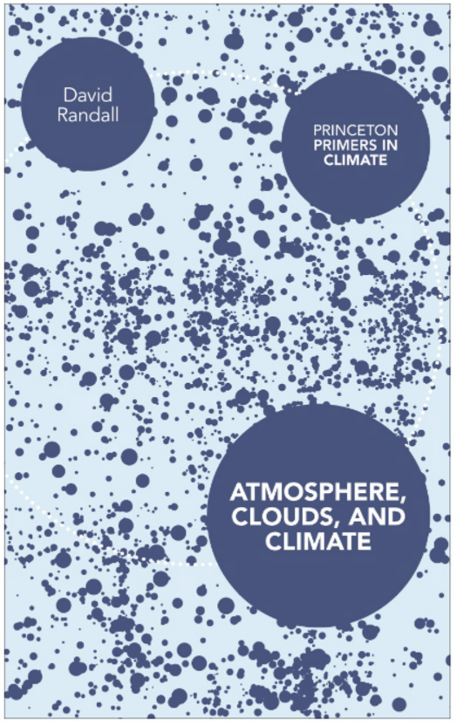
Atmosphere, Clouds, and Climate Introduction

AOSC 680

Ross Salawitch

Class Web Sites:
http://www2.atmos.umd.edu/~rjs/class/fall2024
https://umd.instructure.com/courses/1367293



Lecture 9

26 September 2024

Review: Importance of Oceans

1 2 3 4

Complete the blank in this phrase from the Houghton reading:

"any accurate simulation of likely climate change, especially of its regional variations, must included a description of the structure and dynamics of _____"

and then succinctly summarize (i.e., 2 to 4 sentences) the science behind this phrase.

Your Answer:

Blank: Ocean

Proper representation of the structure and dynamics of the ocean is essential for accurate projection of climate changes because of the following three reasons.

1. The ocean and atmosphere are a strongly coupled system, and therefore both components can influence the state of the other component by transport of heat, mass, and momentum (e.g. the ocean supply moisture to the atmosphere which can condense to release latent heat (heat and mass transport), and the atmosphere can drive ocean circulation through wind stress on the ocean surface (momentum transport).
2. Heat capacity of the ocean is much larger in comparison to the atmosphere. Therefore, the variation in atmospheric temperature is strongly constrained by the less variable ocean temperature.
3. Internal circulation of the ocean transports heat throughout the climate system. The amount of redistributed heat from this circulation is of comparable scale to those of incident solar radiation in the regions of the north Atlantic Ocean.

Additional Comments:

Wonderful reply: thanks for including so much detail. Excellent to mention "therefore both components can influence the state of the other component by transport of heat, mass, and momentum (e.g. the ocean supply moisture to the atmosphere which can condense to release latent heat (heat and mass transport), and the atmosphere can drive ocean circulation through wind stress on the ocean surface (momentum transport)", which is central to our class (+1).

Understanding how atmospheric / oceanic interactions will evolve over time is a topic at the forefront of climate science.

Review: Importance of Clouds

Cloud radiative forcing

A concept helpful in distinguishing between the two effects of clouds mentioned in the text is that of cloud radiative forcing (CRF). Take the radiation leaving the top of the atmosphere above a cloud; suppose it has a value R. Now imagine the cloud to be removed, leaving everything else the same; suppose the radiation leaving the top of the atmosphere is now R'. The difference $R' - R$ is the cloud radiative forcing. Note that a positive $R' - R$ implies that the presence of the cloud leads to a cooling of the Earth–atmosphere system. It can be separated into solar radiation and thermal radiation components that generally act in opposite senses, each typically of magnitude between 50 and 100 W m^{-2}. On average, it is found that clouds tend slightly to cool the Earth–atmosphere system.

A map of cloud radiative forcing (Figure 5.16a) deduced from satellite observations illustrates the large variability in CRF over the globe with both positive and negative values. It is also helpful to study separately the shortwave and longwave components of the atmosphere's radiation budget (Figure 5.16b), the variations of which are dominated by variations in cloud cover and type. Model simulations are able to capture the overall pattern of these variations; the big challenge is to simulate the changing pattern with adequate detail and accuracy (see Question 8). It is through careful comparisons with observations that progress in the understanding of cloud feedback will be achieved.

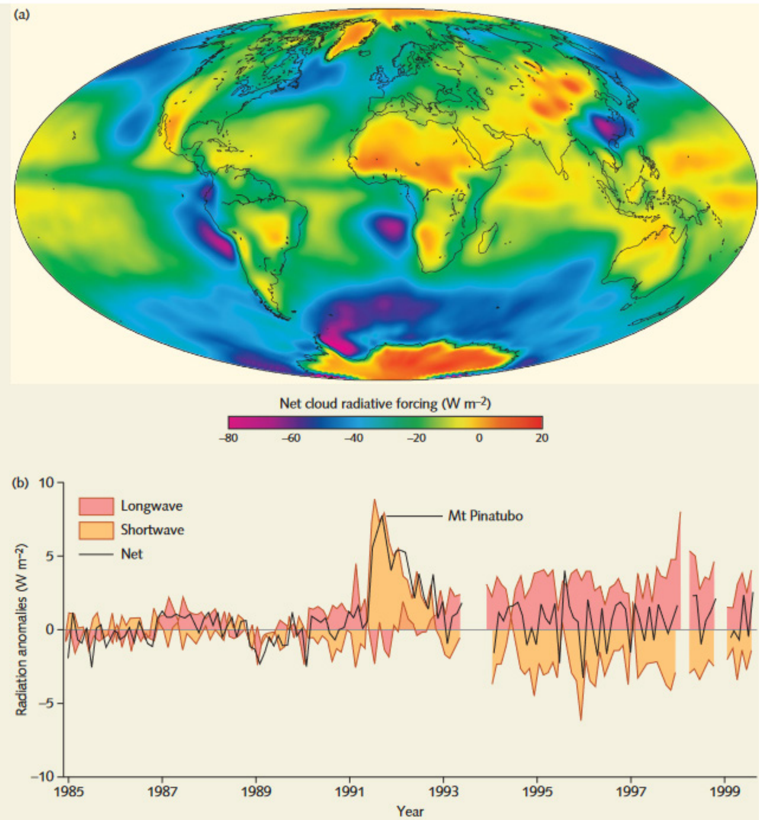
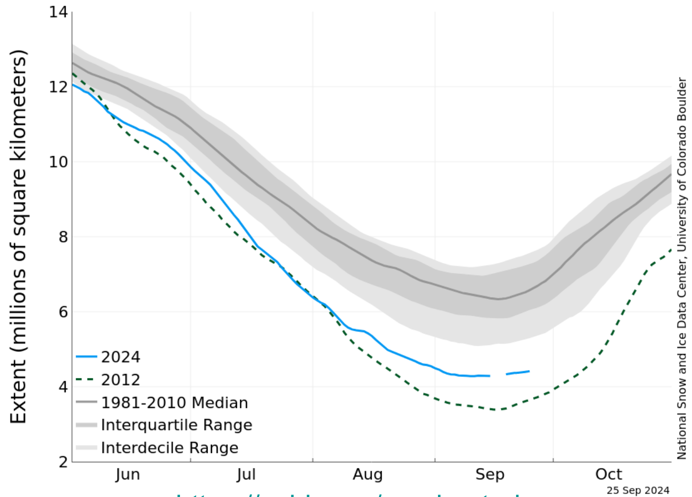


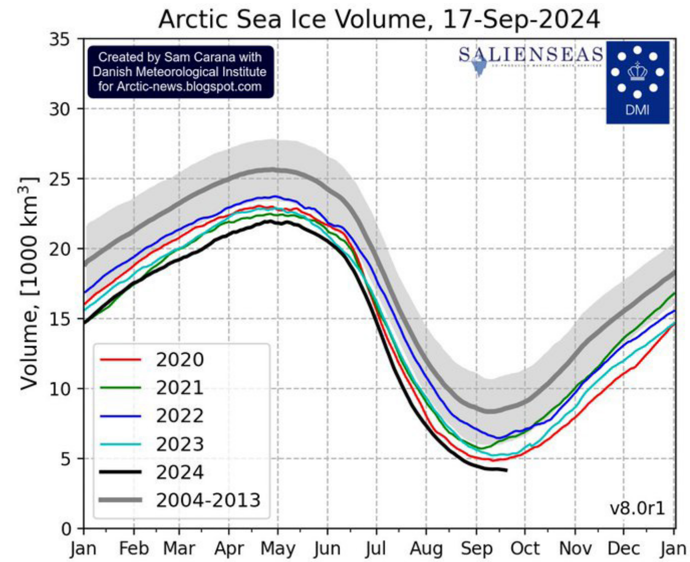
Figure 5.16 (a) Annual mean net cloud radiative forcing (CRF) for the period March 2000 to February 2001 as observed by the CERES instrument on the NASA Terra satellite. (b) Comparison of the observed longwave (pink/red), shortwave (orange) and net radiation at the top of the atmosphere for the tropics (20°N–20°S) as deviation from the mean for 1985–99; data from the ERBE instrument on the ERBS satellite and the CERES instrument on the TRMM satellite. Note the influence of the eruption of Pinatubo volcano.

