P

PLATE TECTONICS AND CLIMATE CHANGE

The horizontal and vertical displacements associated with plate tectonics play a fundamental role in climate change over a wide range of timescales. The solid-earth surface is in direct contact with the atmosphere and oceans and its evolving character affects balances of incoming and outgoing radiation, atmospheric circulation, ocean currents, and the location of elevated terrain suitable for glaciers and ice sheets. Tectonic processes also have important indirect climatic effects through their control on geochemical cycling and the composition of the atmosphere and ocean. This entry provides an introduction to the more direct, physical effects of tectonics on the climate system. While touched on briefly, the less direct climatic effects of volcanism and chemical processes are left to other entries in this volume. For a historical perspective on many of the ideas presented here, the interested reader should refer to the comprehensive survey by Hay (1996). Other excellent resources include an in-depth discussion of climate forcing mechanisms and climate modeling strategies relevant to tectonic timescales (Crowley and North, 1996), and edited volumes focusing on the climatic and geochemical effects of mountain uplift (Ruddiman, 1997) and uncertainties related to tectonic reconstructions and climate forcing (Crowley and Burke, 1998).

Shifting continents

While the German astronomer and climatologist Alfred Wegener (1880–1930) is usually credited with founding the modern theory of plate tectonics, the concept of a fundamental link between climate and the changing distribution of continents and oceans dates back at least to Lyell. He speculated that a concentration of polar land area should cool the Earth and that changing land distributions were somehow responsible for the deposition of Carboniferous coal beds and the obvious differences between Mesozoic and Cenozoic fossil assemblages (Lyell, 1830). Today, we acknowledge that tectonic processes can indeed have local to global-scale climatic impacts. Those impacts can be direct or indirect, and can ~~include~~ horizontal displacements, vertical displacements, or both.

Perhaps the most fundamental, direct link between plate tectonics and climate comes from the slowly evolving global distribution of continental blocks and fragments (Figure P64). As tectonic plates move, so do the subaerial continents and this redistribution of landmass has an important effect on the spatial heterogeneity of the Earth's energy balance via differences in the albedos (reflectivities) and thermal properties of land versus ocean.

Albedo and thermal properties of land and sea

As a whole, the albedo of today's Earth is ~0.31, with the atmosphere and clouds responsible for most (about 85%) of the energy reflected back to space. While the atmosphere and clouds dominate planetary albedo (see entry on *Albedo feedback*), the surface is the dominant absorber of solar energy, responsible for ~65% of the total absorbed solar radiation and for transferring energy to the atmosphere through long-wave radiation, and fluxes of sensible and latent heat. Consequently, variations in surface albedo have important effects on atmospheric dynamics and climate. This is particularly true in mostly cloud-free regions and polar latitudes, which on today's Earth are covered by highly reflective snow and ice. Surface albedos over open, ice-free ocean are generally much lower than those over land (Table P5). Thus, at the global scale, changes in the latitudinal distribution of land can have a significant effect on zonally-averaged net radiation balance. For example, an Earth with polar continents covered by perennial snow and ice will have higher surface albedo than an Earth with a polar geography dominated by open water. Because the need for poleward heat transport (the ultimate driver of winds and ocean currents) is determined by the latitudinal net radiation gradient, major changes in the distribution of continents are likely to have significant climatic consequences.

Like albedo, the thermal properties of open water are also very different from those of land. Water has very high specific heat (about five times greater than soil and rock) and the upper ~10–150 m of the oceans are well mixed by winds and convection. This allows seasonal temperature changes to penetrate much deeper into the ocean (tens to hundreds of meters) than into immobile soil and rock on land (~1 m, by conduction). Consequently, sea surface temperatures (SSTs)

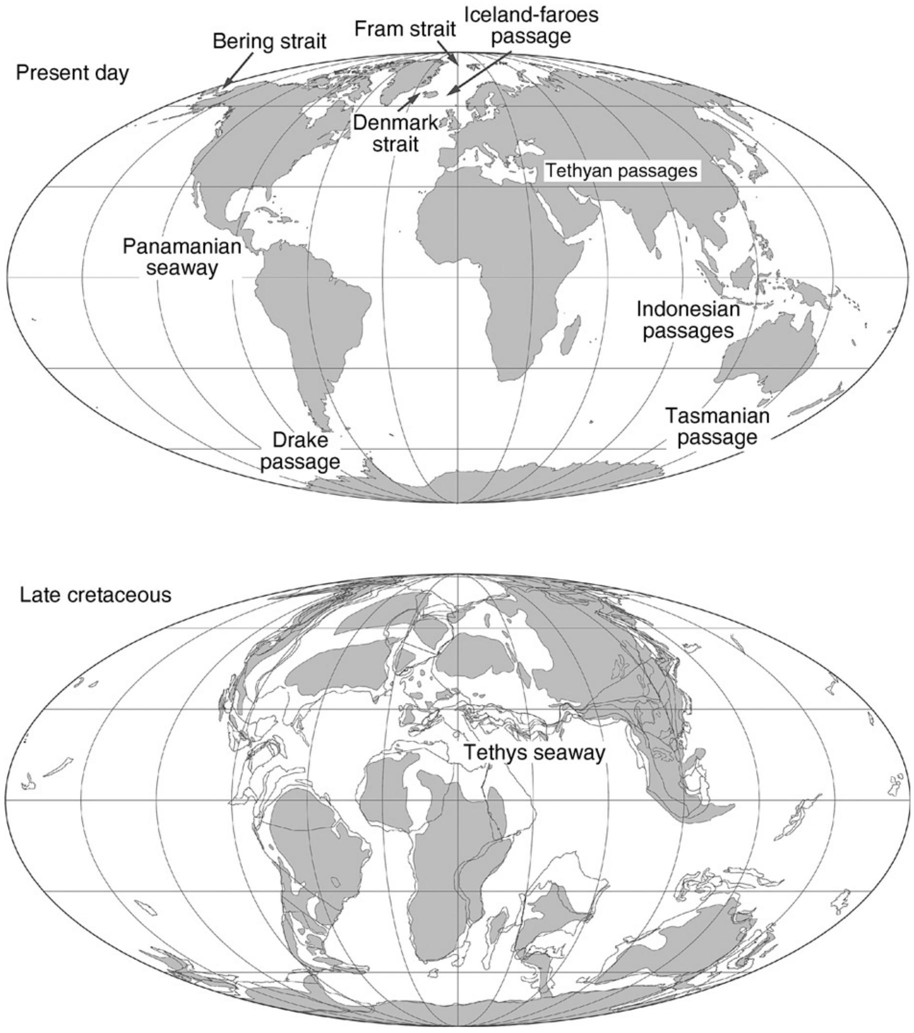


Figure P64 A comparison of modern and late Cretaceous (80 Ma) geography (after DeConto et al., 1999). The approximate locations of Cenozoic gateways discussed in the text are shown on the present day map (*top*). The outlines of tectonic blocks and subaerial land areas (*shaded*) are shown on the Late Cretaceous map (*bottom*). Note the low latitude, circum-global ocean passage (Tethys Seaway), closed Southern Ocean gateways, narrow Atlantic Basin, and flooded continental interiors characteristic of the mostly ice-free Cretaceous and early Cenozoic.

are relatively slow to respond to seasonal changes in insolation. In the hemisphere experiencing winter, relatively warm surface waters suppress extreme temperature swings and provide the atmosphere with a source of moisture and diabatic heating. In contrast, the smaller heat capacity of land combined with relatively high albedo allows much greater seasonality, particularly in the interiors of the larger continents (Figure P65).

