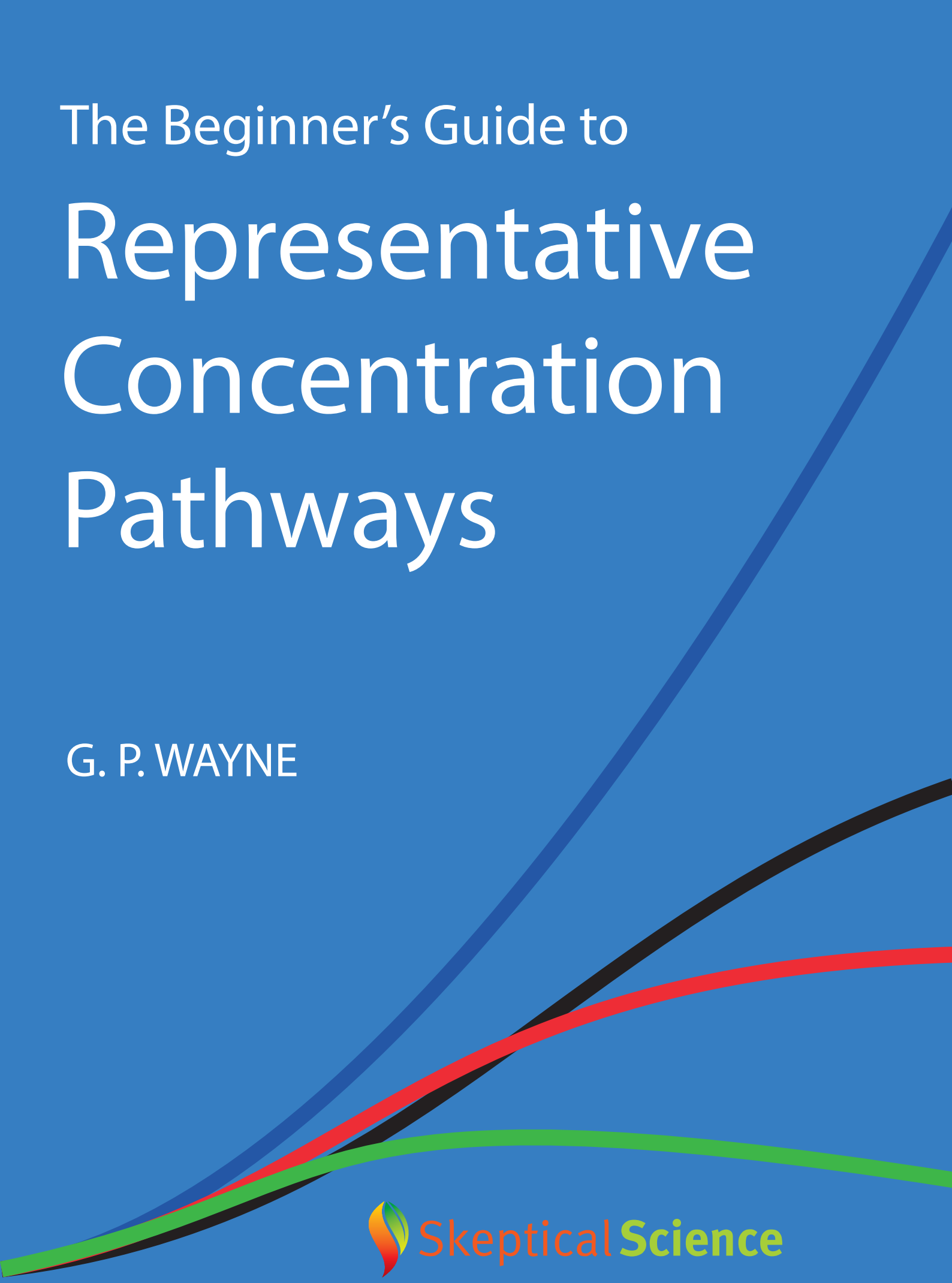
It’s also a warning of what could come — including what could put human life in harm’s way.



The Beginner's Guide to

Representative Concentration Pathways

G. P. WAYNE



Skeptical Science

The Beginner's Guide to Representative Concentration Pathways

By Graham Wayne **Version 1.0, August 2013**

Introduction

Many factors have to be taken into account when trying to predict how future global warming will contribute to climate change. The amount of future greenhouse gas emissions is a key variable. Developments in technology, changes in energy generation and land use, global and regional economic circumstances and population growth must also be considered.

So that research between different groups is complementary and comparable, a standard set of ***scenarios*** are used to ensure that starting conditions, historical data and projections are employed consistently across the various branches of climate science.

The Intergovernmental Panel on Climate Change (IPCC) Fifth Assessment Report (AR5) is due for publication in 2013-14. Its findings will be based on a new set of scenarios that replace the Special Report on Emissions Scenarios (SRES) standards employed in two previous reports. The new scenarios are called **Representative Concentration Pathways (RCPs)**. There are four pathways: RCP8.5, RCP6, RCP4.5 and RCP2.6 - the last is also referred to as RCP3-PD. (The numbers refer to forcings for each RCP; PD stands for *Peak* and *Decline*).

> "The name "representative concentration pathways" was chosen to emphasize the rationale behind their use. RCPs are referred to as pathways in order to emphasize that their primary purpose is to provide time-dependent projections of atmospheric greenhouse gas (GHG) concentrations. In addition, the term pathway is meant to emphasize that it is not only a specific long-term concentration or radiative forcing outcome, such as a stabilization level, that is of interest, but also the trajectory that is taken over time to reach that outcome. They are representative in that they are one of several different scenarios that have similar radiative forcing and emissions characteristics". ***(IPCC Expert Meeting Report, Towards New Scenarios For Analysis Of Emissions, Climate Change, Impacts, And Response Strategies, IPCC 2007)***

This guide to Representative Concentration Pathways assumes no prior knowledge.

In **Part 1** we explore their historical background, explain why scenarios are necessary, and who uses them. Readers already familiar with the background may wish to skip this section.

Part 2 starts with an examination of the demand for new scenarios, and why they were deemed necessary. The aims and requirements of stakeholders are described, how the development teams were selected, and the process by which the RCPs were created, checked and validated.

In **Part 3** we take a look at the scenarios in detail, consider the technical aspects, the differences between the four RCPs, and how they compare to earlier SRES scenarios.

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Part 1: An introduction to scenarios

Why are scenarios necessary?

> **"Scenarios of different rates and magnitudes of climate change provide a basis for assessing the risk of crossing identifiable thresholds in both physical change and impacts on biological and human systems".**
>
> *Source: "Towards New Scenarios for Analysis of Emissions, Climate Change, Impacts, and Response Strategies", IPCC Technical Summary, 2007*

There are many climate modelling teams around the world. If they all used different metrics, made different assumptions about baselines and starting points, then it would be very difficult to compare one study to another. In the same way, models could not be validated against other different, independent models, and communication between climate modelling groups would be made more complex and time-consuming.

Another problem is the cost of running models. The powerful computers required are in short supply and great demand. Simulation programming that had to start from scratch for each experiment would be wholly impractical. Scenarios provide a framework by which the process of building experiments can be streamlined.

In order to address these issues, in 1992 the Intergovernmental Panel on Climate Change (IPCC) published the first set of climate change scenarios, called **IS92**. In year 2000 the IPCC released a second generation of projections, collectively referred to as the Special Report on Emissions Scenarios **(SRES)**. These were used in two subsequent reports; the Third Assessment Report **(TAR)** and Assessment Report Four **(AR4)** and have provided common reference points for a great deal of climate science research in the last decade.

In 2007, the IPCC responded to calls for improvements to SRES by catalysing the process that produced the Representative Concentration Pathways (RCPs). The RCPs are the latest iteration of the scenario process, and are used in the next IPCC report - Assessment Report Five **(AR5)** in preference to SRES. Here's how the IPCC describes the scenarios (emphasis added):

> "In climate change research, scenarios describe plausible trajectories of different aspects of the future that are constructed to investigate the potential consequences of anthropogenic climate change. Scenarios represent many of the major driving forces - including processes, impacts (physical, ecological, and socioeconomic), and potential responses that are important for informing climate change policy. They are used to hand off information from one area of research to another (e.g., from research on energy systems and greenhouse gas emissions to climate modeling). They are also used to explore the implications of climate change for decision making (e.g., exploring whether plans to develop water management infrastructure are robust to a range of uncertain future climate conditions). **The goal of working with scenarios is not to predict the future but to better understand uncertainties and alternative futures, in order to consider how robust different decisions or options may be under a wide range of possible futures".**
>
> *Source: IPCC Scenario Process for AR5*

Who uses climate scenarios?

There are several primary groups who study the effects of climate change. Climate Model (CM) groups study the effects of global warming on the climate itself and how our emissions affect the environment. Integrated Assessment Model (IAM) groups combine information from diverse fields of study, primarily to assess the relationship between emissions and socio-economic scenarios. A third group studies Impacts, Adaptation and Vulnerabilities (IAV), often at regional scales, drawing on disciplines and research traditions including social sciences, economics, engineering, and the natural sciences.

This graphic from the IPCC shows the relationships between the various groups, and their main areas of exploration:

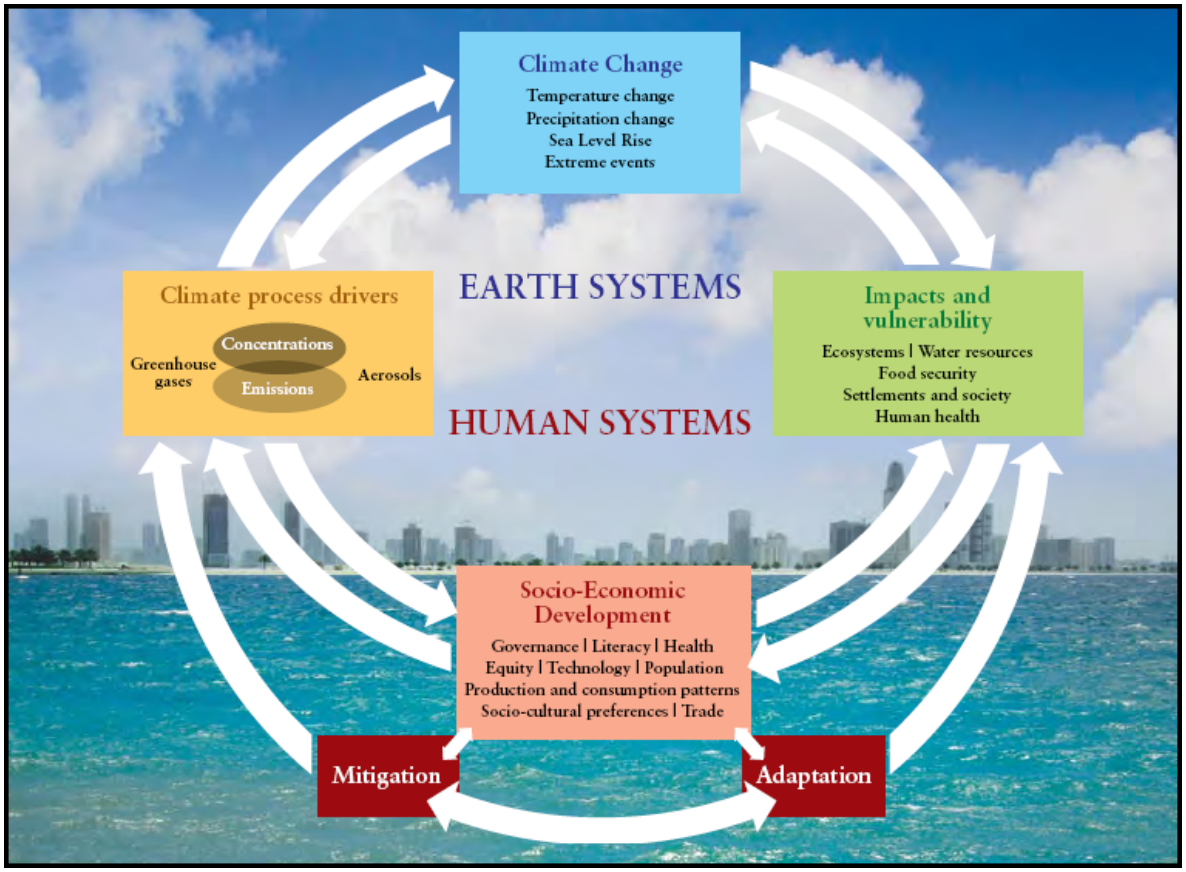


Figure 1. *From the IPCC Expert Meeting Report: Towards New Scenarios - Technical Summary*

As well as the groups mentioned above, scenarios are used extensively by scientists, policy makers, NGOs and commentators as a common framework through which they can discuss climate change, exchange ideas and communicate with each other effectively.

What do RCPs Consist Of ?

A RCP scenario basically consists of numbers - a prodigious amount of them. RCP data is in tables - if you're familiar with a spreadsheet, the format is somewhat similar.

For each category of emissions, an RCP contains a set of starting values and the estimated emissions up to 2100, based on assumptions about economic activity, energy sources, population growth and other socio-economic factors. (The data also contain historic, real-world information). While socio-economic projections were drawn from the literature in order to develop the emission pathways, the database does not include socio-economic data.

Modellers download the database sets to *initialise* their models, which jump-starts what would otherwise be a very lengthy process - one that each modelling team would have to attempt, thus duplicating effort. RCPs and previous scenarios were created exactly to avoid such duplication, and the inevitable initialisation inconsistencies that would ensue.

A quick look at the RCP database screenshot below shows how many emissions categories are addressed by the RCPs. Each RCP contains the same categories of data, but the values vary a great deal, reflecting different emission trajectories over time as determined by the underlying socio-economic assumptions (which are unique to each RCP).

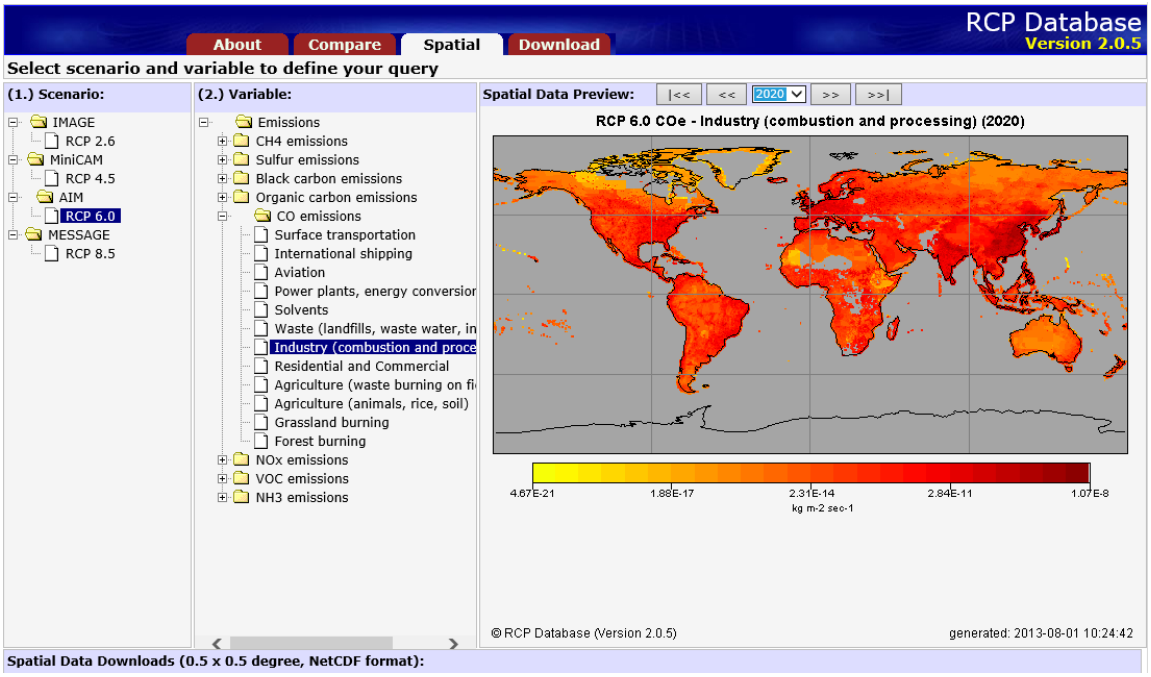


***Figure 2**: RCP on-line database showing RPC6 spatial data for industry COe emissions for the year 2020.*

High-resolution data is generated for a world divided into 'cells' measuring half a degree of latitude and longitude - 518,400 cells in total. The RCP database web interface provides only a preview of the data, which can comprise far more detail than a graphic can show. It is, however, a starting point for researchers, who can evaluate the data graphically before downloading it. (As an alternative, the Compare option allows researchers to plot a graph of trajectories for all four RCPs - some the graphs shown later in this guide were produced using the RCP on-line facility).