Houghton, *Global Warming*, 5th edition

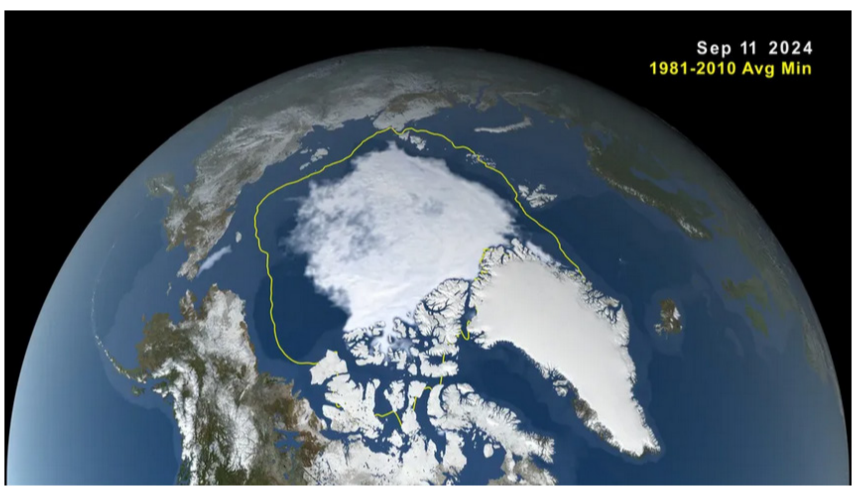
Announcement: Arctic Sea Ice



https://nsidc.org/sea-ice-today



https://twitter.com/SamCarana/status/1836388276531610110



This image, taken from a data visualization, shows Arctic sea ice minimum extent on September 11, 2024. The yellow boundary shows the minimum extent averaged over the 30-year period from 1981 to 2010. Download high-resolution video and images from NASA's Scientific Visualization Studio: svs.gsfc.nasa.gov/5382

NASA's Scientific Visualization Studio/Trent L. Schindler

> “ Today, the overwhelming majority of ice in the Arctic Ocean is thinner, first-year ice, which is less able to survive the warmer months. There is far, far less ice that is three years or older now,

NATHAN KURTZ

Chief, NASA's Cryospheric Sciences Laboratory

https://www.nasa.gov/earth/arctic-sea-ice-near-historic-low-antarctic-ice-continues-decline/

Figure 1.2. A photograph of the atmosphere as seen from space; the Sun is just below the horizon.

If the figure were in color, you would see that the lower atmosphere looks red because the blue photons emitted by the Sun are scattered away from the line of sight. A few clouds can be seen as dark blobs, blocking the Sun's rays.

Source: https://commons.wikimedia.org/wiki/File:Sunset_from_the_ISS.JPG

Overall, considering both solar and infrared radiation, the atmosphere is radiatively cooled. The radiative cooling is balanced by the latent heat released when the water evaporated from the ocean recondenses to form clouds. In this and other ways, the Earth's energy and water cycles are closely linked. Chapters 3, 5, and 6 explore these links in more detail.

AGU Advances

Research Article | Open Access

How Moisture Shapes Low-Level Radiative Cooling in Subsidence Regimes

B. Fildier, C. Muller, R. Pincus, S. Fueglistaler

First published: 16 May 2023 | https://doi.org/10.1029/2023AV000880

Key Points

- New theory is developed for the shape and magnitude of low-level longwave cooling peaks
- Low-level cooling scales with the boundary-layer-to-free-troposphere ratio in relative humidity
- Elevated intrusions of moist air in mid-levels can significantly damp low-level cooling

With rising SSTs, enhanced pattern's lifetime would promote patterns with large shallow cloud fractions and the dryness of their clear-sky surroundings. If at play, this would imply that the Earth's subtropics may reflect more sunlight in the future, and simultaneously allow the surface to cool more efficiently to space, a negative feedback on global warming.

https://agupubs.onlinelibrary.wiley.com/doi/10.1029/2023AV000880

The term "scale" has already been used several times in this book. Atmospheric processes often have characteristic space and/or time scales. The concept of scale is particularly useful in connection with the winds and ocean currents. Spatial and temporal scales tend to increase or decrease together. Turbulent eddies are meters to hundreds of meters across and last for seconds to minutes. Thunderstorms are a few kilometers across and last for an hour or two. Winter storms can be thousands of kilometers across and last for several days. Monsoons span thousands of kilometers and persist for months. These diverse scales of motion can strongly interact with each other. Large-scale weather systems can excite small-scale turbulence. The turbulence, in turn, exerts drag forces and other influences on the large-scale weather. Such "scale interactions" are extremely important and will come up repeatedly throughout this book.

The troposphere is roughly 10 km thick, and this constrains the vertical scale of most weather phenomena. Thus, phenomena of large horizontal scale will have a constrained vertical scale, causing them to be similar to a pancake. However, phenomena with smaller horizontal scale can have aspect ratios (width/height) of about one; namely, their characteristics are **isotropic**.

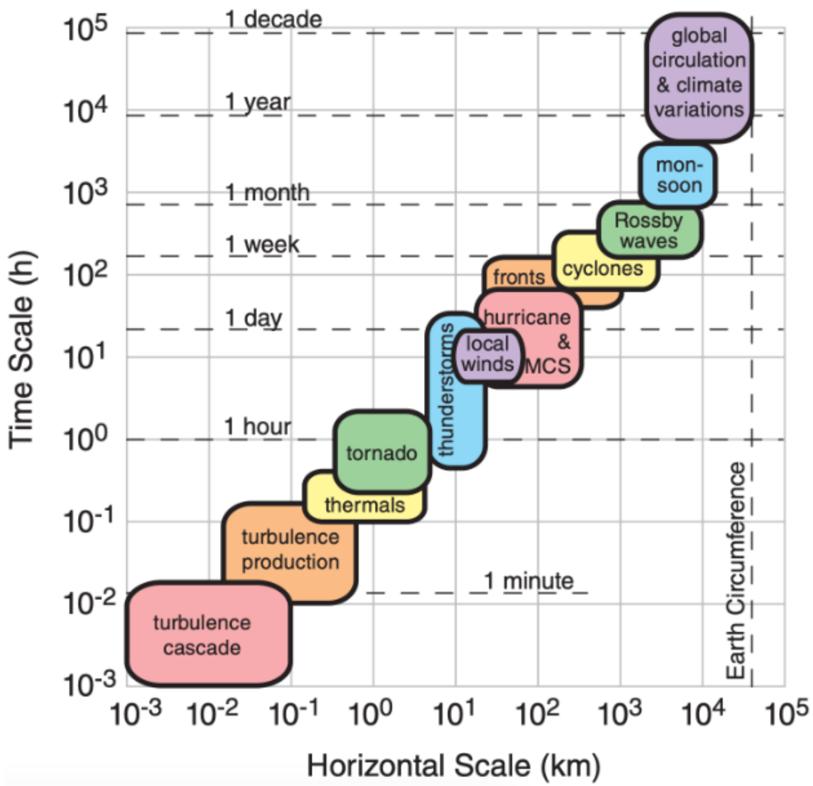


Figure 10.24 Typical time and spatial scales of meteorological phenomena. MCS = Mesoscale Convective System (see the thunderstorm chapter).