Monsoons and the intertropical convergence zone (ITCZ)

The seasonal thermal contrast between land and ocean mentioned above is generally considered to be responsible for the well-known monsoonal circulation systems driven by seasonally alternating low-level pressure patterns over land and sea. The best example on today's Earth is the Asian monsoon system (Figure P65). Over central Asia, seasonal mean temperature differences can exceed 50 °C. During boreal winter, cold (dense), sinking air contributes to high atmospheric surface pressure (the Siberian High). Lower atmospheric pressure over the warmer Indian Ocean produces a meridional pressure gradient and northeasterly flow over southeastern Asia, Indonesia and the northern Indian Ocean. During the summer months, continental heating reverses the pressure gradient and the regional wind field. The resulting southwesterly winds (Southwest Monsoon) bring moisture-laden air and rain northward across India, the foothills of the Himalaya, and parts of China and Indonesia. This seasonal reversal of the wind field also produces dramatic changes in the dominant Indian Ocean currents. For example, during boreal summer, a strong low-level atmospheric jet, in part deflected northeastward by east African highlands, drives the northeastward flowing Somali Current, vigorous coastal upwelling, and high ocean productivity along the coasts of Somalia and Oman. During the Northeast Monsoon, the Somali Current reverses, primary productivity in the Arabian Sea slows, and the entire Indian Ocean equatorial current system reorganizes.

Table P5 Albedos of different cloud and surface types

Cloud/surface type	Range of values	Typical values
Cumulonimbus clouds		0.9
Stratocumulus clouds		0.6
Cirrus clouds	0.4–0.5	0.45
*Water (low wind)	0.02–0.12	0.07
Water (high wind)	0.10–0.20	0.12
Bare land surfaces		
Moist dark soil	0.05–0.15	0.10
Moist gray soil	0.10	0.15
Dry soil	0.20–0.35	0.20
Wet sand	0.20–0.30	0.25
Dry light sand	0.30–0.40	0.35
Vegetation		
Low canopy vegetation	0.10–0.20	0.17
Dry vegetation	0.20–0.30	0.25
Evergreen forest	0.10–0.20	0.12
Deciduous forest	0.15–0.25	0.17
Forest with snow cover	0.20–0.35	0.25
Sea ice	0.25–0.40	0.30
Sea ice (snow covered)	0.40–0.90	0.70
Old, wet snow	0.40–0.65	0.50
Old, dry snow	0.60–0.75	0.70
Fresh, dry snow	0.70–0.90	0.80

Source: Compiled from Barry and Chorley (1998); Hartmann (1994); Houghton (1985).
*At the low solar zenith angles typical of tropical latitudes, the albedo of calm water can be as low as ~0.02, much lower than typical land-surface albedos. Albedos over open water increase sharply as solar zenith angles approach 90°.

While definitions of the monsoons and the ITCZ are somewhat intertwined, the ITCZ is traditionally considered to be the region of low-level convergence and convective precipitation near the equator, where the trade winds meet. As the seasons progress, uneven interhemispheric heating moves the mean latitudinal position of the ITCZ into the hemisphere experiencing summer (Figure P65). This seasonal shift in the ITCZ is exaggerated over the continents. On today's Earth, the Northern Hemisphere contains a greater fraction of total land area, which contributes to the average position of the ITCZ being several degrees north of the equator, as does the orientation of the coastlines of western tropical Africa and South America (Philander et al., 1996). Because the northeast and southeast trade winds generally converge north of the equator, the Southeast Trades can cross the equator where the effects of the Earth's rotation are minimal. This has important consequences for the latitudinal position of the major zonal equatorial ocean currents and zones of oceanic convergence, divergence, and upwelling, which, like the winds, are also asymmetric with respect to the equator (Gill, 1982).

Alternatively, monsoons have been described as convergence zones more than 10° away from the equator that do not necessarily require a strong land-sea thermal contrast (Chao and Chen, 2001). In this interpretation, the Earth's rotation is considered the primary cause of the monsoons, with land-sea contrast playing a lesser, modifying role. This notion is supported by numerical climate modeling studies showing the existence of monsoons even in the absence of continents. In simulations with the Asian and Australian continents, which are replaced by open ocean, monsoonal circulation patterns still develop over roughly the same regions as today's Asian, Indian and Australian monsoons occur. Land-sea contrast appears to play a more fundamental role in monsoons over Africa, South America and Mexico, however (Chao and Chen, 2001).

Continentality

At times in the geologic past, episodic convergence of major continental plates formed giant supercontinents (e.g., Rodinia, Pangea, Gondwana, and Laurasia). Enhanced *continentality* likely produced extreme seasonal temperature swings and aridity in the continental interiors, and invigorated monsoonal circulation patterns in many coastal locations (Crowley et al., 1989; Kutzbach and Gallimore, 1989). The Permo-Triassic convergence of Laurasia and Gondwana provides a well-known example of a single supercontinent, *Pangea,* and a giant *Panthalasia Ocean*. In central Pangea, mean summer temperatures were at least 6–10 °C warmer than today's continental interiors, with daytime high temperatures reaching 50 °C in some locations (Crowley and North, 1996). Extreme aridity dominated the continental interiors because most atmospheric water vapor would have been lost to precipitation near the continental margins and because high summer temperatures would have increased evaporation. This continental aridity effect would have been exacerbated downwind of mountain ranges (see *Rain shadows* below). Estimates of net precipitation minus evaporation over Pangea are only about half that of modern land areas and these conditions must have played an important role in the distribution of terrestrial ecosystems (Crowley and North, 1996).

Pangean climate simulations using global climate models (GCMs) exhibit enhanced land-ocean pressure gradients, resulting in "mega-monsoons" (Kutzbach and Gallimore, 1989) analogous to today's Asian monsoon system described above, albeit more severe. Such mega-monsoons were likely strongest in the hemisphere with more land area in the subtopics (Gibbs and Kump, 1994) and would have been highly sensitive to the location of mountains and high plateaus (Hay and Wold, 1998). Monsoonal circulation systems have also been shown to be sensitive to orbital forcing (see *Astronomical theory of climate change*), with the summer monsoon becoming invigorated in the hemisphere experiencing increased insolation (Kutzbach and Liu, 1997). This sensitivity was likely enhanced over the largest continents and may account for some of the rhythmic sedimentary sequences (cyclothems) seen in the late Paleozoic and early Mesozoic sedimentary record.