For example, here are two graphic representations of RCP6 spatial COe emissions for the years 2010 and 2100:

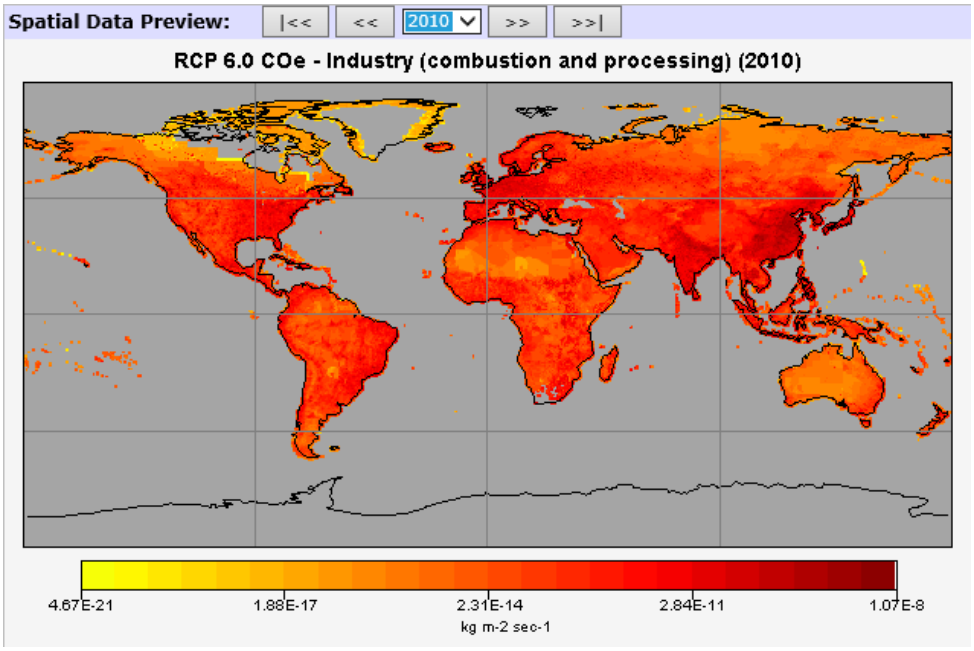


***Figure 3**: RCP on-line database graphic showing RPC6 spatial data for industry COe emissions for the year 2010…*

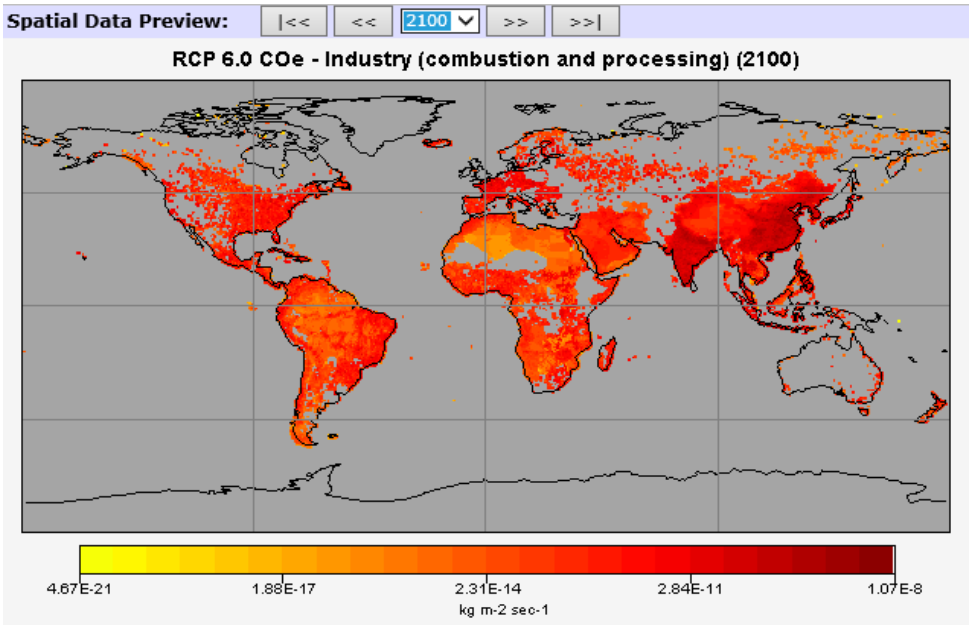


***Figure 4**: …and here's the comparison graphic showing the projected RCP6 emissions in the year 2100.*

By using the all the data available for the intervening years, a trajectory can be given for any specific emissions. Each RCP plots a different emissions trajectory (pathway) and cumulative emission concentration in 2100.

The deliverable is a download from a central repository. Scientists can preview and download data on emissions, concentrations, radiative forcing and land use, in regional and gridded form, following different trajectories over similar timescales. These data sets can then be incorporated into any modelling exercise, providing consistent parameters for each emissions trajectory, and a consistent foundation for all climate modelling teams anywhere in the world. The database is also open to the public and can be accessed without charge using any browser: http://tntcat.iiasa.ac.at:8787/RcpDb/dsd?Action=htmlpage&page=welcome

Part 2: Creating new scenarios

> **"New sets of scenarios for climate change research are needed periodically to take into account scientific advances in understanding of the climate system as well as to incorporate updated data on recent historical emissions, climate change mitigation, and impacts, adaptation, and vulnerability".**
>
> *Source: IPCC Scenario Process For AR5*

Science always seeks to improve its knowledge, its skills, its tools and methods. As scientists make new discoveries, these have to be incorporated in the scenarios they use. Computers improve at tremendous rates, so models that would have been impossible to run 20 years ago are now not only feasible, but desirable. Scientists and decision makers investigating climate change want faster results, more resolution, more detail, better validation and easier data exchange between related and disparate lines of research. These demands inform an on-going requirement to improve the scenarios.

Scenarios don't just contain reference data; they also specify a *process*. The previous IPCC scenarios like SRES were run in sequence (see graphic below). This resulted in protracted development and delivery times. According to the IPCC: *"Lags in the development process meant that it was often many years until climate and socioeconomic scenarios were available for use in studies of impacts, adaptation, and vulnerability"*.

Not only that, but changes to prior processes in a sequential model meant going back and re-running the simulation in order to incorporate the new or changed data.

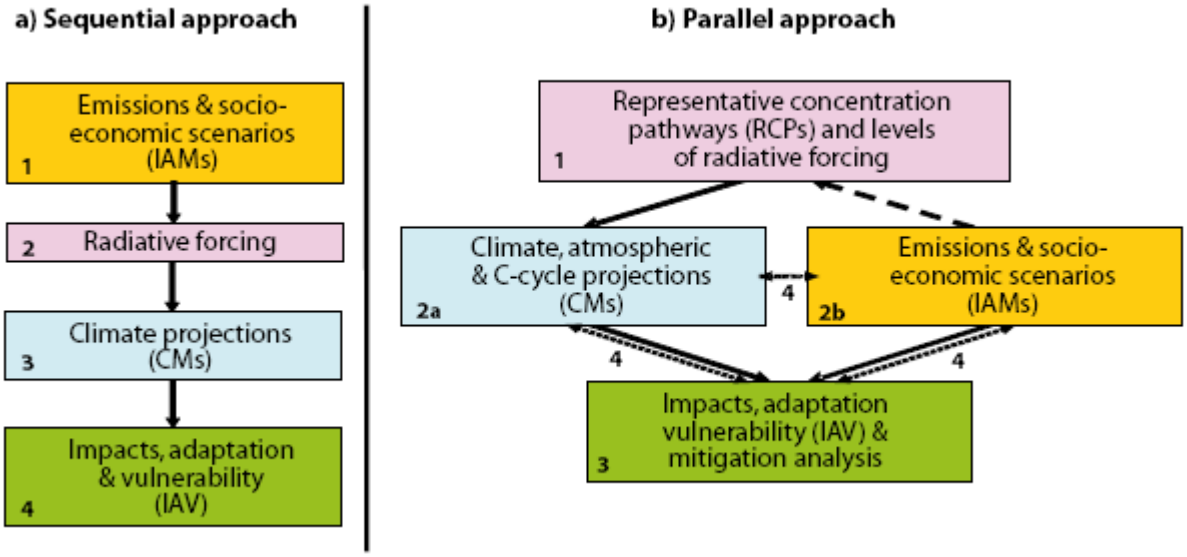


***Figure 5. Approaches to the development of global scenarios**: (a) previous sequential approach; (b) proposed parallel approach. Numbers indicate analytical steps (2a and 2b proceed concurrently). Arrows indicate transfers of information (solid), selection of RCPs (dashed), and integration of information and feedbacks (dotted). Source: Moss et al. (2008).*

The new Representative Concentration Pathways employ a process intended to make the modelling less time-consuming, more flexible, with a reduced economic cost of computation:

> "In the new process… emissions and socioeconomic scenarios are developed in parallel, building on different trajectories of radiative forcing over time…Rather than starting with detailed socio-economic storylines to generate emissions and then climate scenarios, the new process begins with a limited number of alternative pathways (trajectories over time) of radiative forcing levels (or CO2-equivalent concentrations) that are both representative of the emissions scenario literature and span a wide space of resulting greenhouse gas concentrations that lead to clearly distinguishable climate futures.
>
> "These radiative forcing trajectories were thus termed "Representative Concentration Pathways" (RCPs). The RCPs are not associated with unique socioeconomic assumptions or emissions scenarios but can result from different combinations of economic, technological, demographic, policy, and institutional futures".
>
> *(Source: IPCC Scenario Process for AR5)*

Design Criteria

Four design criteria were agreed for the RCPs, as described in Moss et.al. 2008 and quoted here from van Vuuren 2011:

1. The RCPs should be based on scenarios published in the existing literature, developed independently by different modeling groups and, as a set, be 'representative' of the total literature, in terms of emissions and concentrations (see further in this section); At the same time, each of the RCPs should provide a plausible and internally consistent description of the future;
2. The RCPs should provide information on all components of radiative forcing that are needed as input for climate modeling and atmospheric chemistry modeling (emissions of greenhouse gases, air pollutants and land use). Moreover, they should make such information available in a geographically explicit way;
3. The RCPs should have harmonized base year assumptions for emissions and land use and allow for a smooth transition between analyses of historical and future periods;
4. The RCPs should cover the time period up to 2100, but information also needs to be made available for the centuries thereafter.
5.

Working with the Stakeholders

Early in the process of developing new scenarios, the IPCC decided to act only as a catalyst for the process, inviting the research community to develop the scenarios. The subsequent RCP development process was led by the research community (the Integrated Assessment Model Consortium) at the request of the IPCC, but independent of them. A meeting was convened in September 2007 to discuss and agree the way forward:

> "The meeting brought together over 130 participants, including users of scenarios and representatives of the principal research communities involved in scenario development and application. The representatives of the scenario user community included officials from

national governments, including many participating in the United Nations Framework Convention on Climate Change (UNFCCC), international organizations, multilateral lending institutions, and nongovernmental organizations (NGOs). The principal research communities represented at the expert meeting were the integrated assessment modeling (IAM) community; the impacts, adaptation, and vulnerability (IAV) community; and the climate modeling (CM) community. Because of this broad participation, the meeting provided an opportunity for the segments of the research community involved in scenario development and application to discuss their respective requirements and coordinate the planning process".

Source: IPCC Expert Meeting Report

The four IAM groups responsible for the four published scenarios that were selected as "predecessors" of the RCPs, generated the basic data sets from which the final RCPs were developed.

Over the following two years, a unique collaborative effort between integrated assessment modellers, climate modellers, terrestrial ecosystem modellers and emission inventory experts led to the agreement and specification of the four Representative Concentration Pathways (RCPs).

Improvements over SRES

Perhaps the most innovative aspect of the RCPs is that instead of starting with socio-economic 'storylines' from which emission trajectories and climate impacts are projected (the SRES methodology), RCPs each describe an emission trajectory and concentration by the year 2100, and consequent forcing. Each trajectory represents a specific synthesis drawn from the published literature. From this 'baseline', researchers can then test various permutations of social, technical and economic circumstances. These permutations are called '**narratives'**, equivalent to the '**storylines'** employed in SRES.

"As stand alone products, the RCPs have limited usefulness to other research communities. First and foremost, they were selected with the sole purpose of providing data to climate models, taking into consideration the limitations in climate models differentiating levels of radiative forcing. They lack associated socioeconomic and ecological data. They were developed using idealized assumptions about policy instruments and the timing of participation by the international community.

"Therefore, there is a need to develop socioeconomic and climate impact scenarios that draw on the RCPs and associated climate change projections in the scenario process. Referencing the RCP and climate change projections has two potential benefits; they would facilitate comparison across research results in the CM, IAM, and IAV communities, and facilitate use of new climate modeling results in conjunction with IAV research.

"The *parallel phase* has several components. Within CMIP5, CM teams are using the RCPs as an input for model ensemble projections of future climate change. These projections will form the backbone of the IPCC's Working Group I assessment of future climate change in the 5th Assessment Report (AR5). The IAM community has begun exploring new socioeconomic scenarios and producing so-called RCP replications that study the range of socioeconomic scenarios leading to the various RCP radiative forcing levels. In the

meantime, IAV analyses based on existing emission scenarios (SRES) and climate projections (CMIP3) continue.

"In the *integration phase*, consistent climate and socioeconomic scenarios will inform IAM and IAV studies. For example, IAV researchers can use the new scenarios to project impacts, to explore the extent to which adaptation and mitigation could reduce projected impacts, and to estimate the costs of action and inaction. Also, mitigation researchers can use the global scenarios as "boundary conditions" to assess the cost and effectiveness of local mitigation measures, such as land-use planning in cities or changes in regional energy systems.

"These scenarios need to supply quantitative and qualitative narrative descriptions of potential socioeconomic and ecosystem reference conditions that underlie challenges to mitigation and adaptation. And they have to be flexible enough to provide a framework for comparison within which regional or local studies of adaptation and vulnerability could build their own narratives. The defining socioeconomic conditions of these scenarios have been designated **Shared Socioeconomic reference Pathways (SSPs)**."
Source: A framework for a new generation of socioeconomic scenarios for climate change impact, adaptation, vulnerability, and mitigation research; Arnell, Kram, Carter et.al.