Larger-scale meteorological phenomena tend to exist for longer durations than smaller-scale ones. Fig. 10.24 shows that time scales τ and horizontal length scales λ of many meteorological phenomena nearly follow a straight line on a log-log plot. This implies that

$$\tau/\tau_0 = (\lambda/\lambda_o)^b \qquad (10.53)$$

where $\tau_o \approx 10–3$ h, $\lambda_o \approx 10^{-3}$ km, and $b \approx 7/8$.

https://geo.libretexts.org/Bookshelves/Meteorology_and_Climate_Science/Practical_Meteorology_%28Stull%29/10%3A_Atmospheric_Forces_and_Winds/10.05%3A_Section_6-

Table 1.1
A partial list of the well-mixed gases that make up "dry air," in order of their mass fraction of the atmosphere. Oxygen is present in significant quantities only because of the existence of life on Earth. Three of the six leading constituents are noble gases (argon, neon, and helium). Additional gases (not listed) are present in smaller amounts.

Gas	*Molecular form*	*Molecular mass g mol^{-1}*	*Volume fraction ppmv*	*Mass fraction of the "dry" portion of the atmosphere*
Nitrogen	N_2	28.0	781,000	0.755
Oxygen	O_2	32.0	209,000	0.231
Argon	Ar	39.4	9,340	0.0127
Carbon dioxide	CO_2	44.0	390	5.92×10^{-4}
Neon	Ne	20.2	18	1.26×10^{-5}
Helium	He	4.0	5	6.90×10^{-7}
Methane	CH_4	16.0	2	1.10×10^{-7}

Any surprise?

THE COMPOSITION OF THE ATMOSPHERE

The atmosphere is big. Its total mass is about 5×10^{21} g.

The most abundant atmospheric constituents are nitrogen and oxygen. They are very well mixed throughout almost the entire atmosphere, so that their relative concentrations are essentially constant in space and time.

The result is that, except for ozone, the concentrations (*not* the densities) of the gases that make up dry air, for example, nitrogen, oxygen, argon, and carbon dioxide, are observed to be nearly homogeneous, both horizontally and vertically, up to an altitude of about 100 km. Above that level, diffusive separation does become noticeable.

Anyone confused ?

CO_2 versus Latitude and Altitude

M. Diallo et al.: Global distribution of CO_2 in the upper troposphere–stratosphere

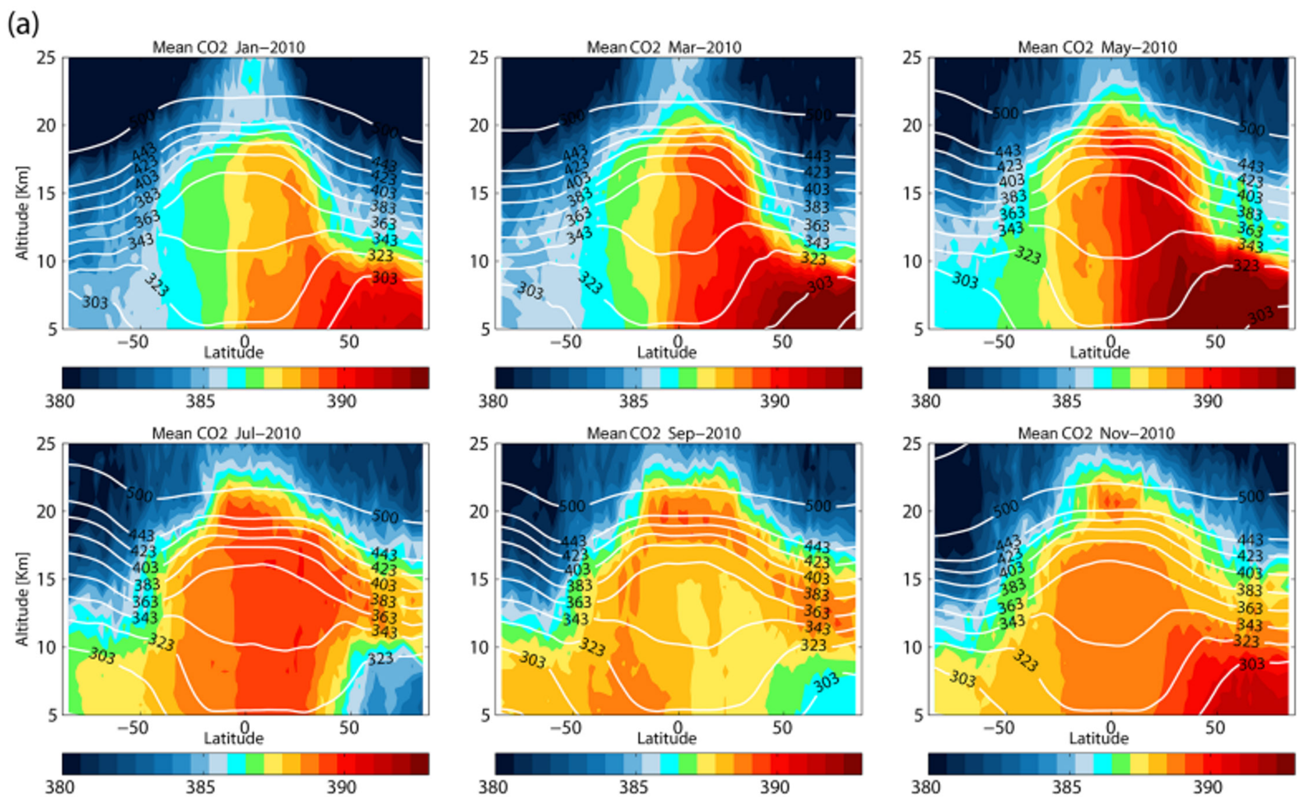


Figure 6. **(a)** Global distribution of the seasonal cycle of the reconstructed monthly mean CO_2 (in ppmv) in the upper troposphere and the lower stratosphere from 5 to 25 km for the odd months of 2010. **(b)** Same as **(a)** but for the even months of 2010 and the altitude range from 5 to 45 km. CO_2 calculated on model levels is first interpolated to altitude levels using the latitude dependency of the zonally and monthly averaged geopotential. **(c)** The standard error of the mean CO_2 over the 2000–2010 period. The white contours show the isentropic surfaces.

Diallo et al., ACP, 2017

CH_4 versus Latitude and Altitude

M. K. van Aalst et al.: Nudged GCM tracer transport in the 1999/2000 Arctic winter

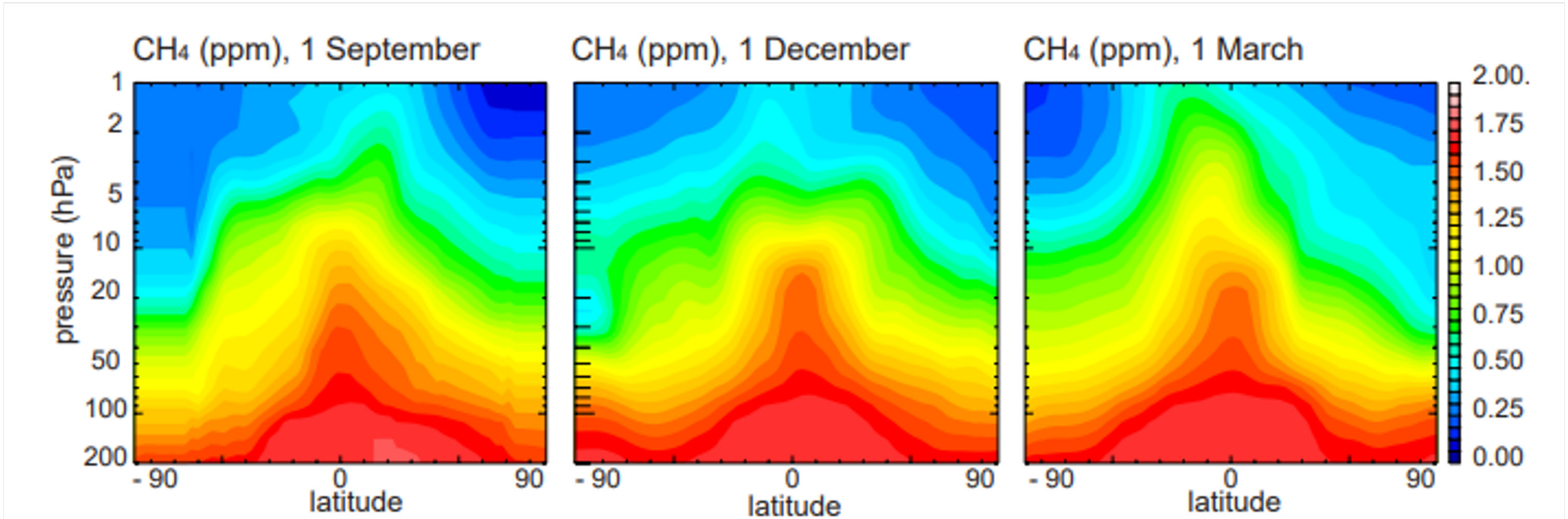


Fig. 1. Latitude versus pressure (from 200 to 1 hPa) zonal mean cross-sections of MA-ECHAM4 CH_4