Extreme seasonality associated with large continents also affects the potential for widespread glaciation. While extreme cold winters provide many opportunities for winter snow accumulation, summer ablation is the critical factor in maintaining perennial snow cover leading to the growth of ice sheets. Increased seasonality and a lack of precipitation in the interior of ancient supercontinents would have been generally unfavorable for the growth of continental-scale ice sheets in mid-high latitudes (Crowley et al., 1989), unless accompanied by global cooling and/or uplift. For example, the Carboniferous glaciation of the Gondwanan supercontinent is thought to coincide with a time of broad epeirogenic uplift (Gonzalez-Bonorino and Eyles, 1995) combined with relatively low levels of atmospheric CO_2 (Berner and Kothavala, 2001), an important contributor to the heat-trapping greenhouse effect. In more recent times, the Cenozoic separation of Southern Hemisphere

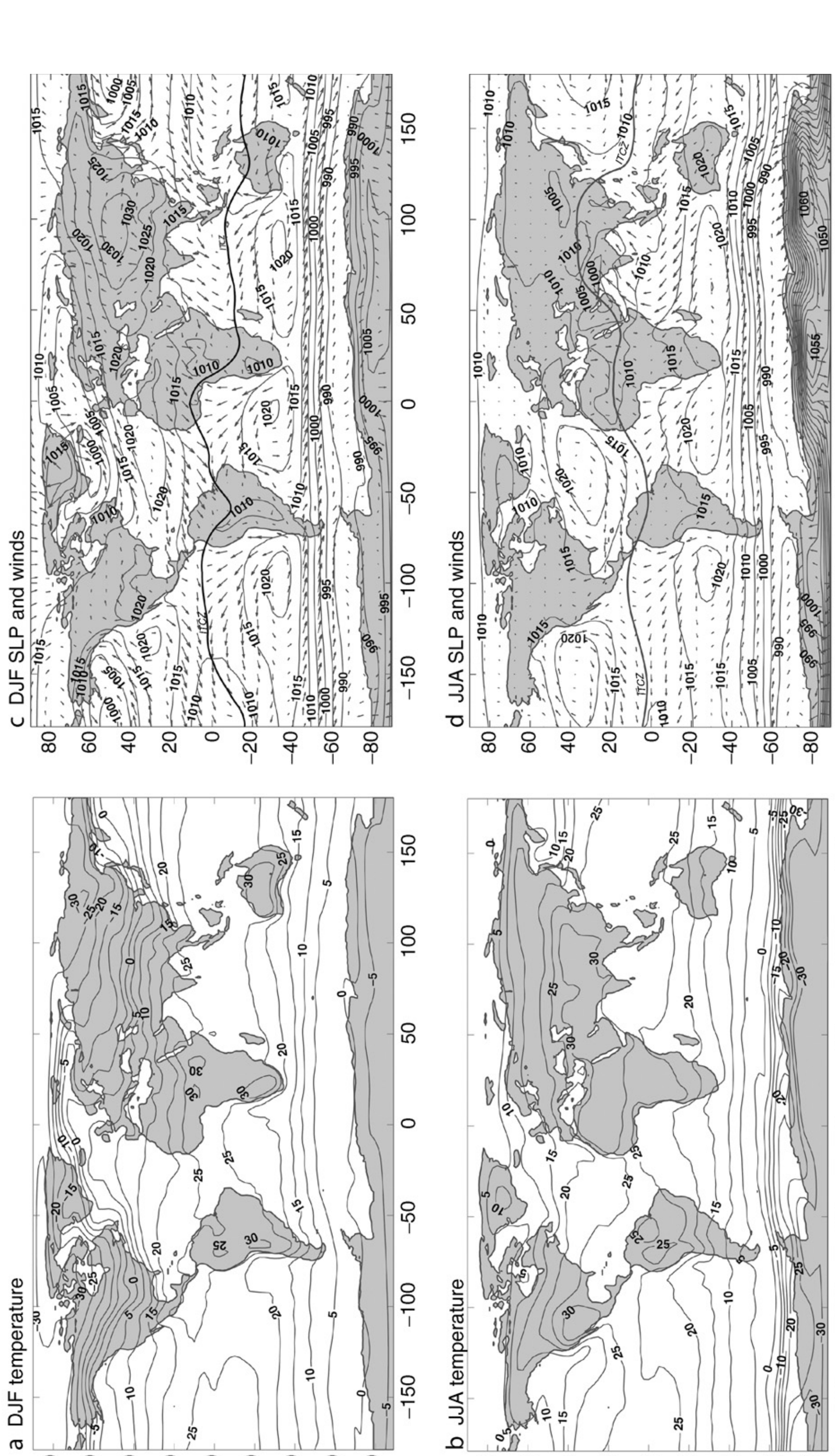


Figure P65 Seasonal surface air temperature (**a** and **b**), sea level pressure, and winds (**c**, **d**). Modern DJF (December, January, February) and JJA (June, July, August) averages are calculated from 1971–2000 NCEP/NCAR reanalysis data. Seasonal temperatures (**a**, **b**) are shown in °C with 5 °C contours. Sea level pressure (**c**, **d**) is shown in hPa with 5 hPa contours. The *thick black line* (**c**, **d**) shows the approximate seasonal location of the ITCZ (see text for discussion). *Vectors* show wind velocity, with the longest vectors representing winds of ~13 m s^{-1}.

continents and associated reductions in continentality and seasonality, along with declining CO_2, may have contributed to the summer cooling requisite for the first extensive glaciation of Antarctica in the Paleogene (DeConto and Pollard, 2003a,b; Oglesby, 1991).

The polar latitudes' sensitivity to geography is clearly illustrated by the very different climatic regimes of today's north versus south polar regions. Increasing concentration of land area in high northern latitudes has long been considered a contributor to the overall global cooling trend through the Cenozoic, eventually culminating in Northern Hemisphere glaciation (Crowell and Frakes, 1970; Donn and Shaw, 1977). A number of early paleoclimate modeling studies using GCMs explored the potential global impacts of latitudinal shifts in land area (Barron et al., 1984; Hay et al., 1990a). Like the simulations shown in Figure P66, these studies showed that an Earth with a concentration of equatorial continents and open polar oceans is indeed warmer and has lower equator-to-pole temperature gradients than an Earth with a concentration of polar continents and an equatorial ocean (Figure P66). Subsequent modeling studies focusing on the extreme warmth of the Cretaceous period (Barron et al., 1993) showed that the effect of paleogeography alone does not account for all of the warmth characteristic of much of the Mesozoic and early Cenozoic. This work concluded that some other climate forcing factor in addition to paleogeography (probably elevated concentrations of atmospheric CO_2) must have contributed to the warmth of the Cretaceous and other warm periods in Earth history, as speculated a century earlier (Arrhenius, 1896). While the models used in these studies were crude by current standards, they showed that changes in continental configuration could produce significant climate change, albeit smaller than the full range of climate variability recognized in the geologic record. They also showed that small tectonic changes associated with the movement of individual continental blocks and fragments have only limited, local effects (Hay, 1996) and could only produce major climatic change if combined with amplifying feedbacks.