For the first time, policy decisions can be tested; previous scenarios were describes as 'no-policy', meaning the scenarios did not respond to changes driven by political or legislative inputs, so mitigation or adaptation strategies could not be incorporated.

The principle difference in approach is that previously, SRES specified the socio-economic circumstances for each scenario, which essentially 'locked in' the options for socio-economic change (and led to a proliferation of SRES scenarios - 40 in total, each a slightly different variation on common socio-economic variables). Models were programmed to generate emissions and subsequent climate scenarios. The socio-economic variables of the SRES scenarios were socially and policy-proscriptive, inflexible in a way emissions and climate change outcomes were not.

By fixing the emissions trajectory and the warming, RCPs come at the problem the other way round. Socio-economic options become flexible and can be altered at will, allowing considerably more realism by incorporating political and economic flexibility at regional scales. Policy decisions on mitigation and adaptation can be tested for economic efficacy, both short and long term. Researchers can test various socio-economic measures against the fixed rates of warming built into the RCPs, to see which combinations of mitigation or adaptation produce the most timely return on investment and the most cost-effective response.

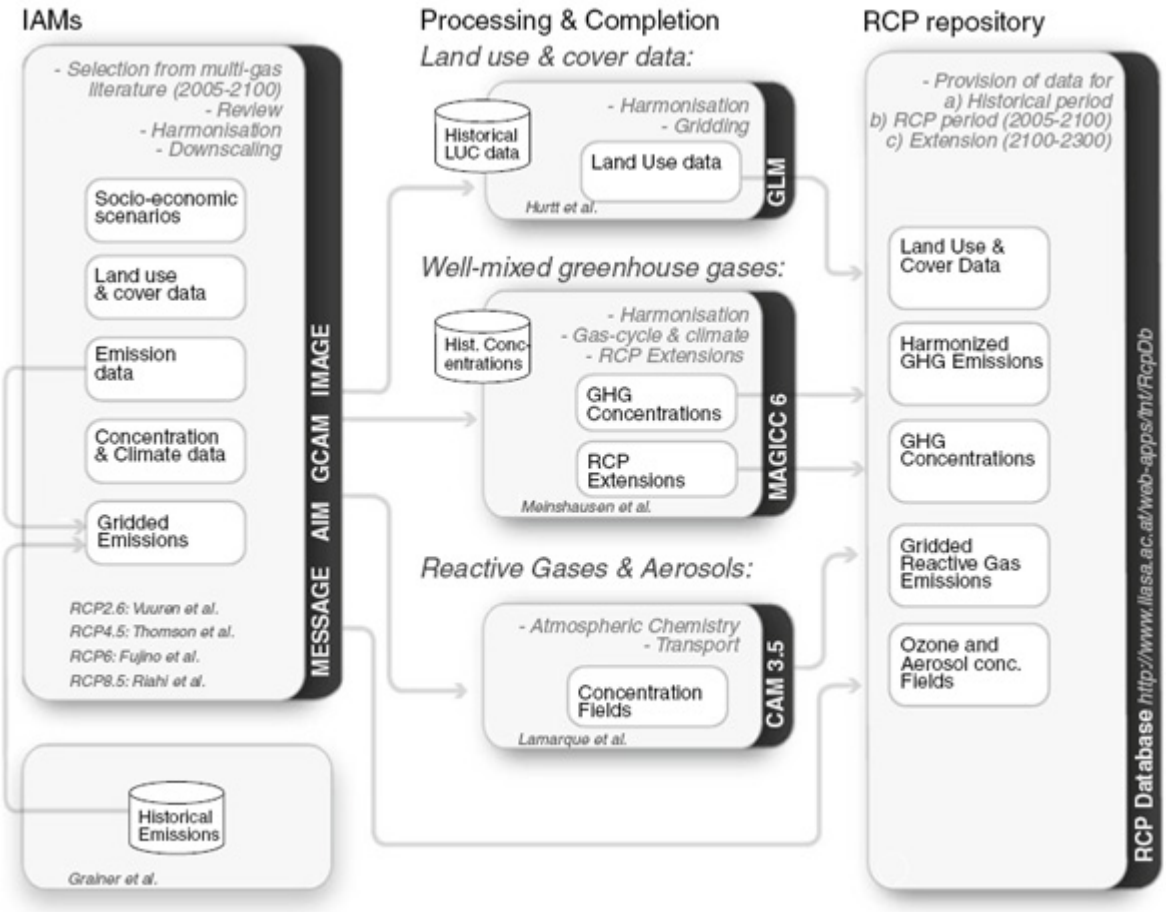


***Figure 6: Overview of the RCP development process**, adapted from van Vuuren et.al. 2011*

Development Aims and Products

There were five end-products expected from development process:

1. Four Representative concentration pathways (RCPs). Four RCPs…produced from IAM scenarios available in the published literature: one high pathway for which radiative forcing reaches >8.5 W/m2 by 2100 and continues to rise for some amount of time; two intermediate "stabilization pathways" in which radiative forcing is stabilized at approximately 6 W/m2 and 4.5 W/m2 after 2100; and one pathway where radiative forcing peaks at approximately 3 W/m2 before 2100 and then declines. These scenarios include time paths for emissions and concentrations of the full suite of GHGs and aerosols and chemically active gases, as well as land use/land cover…

2. RCP-based climate model ensembles and pattern scaling. Ensembles of gridded, time-dependent projections of climate change produced by multiple climate models including atmosphere–ocean general circulation models (AOGCMs), Earth system models (ESMs), Earth system models of intermediate complexity, and regional climate models will be prepared for the four long-term RCPs, and high-resolution, near-term projections to 2035 for the 4.5 W/m2 stabilization RCP only.

3. New IAM scenarios. New scenarios will be developed by the IAM research community in consultation with the IAV community exploring a wide range of dimensions associated with anthropogenic climate forcing…Anticipated outputs include alternative socioeconomic driving

forces, alternative technology development regimes, alternative realizations of Earth system science research, alternative stabilization scenarios including traditional "not exceeding" scenarios, "overshoot" scenarios, and representations of regionally heterogeneous mitigation policies and measures, as well as local and regional socioeconomic trends and policies…

4. Global narrative storylines. These are detailed descriptions associated with the four RCPs produced in the preparatory phase and such pathways developed as part of Product 3 by the IAM and IAV communities. These global and large-region storylines should be able to inform IAV and other researchers.

5. Integrated scenarios. RCP-based climate model ensembles and pattern scaling (Product 2) will be associated with combinations of new IAM scenario pathways (Product 3) to create combinations of ensembles. These scenarios will be available for use in new IAV assessments. In addition, IAM research will begin to incorporate IAV results, models, and feedbacks to produce comprehensively synthesized reference.

(End-product definitions excerpted from the IPCC report "Towards New Scenarios...")

Beyond 2100 - Extended Concentration Pathways (ECP)

During the consultation phase, modelling communities made clear their interest in exploring longer-term processes. To facilitate these investigations, a single extension was developed for each RCP, extending the scenarios up to the year 2300. These data form the Extended Concentration Pathways (ECPs).

Since socio-economic factors cannot be predicted reliably over long timescales, the ECPs were developed using simple rules to extend GHG concentrations, emissions and land-use data series. The ECPs are intended only as the basis for long-term climate simulations.

A supplemental RCP called SCP6to4.5 was also developed with a peak at 2100, followed by a decline, to facilitate specific investigations into physical asymmetries and reversibility of climate, carbon cycle, and biophysical impacts systems (e.g. ecosystems, sea level rise).

Parameter	ECP	Generic rule
CO_2 and other well-mixed GHGs	ECP8.5	Follow stylized emission trajectory that leads to stabilization at 12 W/m^2
	ECP6 ECP4.5	Stabilize concentrations in 2150 (around 6.0 W/m^2) Stabilize concentrations in 2150 (around 4.5 W/m^2)
	ECP3PD	Keep emissions constant at 2100 level
	SCP6to4.5	Return radiative forcing of all gases from RCP6.0 to RCP4.5 levels by 2250
Reactive gases	All ECPs	Keep constant at 2100 level
	SCP6to4.5	Scale forcing of reactive gases with GHG forcing
Land use	All ECPs	Keep constant at 2100 level

Table 1: Basic rules for deriving extended concentration pathways *(van Vuuren et.al. 2011)*

Part 3: RCP technical summary

This section contains a summary of the key metrics and assumptions that define the RCP architecture: emissions trajectories and concentrations, energy use, population, air pollutants and land use, and the consequent radiative forcing and temperature anomalies specified by each of the four RCP pathways.

The data employed in the development of the RCPs is drawn from the published literature. Each RCP was developed by an Integrated Assessment Modeling (IAM) group, whose published scenario papers were consistent with the base criteria for a particular RCP. Each team then surveyed and created synthesis data sets from available representative studies, which were reviewed repeatedly by different stakeholders. The final agreed set of RCPs was published:

Pathway	Paper	Model
RCP 2.6	Van Vuuren et al. 2007a; van Vuuren et al. 2006	IMAGE
RCP 4.5	Clarke et al. 2007; Smith and Wigley 2006; Wise et al. 2009	GCAM
RCP6	Fujino et al. 2006; Hijioka et al. 2008	AIM
RCP8.5	Riahi et al. 2007	MESSAGE

Table 2: RCP-specific publications and model group responsible.

For comprehensive discussions of development methodologies and complete technical information on any RCP, please see the further reading section at the end of this guide.

An important note about socio-economic data

The underlying assumptions about socio-economic trajectories and priorities are not consistent between the RCPs. This quote from van Vuuren 2011 makes clear this point (emphasis added):

> "The RCPs were selected from the existing literature on the basis of their emissions and associated concentration levels. This implies that the socio-economic assumptions of the different modeling teams were based on individual model assumptions made within the context of the original publication, and that **there is no consistent design behind the position of the different RCPs relative to each other for these parameters.**
>
> Scenario development after the RCP phase will focus on developing a new set of socio-economic scenarios. Therefore, socio-economic parameters have not been included in the RCP information available for download. Still, this information does form part of the underlying individual scenario development, and thus provides useful information on internal logic and the plausibility of each of the individual RCPs".
>
> ***van Vuuren et. al, 2011***

Socio-economic data does not form any part of the RCP database. Please note that in this guide, as in van Vuuren 2011, the primary socio-economic characteristics are discussed here only in the context of the RCP development.

RCP Primary Characteristics

RCP 8.5 was developed using the MESSAGE model and the IIASA Integrated Assessment Framework by the International Institute for Applied Systems Analysis (IIASA), Austria. This RCP is characterized by increasing greenhouse gas emissions over time, representative of scenarios in the literature that lead to high greenhouse gas concentration levels (Riahi et al. 2007).

RCP6 was developed by the AIM modeling team at the National Institute for Environmental Studies (NIES) in Japan. It is a stabilization scenario in which total radiative forcing is stabilized shortly after 2100, without overshoot, by the application of a range of technologies and strategies for reducing greenhouse gas emissions (Fujino et al. 2006; Hijioka et al. 2008).

RCP 4.5 was developed by the GCAM modeling team at the Pacific Northwest National Laboratory's Joint Global Change Research Institute (JGCRI) in the United States. It is a stabilization scenario in which total radiative forcing is stabilized shortly after 2100, without overshooting the long-run radiative forcing target level (Clarke et al. 2007; Smith and Wigley 2006; Wise et al. 2009).

RCP2.6 was developed by the IMAGE modeling team of the PBL Netherlands Environmental Assessment Agency. The emission pathway is representative of scenarios in the literature that lead to very low greenhouse gas concentration levels. It is a "peak-and-decline" scenario; its radiative forcing level first reaches a value of around 3.1 W/m2 by mid-century, and returns to 2.6 W/m2 by 2100. In order to reach such radiative forcing levels, greenhouse gas emissions (and indirectly emissions of air pollutants) are reduced substantially, over time (Van Vuuren et al. 2007a).
(Characteristics quoted from van Vuuren et.al. 2011)

RCP Information, Data Types and Resolutions

The following table shows the data types available for the RCPs, the sectors by which emissions are broken down, and the geographical resolution of the information:

Available information from RCPs and resolution		
	Resolution (sectors)	**Resolution (geographical)**
Emissions of greenhouse gases		
CO_2	Energy/industry, land	Global and for 5 regions
CH_4	12 sectors	0.5°×0.5° grid
N_2O, HFCs, PFCs, CFCs, SF_6	Sum	Global and for 5 regions
Emissions aerosols and chemically active gases		
SO_2, Black Carbon, Organic Carbon, CO, NO_x, VOCs, NH_3	12 sectors	0.5°×0.5° grid
Speciation of VOC emissions		0.5°×0.5° grid
Concentration of greenhouse gases		
(CO_2, CH_4, N_2O, HFCs, PFCs, CFCs, SF_6)		Global
Concentrations of aerosols & chemically active gases		
(O_3, Aerosols, N deposition, S deposition)		0.5°×0.5° grid
Land-use/land-cover data	Cropland, pasture, primary vegetation, secondary vegetation, forests	0.5°×0.5° grid with subgrid fractions, (annual maps and transition matrices including wood harvesting)

Table 3: from van Vuuren et.al. 2011

Emissions and concentrations, forcings and temperature anomalies

Each Representative Concentration Pathway (RCP) defines a specific emissions trajectory and subsequent radiative forcing (a radiative forcing is a measure of the influence a factor has in altering the balance of incoming and outgoing energy in the Earth-atmosphere system, measured in watts per square metre):

Name	Radiative forcing	CO_2 equiv (p.p.m.)	Temp anomaly (°C)	Pathway	SRES temp anomaly equiv
RCP8.5	8.5 Wm^2 in 2100	1370	4.9	Rising	SRES A1F1
RCP6.0	6 Wm^2 post 2100	850	3.0	Stabilization without overshoot	SRES B2
RCP4.5	4.5 Wm^2 post 2100	650	2.4	Stabilization without overshoot	SRES B1
RCP2.6 (RCP3PD)	3Wm^2 before 2100, declining to 2.6 Wm^2 by 2100	490	1.5	Peak and decline	None

Table 4: from Moss et.al. 2010. *Median temperature anomaly over pre-industrial levels and SRES comparisons based on nearest temperature anomaly, from Rogelj et.al. 2012*

Radiative Forcings

The graph below shows radiative forcing trajectories for the four RCPs, the other candidate scenarios that informed the final versions, and the modelling group associated with each.

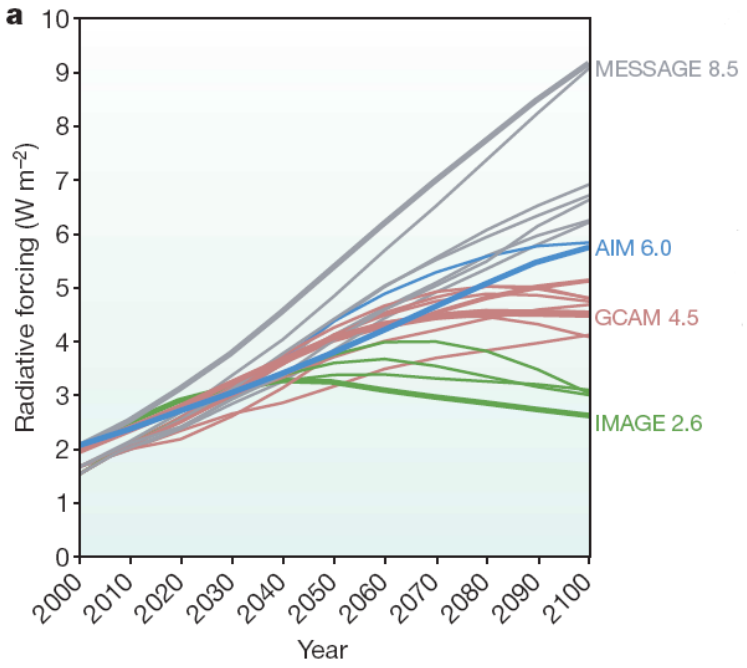


Figure 7: Changes in radiative forcing relative to pre-industrial conditions. *Bold coloured lines show the four RCPs; thin lines show individual scenarios from approximately 30 candidate RCP scenarios that provide information on all key factors affecting radiative forcing… (Moss et.al., 2010)*

The forcing trajectories are consistent with socio-economic projections unique to each RCP. For example, RCP2.6 (RCP3PD) assumes that through drastic policy intervention, greenhouse gas emissions are reduced almost immediately, leading to a slight reduction on today's levels by 2100. The worst case scenario - RCP8.5 - assumes more or less unabated emissions.