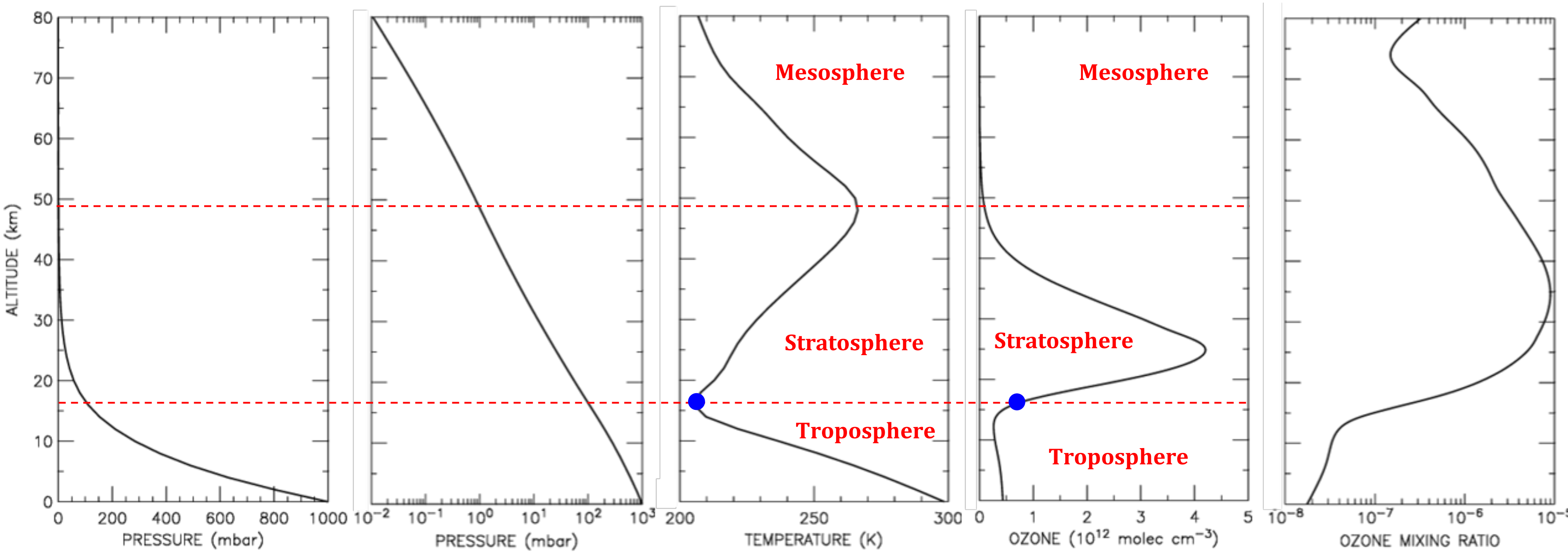
van Aalst et al., ACP, 2004

Chapter 1, Atmosphere, Clouds, and Climate

First point sentence quite important. Next sentence a bit of an overstatement, IMHO

The liquid and ice particles that make up clouds are typically "nucleated" on aerosols; the variable abundance of these cloud-nucleating particles, which are called cloud condensation nuclei (CCN), is a factor influencing the formation, number, and size of cloud particles. The number of CCN does not strongly influence the probability of cloud formation or the area-averaged rate of precipitation, under realistic conditions. The fascinating complexities of clouds are discussed further in later chapters.

Above the tropopause is the stratosphere. It is meteorologically quiet, compared to the troposphere. A distinctive property of the stratosphere is that the temperature generally increases upward there, as can be seen in Figure 1.1. The upward temperature increase is caused by heating of the middle and upper stratosphere due to the absorption of UV radiation from the Sun. The UV is actually absorbed by ozone (O_3), which is thereby converted into molecular oxygen (O_2) and atomic oxygen (O). The O_2 and O then recombine (in the presence of other species) to create another ozone molecule, so that there is no net loss of ozone. The net effect is a continual destruction and regeneration of ozone, accompanied by a heating of the air. The ozone cycle is an essential characteristic of the stratosphere.



Chapman Chemistry

• Production of ***stratospheric*** O_3 initiated when O_2 is photodissociated by UV sunlight

• O_3 formed when resulting O atom reacts with O_2 :

$$h\nu + O_2 \rightarrow O + O \qquad (1)$$
$$O + O_2 + M \rightarrow O_3 + M \qquad (2)$$

• O_3 removed by photodissociation (UV sunlight) or by reaction with O :

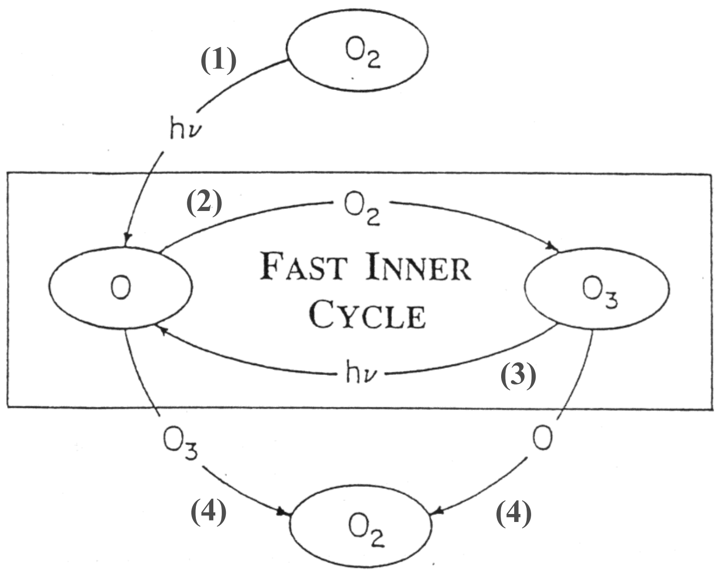
$$h\nu + O_3 \rightarrow O + O_2 \qquad (3)$$
$$O + O_3 \rightarrow O_2 + O_2 \qquad (4)$$

This reaction sequence was first worked out in the 1930s by Sydney Chapman,
an English mathematician and geophysicist

Chapman Chemistry

• The cycling between O and O_3 (rxns 2 and 3) occurs *much* more rapidly than leakage into (rxn 1) or out of the system (rxn 4)

• The sum O + O_3 is commonly called *"odd oxygen"*

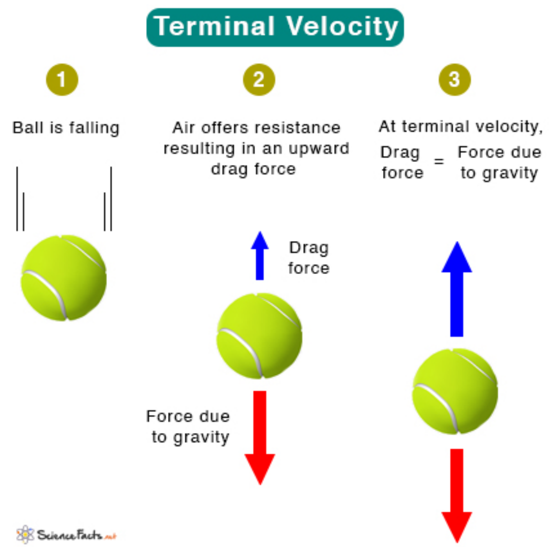


Rxn (1) produces two *odd oxygen* molecules
Rxn (4) consumes two *odd oxygen* molecules

and reactions 2 and 3 recycle *odd oxygen* molecules

The most obvious liquid and solid particles in the atmosphere are the liquid water drops and ice crystals that make up clouds. As a matter of terminology, however, atmospheric scientists conventionally reserve the term "aerosol" to refer to the wide variety of *noncloud* liquid and solid particles in the air. Cloud particles range in size from about 10 microns (a micron is 10^{-6} meter) to a centimeter or so for large rain drops and snowflakes. Particles smaller than about 0.1 mm fall slowly because their motion is strongly limited by drag, so to a first approximation they can be considered to move with the air like a gas. Particles that fall more quickly are said to "precipitate"; their fall speeds are close to the "terminal velocity" at which their weight is balanced by aerodynamic drag. The drag increases with the density of the air, and with the square of the fall speed. Large rain drops fall at about 5 m s^{-1}, relative to the air. If such drops find themselves in an updraft with a speed faster than 5 m s^{-1}, they will actually be carried upward by the air.