In addition to latitudinal displacements, the evolving zonal (east-west) distribution of land can also have important climatic consequences. In tropical latitudes, the modern pan-Pacific atmospheric pressure pattern is dominated by the *Walker Circulation*. The Walker Circulation is driven by the longitudinal distribution of diabatic heating over land and sea, with Africa, South America, and the warm waters surrounding Indonesia providing sources of heating. On today's Earth, easterly trade winds drive warm surface waters westward, forming the Western Pacific Warm Pool, where surface waters "pile-up" against the Indonesian archipelago. Consequently, the equatorial thermocline (the layer in the ocean below the surface mixed layer where temperature decreases rapidly with depth) slopes down toward the west and shoals on the eastern margin of the basin to produce the $\sim 5\,^{\circ}C$ zonal SST difference across the Pacific. On the warmer, Indonesian side of the Pacific, latent and diabatic heating contributes to convection, low surface pressure, low-level convergence, and high pressure aloft. The upper-level, westerly pressure gradient is balanced by a sinking branch of the *Walker Circulation* and corresponding high surface pressure near the west coast of South America – a region typically dominated by upwelling and relatively cool sea surface temperatures.

Thus, the distribution of landmasses and restriction between the Pacific and Indian Oceans provided the basic framework for the Walker Circulation and related atmosphere-ocean dynamics like ENSO (El Niño Southern Oscillation – the dominant mode of modern Pacific climate variability). During an El Niño event, the western tropical Pacific cools and the core of warmest surface waters moves eastward, inhibiting upwelling along the west coast of South America. This zonal redistribution of SSTs perturbs the Walker Circulation and this is reinforced by atmosphere-ocean feedbacks involving the trade winds (Bjerknes, 1969). The 2–7-year quasi-periodicity of the modern ENSO cycle is, in part, modulated by equatorially-bound Kelvin waves in the ocean's interior, propagating from west to east in the upper thermocline, and westward traveling Rossby waves. On the eastern side of the basin, where the thermocline is already near the surface, these long waves can have a significant impact on SSTs and hence the atmosphere (Battisti and Hirst, 1989).

The long-term evolution of the Warm Pool has likely been influenced by the progressive closure of the eastern Tethys Sea (Figure P64) and restriction of the Indonesian Seaway through the Neogene (Cane and Molnar, 2001). While the

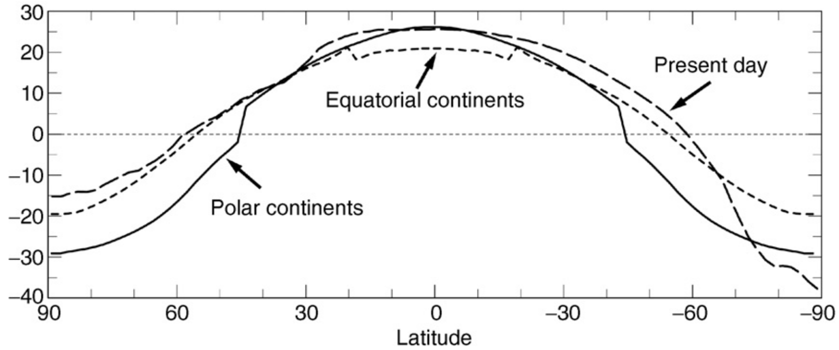


Figure P66 Zonally-averaged, mean annual surface air temperature over the modern Earth (*long dashes*) compared with GCM simulations of a planet with two polar continents (*solid line*) and a planet with a single equatorial supercontinent (*short dashes*). The climate simulations used the current (2008) version of the GENESIS V.3 GCM (Thompson and Pollard, 1997), preindustrial CO_2 (280 ppmv), and a mean orbital configuration (zero eccentricity and an obliquity of 23.5°). Total land areas in the polar and equatorial geographies are equal and close to modern. All land points were assigned an elevation of 800 m and a generic (mixed deciduous and evergreen) vegetation. Note the similarity of north polar temperatures in the modern versus equatorial continent (polar ocean) scenarios, and south polar temperatures in the modern versus polar continents scenarios.

character and timing of long wave-modulated atmosphere-ocean dynamics are expected to be different in a world with a wider Pacific basin, climate model simulations and geological evidence show the presence of ENSO-like variability as early as the Eocene (Huber and Caballero, 2003). This would suggest atmosphere-ocean oscillations like ENSO are robust features of the climate system that operate over a wide range of paleogeographic boundary conditions.

Eustasy

Tectonically-driven changes in global mean sea level (eustasy) and the associated flooding or exposure of low-lying continental areas also have important climatic consequences. The effects mainly stem from the changes in albedo and surface-atmospheric heat and moisture exchange associated with sub-aerial versus water-covered surfaces, although indirect effects on the marine carbon cycle and atmospheric CO_2 may also be important (Gibbs and Kump, 1994). Over 10^6-year and longer time scales, eustasy is thought to be dominated by tectonic influences on the volume of the ocean basins (Hays and Pitman, 1973). Increases in the total length of mid-ocean ridges, high sea floor spreading rates producing warm, lower-density ocean crust, and the emplacement of Large Igneous Provinces (LIPs) can all displace ocean water and raise sea level (Kominz, 1984). Increased spreading rates have also been associated with increased volcanic outgassing and high levels of atmospheric CO_2 (Larson and Erba, 1999), although the linkages between sea floor spreading and atmospheric composition remain equivocal (Conrad and Lithgow-Bertelloni, 2007; Rowley, 2002).

The Cretaceous period offers one of the best examples of a warm “greenhouse” climate during a time of high sea level, when a combination of mostly ice-free poles and high sea floor spreading rates produced sea levels $\sim$100 m higher than today (Haq et al., 1987; Larson, 1991). More than 20% of the continents were flooded (Figure P64), forming vast epicontinental seas (Hay et al., 1999). Climate modeling studies (Barron et al., 1993; DeConto et al., 1999; Otto-Bliesner et al., 2002) have shown that the combination of high greenhouse gas concentrations, continental positions, and high sea levels all contributed to the overall warmth of the Cretaceous. The ameliorating effects of open water in epicontinental seas and inland lakes likely reduced seasonality and contributed to the apparent winter-warmth of many continental locations (Sloan, 1994; Valdes et al., 1996). As discussed below, however, a satisfactory explanation for the extreme warmth of the polar regions during the Cretaceous and other warm climate intervals remains elusive.

Ocean gateways

The oceans transport vast amounts of water, heat, and salt across entire ocean basins and from low to high latitudes. As the ocean basins and the *gateways* between them evolve over tectonic timescales, so does ocean circulation. These tectonically-forced changes in ocean circulation have long been thought to play a key role in some of the major climatic events and transitions recognized in the geological record. While the timing of some tectonic gateway events broadly correspond with major paleoenvironmental changes (e.g., the ocean anoxic events (OAEs) of the Cretaceous (Leckie et al., 2002), the onset of Antarctic glaciation in the earliest Oligocene (Kennett, 1977; Livermore et al., 2004), and the onset of Northern Hemisphere glacial cycles in the Pliocene (Haug and Tiedemann, 1998)), the actual role of the ocean in these changes remains equivocal and likely involves a complex web of both direct and indirect effects.