RCP Emission Trajectories

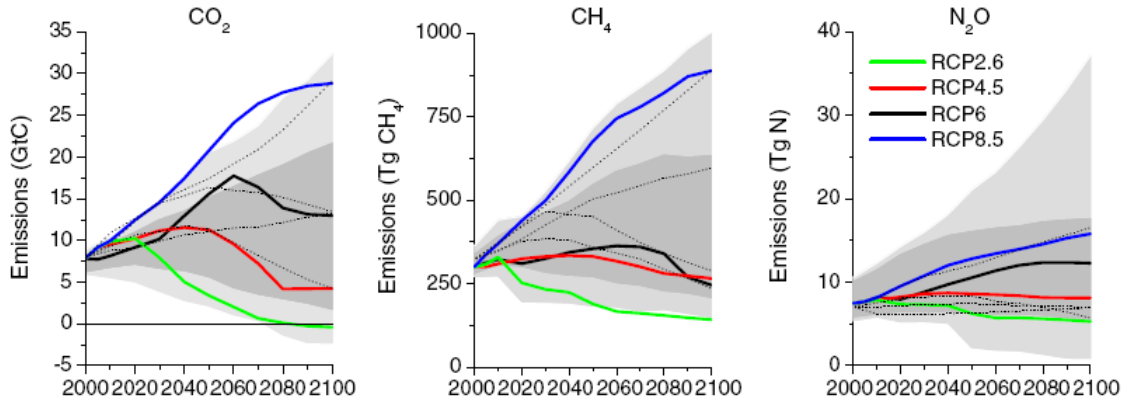


Figure 8: Emissions of main greenhouse gases across the RCPs. *Grey area indicates the 98th and 90th percentiles (light/dark grey) of the literature…The dotted lines indicate four of the SRES marker scenarios. Note that the literature values are not harmonized (from van Vuuren et.al. 2011)*

"The CO2 emissions of the four RCPs correspond well with the literature range, which was part of their selection criterion (Fig. 8). The RCP8.5 is representative of the high range of non-climate policy scenarios. Most non-climate policy scenarios, in fact, predict emissions of the order of 15 to 20 GtC by the end of the century, which is close to the emission level of the RCP6. The forcing pathway of the RCP4.5 scenario is comparable to a number of climate policy scenarios and several low-emissions reference scenarios in the literature, such as the SRES B1 scenario. The RCP2.6 represents the range of lowest scenarios, which requires stringent climate policies to limit emissions.

"The trends in CH4 and N2O emissions are largely due to differences in the assumed climate policy along with differences in model assumptions (Fig. 8). Emissions of both CH4 and N2O show a rapidly increasing trend for the RCP8.5 (no climate policy and high population). For RCP6 and RCP4.5, CH4 emissions are more-or-less stable throughout the century, while for RCP2.6, these emissions are reduced by around 40%. The low emission trajectories for CH4 are a net result of low cost emission options for some sources (e.g. from energy production and transport), and a limited reduction for others (e.g. from livestock)" ***(van Vuuren et.al. 2011)***

Greenhouse Gas Concentrations

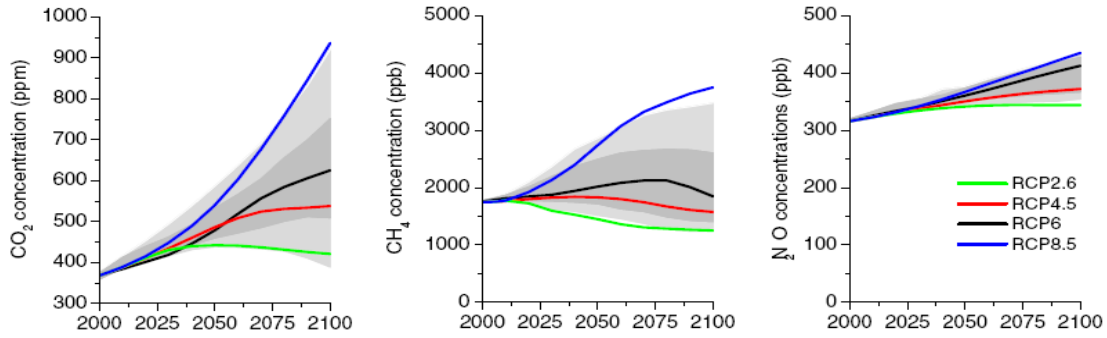


Figure 9: Trends in concentrations of greenhouse gases (van Vuuren 2011)*. Grey area indicates the 98th and 90th percentiles (light/dark grey) of the recent EMF-22 study (Clarke et al. 2010)*

"The greenhouse gas concentrations in the RCPs closely correspond to the emissions trends discussed earlier . For CO2, RCP8.5 follows the upper range in the literature (rapidly increasing

concentrations). RCP6 and RCP4.5 show a stabilizing CO2 concentration (close to the median range in the literature). Finally, RCP2.6 has a peak in CO2 concentrations around 2050, followed by a modest decline to around 400 ppm CO2, by the end of the century. For CH4 and N2O, the order in which the RCPs can be placed are also a direct result of the assumed level of climate policy. The trends in CH4 concentrations are more pronounced, as a result of the relatively short lifetime of CH4. Emission reductions, as in the RCP2.6 and RCP4.5, therefore, may lead to an emission peak much earlier in the century. For N2O, in contrast, a relatively long lifetime and a modest reduction potential imply an increase in concentrations, in all RCPs. For both CH4 and N2O, the concentration levels correspond well with the range in the literature. Further information on the calculations of concentration can be found in Meinshausen et al. (2011b)*" **(van Vuuren et.al. 2011).***

Atmospheric Air Pollutants

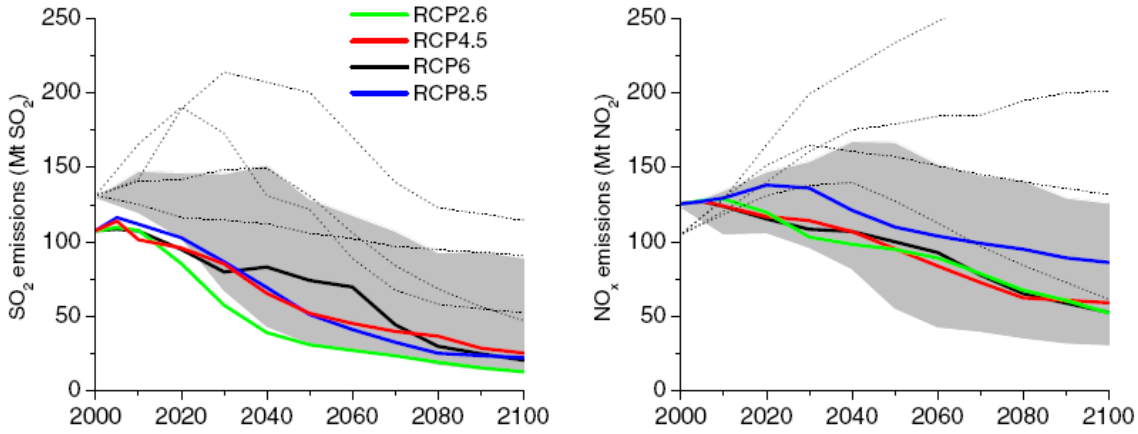


***Figure 10: Emissions of SO_2 and NO_X across the RCPs**. Grey area indicates the 90th percentile of the literature (only scenarios included in Van Vuuren et al. 2008b, i.e. 22 scenarios; the scenarios were also harmonized for their starting year—but using a different inventory). Dotted lines indicate SRES scenarios. The different studies use slightly different data for the start year. (van Vuuren et.al. 2011)*

"The RCPs generally exhibit a declining trend of air polluting emissions. The emission trends for air pollutants are determined by three factors: the change in driving forces (fossil- fuel use, fertilizer use), the assumed air pollution control policy, and the assumed climate policy (as the last induces changes in energy consumption leading to changes (generally reductions) in air polluting emissions). We have illustrated the trends in air pollutants by looking at SO2 and NOx (Fig. 10). In general, similar trends can be seen for other air pollutants.

"All RCPs include the assumption that air pollution control becomes more stringent, over time, as a result of rising income levels. Globally, this would cause emissions to decrease, over time—although trends can be different for specific regions or at particular moments in time. A second factor that influences the results across the RCPs is climate policy. In general, the lowest emissions are found for the scenario with the most stringent climate policy (RCP2.6) and the highest for the scenario without climate policy (RCP8.5), although this does not apply to all regions, at all times". ***(van Vuuren et.al. 2011).***

Radiative Forcing Trends

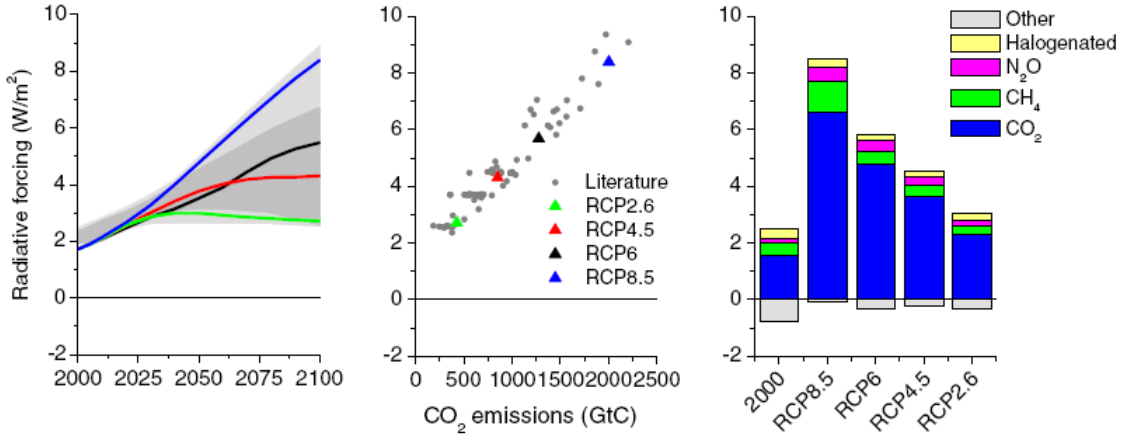


Figure 11: Trends in radiative forcing (left), cumulative 21st centuryCO2 emissions vs 2100 radiative forcing (middle) and 2100 forcing level per category (right). *Grey area indicates the 98th and 90th percentiles (light/dark grey) of the literature. The dots in the middle graph also represent a large number of studies. Forcing is relative to pre-industrial values and does not include land use (albedo), dust, or nitrate aerosol forcing (van Vuuren 2011).*

Population and GDP

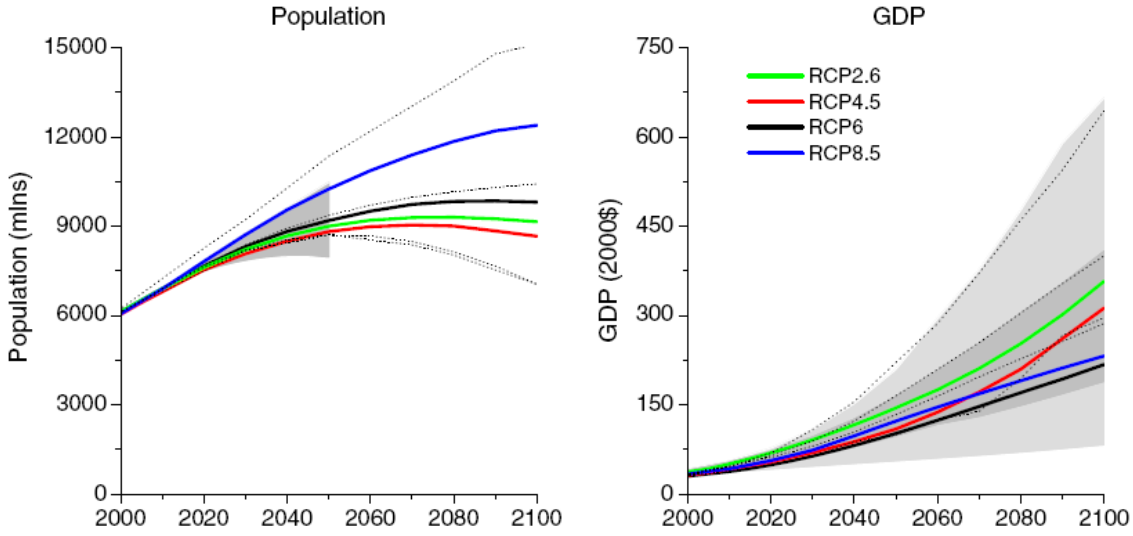


Figure 12: Population and GDP projections of the four scenarios underlying the RCPs (van Vuuren et.al. 2011). *Grey area for population indicates the range of the UN scenarios (low and high) (UN 2003). Grey area for income indicates the 98th and 90th percentiles (light/dark grey) of the IPCC AR4 database (Hanaoka et al. 2006). The dotted lines indicate four of the SRES marker scenarios*

"The population and GDP pathways underlying the four RCPs are shown in Fig. 12. The figure also shows, as reference, the UN population projections and the 90th percentile range of GDP scenarios in the literature on greenhouse gas emission scenarios. Figure 12 shows the RCPs to be consistent with these two references. It should be noted that, with one exception (RCP8.5), the modeling teams deliberately made intermediate assumptions about the main driving forces (as illustrated by their position in Fig. 12)…In contrast, the RCP8.5 was based on a revised version of the SRES A2 scenario; here, the storyline emphasizes high population growth and lower incomes in developing countries". ***(van Vuuren et.al. 2011).***