Super important text !



https://www.sciencefacts.net/wp-content/uploads/2022/06/Terminal-velocity.jpg
https://www.sciencefacts.net/terminal-velocity.html

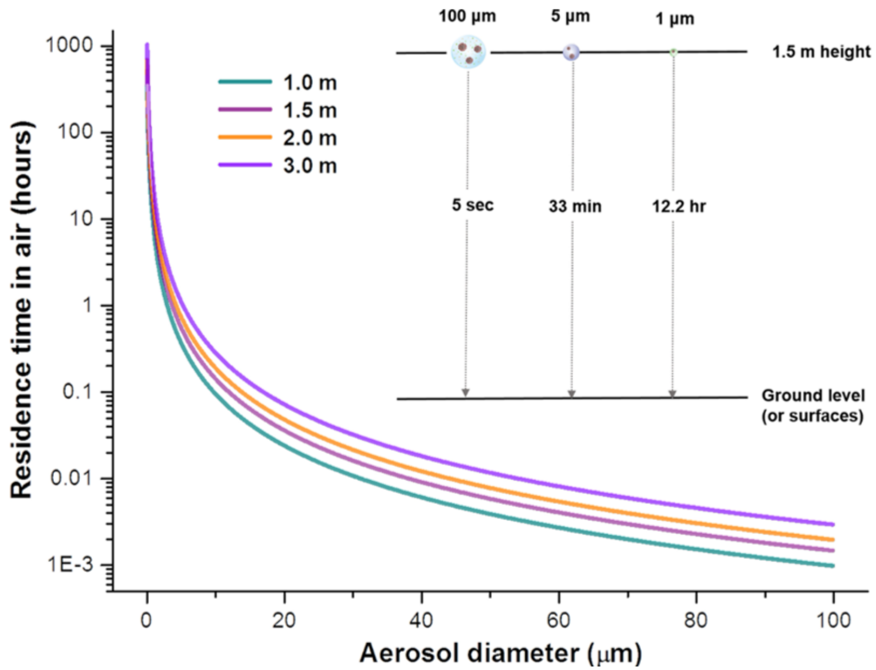


Fig. 3. How long can aerosols linger in air?
Residence time of aerosols of varying size in still air can be estimated from Stokes' law for spherical particles (*116*). For example, the time required for an aerosol of 100, 5, or 1 µm to fall to the ground (or surfaces) from a height of 1.5 m is 5 s, 33 min, or 12.2 hours, respectively.

https://www.science.org/doi/10.1126/science.abd9149

NATURE · VOL 339 · 15 JUNE 1989

Denitrification in the Antarctic stratosphere

R. J. Salawitch, G. P. Gobbi*, S. C. Wofsy & M. B. McElroy

Department of Earth and Planetary Sciences and Division of Applied Sciences, Harvard University, 29 Oxford Street, Cambridge, Massachusetts 02138, USA

Rapid loss of ozone over Antarctica in spring requires that the abundance of gaseous nitric acid be very low. Nitric acid is removed from the gas phase in the lower stratosphere at temperatures below about 195 K through the formation of crystalline nitric acid trihydrate[1–4], and below 188 K in association with ice crystals[5–9]. Precipitation of particulate nitric acid has been assumed to occur in association with large ice crystals, requiring significant removal of H_2O and temperatures well below the frost point. However, stratospheric clouds exhibit a bimodal size distribution in the Antarctic atmosphere, with most of the nitrate concentrated in particles with radii ≥1 µm (refs 10, 11). Here we argue that the bimodal size distribution sets the stage for efficient denitrification, with nitrate particles either falling on their own or serving as nuclei for the condensation of ice. Denitrification can therefore occur without significant dehydration: it is unnecessary for temperatures to drop significantly below the frost point.

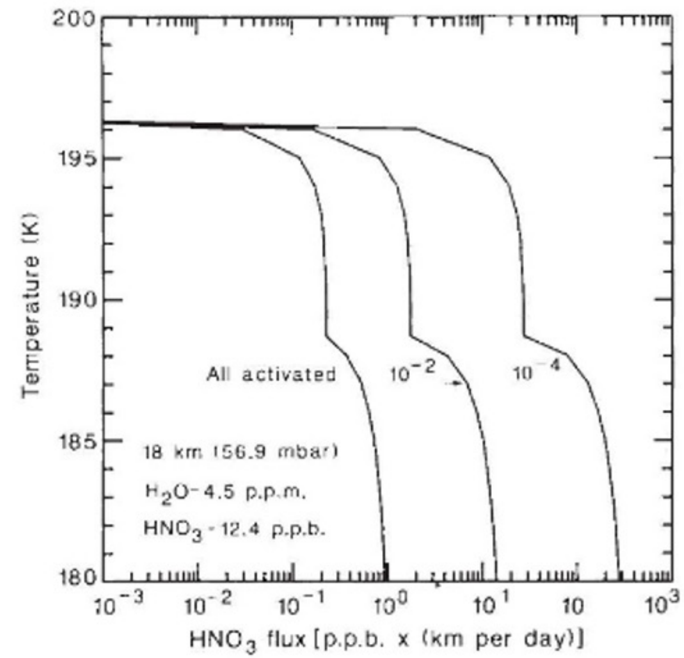


FIG. 1 The flux of nitric acid in sedimenting particles is shown as a function of temperature for various numbers of activated nuclei, expressed as a fraction of the initial condensation nuclei (assumed density 10 cm^{-3}). The air parcel was assumed to be at an altitude of 18 km and to maintain thermodynamic equilibrium with its environment at all times. Condensation of NAT begins at 196.3 K; the same particles activated during the growth of NAT were assumed to be activated when water ice forms (188.7 K). The analysis focuses on vapour pressures and particle densities in clouds after condensation, when vapour and solid phases will be approximately at equilibrium. Therefore the lower temperatures ($\Delta T \approx 1$ K) required to activate background aerosols are not explicitly considered.

https://www.nature.com/articles/339525a0

Airborne transmission of respiratory viruses

CHIA C. WANG, KIMBERLY A. PRATHER, JOSUÉ SZNITMAN, JOSE L. JIMENEZ, SEEMA S. LAKDAWALA, ZEYNEP TUFEKCI, AND LINSEY C. MARR

Authors Info & Affiliations

SCIENCE • 27 Aug 2021 • Vol 373, Issue 6558 • DOI: 10.1126/science.abd9149

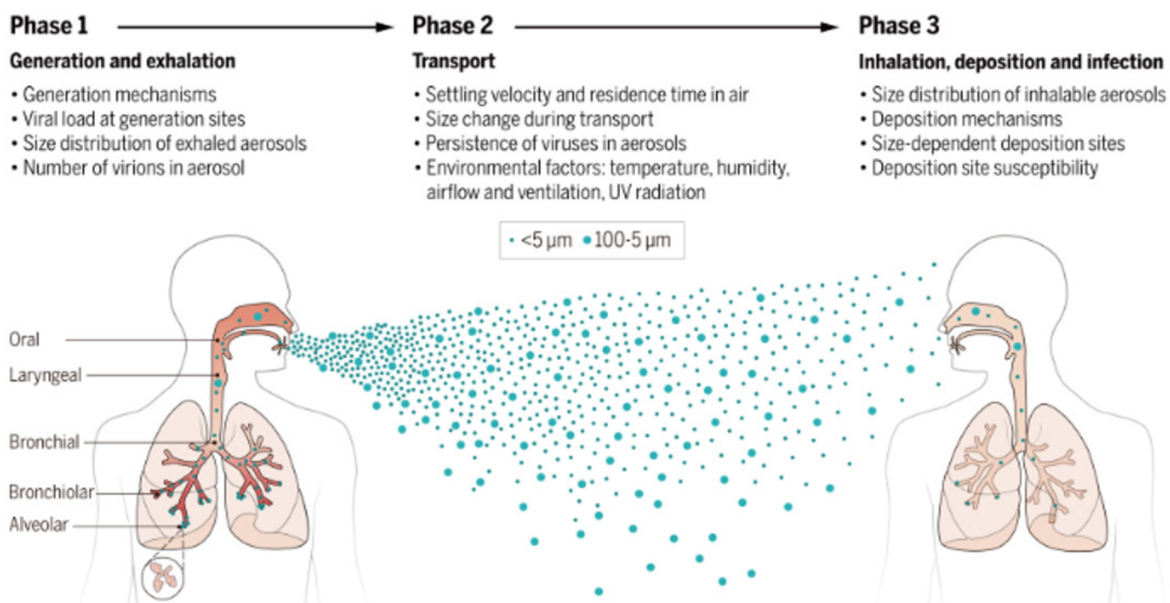


Fig. 1. Airborne transmission of respiratory viruses.