Ocean heat transport

The link between tectonically forced changes in ocean circulation and global climate is usually attributed to changes in the ocean’s contribution to poleward heat transport (Covey and Barron, 1988; Rind and Chandler, 1991). While modern estimates of atmospheric and oceanic heat transport remain poorly constrained, it is generally believed the oceans contribute about half of the total heat transport required to maintain the Earth’s meridional energy balance. Because of the large equatorward latent heat flux associated with the lower limb of the Hadley circulation, the atmosphere contributes little heat transport out of the tropics, where the oceans do most of the work (maximum poleward ocean heat transport of about 2×10^{15} watts occurs at about 20–25° North and South (Peixoto and Oort, 1992; Trenberth and Solomon, 1994)). While ocean circulation contributes little direct poleward heat transport in high latitudes, it plays an important role in polar climate via its influence on atmospheric teleconnections to the tropics (Cane and Evans, 2000) and its control on seasonal distributions of sea ice, which affects albedo and energy transfer between the ocean and atmosphere (Rind et al., 1995).

Because changes in the physiognomy of ocean basins and/or the opening or closure of gateways alters both the wind-driven (surface) and density-driven (deep) components of the ocean’s meridional overturning (Bice et al., 1998; Poulsen et al., 2001), it may be reasonable to assume that tectonically-forced changes in ocean circulation can have profound climatic consequences (Covey and Barron, 1988). Conversely, theoretical arguments suggest the potential effects of ocean circulation on total poleward heat transport are inherently limited. It has been hypothesized that the total energy transport by the atmosphere-ocean system remains roughly unchanged, so if the efficiency of either the atmosphere or ocean is reduced, the other compensates (Stone, 1978). Furthermore, ocean heat transport is proportional to the product of the temperature change and mass transport of water advected into a given region. Thus, the ocean’s potential to transport heat would have been limited during times in the past when the temperature difference between low and high latitudes (and surface and deep waters) was much smaller than today, such as during the Cretaceous and early Eocene (Hay and DeConto, 1999; Sloan et al., 1995).

The limited potential of the oceans to maintain the warm, ice free polar conditions characteristic of most of the Phanerozoic is generally supported by numerical climate model simulations of paleoclimates using Mesozoic and early Cenozoic paleogeographies and high greenhouse gas concentrations (Brady et al., 1998; Huber and Sloan, 2001; Otto-Bliesner et al., 2002). While these simulations do produce relatively warm climates with somewhat reduced equator-to-pole temperature gradients relative to today, they fail to produce the dramatic increases in ocean heat transport required to explain all of the polar warmth of these intervals, provided the oceans were the primary mechanism for polar warming (Lyle, 1997). Several alternative mechanisms have been proposed to account for extreme polar warmth during these periods, including increased atmospheric latent heat transport (Hay and DeConto, 1999), increased tropical cyclone activity (Emanuel, 2001) and polar stratospheric clouds (Sloan and Pollard, 1998), although the potential climatic effects of these mechanisms are speculative.

The underestimation of polar warmth in model simulations of these ancient climates remains an important problem, because it implies that climate model simulations of future climates may be underestimating future polar warming in response to anthropogenic increases in greenhouse gas concentrations.

While the global-scale impact of tectonically forced changes in ocean circulation is debatable, the breakup of Pangea certainly had profound effects on both the wind-driven current system and location(s) of deepwater formation. Such changes likely had important regional climatic effects that could have triggered indirect forcing mechanisms with global consequences. For example, tectonically forced changes in ocean circulation can impact atmospheric CO_2 via changes in ocean overturning, primary productivity, and the biological pump. Futhermore, changes in ocean circulation contributing to deep sea warming have the potential to trigger the release of methane (another important greenhouse gas that oxidizes to form CO_2) from frozen, temperature-sensitive gas hydrates (clathrates) stored in deep sea sediments. Such a feedback mechanism has been proposed to have contributed to the dramatic global warming event known as the Paleocene-Eocene Thermal Maximum (Dickens et al., 1997).

The oldest surviving ocean crust is less than 200 million years old and detailed paleogeographic reconstructions become increasingly difficult to assemble with increasing age (Crowley and Burke, 1998). Despite the inherent limitations, the timing of major late Mesozoic and Cenozoic gateway events are becoming better constrained, including the opening of the North and South Atlantic basins, the closure of the ancient Tethys Ocean, the opening of the Southern Ocean gateways, the restriction of the Indonesian Seaway, and the closure of the Panamanian Isthmus (Figure P64). Among these, the closure of the Tethys and the opening of the Southern Ocean may be the most profound because they produced the late Cenozoic world as we know it today – with a single high-latitude circum-global passage (Southern Ocean) rather than a tropical circum-global passage (Tethys Ocean) as existed during the relative global warmth of the late Mesozoic and early Cenozoic.

Tethys Ocean

The Tethys (Figure P64) formed sometime in the late Jurassic, reached its zenith during the Cretaceous, and remained at least partially open until the Miocene (Hay et al., 1999). The modern Mediterranean Basin is the last remaining expression of the ancient Tethys, which both reduced Eurasian continentality and provided a low latitude ocean passage between the Indian and Atlantic Ocean basins. The Cenozoic retreat of the Tethys and Paratethys Seas, and the associated increase in Eurasian continentality and aridity, may have had climatic effects comparable in magnitude to the uplift of the Himalaya and Tibetan Plateau (Ramstein et al., 1997).

Like the modern Mediterranean, Tethyan (sub-tropical) latitudes were likely dry and dominated by strong net evaporation. Relatively warm and saline (high density) deep-water masses could have formed there. Given the size of the ancient Tethys, such water masses have been proposed to have driven a thermohaline circulation system essentially the opposite of today's, with warm, but dense (high salinity) deep waters sinking in low latitudes and flowing polewards (Brass et al., 1982; Chamberlin, 1906). The potential climatic impacts of warm saline deep and bottom water formation remains equivocal, however. Numerical ocean models have not been able to maintain a stable mode of circulation with deep convection in low latitudes (Bice, 1997; Brady et al., 1998). Furthermore, even if such a mode of circulation did persist, it is unlikely that a reversal of the thermohaline component of the meridional overturning circulation would significantly increase ocean heat transport on an Earth with warm polar temperatures already in place.

While the global importance of Tethyan deepwater formation is debatable, closure of the Tethys likely did have a number of important impacts on the evolution of Cenozoic oceans and climate. For example, a recent modeling study of Cenozoic ocean circulation (von der Heydt and Dijkstra, 2006) showed that the opening of Southern Ocean gateways combined with progressive closure of the Tethys could have induced a flow reversal (from westward to eastward) between the Atlantic and Pacific oceans through the Central American Seaway. Today, net fresh water flux out of the Atlantic basin maintains a relatively saline Atlantic and fresh Pacific. In turn, dominant sources of deepwater formation are limited to the Atlantic. Prior to the closure of the Tethyan and Panamanian seaways, inter-basin connectivity would have reduced this salinity contrast, possibly allowing locations of major deepwater formation in both the North Atlantic and North Pacific. Another ocean modeling study of an open versus closed circum-equatorial passage suggest an open Tethys would have increased upwelling of cold deepwaters in low latitudes, possibly helping to moderate tropical temperatures during warmer Mesozoic and early Cenozoic climate intervals (Hotinski and Toggweiler, 2003).