Energy and oil consumption

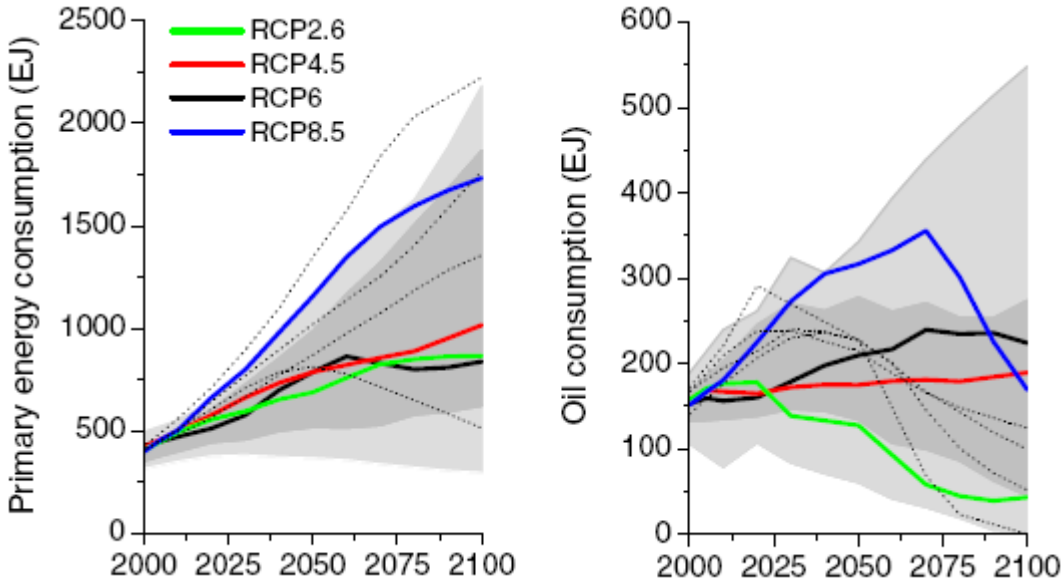


Figure 13: Development of primary energy consumption (direct equivalent) and oil consumption for the different RCPs (van Vuuren et.al. 2011). *The grey area indicates the 98th and 90th percentiles (light/dark grey) (AR4 database (Hanaoka et al. 2006) and more recent literature (Clarke et al. 2010; Edenhofer et al. 2010). The dotted lines indicate four of the SRES marker scenarios*

"For energy use, the scenarios underlying the RCPs are consistent with the literature— with the RCP2.6, RCP4.5 and RCP6 again being representative of intermediate scenarios in the literature (resulting in a primary energy use of 750 to 900 EJ in 2100, or about double the level of today). The RCP8.5, in contrast, is a highly energy-intensive scenario as a result of high population growth and a lower rate of technology development". ***(van Vuuren et.al. 2011).***

Energy sources at years 2000 and 2100

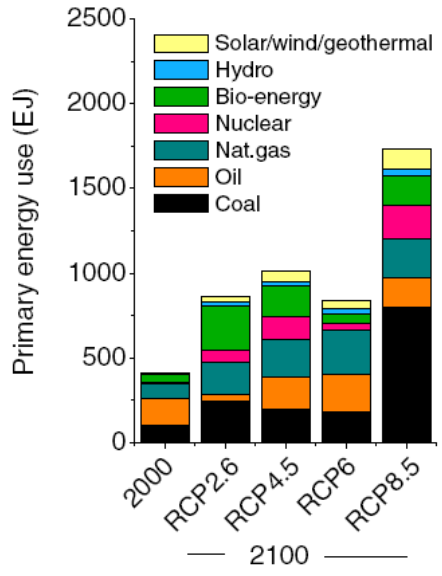


Figure 14: Energy sources by sector (van Vuuren et.al. 2011)

"In terms of the mix of energy carriers, there is a clear distinction across the RCPs given the influence of the climate target. Total fossil- fuel use basically follows the radiative forcing level of the scenarios; however, due to the use of carbon capture and storage (CCS) technologies (in particular in the power sector), all scenarios, by 2100, still use a greater amount of coal and/or natural gas than in the year 2000. The use of oil stays fairly constant in most scenarios, but declines in the RCP2.6 (as a result of depletion and climate policy).

The use of non-fossil fuels increases in all scenarios, especially renewable resources (e.g. wind, solar), bio-energy and nuclear power. The main driving forces are increasing energy demand, rising fossil-fuel prices and climate policy. An important element of the RCP2.6 is the use of bio-energy and CCS, resulting in negative emissions (and allowing some fossil fuel without CCS by the end of the century)". ***(van Vuuren et.al. 2011).***

Land Use

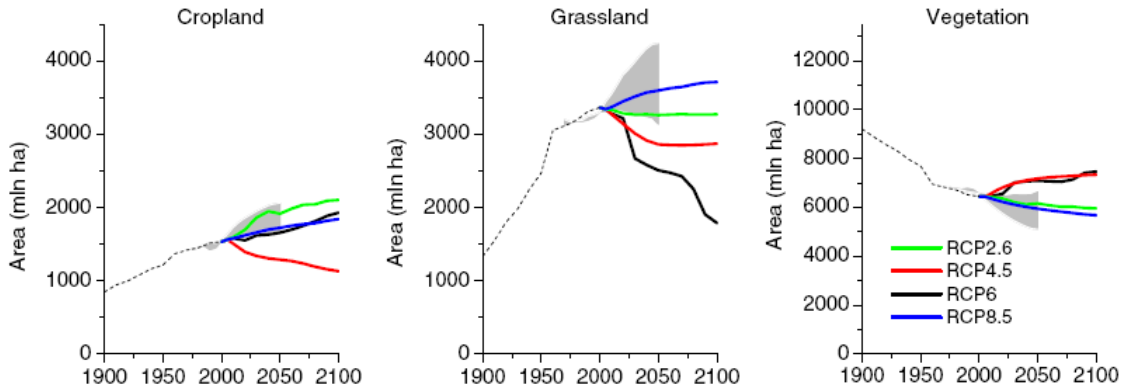


Figure 15: Land use (crop land and use of grass land) across the RCPs. *Grey area indicates the 90th percentile of scenarios reported in the literature (taken from Smith et al. 2010). Vegetation is defined as the part not covered by cropland or anthropogenically used grassland. (van Vuuren et.al. 2011)*

"A crucial element of the new scenarios is land use. Land use influences the climate system in many different ways including direct emissions from land-use change, hydrological impacts, biogeophysical impacts (such as changes in albedo and surface roughness), and the size of the remaining vegetation stock (influencing CO2 removal from the atmosphere). Historically, cropland and anthropogenic use of grassland have both been increasing, driven by rising population and changing dietary patterns. There are far fewer land-use scenarios published in the literature than emission or energy-use scenarios. Moreover, far less experience exists with scenario projections (Rose et al. 2011; Smith et al. 2010). Most projections focus on a shorter time period (up to 2030 or 2050) and show an increasing demand for cropland and pasture.

"The limited experience in global land-use modeling as part of integrated assessment work is also reflected in the RCP development process. Compared to emission modeling, definitions of relevant variables and base year data differ more greatly across the IAMs for the land use components.

"The RCPs cover a very wide-range of land-use scenario projections. This is illustrated by the trends shown in Fig. 15 (i.e. after harmonization). The use of cropland and grasslands increases in RCP8.5, mostly driven by an increasing global population. Cropland also increases in the RCP2.6, but largely as a result of bio-energy production. The use of grassland is more-or-less constant in the RCP2.6, as the increase in production of animal products is met through a shift from extensive to more intensive animal husbandry. The RCP6 shows an increasing use of cropland but a decline in pasture. This decline is caused by a similar trend as noted for RCP2.6, but with a much stronger implementation. Finally, the RCP4.5 shows a clear turning point in global land use based on the assumption that carbon in natural vegetation will be valued as part of global climate policy. As a result of reforestation programs, the use of cropland and grassland decreases, following considerable yield increases and dietary changes". ***(van Vuuren et.al. 2011).***

Comparisons with SRES equivalents

Main similarities and differences between temperature projections for SRES scenarios and RCPs.		
RCP	**SRES scenario with similar median temp increase by 2100**	**Particular differences**
RCP3PD	None	The ratio between temperature increase and net radiative forcing in 2100 is 0:88 C(Wm^{-2})$^{-1}$ for RCP3-PD, whereas all other scenarios show a ratio of about 0:62 C(Wm^{-2})$^{-1}$; that is, RCP3-PD is closer to equilibrium in 2100 than the other scenarios.
RCP4.5	SRES B1	Median temperatures in RCP4.5 rise faster than in SRES B1 until mid-century, and slower afterwards.
RCP6	SRES B2	Median temperatures in RCP6 rise faster than in SRES B2 during the three decades between 2060 and 2090, and slower during other periods of the twenty-first century.
RCP8.5	SRES A1FI	Median temperatures in RCP8.5 rise slower than in SRES A1FI during the period between 2035 and 2080, and faster during other periods of the twenty-first century.

Table 5: From Rogelj et.al. 2012

Extended Concentration Pathway Emissions and Forcing

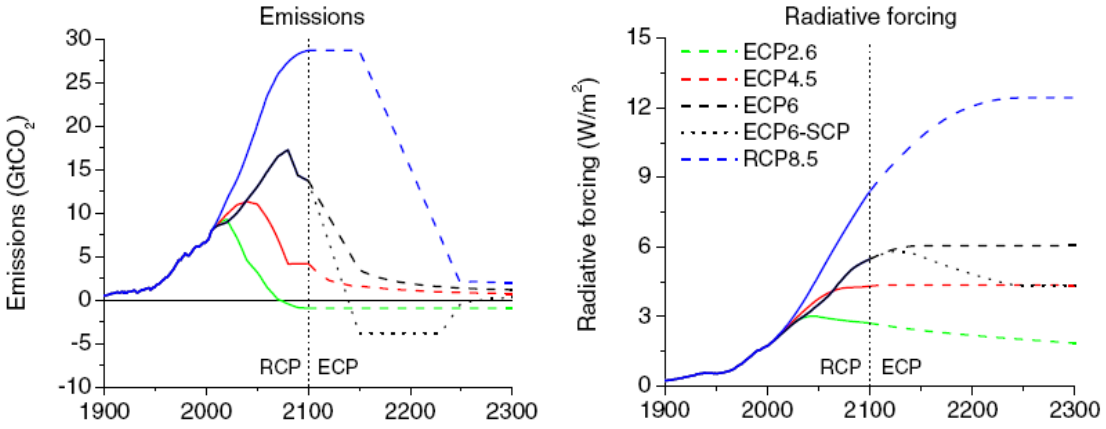


Figure 16: Extension of the RCPs (radiative forcing and associated CO2 emissions) from van Vuuren 2011. *(ECP is Extended Concentration Pathway). The SCP6to4.5 (supplementary concentration pathway) shows an alternative extension for RCP6…(Meinshausen et al. 2011b)*

Figure 16 shows the CO2 emissions and radiative forcing trajectories for each of the four extensions of the RCPs (ECPs). These extensions have not been based on integrated assessment modeling, but on simple extension rules consistent with the rationale of each of the RCPs to which they connect. This has resulted in a set of extended concentration pathways to be used for climate model runs.

Further Reading

Summary Publications

Towards New Scenarios for Analysis of Emissions, Climate Change, Impacts, and Response Strategies, IPCC Expert Meeting Report, 19–21 September, 2007

The representative concentration pathways: an overview, van Vuuren et. al 2011, Climatic Change (2011) 109:5–31, DOI 10.1007/s10584-011-0148-z

The next generation of scenarios for climate change research and assessment, Moss et.al. 2010, Nature, doi:10.1038/nature08823

Global warming under old and new scenarios using IPCC climate sensitivity range estimates, Rogelj, Meinshausen and Knutti 2012, Nature Climate Change, DOI: 10.1038/NCLIMATE1385

IPCC Data Distribution Centre on-line documentation of the scenario process, including documentation of narrative development and usage: http://sedac.ipcc-data.org/ddc/ar5_scenario_process/index.html

RCP8.5 specific papers (MESSAGE):

Scenarios of long-term socio-economic and environmental development under climate stabilization. Riahi K, Grübler A, Nakicenovic N (2007) Technol Forecast Soc Chang 74:887–935

RCP-8.5: exploring the consequence of high emission trajectories. Riahi K, Krey V, Rao S, Chirkov V, Fischer G, Kolp P, Kindermann G, Nakicenovic N, Rafai P (2011) Climatic Change. doi: 10.1007/s10584-011-0149-y

RCP 6 specific papers (AIM):

Multigas mitigation analysis on stabilization scenarios using aim global model. Fujino J, Nair R, Kainuma M, Masui T, Matsuoka Y (2006) The Energy Journal Special issue #3:343–354

A emission pathway to stabilize at 6 W/m2 of radiative forcing. Masui T, Matsumoto K, Hijioka Y, Kinoshita T, Nozawa T, Ishiwatari S, Kato E, Shukla PR, Yamagata Y, Kainuma M (2011) Climatic Change. doi: 10.1007/s10584-011-0150-5

Global GHG emission scenarios under GHG concentration stabilization targets. Hijioka Y, Matsuoka Y, Nishimoto H, Masui T, Kainuma M (2008) J Glob Environ Eng 13:97–108

RCP 4.5 specific papers (GCAM):

RCP4.5: a pathway for stabilization of radiative forcing by 2100. Thomson AM, Calvin KV, Smith SJ, Kyle GP, Volke A, Patel P, Delgado-Arias S, Bond-Lamberty B, Wise MA, Clarke LE et al (2011) Climatic Change. doi: 10.1007/s10584-011-0151-4

Scenarios of greenhouse gas emissions and atmospheric concentrations. Clarke LE, Edmonds JA, Jacoby HD, Pitcher H, Reilly JM, Richels R (2007) Sub-report 2.1a of Synthesis and Assessment Product 2.1. Climate Change Science Program and the Subcommittee on Global Change Research, Washington DC

MultiGas forcing stabilization with minicam. Smith SJ, Wigley TML (2006) The Energy Journal Special issue #3:373–392

Implications of limiting CO2 concentrations for land use and energy. Wise M, Calvin K, Thomson A, Clarke L, Bond-Lamberty B, Sands R, Smith SJ, Janetos A, Edmonds J, (2009) Science 324:1183-1186

RCP 2.6 (RCP3-PD) specific papers (IMAGE):

RCP2.6: Exploring the possibility to keep global mean temperature change below 2°C. Van Vuuren DP, Stehfest E, Den Elzen MGJ, Deetman S, Hof A, Isaac M, Klein Goldewijk K, Kram T, Mendoza Beltran A, Oostenrijk R et al (2011a) Climatic Change. doi: 10.1007/s10584-011-0152-3

Long-term multi-gas scenarios to stabilise radiative forcing—exploring costs and benefits within an integrated assessment framework. Van Vuuren DP, Eickhout B, Lucas PL, den Elzen MGJ (2006) Energ J 27:201–233

Stabilizing greenhouse gas concentrations at low levels: an assessment of reduction strategies and costs. Van Vuuren DP, Den Elzen MGJ, Lucas PL, Eickhout B, Strengers BJ, Van Ruijven B, Wonink S, Van Houdt R (2007a) Clim Chang 81:119–159

RCP Development papers:

Evolution of anthropogenic and biomass burning emissions at global and regional scales during the 1980–2010 period. Granier C, Bessagnet B, Bond T, D'Angiola A, van der Gon HG, Frost G, Heil A, Kainuma M, Kaiser J, Kinne S et al (2011) Climatic Change. doi: 10.1007/s10584-011-0154-1. (An assessment of a range of historical emissions data as context for the harmonized emissions used as a common starting point for the future trajectories of the RCPs.)

Global and regional evolution of short-lived radiatively-active gases and aerosols in the Representative Concentration Pathways. Lamarque JF, Page Kyle G, Meinshausen M, Riahi K, Smith S, van Vuuren DP, Conley AJ, Vitt F (2011) Climatic change. doi: 10.1007/s10584-011-0155-0. (Describes how information on air pollutant emissions was used for deriving future concentration and forcing scenarios for aerosols and ozone).

Land use Change and earth system dynamics. Hurtt G, Chini L, Frolking S, Betts R, Edmonds J, Feddema J, Fisher G, Goldewijk KK, Hibbard KA, Houghton R et al (2011) Climatic Change. doi: 10.1007/s10584-011-0153-2. (The paper by Hurtt et al. (2011) is the first to harmonize land-use history data with future scenario data from multiple IAMs to form a single consistent, spatially gridded, set of scenarios on land-use change, to study human impacts on the past, present, and future Earth system).