Phases involved in the airborne transmission of virus-laden aerosols include (i) generation and exhalation; (ii) transport; and (iii) inhalation, deposition, and infection. Each phase is influenced by a combination of aerodynamic, anatomical, and environmental factors. (The sizes of virus-containing aerosols are not to scale.)

https://www.science.org/doi/10.1126/science.abd9149

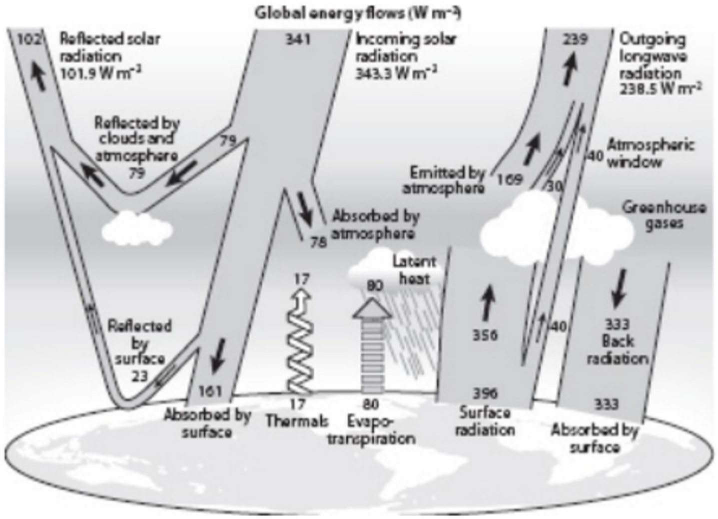


Figure 1.3. An overview of the flow of energy in the climate system.
The global annual mean Earth's energy budget for the March 2000 to May 2004 period (W m^{-2}). The broad arrows indicate the schematic flow of energy in proportion to their importance.
Source: Based on a figure in Trenberth et al. (2009).

Overall, considering both solar and infrared radiation, the atmosphere is radiatively cooled. The radiative cooling is balanced by

What word describes this angle?

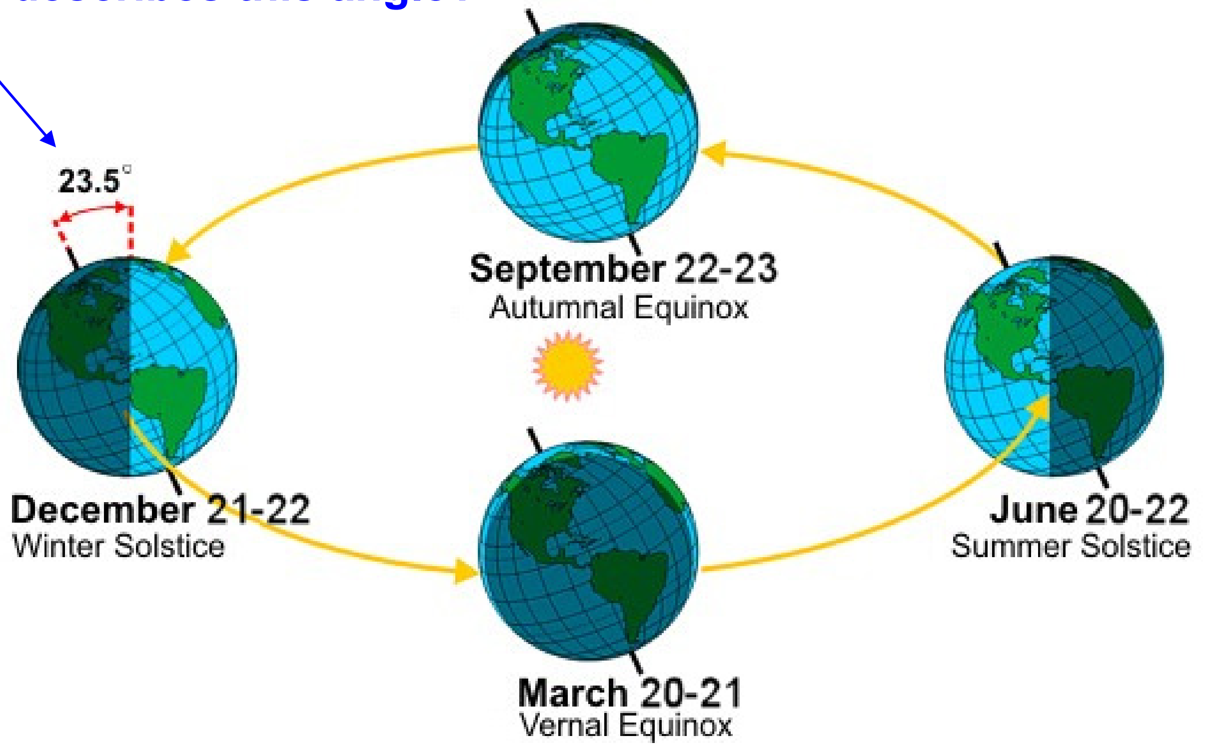


Figure 2.1. Changes in the Sun-Earth geometry as the Earth moves in its orbit.
The Earth's axis is tilted with respect to the plane of its orbit. As the tilted Earth revolves around the Sun, changes in the distribution of sunlight cause the succession of seasons.

Source: https://www.weather.gov/images/cle/Education/EarthOrbit.png

Can someone talk us through this image?

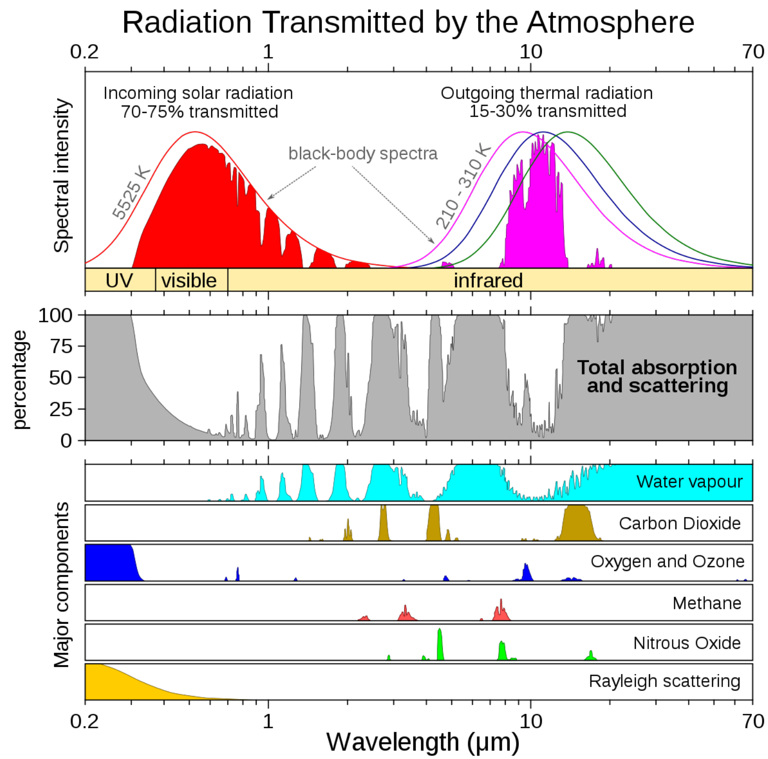


Figure 2.3. The absorption and scattering spectra for major gases in the Earth's atmosphere.