While the actual global climatic effects of these changes in the ocean remain unconstrained, the final closure of the eastern Tethys, ongoing Cenozoic restriction of Indonesia throughflow, and final closure of the Panamanian Isthmus in the Pliocene all contributed to forming the thermal structure of the oceans as we know them today. This includes a trade wind-driven Western Pacific Warm Pool (WPWP), westward dipping thermoclines in the major ocean basins, and cool eastern Pacific tropical SSTs – the essential components of the modern ENSO system (Philander, 1999). Differentiation of the ocean basins during the Cenozoic also influenced the partitioning of salt in the oceans, with fundamental consequences for the location of deepwater formation and ocean circulation in general.

Southern Ocean gateways

The opening of the Southern Ocean passages between Antarctica and Australia (Tasmanian Passage) and Antarctica and South America (Drake Passage) and Antarctic glaciation in the earliest Oligocene is an often-cited example of ocean gateways causing a specific climatic event. As these passages widened and deepened and the Antarctic Circumpolar Current and Polar Frontal Zone developed, reduced poleward ocean heat transport and cooler Southern Ocean temperatures are thought to have cooled Antarctica enough to allow continental glaciation (Kennett, 1977; Robert et al., 2001). While the earliest Oligocene glaciation of East Antarctica broadly coincides with the timing of Tasmanian gateway development (Stickley et al., 2004), estimates of the opening of Drake Passage range between about 45 and 22 Ma (Barker and Burrell, 1977; Lawver and Gahagan, 1998; Scher and Martin, 2006), clouding the direct cause and effect relationship between the gateways, cooling, and glaciation.

Several ocean modeling studies (Mikolajewicz et al., 1993; Toggweiler and Bjornsson, 2000) have reported reductions in Southern Ocean heat transport and cooler SSTs ($\sim$1–4 °C) in response to opening Southern Ocean gateways. In contrast, a recent study using a fully coupled atmosphere-ocean GCM showed little change in ocean heat transport or the climate of the Antarctic interior associated with the opening of the Tasmanian gateway (Huber et al., 2004). Even when forced with an unrealistically large (25%) change in southward ocean heat transport, a coupled GCM-ice sheet model showed that the effect on the timing of Antarctic glaciation caused by the gateways is small (DeConto and Pollard, 2003b) relative to the effects of declining Cenozoic CO_2 (Pagani et al., 2005). When considered together, these results suggests some other forcing mechanisms and/or feedbacks, perhaps related to the carbon cycle, may bear greater responsibility for Antarctic cooling and glaciation than the direct physical effects of the gateways on the oceans (Zachos and Kump, 2005).

The importance of atmospheric CO_2 concentrations relative to ocean gateways has also been recognized in modeling studies of other periods. For example, in ocean simulations testing the effects of the mid-Cretaceous opening of the gateway between the North and South Atlantic (Poulsen et al., 2001), very warm, saline conditions dominate the North Atlantic prior to the opening. While connecting these basins produces significant changes in regional ocean circulation, no fundamental change in the mode of thermohaline circulation occurs, and globally averaged SSTs change only by $\sim$0.2 °C, which is insignificant relative to the warming caused by the levels of greenhouse concentrations presumed to have existed at that time.

Central American passage and Indonesian Seaway

The Neogene closure of the Central American passage and ongoing restriction of the Indonesian Seaway have also been implicated in climate change, including the onset of Northern Hemisphere glaciation and African aridification (Cane and Molnar, 2001; Mikolajewicz et al., 1993). In the case of the Panamanian Seaway, restricted inter-basin water mass exchange was thought to have invigorated the Gulf Stream component of the North Atlantic subtropical gyre, increasing the moisture available for advection to high latitudes and growing ice sheets (Keigwin, 1982). More recent studies, however, have shown that closure of the Isthmus of Panama likely predates the onset of Northern Hemisphere glaciation, pointing to other possible forcing mechanisms (Haug et al., 2005).

Between the Pacific and Indian Oceans, the northward movement of Australia and New Guinea through the Cenozoic, combined with the emergence of individual volcanic islands, has led to the progressive restriction of Indonesian throughflow and a change in the source of surface waters entering the Indian Ocean from predominantly warm South Pacific waters to cooler North Pacific waters. The resulting cooling of the Indian Ocean has been implicated in African aridification $\sim$3–5 Ma, with possible impacts on Hominid evolution. Furthermore, the restriction would have increased the zonal temperature gradient across the Pacific, with potentially important teleconnections to high northern latitudes (Cane and Molnar, 2001).

Other significant gateway events during the Cenozoic include the opening of Fram Strait, which has allowed a deep water passage between the Arctic and North Atlantic basins since sometime in the late Miocene (Kristoffersen, 1990), and the Oligocene-Miocene deepening of the Denmark Strait and Iceland-Faeroes passages (Wold et al., 1993). The Denmark Strait and Iceland-Faeroes passages are of particular importance because they allow the outflow of deep waters formed in the Greenland and Norwegian Seas to flow into the North Atlantic, then becoming North Atlantic Deep Water– an important component of the so-called “global ocean conveyor” (Broecker, 1991).

The relatively narrow and shallow ($<$50 m) Bering Strait provides the only modern connection between the Arctic and the Pacific. While the gateway lies on continental crust, the low elevation of the area is attributed to regional extension and faulting associated with subduction of the Pacific Plate and rotation of the Bering Block (Mackey et al., 1997). The Bering Strait has allowed an intermittent shallow-water connection between the Arctic and Pacific basins since the late Miocene, with the Strait and surrounding area becoming subaerial during glacial periods. Today, only about 1 Sv (10^6 m^3 s^{-1}) of water passes through the strait into the Arctic basin; however, it has had an important impact on Plio-Pleistocene climate via its influence on North Pacific and Arctic salinity, sea ice, and the effect of repeated flooding/exposure of the Bering and Chukcki shelves on regional albedo and moisture availability for Beringian ice sheets (Brigham-Grette, 2001).

While the global effects of tectonically forced changes in ocean circulation remain equivocal, numerous ocean modeling studies have indeed shown that changes in ocean circulation in response to evolving basin configuration do have important regional impacts. Indeed, ocean currents are often attributed to the maintenance of specific terrestrial climate patterns, especially in maritime locations. Perhaps the most obvious example is the transport of warm subtropical surface waters in the North Atlantic Subtropical Gyre (via the Gulf Stream and North Atlantic Current), which warms the waters adjacent to northwestern Europe. The advection of warm waters into this region, in part driven by the North Atlantic Deep Water formation noted above, has long been presumed to be the primary reason for Northern Europe’s equability relative to North American climates at the same latitude. This paradigm has recently been challenged, however, with the role of topographically-forced atmospheric planetary waves (see below) and prevailing North Atlantic wind patterns possibly being more important in the transport of heat to Northern Europe (via the atmospheric transport of heat stored seasonally in North Atlantic surface waters) than previously considered (Seager et al., 2002).

Mountain uplift

Tectonic uplift is often cited as an important contributor to long-term climatic change (Birchfield et al., 1982; Raymo and Ruddiman, 1992; Ruddiman and Kutzbach, 1991). Because modern orography appears to be anomalously high relative to the warmer climatic intervals that dominated the Mesozoic and early Cenozoic, relatively recent mountain building events, like the collision of India with Asia that formed the Himalayas beginning $\sim$40 Ma, are often implicated as primary contributors to Cenozoic cooling and Northern Hemispheric glaciation. In fact, a number of the world’s mountainous regions have been described as having undergone significant uplift during the late Cenozoic (see Hay et al., 2002 and references therein) and speculated linkages between those uplifts and Cenozoic cooling date as far back as the nineteenth century (Dana, 1856).

