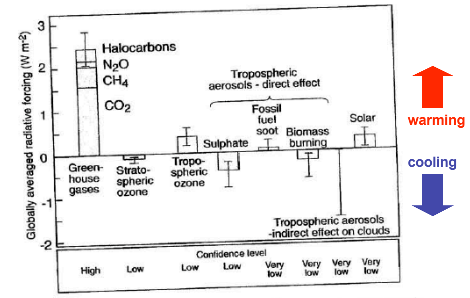
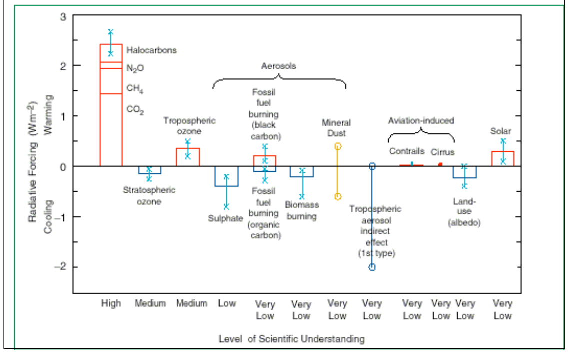
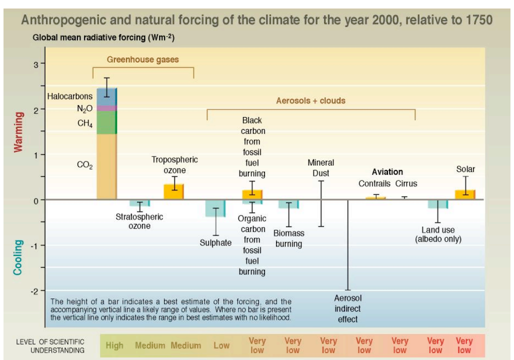


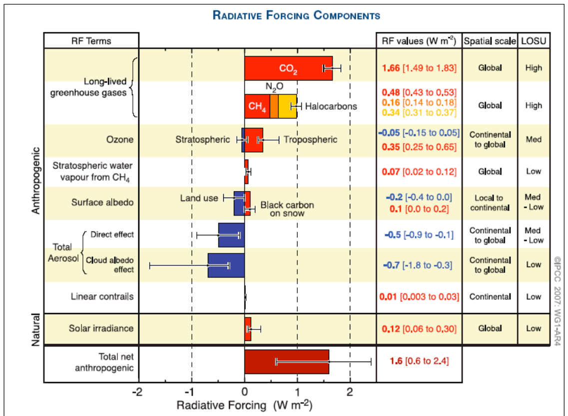
Figure 4.12. Estimate of the globally and annually averaged anthropogenic radiative forcings due to changes in the concentrations of greenhouse gases and aerosols from preindustrial times to 1995 and to natural changes in solar output from 1850. The height of the rectangular bar indicates a midrange estimate of the forcing, and the error bar shows an estimate of the uncertainty range. A subjective confidence level that the actual forcing lies within the error bar is indicated by the "confidence level." [From *Climate Change 1995*, Ed. J. T. Houghton et al., Cambridge University Press (1996).]

IPCC



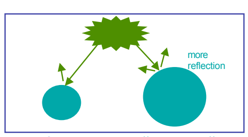


IPCC Synthesis Report

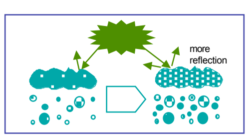


How do Aerosols cool?

Aerosol **direct** effects cause cooling by reflecting more light (e.g. smog).



Aerosol **indirect** effects cause cooling by clouds that reflect more light (e.g. tracks).



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ESSAY

BEYOND THE IVORY TOWER

The Scientific Consensus on Climate Change

Naomi Oreskes

Without substantial disagreement, scientists find human activities are heating the Earth's surface.