For both panels of the figure, the values along the horizontal axis are the wavelengths of the radiation, with a logarithmic scale. Recall that visible light has wavelengths in the range 0.4 to 0.8 μm, so it is concentrated on the left-hand side of the axis. In the upper panel, the vertical axis shows the percentage of a particular wavelength that is absorbed or scattered back to space by atmospheric gases. The lower panel shows the contributions to absorption and scattering from various constituents, namely water vapor, carbon dioxide, oxygen and ozone, methane, nitrous oxide, and other gases (mainly nitrogen). Rayleigh scattering is the scattering of radiation by gases; it is distinguished from scattering by clouds and aerosols.

Source: https://upload.wikimedia.org/wikipedia/commons/thumb/5/5e/Atmospheric_Transmission-en.svg/1200px-Atmospheric_Transmission-en.svg.png

How does this passage compare to our prior description?

The total solar energy *incident* on the Earth is $S\pi a^2$, where $S \cong 1365\ \mathrm{W\ m^{-2}}$ is the energy per unit area per unit time in the "solar constant," and a is the radius of the Earth. Here πa^2 is the circular cross-sectional area that that the spherical Earth presents to the solar beam. The total solar energy *absorbed* by the Earth, divided by the area of the Earth's spherical surface, can be written as

$$\begin{aligned} S_{abs} &= S\left(\frac{\pi a^2}{4\pi a^2}\right)(1-\alpha) \\ &= \frac{1}{4}S(1-\alpha) \cong 240\ \mathrm{W\,m^{-2}} \quad \text{(annual mean).} \end{aligned} \tag{2.1}$$

Here S_{abs} is the average absorbed solar energy per unit area and per unit time and α is the *planetary albedo*, defined as the globally averaged energy scattered back to space divided by the globally averaged energy coming in from the Sun. The reflected solar radiation, per unit area, is $S\alpha$. The Earth's albedo is close to 0.3, independent of season; this number has been accurately known only since the advent of satellite data in the 1970s.

Can someone talk us through this table?

Table 2.1
Components of the globally and annually averaged surface radiation budget. A positive sign means that the surface is warmed.

Absorbed solar (SW)	161 W m^{-2}
Downward infrared (LW↓)	333 W m^{-2}
Upward infrared (LW↑)	–396 W m^{-2}
Net longwave (LW)	–63 W m^{-2}
Net radiation (SW + LW)	98 W m^{-2}

What is the value and meaning of the bulk emissivity of Earth's atmosphere?

Much less infrared energy leaves the atmosphere to space (240 W m^{-2}) than enters the atmosphere from below (390 W m^{-2}). For the real Earth, the OLR is only about as large as the upward emission of infrared by the Earth's surface . Writing

$$OLR = \varepsilon_B \sigma_{SB} T_S^4, \tag{2.3}$$

we can say that the *bulk emissivity of the Earth*, denoted by ε_B, is We can also write $OLR = \sigma_{SB} T_{\text{brightness}}^4$, where $T_{\text{brightness}}$ = 255 K is the Earth's brightness temperature. It follows that $\varepsilon_B = \frac{T_{\text{brightness}}^4}{T_S^4}$. The bulk emissivity will come up again later.

If the Earth had no atmosphere, ε_B would be equal to one. The departure of ε_B from unity, that is, 1-ε_B, is a measure of the opacity of the atmosphere to infrared radiation. The actual value of ε_B is influenced by several factors, including the composition of the atmosphere (gases, clouds, and aerosols) and the temperature sounding (i.e., the vertical profile of atmospheric temperature).

What do folks think of the simplified two layer model?

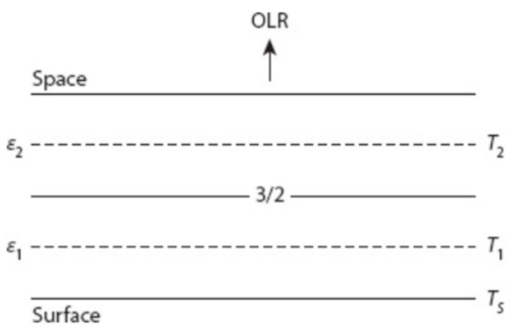


Figure 2.4. An idealized model in which the atmosphere is represented by just two layers, with different temperatures and emissivities.

The lower layer is denoted by subscript 1, and the upper layer by subscript 2. The interface between the two layers is denoted by subscript 3/2.

After substitution from (2.4)–(2.7), these two equations can be used to solve for the temperatures of the two model layers. The results are

$$\sigma_{SB}T_1^4 = \left[\frac{2+(1-\varepsilon_1)\varepsilon_2}{4-\varepsilon_1\varepsilon_2}\right]\sigma_{SB}T_S^4 \text{ and}$$
$$\sigma_{SB}T_2^4 = \left(\frac{2-\varepsilon_1}{4-\varepsilon_1\varepsilon_2}\right)\sigma_{SB}T_S^4. \quad (2.9)$$

The first of these is valid provided that $\varepsilon_1 \neq 0$, and the second provided that $\varepsilon_2 \neq 0$. By trying some numerical values for ε_1 and ε_2, and choosing a surface temperature, you can calculate the temperatures T_1 and T_2. For example, putting T_S equal to the observed globally averaged value of 288 K, and setting $\varepsilon_1 = \varepsilon_2 = 0.5$, we find that $T_1 = 253$ K and $T_2 = 229$ K. These are comparable to the observed temperatures in the middle and upper troposphere, respectively.

Substituting from Equation (2.9) back into Equation (2.5), we find, after some algebra, that

$$OLR = \left\{\frac{(2-\varepsilon_1)(2-\varepsilon_2)+2\varepsilon_1\varepsilon_2}{4-\varepsilon_1\varepsilon_2}\right\}\sigma_{SB}T_S^4. \quad (2.10)$$

Comparing Equation (2.10) to Equation (2.3), we see that the expression in curly braces, in Equation (2.10), is the bulk emissivity. Using $\varepsilon_1 = \varepsilon_2 = 0.5$, we obtain a bulk emissivity of 0.73. Many additional results can be worked out, but we will stop here.

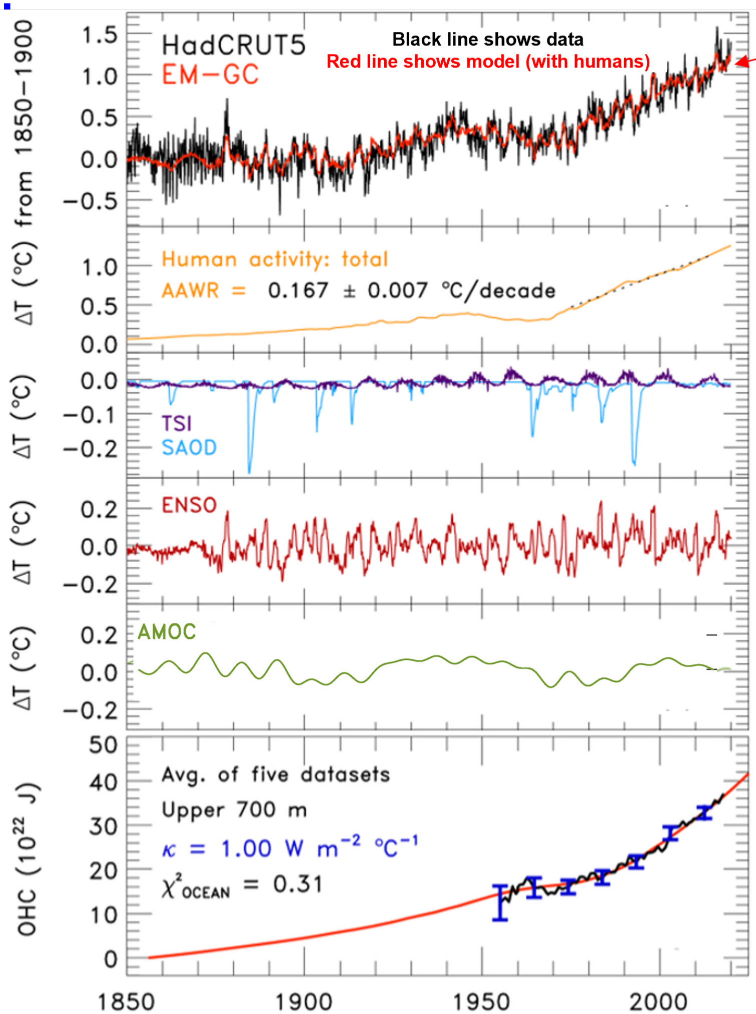
Let's look at the solar model, in terms of prior class material:

Then (2.14) reduces to

$$\frac{\Delta T_S}{T_o} \cong \frac{1}{4}\frac{\Delta S}{S_o} \qquad (2.15)$$

This says that the fractional change in surface temperature is equal to one-fourth of the fractional change in solar output. From measurements, we know that the globally averaged surface temperature is currently 288 K. Using this value for T_O, we find that a 1% change in solar output will lead to a 0.72 K change in surface temperature. Observed fluctuations of solar output over the past 30 years are about 0.1% in magnitude, so the expected temperature variations of T_S due to changes in the Sun are less than one-tenth of a kelvin.