With the exception of the ice sheets covering Greenland and Antarctica, the most prominent topographic feature on today’s Earth is the Himalayan-Tibetan plateau, covering a vast area

(4×10^6 km^2) at an elevation averaging ~5 km. The only geological intervals known to have had comparable orographic features occurred as a result of the continental collisions that formed Pangea about 320 and 240 Ma (see Ruddiman, 1997). The climatic effects of such features are multifaceted and complex. In addition to providing a barrier to atmospheric flow, mountains and high plateaus produce rain shadows on their leeward slopes and can have important effects on albedo and energy balance. Tectonic setting, including the orientation of land surface slope and the presence of bare soil and/or rock can influence albedo directly and through its control on vegetation. Vegetation is of particular importance because it affects land surface roughness, atmospheric moisture, and the partitioning of sensible and latent heat fluxes between the surface and atmosphere (Dickinson and Henderson-Sellers, 1988). Areas dominated by evergreen vegetation are also likely to maintain much lower albedo and warmer temperatures than bare soil or tundra, which are easily covered by winter snow (Bonan et al., 1992). Because tropospheric temperatures generally decrease with increasing altitude (the globally averaged tropospheric lapse rate is about 6.5 °C km^{-1}), high terrain is more likely to maintain perennial snow cover and glacial ice with high albedos (0.6–0.8).

Uplift and glaciation

In addition to the latitudinal position of continents, epeirogenic uplift is a critical factor in glaciation (Birchfield et al., 1982). This is mainly due to the dependence of both temperature and orographic effects on snowfall. On local to regional scales, glaciers are found where winter snowfall is high and summer temperatures remain cold enough to limit ablation. Conversely, the termination zones of glaciers often occur in low-lying valleys, where temperatures are warmer and summer ablation outpaces winter accumulation. The slope of the land surface also has a mechanical effect on ice flow, although the relative importance of snowline elevation versus flow effects is likely to depend on the amplitude and spacing of individual mountain peaks and troughs (Marshall and Clarke, 2000; Oerlemans, 2002).

Theoretical and numerical modeling studies have shown that tectonic uplift of widespread land areas above the equilibrium snowline can lead to the sudden, non-linear growth of glaciers and ice sheets, possibly leading to continental-scale glaciation. The rapid non-linear response is caused by albedo and height-mass balance feedbacks (North et al., 1983; Weertman, 1961) associated with the high reflectivity of snow and ice and the geometrical effect of a growing ice cap and its rapidly expanding net accumulation zone. The spatial scale of uplift appears to be a key factor in the potential response. For example, climate-ice sheet modeling studies have shown that broad uplift of an initially ice-free Antarctic interior can trigger sudden glaciation, provided the continent is already near a glaciation threshold (DeConto and Pollard, 2003b). In contrast, ice modeling studies of the effects of the Transantarctic Mountains on the East Antarctic Ice Sheet have shown that the uplift of individual mountains or the development of mountain troughs has limited influence (Kerr and Huybrechts, 1999).

Mountains, high plateaus, and atmospheric planetary waves

In mid-latitudes, zonal distributions of major landmasses and sea surface temperatures contribute to the position of standing atmospheric planetary waves via their influence on dominant low-level pressure patterns. Large mountains belts (Himalaya and Rocky Mountains) and high plateaus also provide direct physical barriers to atmospheric flow, which induce these long planetary *Rossby* waves in the upper-level westerlies, recognizable by the well-known meridional meanders in the polar and sub-tropical jet streams. Rossby waves form as a consequence of the conservation of absolute vorticity. When a parcel of air (or water) is displaced from its original latitude or changes its thickness as it passes over a mountain range, it responds by changing its latitude (planetary vorticity) and/or its relative vorticity (tendency to rotate). The resulting oscillations in the mean flow about a given latitude produce long planetary waves which move westward relative to the flow. In the fast flow of the upper troposphere, the waves appear to move slowly eastward relative to the Earth's surface. In the Northern Hemisphere, the longest of these waves tend to become stationary or *locked* relative to major orographic features. Wave crests (ridges) form over the Rocky Mountains and Tibet where the westerlies are displaced poleward before turning equatorward in the lee of the mountains, forming atmospheric troughs at ~70 °W and ~150 °E (Figure P67). While the amplitude of the long waves is highly variable, higher elevations tend to increase the meridional component of flow (Nigam et al., 1988). Thus, atmospheric planetary waves in the Northern Hemisphere tend to exhibit higher amplitudes than those in the ocean-dominated Southern Hemisphere, where the upper-level flow is more zonal.

Because vorticity associated with curvature in the waveform produces convergence downwind of the wave's crest (ridge) and divergence downwind of the trough, high pressure and

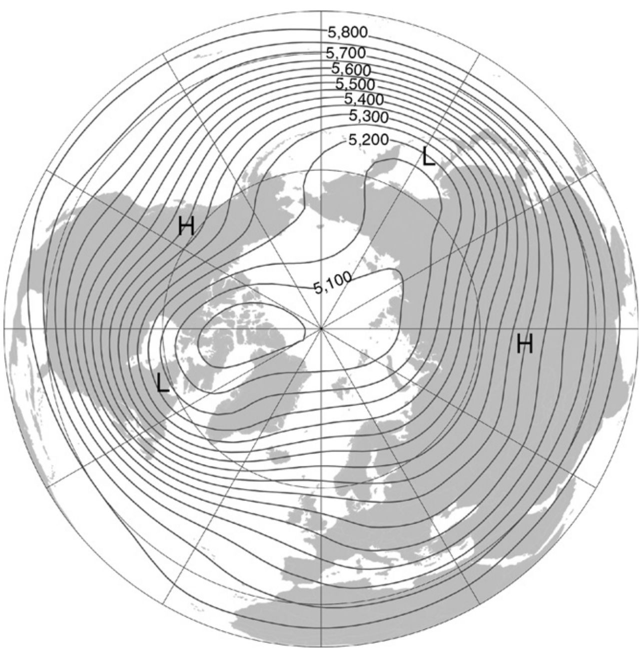


Figure P67 Seasonally averaged (DJF) 500 hPa geopotential heights over the Northern Hemisphere. Heights (shown in meters with 100 m contours) are calculated from 1971–2000 NCEP/NCAR reanalysis data, showing the time-averaged position of the dominant planetary wave pattern (see text for discussion).

generally fair conditions tend to be maintained eastward of the ridge. Conversely, cyclonic flow, low pressure, and storminess are found east of the trough. Thus, the time-averaged position of the waves has a fundamental effect on regional climates, storm tracks, and precipitation patterns. For example, the general aridity of the American interior and storminess of the U.S. East Coast can, at least in part, be attributed to the effect of the Rocky Mountains on the planetary wave pattern (Manabe and Broccoli, 1990). Further downstream, some of Western Europe's winter warmth can be attributed to the generally southwesterly flow over the relatively warm eastern North Atlantic, which is ultimately controlled by the standing wave pattern fixed by the Rockies (Seager et al., 2002). The standing wave pattern forced by the Rockies and Himalayas has also been implicated in establishing the conditions necessary for the onset of Northern Hemisphere glacial cycles in the Pliocene by allowing increased moisture advection from the warmth of the Gulf of Mexico and the western Atlantic to sites of Laurentide and Scandinavian Ice Sheet nucleation (Ruddiman and Kutzbach, 1989).

