now, and perhaps also by the effects of a largely water-covered earth (meaning lots of water vapor in the atmosphere).

• ***Volcanic eruptions***. Major individual explosive eruptions inject dust and sulfur gases (especially sulfur dioxide) into the stratosphere, the latter forming sulfuric acid droplets.

Equatorial eruption plumes spread into both hemispheres, whereas plumes from eruptions in mid-to high latitudes are confined to that hemisphere. Observational evidence from the past 100 years emonstrates that major eruptions can be associated with global averaged cooling of several tenths of a degree C in the year following the event and much larger changes on a regional to hemispheric basis.

The cooling is primarily from the sulfuric acid droplets which reflect solar radiation. Dust also causes surface cooling by absorbing solar radiation in the stratosphere, but compared to the sulfuric acid these effects are short-lived (weeks to months) Stratospheric aerosols may also cause brilliant sunsets

• ***Human-induced*** *changes in atmospheric composition and land cover*. The effect of greenhouse gases such as carbon dioxide and methane on the radiation budget has already been introduced. The observed buildup of these gases since the dawn of the industrial age represents a positive forcing. Human activities have also led to a buildup of tropospheric aerosols, which induce a partly compensating cooling. Changes in land use and land cover have also led to a small increase in surface albedo that promotes cooling.

Climate feedbacks

Building on the framework of radiative forcing, consider further the change in global average surface temperature resulting from increasing the atmospheric concentration of carbon dioxide. As just discussed, because of the imposed perturbation, more of the longwave radiation emitted upward from the surface is absorbed by the atmosphere, and directed back towards the surface. The result is a radiation imbalance at the top of the atmosphere – net solar radiation entering the top of the atmosphere exceeds the longwave loss to space. The climate forcing from adding carbon dioxide is hence positive. Now consider the feedbacks. The most important of these is the water vapor feedback. Warming results in more evaporation, and a warmer atmosphere can carry more water vapor. However, water vapor is also a greenhouse gas, so it causes further warming. Some of the earth's snow cover and sea ice will melt, reducing the earth's surface albedo, also causing further warming. These are examples of positive feedbacks, as they amplify the global surface temperature change induced by the climate forcing. If the carbon dioxide concentration in the atmosphere were lowered, thereby imposing a negative climate forcing, the positive feedbacks would foster further cooling.

A fascinating aspect of the global climate system is that positive feedbacks dominate. For example, one of the responses to increasing greenhouse gases could be an increase in cloud cover, which through increasing the planetary albedo would represent a negative feedback. However, this and other potential negative feedbacks would only appear to be capable of slowing the rate of warming, not reversing it.

While climate feedbacks can be either positive or negative, they can also be broadly differentiated regarding how quickly they operate. In the framework of global radiative forcing appropriate to understanding human-induced global climate change, it is the fast feedbacks which are relevant.

The most important are changes in water vapor and albedo (mentioned above). Both can operate over timescales of days and even less. Cloud cover can also change very quickly (hours). Examples of slow feedbacks are changes in the extent of continental ice sheets (influencing planetary albedo) and greenhouse gases during the Pleistocene in response to Milankovich periodicities. Records from ice cores show that these glacialinterglacial cycles were nearly coincident with fluctuations in both atmospheric carbon dioxide (±50ppm) and methane (±150ppb).

Climate response

How much does the global mean surface temperature change in response to a radiative forcing of a given magnitude? How long does it take for the change to occur? These are among the most important, pressing questions in climate change science.

The first question deals with the issue of equilibrium climate sensitivity. In the IPCC framework, equilibrium climate sensitivity is the equilibrium change in annual mean global averaged surface air temperature following a doubling of the atmospheric equivalent carbon dioxide. Doubling the carbon dioxide concentration equates to a radiative forcing (top of atmosphere radiation imbalance) of about $4 W\ m^{-2}$. In response to this doubling the surface and atmosphere would warm up. Eventually, radiative balance would be restored again with a new and higher surface temperature. Estimates of equilibrium climate sensitivity obtained from the current generation of global climate models range from $2–4.5^{o}C$, with a best estimate of $3.0^{o}C$. The uncertainly lies largely in the spread of model estimates of the climate feedbacks, particularly in the cloud feedbacks. Cloud feedbacks are complex and hard to model. Negative feedbacks may operate when increased global heating leads to greater evaporation and greater amounts of highaltitude cloud cover, which reflect more incoming solar radiation. However, other types of clouds, and clouds in the polar regions, can induce surface warming Expressed in a more convenient fashion, the best estimate of 3^oC for carbon dioxide doubling equates to 0.75oC global mean surface temperature increase per $W\ m^{-2}$ of forcing. It is stressed that the climate simulations used to obtain these sensitivity numbers only deal with the fast feedbacks. If there were no feedbacks present in the climate system, the climate sensitivity would be only about 0.30^oC per $W\ m^{-2}$.