Let’s look at the solar model, in terms of prior class material: Lecture 2:



$$\Delta T_{MDL\ i} = (1+\gamma)\left(\frac{GHG\ RF_i + LUC\ RF_i + Aerosol\ RF_i}{\lambda_p}\right) + C_0 + C_1 \times SOD_{i-6} + C_2 \times TSI_{i-1} + C_3 \times ENSO_{i-2} + C_4 \times AMOC_i - \left(\frac{Q_{OCEAN_i}}{\lambda_p}\right)$$

where:

- i denotes month
- $\lambda_p = 3.2\ \mathrm{W\ m^{-2}\,°C^{-1}}$
- $1+\gamma = \{1 - \lambda_\Sigma/\lambda_p\}^{-1}$
- GHG RF = RF due to all anthropogenic GHGs
- LUC RF = RF due to Land Use Change
- Aerosol RF = RF due to Tropospheric Aerosols
- SOD = Stratospheric Optical Depth
- TSI = Total Solar Irradiance
- ENSO = El Niño Southern Oscillation
- AMOC = Atlantic Meridional Overturning Circulation
- Q_{OCEAN} = Ocean heat export = $\kappa(1+\gamma)\{\Delta T_{MDL\ i} - \Delta T_{OCEAN\ SURFACE\ i}\}$

CRU: Climate Research Unit of East Anglia, United Kingdom
EM-GC: Empirical Model of Global Climate, Univ of Maryland

Canty *et al.*, 2013 https://www.atmos-chem-phys.net/13/3997/2013/acp-13-3997-2013.html
McBride *et al.*, 2021 https://esd.copernicus.org/articles/12/545/2021
Nicholls *et al.*, 2021 https://agupubs.onlinelibrary.wiley.com/doi/full/10.1029/2020EF001900
Figure provided by Laura McBride.

Let's look at the GHG model, in terms of prior class material:

As a second example, consider a simplified "global warming" case in which

$\Delta S = 0$, that is, there are no changes in the Sun's output
$\Delta\alpha = 0$, that is, there is no change in the planetary albedo

Then (2.14) reduces to

$$\frac{\Delta T_S}{T_0} \cong -\frac{1}{4}\frac{\Delta\varepsilon}{\varepsilon_0}. \qquad (2.16)$$

Equation (2.16) says that the change in the surface temperature is entirely due to a change in the bulk emissivity. The minus sign appears because a decrease in ε makes it harder for the Earth to emit infrared, and so leads to a warming. An increase in atmospheric CO_2 leads to a decrease in the bulk emissivity. It is known from the measured optical properties of CO_2 that, for the current climate, a doubling of CO_2 relative to its preindustrial concentration would reduce the OLR by 4 W m^{-2}, so that $\Delta\varepsilon\sigma T_0^4 \cong -4$ W m^{-2}. We also know, from satellite observations, that the OLR is $\varepsilon_0\sigma T_0^4 = 240$ W m^{-2}. Forming the ratio, we find that $-\left(\frac{\Delta\varepsilon}{\varepsilon_0}\right) = \frac{4}{240} \cong 0.017$. This means that doubling CO_2 creates a 1.7% perturbation to the OLR.

You should think of this reduction in the OLR as happening "instantaneously"; imagine that we could somehow double CO_2 without changing the temperature, or the albedo, or anything else. The system would then evolve so as to make the OLR increase again, reestablishing global energy balance. It would accomplish this by warming up, that is, by increasing T_0. Using the current globally averaged surface temperature of 288 K, we find from (2.13) that doubling CO_2 leads to a 1.2 K increase in the surface temperature—less than a 0.5% warming.

Let's look at the GHG model, in terms of prior class material:
Lecture 4:

$$\Delta T \approx \frac{1}{4\,\sigma\,T^3}\,\Delta RF$$

So: $$\lambda = \frac{1}{4\,\sigma\,T^3}$$

Earth's atmosphere is slightly more complicated, than a pure black body, as explained at:
http://web.archive.org/web/20211127135550/http://zebu.uoregon.edu/ph311/lec06.html

$\lambda_P \approx$ **0.3 K / (W m^{-2})**

Here: P refers to Planck Response Function

$$\Delta T \approx 0.3\,\frac{K}{W\,m^{-2}}\,\Delta RF$$

$$\Delta RF_{\text{Doubling CO2}} \approx 5.35\ W\,m^{-2}\,\ln\left(\frac{CO_2^{\,Final}}{CO_2^{\,Initial}}\right) = 3.7\ W\,m^{-2}$$

$$\Delta T \approx 0.3\,\frac{K}{W\,m^{-2}}\,\Delta RF = 0.3 \times 3.7\ W\,m^{-2} = \boxed{}$$

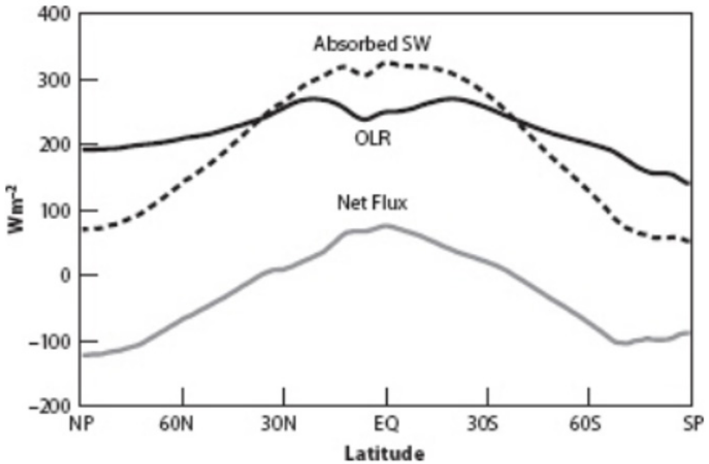


Figure 2.5. The zonally and annually averaged absorbed solar radiation (dashed), outgoing longwave radiation (solid black), and net radiation at the top of the atmosphere (gray), as observed from satellites.
The data are discussed by Wielicki et al. (1996).

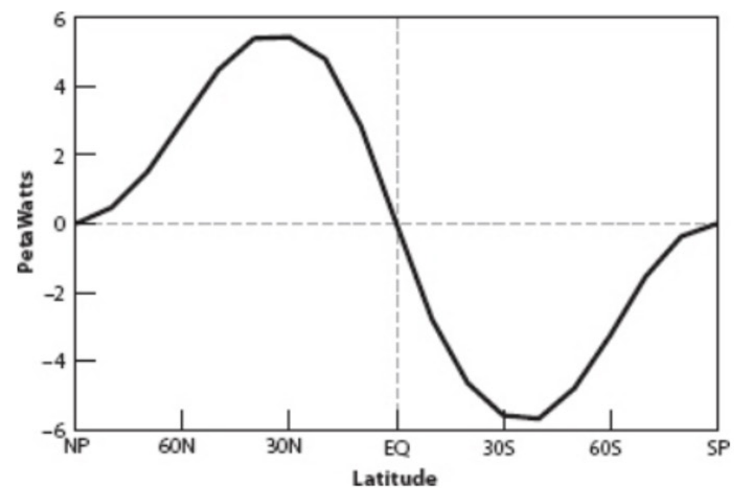


Figure 2.6. The poleward energy transport by the atmosphere and ocean combined, as inferred from the observed annually averaged net radiation at the top of the atmosphere.
A petawatt is 10^{15} W