The RCP greenhouse gas concentrations and their extensions from 1765 to 2300. Meinshausen M, Smith SJ, Calvin K, Daniel JS, Kainuma MLT, Lamarque J-F, Matsumoto K, Montzka SA, Raper S, Riahi K et al (2011b) Climatic Change. doi: 10.1007/s10584-011-0156-z . (Describes how the IAM's emission projections of long-lived greenhouse gases were harmonized and used for calculating concentration trajectories for these gases, including the Extended Concentration Pathways (ECPs) from 2100 to 2300.

Acknowledgements

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This guide quotes extensively from **The representative concentration pathways: an overview**, van Vuuren et. al. 2011, a special report first published by Climatic Change, and made available through Springerlink.com as an Open Access document, distributed under the terms of the Creative Commons Attribution Non-commercial License which permits any non-commercial use, distribution, and reproduction in any medium, provided the original author(s) and source are credited.

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PDF Cover artwork by John Cook.

RCP4.5: A Pathway for Stabilization of Radiative Forcing by 2100

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ABSTRACT

Representative Concentration Pathway (RCP) 4.5 is a scenario that stabilizes radiative forcing at 4.5 Watts per meter squared in the year 2100 without ever exceeding that value. Simulated with the Global Change Assessment Model (GCAM), RCP4.5 includes long-term, global emissions of greenhouse gases, short-lived species, and land-use-land-cover in a global economic framework. RCP4.5 was updated from earlier GCAM scenarios to incorporate historical emissions and land cover information common to the RCP process and follows a cost-minimizing pathway to reach the target radiative forcing. The imperative to limit emissions in order to reach this target drives changes in the energy system, including shifts to electricity, to lower emissions energy technologies and to the deployment of carbon capture and geologic storage technology. In addition, the RCP4.5 emissions price also applies to land use emissions; as a result, forest lands expand from their present day extent. The simulated future emissions and land use were downscaled from the regional simulation to a grid to facilitate transfer to climate models. While there are many alternative pathways to achieve a radiative forcing level of 4.5 W m^{-2}, the application of the RCP4.5 provides a common platform for climate models to explore the climate system response to stabilizing the anthropogenic components of radiative forcing.

Keywords: climate change mitigation; Integrated Assessment; Global Change Assessment Model; Representative Concentration Pathway; scenario.

1. INTRODUCTION

Representative Concentration Pathway (RCP) 4.5 is a scenario of long-term, global emissions of greenhouse gases, short-lived species, and land-use-land-cover which stabilizes radiative forcing at 4.5 Watts per meter squared (W m^{-2}, approximately 650 ppm CO_2-equivalent) in the year 2100 without ever exceeding that value. The defining characteristics of this scenario are enumerated in Moss et al. (2008, 2010). RCP 4.5 is based on the MiniCAM Level 3 stabilization scenario reported in Clarke, et al. (2007) with additional detail on the non-CO_2 and pollution control assumptions in Smith and Wigley (2006), and incorporating updated land use modeling and terrestrial carbon emissions pricing assumptions as reported in Wise et al (2009a,b).

Unlike the scenarios developed by the IPCC and reported in Nakicenovic et al. (2000), which examined possible global futures and associated greenhouse-related emissions in the absence of measures designed to limit anthropogenic climate change, RCP 4.5 is a stabilization scenario and assumes that climate policies, in this instance the introduction of a set of global greenhouse gas emissions prices, are invoked to achieve the goal of limiting emissions, concentrations and radiative forcing.

While RCP 4.5 is based on the "MiniCAM Level 3" scenario reported in Clarke et al. (2007), RCP 4.5 differs in several important regards. First, the Clarke et al. (2007) scenario considered a slightly different definition of radiative forcing than RCP 4.5. In Clarke et al. (2007), radiative forcing is defined in terms of a suite of six greenhouse gases, carbon dioxide (CO_2), methane (CH_4), nitrous oxide (N_2O), hydrofluorocarbons (HFCs), perfluorocarbons (PFCs) and sulfur hexafluoride (SF_6). RCP 4.5 considered the influences of a broader set of anthropogenic emissions including CO_2, CH_4, N_2O, HFCs, PFCs, and SF_6, but also chemically active gases such as carbon monoxide (CO) and volatile organic compounds (VOCs). Importantly, RCP 4.5 considers the influence of sulfur aerosols, as well as black and organic carbon. The Clarke et al. (2007) scenario stabilized radiative forcing at approximately 4.7 W m^{-2} while RCP stabilizes radiative forcing at 4.5 W m^{-2}. For CO_2, the most

important anthropogenically released greenhouse gas, year 2100 concentrations are somewhat higher in the Clarke et al. (2007) scenario, approximately 550 ppm CO_2, than in RCP 4.5, approximately 525 ppm CO_2. RCP 4.5 also employed updated historical data series, calibration to the year 2000 consensus emissions inventories from Lamarque et al. (2010), and a new representation of residue biomass supply (Gregg and Smith 2010). More importantly, RCP 4.5 employed a more sophisticated land-use and land-cover model (Wise et al. 2009a,b) than was available for use in Clarke et al. (2007). Perhaps the most important difference between Clarke et al. (2007) and RCP 4.5 is the downscaling of emissions and land-use-land-cover from the 14 geopolitical GCAM regions to a 0.5 degree grid for the RCPs in order to enable use of the scenario in global climate models. Finally, RCP 4.5 takes advantage of one technology that was not modeled in Clarke et al., namely combining bioenergy production with CO_2 capture and geologic storage (CCS). This technology combination is capable of producing final energy such as electricity with net-negative carbon emissions.

Because the RCPs are based on scenarios documented in the open literature, each reflects a different set of underlying socioeconomic assumptions. RCP 4.5 is a stabilization scenario and thus assumes the imposition of emissions mitigation policies. RCP 4.5 is derived from its own "reference", or "no-climate-policy", scenario. This reference scenario is unique to RCP 4.5 and differs from RCP 8.5 as well as from the reference scenarios associated with RCP 6.0 and RCP 2.6.

In the remainder of this paper we will discuss the modeling environment employed to develop RCP 4.5, the Global Change Assessment Model (GCAM), which was used to transform the original MiniCAM Level 3 scenario (Clarke et al., 2007) into RCP 4.5. We will then proceed to describe the underlying socioeconomic assumptions that shape RCP 4.5 and its associated reference scenario and discuss the characteristics of RCP 4.5, highlighting the global energy, economic, land use, and land cover systems, as well as the mechanisms employed to limit radiative forcing to 4.5 W m^{-2} and contrast RCP 4.5 to its reference scenario.

2. METHODS

2.1 The Global Change Assessment Model

The GCAM is a global integrated assessment model and a direct descendent of the MiniCAM model (Kim et al., 2006; Clarke et al., 2007; Brenkert et al., 2003). It combines representations of the global economy, energy systems, agriculture and land use, with representation of terrestrial and ocean carbon cycles, a suite of coupled gas-cycle, climate, and ice-melt models. GCAM tracks emissions and concentrations of greenhouse gases and short-lived species including CO_2, CH_4, N_2O, NO_x, VOCs, CO, SO_2, carbonaceous aerosols, HFCs, PFCs, and SF_6.

GCAM is a dynamic recursive economic model driven by assumptions about population size, and labor productivity that determine potential gross domestic product in each of 14 regions at 15 year time steps. GCAM establishes market-clearing prices for all energy, agriculture and land markets such that supplies and demands for all markets balance simultaneously. The GCAM energy system includes primary energy resources, production, energy transformation to final fuels, and the employment of final energy forms to deliver energy services such as passenger kilometers in transport or space conditioning for buildings. GCAM contains detailed representations of technology options in all of the economic components of the system with technology choice determined by market competition.

The agriculture and land use component is fully integrated with the GCAM economic and energy system components. Land is allocated between alternative uses based on expected profitability, which in turn depends on the productivity of the land-based product (e.g. mass of harvestable product per ha), product price, and non-land costs of production (labor, fertilizer, etc.). The productivity of land-based products is subject to change over time based on future estimates of crop productivity change. We adopt assumptions for the next 30 years based on those employed by the UN Food

and Agricultural Organization (Bruinsma, 2003). A full description of the agriculture and land use modeling in GCAM as used for RCP4.5 can be found in Wise et al. (2009a).

The GCAM physical atmosphere and climate are represented by the Model for the Assessment of Greenhouse-Gas Induced Climate Change (MAGICC; Wigley and Raper 1992, 2002; Raper et al., 1996). To construct the RCP4.5 scenario, we use MAGICC version 5.3, which is initialized to the IPCC 4th Assessment Report.[1] The definition of total radiative forcing for the RCP4.5 does not include albedo, nitrate, and mineral dust. These three forcing agents have a fixed future forcing of -0.4 W m^{-2} in MAGICC version 5.3.

The RCP4.5 stabilization scenario is a cost-minimizing pathway. It assumes that all nations of the world undertake emissions mitigation simultaneously and effectively, and share a common global price that all emissions to the atmosphere must pay with emissions of different gases priced according to their hundred-year global warming potentials (Forster et al., 2007). All sectors of the economy are covered, including agriculture and land use emissions. That emissions price also rises over time so as to minimize the present discounted cost of emissions mitigation. The policy also assumes that deployment mechanisms and measurement and monitoring of both fossil fuel and terrestrial carbon are not barriers to implementation of emissions mitigation.

This cost-minimizing price path has two components. Prior to reaching the target, 4.5 W m^{-2}, cost minimization requires that the greenhouse gas emissions price rise at the interest rate, adjusted by the rate of ocean uptake (Edmonds et al., 2008; Clarke et al., 2007; Hotelling, 1931; Peck and Wan, 1996). An emissions price path with this property precludes all opportunities for arbitrage because the marginal cost of abatement is constant across time. The second component of the pathway occurs after the target is reached. At this time, the emissions price is adjusted to ensure that

the radiative forcing level remains at its target. For the RCP4.5, stabilization occurs in 2080; prior to 2080 the emissions price rises at 5% per year, and after 2080 the emissions price is roughly constant.

2.2 Emissions Downscaling

Emissions from GCAM were downscaled using a two-step method, first downscaling to the country level and then mapping to a spatial grid within each country. Detailed emissions from GCAM were aggregated into the 12 RCP reporting sectors for the 14 GCAM regions at each 15-year model time period. Emissions from each sector were downscaled to a country level for 231 countries using the methodology outlined in Van Vuuren et al. (2007). Input data for this step includes the country-level population projection used for the RCP4.5 scenario, a gridded base-year GDP data set from van Vuuren et al. (2007), who combined World Bank GDP information with the GPW gridded population data set (CIESIN & CIAT 2005), and year 2000 gridded emissions data from Lamarque et al. (2010). Gridded GDP and emissions data were used so that base-year GDP and emissions differences could consistently be estimated for any set of countries or regions. A convergence year of 2200 for GDP and emissions intensity calculations (slightly larger than value used by Van Vuuren et al. 2007) was used for all anthropogenic emissions sectors. The pattern of forest and grassland emissions within each region was held constant by setting a high convergence year of 10,000.

The downscaled sectoral emissions for each country were mapped to a 0.5 degree grid using the base-year 2000 gridded emissions data from Lamarque et al. (2010). The relative emissions distribution within each country was held constant over time for each emissions sector. All downscaling calculations were performed with each 0.5 degree grid cell subdivided into 231 countries using 2.5 minute GPW country boundary data (CIESIN & CIAT 2005). After downscaling on a sub-divided grid for each country and

[1] Note that the final concentration pathway values for the RCP4.5 were produced in MAGICC version 6 (see Meinhausen et al., this issue).

emissions sector, emissions were summed to a 0.5 degree resolution for the final data product.

Emissions from international shipping and aircraft do not occur within country boundaries and instead were aggregated to one global emissions figure. Gridded shipping emissions were globally scaled from the RCP consensus year 2000 emissions grid (Lamarque et al. 2010). Aircraft emissions used a time changing pattern from the QUANTIFY B2i emissions scenario (Lee et al. 2010), as the overall pathway for this scenario closely matched the GCAM model output. The QUANTIFY three dimensional emissions pattern was collapsed to two dimensions (latitude and longitude) for 2000, 2025, 2050, and 2100. This pattern was interpolated to decadal intervals and scaled globally to match the GCAM global aviation emissions values.

2.3 Land Use Downscaling

The RCP scenario process is the first to explicitly provide land use projections in addition to future emissions pathways for input to global climate models. Because all four participating integrated assessment models, and all receiving climate models, use different characterizations and definitions of land use types and transitions, a harmonization step was necessary. The harmonization was designed to provide a continuous, consistent set of land use inputs for climate models from 1500 through 2100 with a smooth transition between historical data (1500-2005) and future projections (2005-2100) (see Hurtt et al., this volume).

In the GCAM model results, land use is simulated at the 14 region level and land use changes and transitions are not spatially attributed. In this case, the land use was first downscaled to the 0.5 degree harmonization grid, following the algorithms of the the global land-use model (GLM) (Hurtt et al. 2006), preserving GCAM regional land use area totals and generating smooth spatial patterns in the transition from historical to future states. These downscaling algorithms were developed and implemented by the land use harmonization

team at the University of New Hampshire (Hurtt et al., this volume) and are fully described in Thomson et al. (2010).

3. RESULTS

3.1 GCAM Reference Scenario

Each of the RCPs was produced by a different integrated assessment model; therefore, each has its own reference scenario. Thus, the reference scenario for RCP 4.5 is not RCP8.5 but rather a GCAM reference scenario. The GCAM reference scenario (Clarke et al., 2007) depicts a world in which global population reaches a maximum of more than 9 billion in 2065 and then declines to 8.7 billion in 2100 while global GDP grows by an order of magnitude and global energy consumption triples. The reference scenario includes no explicit policies to limit carbon emissions, and therefore fossil fuels continue to dominate global energy consumption, despite substantial growth in nuclear and renewable energy. Atmospheric CO_2 concentrations rise throughout the century and reach 792 ppmv by 2100, with total radiative forcing approaching 7 W m^{-2} (Fig. 1). Forest land declines in the reference scenario to accommodate increases in land use for food and bioenergy crops. Even with the assumed agricultural productivity increases, crop land increases in the first half of the century due to increases in population and income, which drives an increase in land-intensive meat consumption. After 2050 the rate of growth in food demand slows, in part due to declining population. As a result the amount of cropland and also land use change (LUC) emissions decline as agricultural crop productivity continues to increase.

3.2 4.5 Stabilization Scenario

The RCP4.5 scenario is based on the same population and income drivers as the GCAM reference scenario but applies greenhouse gas emissions valuation policies to stabilize atmospheric radiative forcing at 4.5 W m^{-2} in 2100 (Fig. 1). This imperative to stabilize climate change drives anthropogenic CO_2 emissions downward throughout the next century (Fig. 2). RCP4.5 results in

an atmospheric CO_2 concentration of 526 ppm in 2100[2], compared to 792 ppm in the GCAM reference case. This stabilization is achieved in 2080. At this point in time, total radiative forcing reaches 4.5 W m^{-2} and the emissions price becomes roughly constant. CO_2 emissions also become roughly constant. RCP4.5 depicts declines in overall energy use, as well as declines in fossil fuel use compared to the reference case, while substantial increases in renewable energy forms and nuclear energy both occur (Fig. 3). The proportion of total final energy that is supplied by electricity also increases due to fuel switching in the end-use sectors. The emergence of large-scale carbon dioxide capture and storage (CCS) (Fig. 4) allows continued use of fossil fuels for electricity generation and cement manufacture, among other uses, though total use is lower than in the reference scenario. Bioenergy with CCS is used to produce electricity, providing an energy source that is carbon-negative with respect to the atmosphere. The amount of bioenergy deployed is limited by the availability of dedicated crop and crop residue feedstocks from the land system.