Lecture Ch. 7a

- CAPE
- Stability
- Review of Ch.7 Concepts
  - "Homework" Ch. 7, Prob. 3 for discussion
- Cloud Classification
- Precipitation Processes

Curry and Webster, Ch. 7, 8
For Tuesday: Finish reading Ch. 8

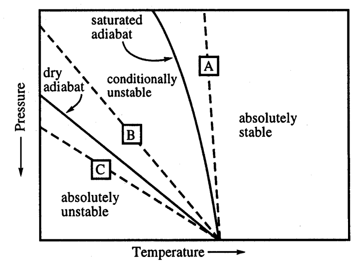

Figure 7.1 Regions of stability, instability, and conditional stability illustrated on an aerological diagram. When the environmental lapse rate is less than the saturated adiabatic lapse rate (e.g., lapse rate A), the atmosphere is absolutely stable. When the environmental lapse rate is greater than the saturated lapse rate, but less than the dry adiabatic lapse rate (e.g., lapse rate B), the atmosphere is conditionally stable. When the environmental lapse rate is greater than the dry adiabatic lapse rate (e.g., lapse rate C), the atmosphere is absolutely unstable.

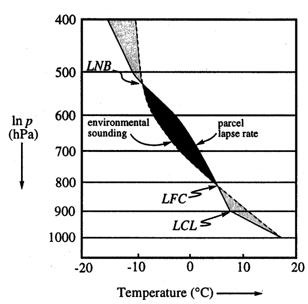

Figure 7.2 Convective instability illustrated on an aerological diagram. The dashed line represents the environment (T) and the solid line represents the parcel (T'). Below 810 mb and above 530 mb, energy is required to lift the parcel. Above 810 mb and below 530 mb, the parcel accelerates freely. The dark shaded area represents the convective available potential energy (CAPE), while the two lightly shaded areas represent the convection inhibition energy (CINE).

CAPE

The amount of energy available for the upward acceleration of a particular parcel is called the *convective available potential energy* (*CAPE*). On a thermodynamic diagram whose area is proportional to energy (e.g., the emagram; see Section 6.8), *CAPE* is proportional to the area enclosed by the two curves that delineate the temperature of a parcel and its environment, as illustrated by the darker shaded region in Figure 7.2. The amount of *CAPE* of a parcel lifted from a height z (at or above the *LFC*) to the *LNB* is given by the vertical integral of the buoyancy force between these levels

$$CAPE(z) = \int_z^{LNB} g\,\frac{\rho - \rho'}{\rho'}\,dz \qquad (7.24)$$

where the units of *CAPE* are J kg^{-1}. If the environment is in hydrostatic equilibrium we can use (1.26) and (1.33) to obtain

$$CAPE(p) = \int_{p(LNB)}^{p(z)} R_d\left(T_v' - T_v\right)\,d(\ln p) \qquad (7.25)$$

CAPE is defined only for parcels that are positively buoyant somewhere in the vertical profile. The term *convection inhibition energy* (*CINE*) is analogous to *CAPE* but refers to a negative area on the thermodynamic diagram.

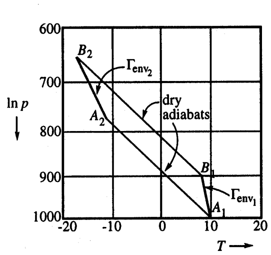

Figure 7.3 An initially stable layer A_1B_1 is made less stable as a result of dry adiabatic ascent.

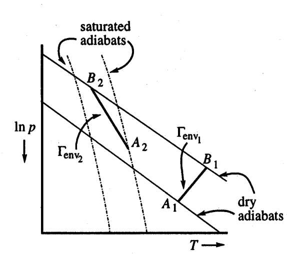

Figure 7.4 Destabilizing an initially stable atmospheric layer. The initially stable and unsaturated inversion layer A_1B_1 is lifted adiabatically. If the bottom of the layer reaches saturation before the top of the layer (as, for example, in an inversion layer in which the mixing ratio decreases with height), further lifting will destabilize the layer. This occurs because the bottom of the layer cools at the much slower saturated adiabatic lapse rate, while the top of the layer continues to cool at the faster dry adiabatic lapse rate.

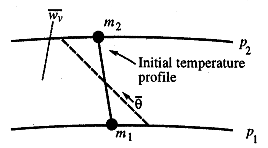

Figure 7.5 Vertical mixing of air parcels, m_1 and m_2, without condensation. Two air parcels, initially at different pressure levels, mix at an intermediate pressure level. The potential temperature of the mixture is a mass-weighted average of the individual parcels' potential temperatures. Mixing of an entire layer results in a constant potential temperature $\overline{\theta}$ throughout the layer. This destabilizes an initially stable layer and stabilizes an initially unstable layer. Because the dry adiabat corresponding to $\overline{\theta}$ does not intersect the average mixing ratio line, $\overline{w_v}$, the mixing process is dry adiabatic and no condensation occurs.