While equilibrium climate sensitivity in the IPCC framework is based on a doubling of atmospheric equivalent carbon dioxide, it appears that the equilibrium temperature response to any radiative forcing is roughly the same. This is an important concept, since it means that to a first approximation, one can linearly add different forcings to obtain a net value from which an equilibrium temperature change can be estimated. It also appears that most of the equilibrium temperature response to a radiative forcing with the fast feedbacks at work occurs over a time span of 30 to 50 years. Most of the time lag is due to the large thermal inertia of the oceans.

The basic issue is that the oceans can absorb and store a great deal of heat without a large rise in the surface (radiating) temperature. Consider what is happening in response to the current radiative forcing from human activities of $1.6Wm^{-2}$. Using the equilibrium climate sensitivity of 0.75 implies that this radiative forcing, if maintained, will eventually yield about 1.2^oC of warming. Over the instrumental record, the global mean temperature has risen by about 0.7^oC, implying another 0.5^oC remaining after the ocean sufficiently heats up. How much has the heat content of the ocean already increased? Based on available hydrographic data from 1955– 998, the world ocean between the surface and 3000m depth gained ~1.6 × 1022 J. Compared with atmospheric kinetic energy (p. 70), this is a very large number.

The Climatic record

1 The geological record

Understanding the significance of climatic trends over the past 100 years requires that they be viewed against the backdrop of earlier conditions. On geological timescales, global climate has undergone major shifts between generally warm, ice-free states and Ice Ages with continental ice sheets. There have been at least seven major Ice Ages through geological time. The first occurred 2500 million years ago (Ma) in the Archean period, followed by three more between 900 and 600Ma, in the Proterozoic. There were two Ice Ages in the Paleozoic era (the Ordovician, 500–430Ma; and the Permo-Carboniferous, 345– 225Ma). The most recent Ice Age began about 34Ma in Antarctica at the Eocene/Oligocene boundary and about three million years ago in northern high latitudes. At present, we are considered to be still within this most recent Ice Age, albeit in the warm part of it known as the Holocene, which began about 11.5ka. While the total volume of land ice today (mostly comprising the Antarctic and Greenland ice sheets) is certainly much smaller than it was at 20ka, it is still substantial compared to other times of the earth's past.

Major Ice Ages and ice-free periods can be linked to a combination of external and internal climate forcing (plate tectonics, greenhouse gas concentrations, solar irradiance). The ice sheets of the Ordovician and Permo-Carboniferous periods formed in high southern latitudes on the former mega-continent of Gondwanaland. Uplift of the western cordilleras of North America and the Tibetan Plateau by plate movements during the Tertiary period (50–2Ma) caused regional aridity to develop in the respective continental interiors. However, geographical factors are only part of the explanation of climate variations. For example, warm high-latitude conditions during the mid- Cretaceous period, about 100Ma, may be attributable to atmospheric concentrations of carbon dioxide three to seven times higher than at present, augmented by the effects of alterations in land–sea distribution and ocean heat transport.

Much more is known about ice conditions and climate forcings through the Quaternary, which began about 2.6 million years ago, comprising the Pleistocene (2.6Ma–11.5 ka) and the Holocene (11.5ka-present) epochs. It is abundantly clear that this most recent Ice Age we live in was far from being uniformly cold. Instead it was characterized by oscillations between glacial and interglacial conditions. Eight cycles of global ice volume are recorded in land and ocean sediments during the last 0.8–0.9Ma, each averaging roughly 100ka, with only 10 percent of each cycle as warm as the twentieth century. Each glacial period was in turn characterized by abrupt terminations.

Because of reworking of sediments, only four or five of these glaciations are identified from terrestrial records. Nevertheless, it is likely that all were characterized by large ice sheets covering northern North America and northern Europe. Sea-levels were also lowered by about 130m due to the large volume of water locked up in the ice. Records from tropical lake basins show that these regions were generally arid at those times. Prior to 0.9Ma the timing of glaciations is more complex. Ice volume records show a dominant 41ka periodicity, while ocean records of calcium carbonate indicate fluctuations of 400ka.

These periodicities are linked to the Milankovich forcings discussed earlier. The precession signature (19 and 23ka) is most apparent in lowlatitude records, whereas that of obliquity (41ka) is represented in high latitudes. However, the 100ka orbital eccentricity signal is generally dominant overall.

The basic idea is that onset of glacial conditions is initiated by Milankovich forcings that yield summer cooling over the northern land masses. This favors survival of snow cover through summer, a feedback promoting further cooling and ice sheet growth, leading to even further cooling through slow feedbacks in the carbon cycle discussed earlier. Onset of an interglacial works the other way, with Milankovich forcings promoting initial warming over the northern land masses, setting feedbacks into motion to give further warming and ice melt.