One important feature influencing the availability of bioenergy feedstocks in the RCP4.5 is the expansion of forests as part of the larger emissions mitigation strategy. The extent of afforestation follows Wise et al., 2009b. This idealized case assumes that all carbon from fossil fuel and land use emissions are charged an equal penalty price, and thus reductions in LUC emissions constitute an available strategy for global emissions mitigation. The GCAM therefore simulates the preservation of large stocks of terrestrial carbon in forests, with some crop and pasture lands converted to bioenergy crops (Fig. 5). Under this policy environment, bioenergy crops still provide an important source of fuel, >50 EJ/yr, to meet global energy demand in 2100. This is accomplished while still providing for the world's dietary need by shifting toward food products with a smaller carbon footprint.

[2] 526 ppmv is the CO_2 concentration from GCAM before harmonization in MAGICC6 (see Meinhausen et al., this issue. The harmonized RCP4.5 CO_2 concentration in 2100 is 538 ppmv.

Carbon prices reach $85 per ton of CO_2 by 2100 (Fig. 6) which transforms the global economy. Electric power generation changes from the largest source of emissions in the world to a system with net negative emissions—made possible by increased reliance on nuclear and renewable energy forms such as wind, solar and geothermal, and the application of CO_2 capture and storage technology to both fossil fuel sources and bioenergy (Figs. 3 and 4). Buildings and industry largely de-carbonize by employing more efficient end-use technologies and by electrifying. Annual land-use change emissions are reduced to 0.13 $GtCO_2$/yr (Fig. 2). Total anthropogenic CO_2 emissions for the RCP4.5 peak around 42 Gt CO_2 per year (Fig. 2) around 2040 and decline to 2080 before leveling off around 15 Gt CO_2 per year for the remainder of the century. Other greenhouse gases respond to both the mitigation price signals in the GCAM as well as to assumed pollution controls. Downscaling of these emissions leads to additional insights about future non-CO_2 emissions that are important considerations in the global climate and atmospheric chemistry models. For example, aggregate emissions and the associated radiative forcing contribution of CH_4 from all sectors are relatively constant at the global level (Fig. 7); however, when downscaled they exhibit geographical shifts (Fig. 8). CH_4 emissions in South America and Africa increase over the century while those from China, India the US and Western Europe decline.

CO_2 constitutes the largest contribution to total radiative forcing in the RCP4.5, followed by CH_4, halocarbons, tropospheric ozone, and N_2O (Fig. 7). The relative proportion of the non-CO_2 components of positive radiative forcing remains constant over time. Sulfate forcing is net negative throughout the century but this influence declines over time largely due to assumed increases in pollution control with income, although there are also indirect sulfur dioxide emissions reductions due to the greenhouse gas mitigation policy (Smith et al. 2005).

3.3 GCAM-Simulation of the Four Pathways

In order to facilitate model intercomparisons and further explore the characteristics of the RCPs, the participating models simulated their assigned RCP as well as the other three defined radiative forcing levels. The GCAM was used to simulate a 2.6 W m^{-2} peak-and-decline scenario for use in the evaluation of low radiative forcing targets during the planning stages of the RCPs (Weyant et al, 2009) and is fully documented in Calvin et al (2009). The GCAM6.0 was simulated as a stabilization following the same methods as the GCAM4.5 but resulting in lower carbon prices and a longer time to stabilization. All three of the mitigation cases with GCAM (2.6, 4.5 and 6.0 W m^{-2}) used the same technology, population and economic assumptions described earlier in this paper and in Clarke et al. (2007). The GCAM reference case with these assumptions and no climate mitigation policy reaches around 7.0 W m^{-2} radiative forcing in 2100. Thus, to reach the RCP level of 8.5 W m^2 required altering some underlying assumptions. Several modifications of underlying assumptions were tested; the case selected for the GCAM8.5 follows the same population and economic drivers as the other GCAM scenarios, but assumes no technological improvement in energy technologies or agricultural productivity. In other words, the GCAM8.5 is the GCAM reference case if all technological development is frozen after 2005. This is a hypothetical experiment in exploring high levels of radiative forcing which also allows exploration of the role of technological development in scenarios.

The CO_2 emissions from the four pathways simulated with GCAM are illustrated in Fig. 9 along with the four official released RCPs. The GCAM2.6 results in even lower emissions of CO_2 than the RCP2.6. Total radiative forcing still declines to 2.6 W m^{-2} as a result of higher CH_4 and N_2O emissions in GCAM than in the IMAGE for this scenario. Conversely, CO_2 emissions for the GCAM8.5 are higher than the RCP8.5 due, but the same radiative forcing is reached due to lower CH_4 and N_2O emissions in the GCAM than in the MESSAGE model for this scenario.

Differences in emissions by gas between the official RCPs and the GCAM replication of the RCP forcing levels can be attributed to any number of forces. First, GCAM employs different population and GDP assumptions than the other three models. GCAM has the smallest population and the second highest GDP of the four models (see van Vuuren et al., this issue). Second, the four models have different assumptions about technological change and resource availability. Third, the GCAM model uses a terrestrial carbon policy that has a significant impact on land use and land-use change emissions. While each of the models consider abatement opportunities in the terrestrial system, the method of attaining these opportunities differs across the four RCP models. The differences listed here are only a subset of differences between the four models. However, they illustrate an important point; namely, there are numerous ways that a given radiative forcing goal can be achieved. The RCP4.5 is only one pathway to stabilize radiative forcing at 4.5 W/m^2.

4. CONCLUSIONS

The RCP4.5 scenario is intended to inform research on the atmospheric consequences of reducing greenhouse gas emissions in order to stabilize radiative forcing in 2100. It is also a mitigation scenario – the transformations in the energy system, land use, and the global economy required to achieve this target are not possible without explicit action to mitigate greenhouse gas emissions. However, there are many possible pathways in GCAM and other integrated assessment models that would also achieve a radiative forcing level of 4.5 Wm^{-2}. For example, simulations with GCAM can reach 4.5 Wm^{-2} even if some technology options, such as CCS or nuclear power, are removed from consideration or even if not all countries enter into an emissions mitigation agreement at the same time (Clarke et al., 2009). Such alternate scenarios have different characteristics – higher emissions prices and different energy system transformations, for example – than the RCP4.5. The pathway discussed here and released as RCP4.5 is cost-minimizing, and therefore invokes all available technology options that can cost-effectively contribute to mitigation.

RCP4.5 aims to achieve stable radiative forcing in 2100; however, this does not imply that greenhouse gas emissions, greenhouse gas concentrations, or the climate system are stable. Radiative forcing is stable from 2080-2100 in the RCP4.5, but emissions and concentrations of greenhouse gases continue to vary in the underlying scenario. As a separate exercise, the greenhouse gas emissions and concentrations of RCP4.5 were extended to 2300 using MAGICC 6 (Meinhausen et al., this volume). This ECP4.5 is also defined as a stabilization path, and therefore radiative forcing is held constant at 4.5 W m^{-2} from 2100-2300. It is important to note that ECP4.5 does not imply an extension of the socioeconomic, energy or land use assumptions underlying RCP4.5 in GCAM. The application of RCP4.5 and ECP4.5 in climate models provides a platform to explore the climate system response to stabilizing the anthropogenic components of radiative forcing.

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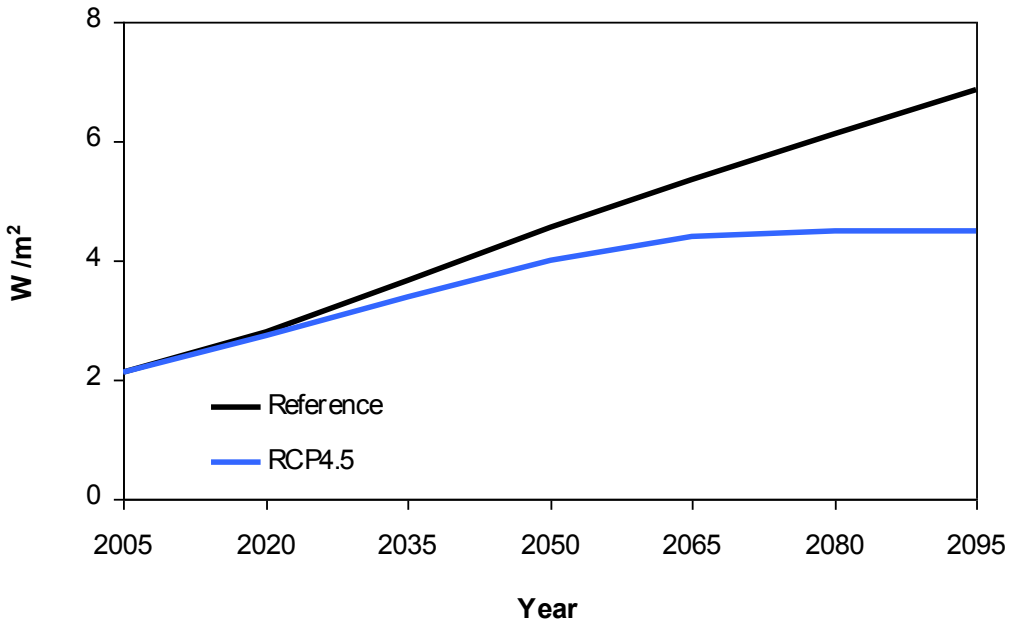


Figure 1: Radiative forcing (W m^{-2}) of the GCAM reference and RCP4.5 scenarios over the model simulation period

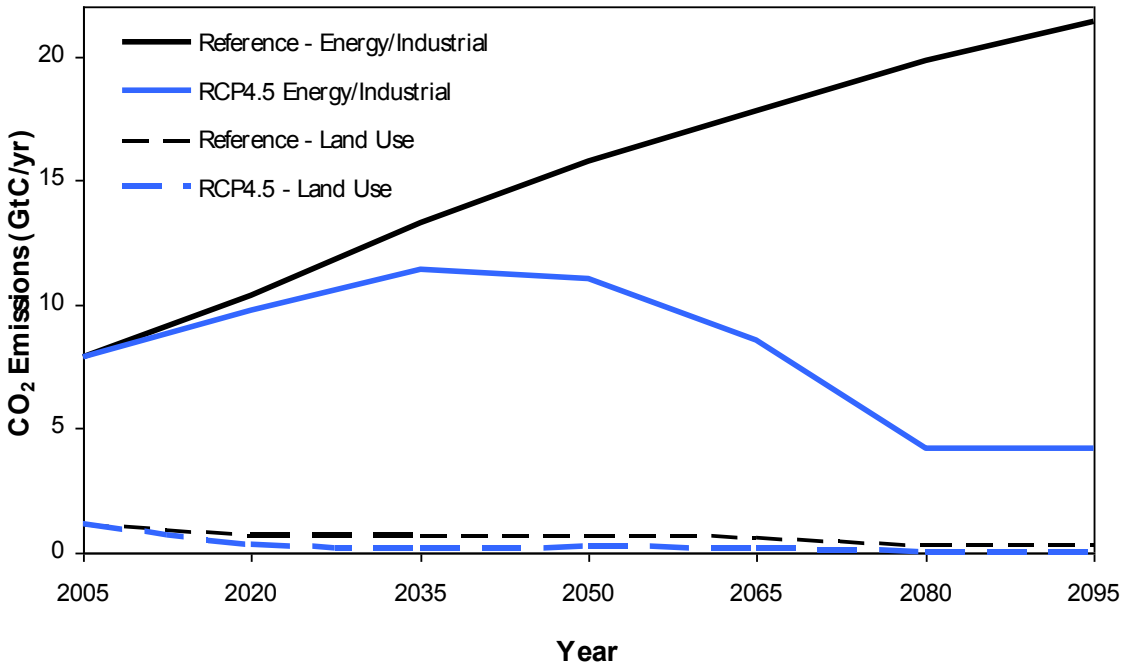


Figure 2: CO_2 emissions from energy and industrial sources and from land use/ land use change in the GCAM reference and RCP4.5 scenarios

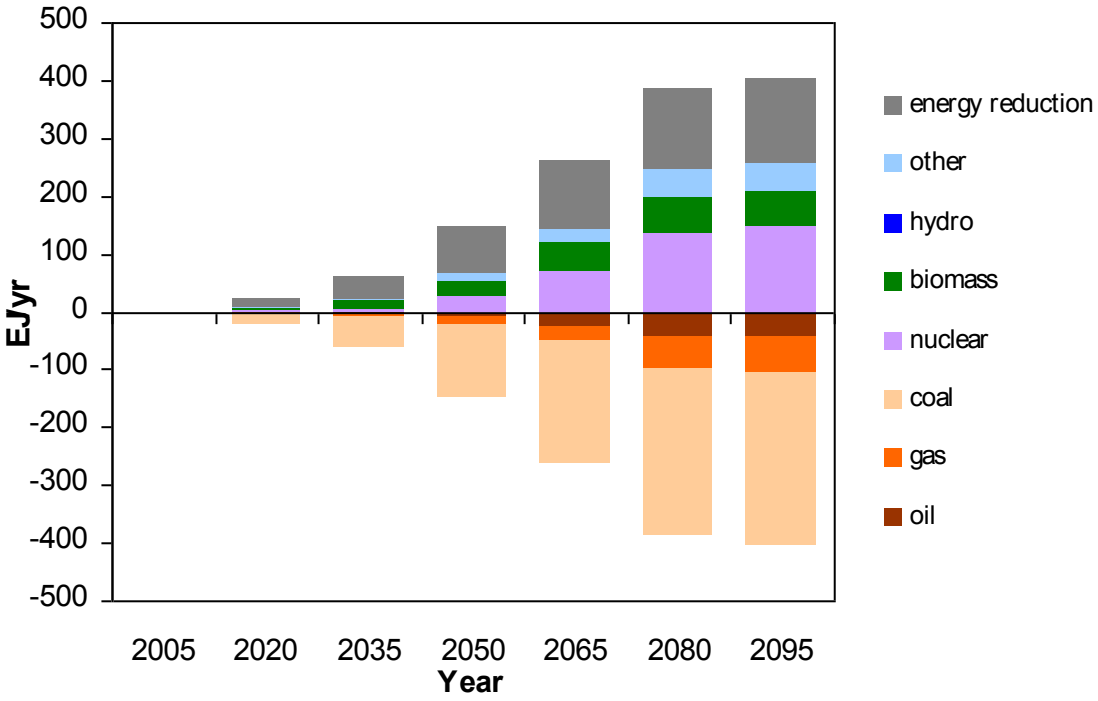


Figure 3: Primary energy sources in RCP4.5 expressed as difference from the reference case. Categories above the x-axis indicate a greater amount of primary energy from that source in the RCP4.5 whereas values below indicate less energy from that primary source in the RCP4.5.