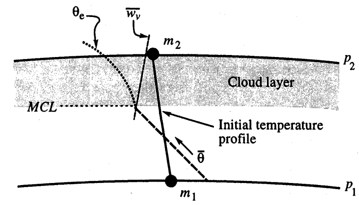

Figure 7.6 Vertical mixing of air parcels, m_1 and m_2, with condensation. If the mixing of two air parcels results in an average potential temperature, $\overline{\theta}$, that intersects the average mixing ratio line, $\overline{w}_v$, then from the level of intersection upwards, condensation will occur and the final temperature distribution will follow a saturated adiabat, θ_e. The lapse rate below the cloud layer moves towards the dry adiabatic lapse rate, while the lapse rate within the cloud layer moves towards the saturated adiabatic lapse rate.

Connecting this course to current research…

Start from two sections of Curry & Webster:

Parameterization of Cloud Microphysical Processes (Section 8.6), pages 241 - 244. Understand the ideas behind the equations on page 242 and note the remarks on page 244.

Cloud-radiation Feedback (Section 13.4), especially the last 2 paragraphs of Section 13.4.1 on pages 368, 369, and the last paragraph of Section 13.4 on page 374.

What do we mean by "parameterization"?

Some physical processes are too poorly understood, and/or they occur on too small space and time scales, so we cannot adequately represent them in global numerical climate models.

For example, clouds and cloud-radiation interactions are important to climate and to modeling climate change.

We don't know how to include these processes correctly, but we cannot afford to omit them.

A working definition of "parameterization":

A parameterization is an algorithm or rule for obtaining the statistical effect, of an ensemble of small-scale processes (e. g., cloud processes), on the large-scale prognostic fields computed explicitly in a model (e. g., wind, pressure, temperature, humidity).

In general, the parameterization must be explicit, in the sense that the statistical effect can be computed as a function of the large-scale variables themselves.

Example: cloud fraction might be a function of humidity.

First, some background.

We define "climate sensitivity" as the equilibrium change in global average surface atmospheric temperature in response to doubling the present atmospheric concentration of carbon dioxide. (There are many alternative definitions).

Many (but not all) reputable models have sensitivities ranging from about 1.5 deg C to 4.5 deg C. This range is an old result. It has not changed in nearly 30 years. Why?

Cloud effects dominate climate sensitivity.

In a typical comparison of coupled models, some of which have the same atmosphere, and some of which have the same ocean, the clear result is that it is the atmospheric model which largely determines the sensitivity, and in particular, it is clouds.

This is another confirmation of something known for a long time. Changing the cloud-radiation scheme in one model replicates the sensitivity spread of many models.

The problem is we don't know which scheme is best.

Cloud algorithms require comprehensive approaches.

Cloud radiative processes are determined BOTH by cloud macrophysics (cloud size, altitude, thickness, etc.) AND by cloud microphysics (water content, phase, particle shapes and size distributions, etc.).

The old procedure of predicting clouds as a simple function of relative humidity, and then assigning radiative properties arbitrarily, is not good enough.

Zeroth-order cloud effects are still unclear.

Models don't agree on even the simplest aspects of cloud changes as climate warms. Some predict cloud amount increases; some predict it decreases.

If in-situ and satellite interpretations can show an observed large-scale secular trend in cloud amount in recent decades, our models must be able to simulate that.

To be believable, climate models must pass tests.

It's unacceptable that models with different sensitivities all manage to reproduce the global mean 20th-century temperature evolution by using dramatically different assumptions on forcing.

The temptation is to treat the 20th-century record of solar variability, volcanism, aerosols and greenhouse gas changes as parameters that can be tuned so the model produces the observed record of surface temperature as a function of time.

Climate sensitivity cannot be only 1 global number

The annual cycle may be masked in average results or thrown away in perpetual-month simulations. In some models, cloud feedback in the longwave (terrestrial) is about the same in all seasons, but the shortwave (solar) cloud feedback actually changes sign over the annual cycle, due to cloud amount changes.

The concept of cloud feedback should also include spatial as well as temporal variability.

It may be a distraction to concentrate on global effects.

We already know that aerosol effects are highly local. Perhaps cloud effects are largely local too.

Even if cloud-radiation effects are not large globally, and we don't know yet if this is true, they may be very important locally.

All this uncertainty about how clouds should be modeled has motivated intense research, and one theme of this research is to observe real clouds.