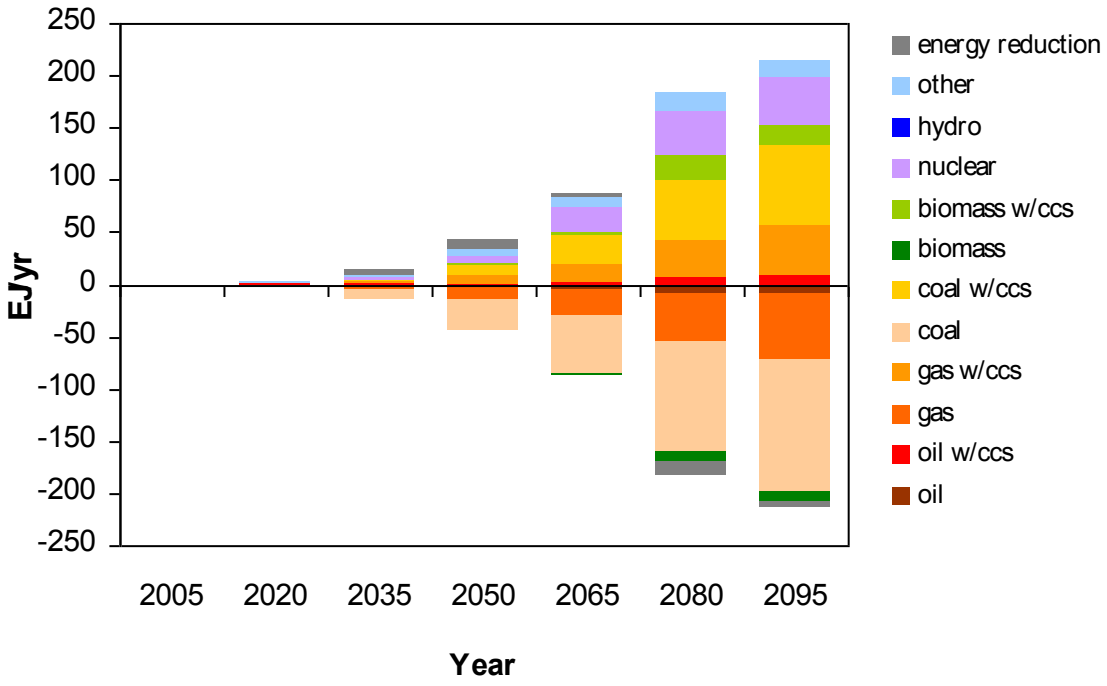


Figure 4: Electricity generation technology in RCP4.5 expressed as difference from the reference case. Categories above the x-axis indicate a greater amount of electricity generation from that technology in the RCP4.5 whereas values below indicate less electricity generation from that technology in the RCP4.5.

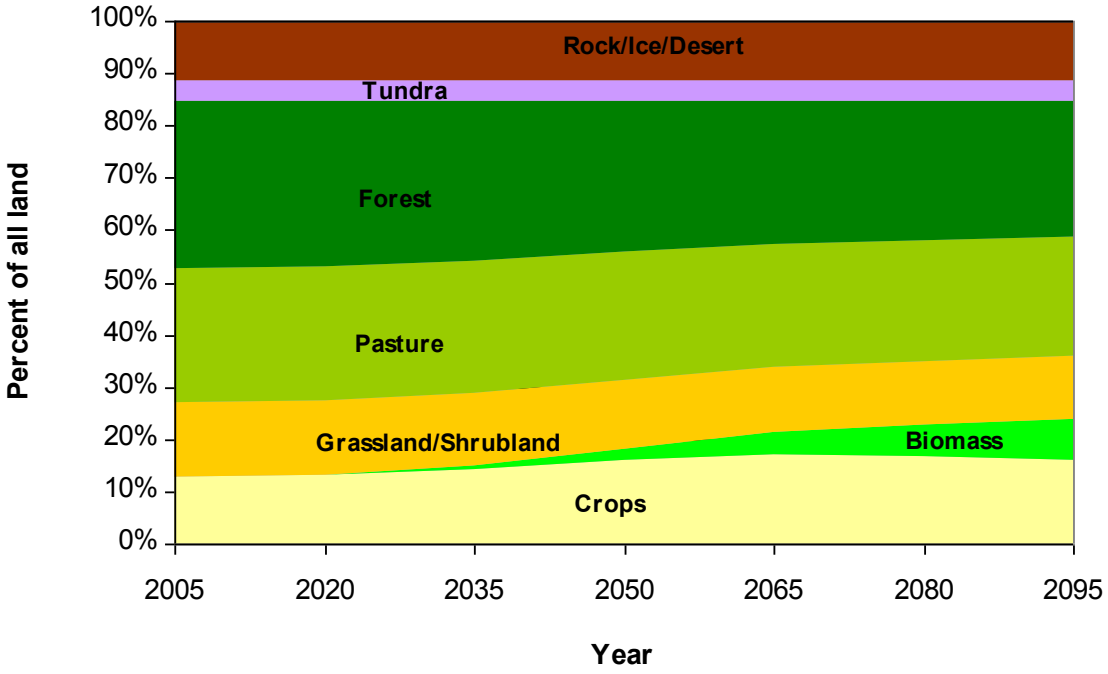


Figure 5: Global land cover over time in the RCP4.5 scenario expressed as a percentage of total global land area.

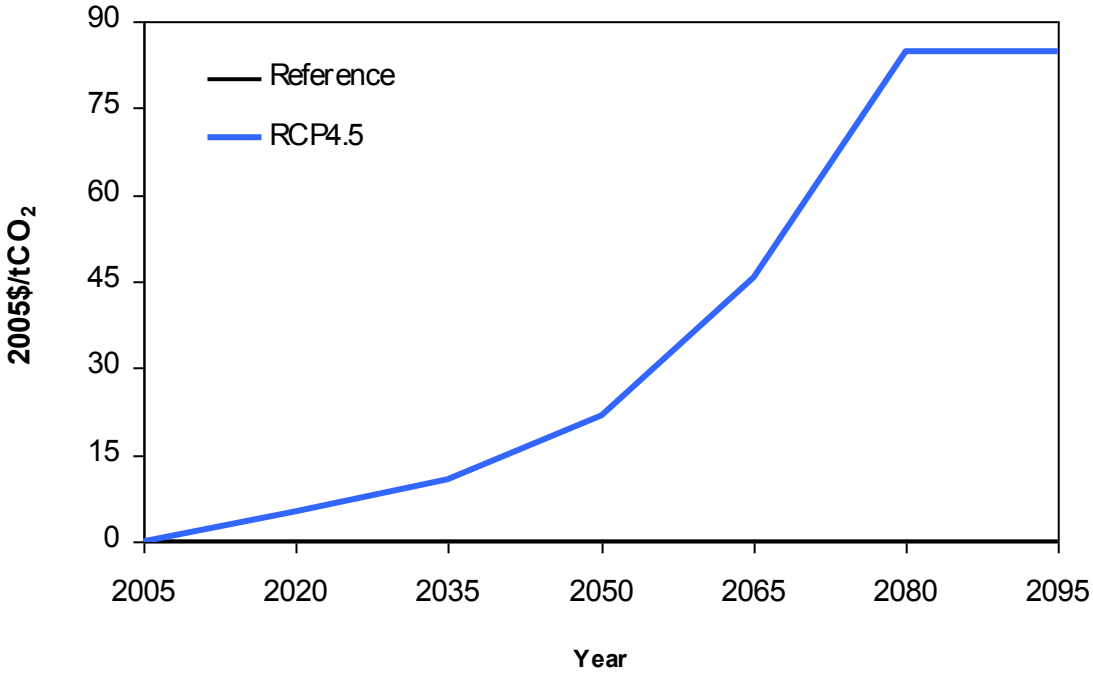


Figure 6: Price of CO_2 per ton (2005$) in the RCP4.5 scenario over the simulation period.

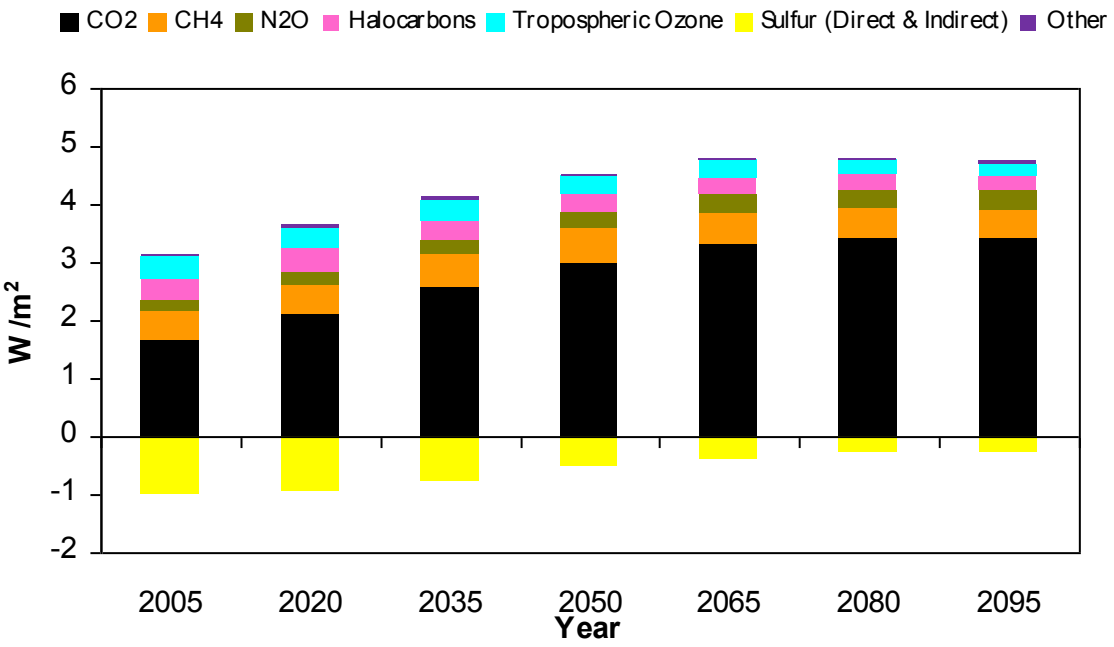
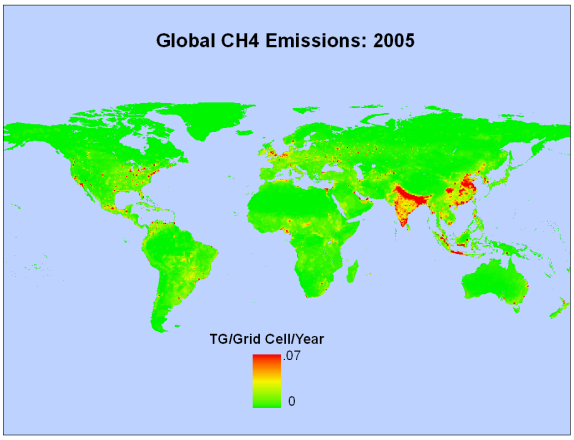
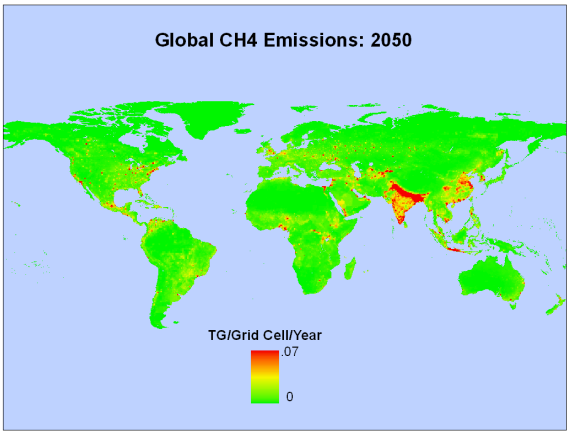


Figure 7: Contributions of the different greenhouse gases to total radiative forcing in the RCP4.5 scenario.


Global CH4 Emissions: 2005
TG/Grid Cell/Year
.07
0


Global CH4 Emissions: 2050
TG/Grid Cell/Year
.07
0

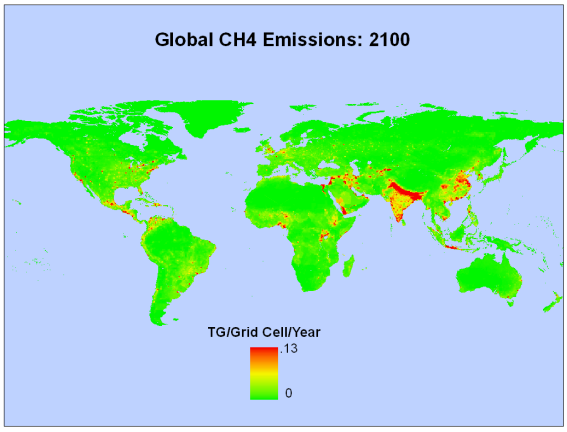


Figure 8: Global CH_4 emissions (in Tg per grid cell per year) downscaled from the GCAM RCP4.5 scenario for 2005, 2050 and 2100.

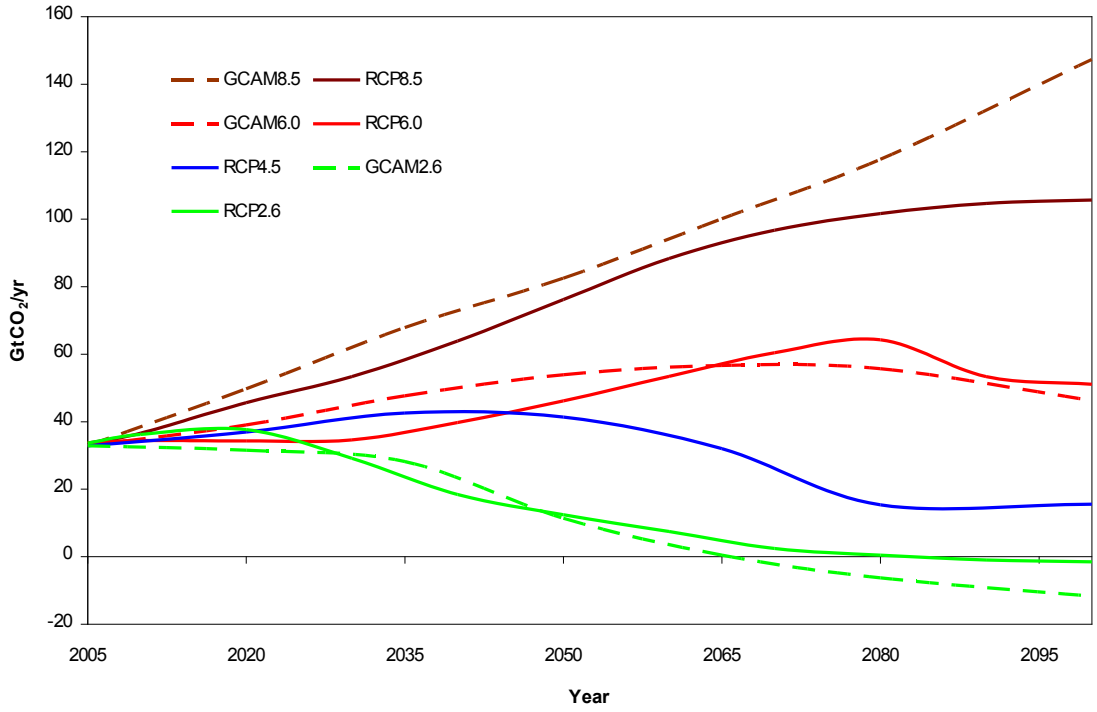


Figure 9: Annual CO_2 emissions ($GtCO_2$) from all anthropogenic sources for the four released RCPs (RCP2.6, RCP4.5, RCP6.0, RCP8.5) and the corresponding pathways simulated with the GCAM model (GCAM2.6, GCAM6.0, GCAM8.5).