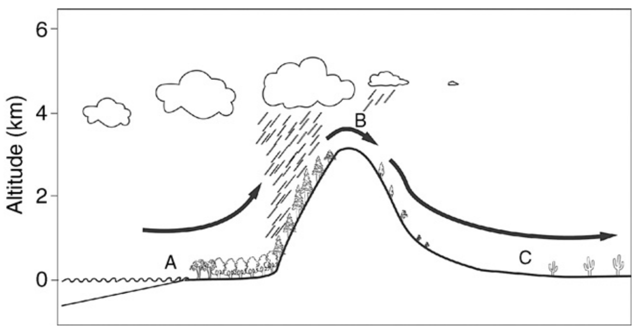


Figure P68 A schematic diagram showing the orographic precipitation and rainshadow effects of a mountain range. The *black arrow* represents air flow. Relationships between altitude, pressure, temperature, and absolute humidity control the distribution of precipitation, temperature and humidity on the windward and leeward side of the mountains. Typical temperatures and relative humidities of air at locations A, B, and C are ~28 °C and 70%, 8 °C and 100%, and 38 °C and 25%, respectively (redrawn from Hay, 1996).

Uplift and monsoons

In addition to their effect on zonal airflow, the lower air densities associated with high elevations amplify seasonal temperature changes and vertical motions associated with monsoonal systems (rising and sinking air in summer and winter, respectively). This effect is thought to be largely responsible for the strength of the modern Asian-Indian monsoon system (Prell and Kutzbach, 1992) and modeling studies have shown that elevations roughly equal to half those of the modern Himalayan-Tibetan elevation are required to produce the strength of the southeast summer monsoon over India (Kutzbach et al., 1993). The Himalayan and Tibetan plateaus have also been implicated as an amplifier of orbital (Milankovitch) variability, whereby a combination of enhanced condensational heating over South Asia and dynamical effects associated with planetary waves significantly increases the monsoon's sensitivity to orbital forcing (Liu et al., 2003).

Rain shadows

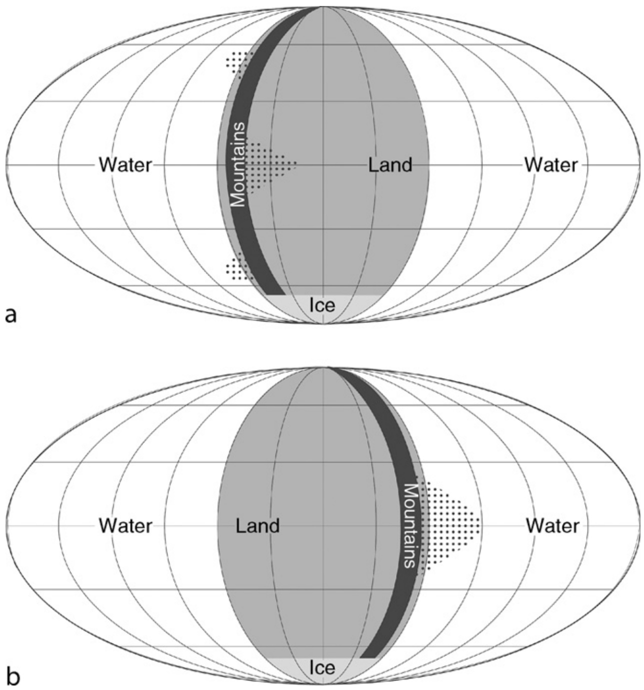
In addition to their widespread radiative and dynamical effects, mountains can have important regional impacts – especially on precipitation. When air is forced upward as it passes over a mountain range, adiabatic cooling induces condensation of available water vapor, resulting in clouds and precipitation (Figure P68). Having lost some of its available water, sinking, dry air on the leeward flanks of a mountain range is adiabatically warmed, which is why arid and warm conditions tend to prevail downwind of mountain ranges. This *rain shadow* effect can extend hundreds of kilometers, with obvious implications for the distribution of vegetation and the formation of climate sensitive sediments, such as evaporites. This drying effect is strongest where air can descend back to (or below) its initial altitude after passing over a mountain range, as can occur in rift valleys. Due to the non-linear relationships between temperature, saturation vapor pressure, and air pressure, an air mass passing over a second mountain range (out of a rift valley, for example) will be warmer and capable of carrying more water vapor than it was before it entered the basin. The resulting net export of fresh water out of rift valleys may account for the wide latitudinal distribution of Phanerozoic evaporite deposits, which extends far outside the subtropical arid zones (see Hay, 1996).

The modern Andes provide an excellent example of the importance of orographic effects on regional precipitation. In mid latitudes dominated by the westerlies, precipitation is enhanced on the western side of the mountain range, while the Patagonian Desert lies to the east. In tropical South America, where the winds are generally easterly, the opposite pattern is found, with enhanced precipitation over the rainforests of the Amazon Basin and dry conditions along the coastal margin of the Andes. For example, the high altitude Atacama Desert in northern Chile, one of the driest places on Earth, lies between the Pacific Ocean and the foothills of the Andes at a latitude of about 20° S and within the influence of the easterlies.

Similar rain shadow effects may have been even more profound on ancient supercontinents, with large seasonal cycles and arid interiors already in place. In GCM simulations using a single, idealized supercontinent extending from the north to south poles and with meridionally-oriented coastal mountain ranges running along the eastern continental margin, tropical precipitation associated with the trade winds and the ITCZ is intercepted on the seaward (eastern) flanks of the mountains (Figure P69). In a simulation with mountains running along the west coast of the supercontinent, continental precipitation is increased in tropical latitudes, while precipitation associated with cyclonic systems embedded in the westerlies falls mainly on the western slopes of the mountains (Hay et al., 1990b).

Mountain uplift or climate change?

In addition to tectonic, geomorphological, and structural studies, evidence of mountain uplift comes from records of continental denudation found in sediment accumulation rates (Hay, 1988) and marine records of $^{87}Sr/^{86}Sr$ (an indicator of chemical weathering (Richter et al., 1992)). Other techniques rely on the physiognomy of fossil vegetation (Royer, 2001), and oxygen isotopes in carbonate rocks (Garzione et al., 2004) to constrain the ancient elevations of specific locations. It should be noted that the term "uplift" can be used to describe the uplift of a region of the Earth's surface (with respect to the geoid) or the



Au1 **Figure P69** The distribution of maximum precipitation rate (stippling) in GCM simulations of a planet with a single supercontinent and a meridional mountain range oriented on either the western (**a**) or eastern (**b**) continental margin (redrawn from Hay, 1996 and based on GCM simulations described in Hay et al., 1990).

uplift of rocks (with respect to the surface). Depending on erosion rates, the uplift of rocks relative to the local surface, sometimes referred to as "exhumation" (England and Molnar, 1990), can be much greater than the uplift relative to the geoid. Erosion and the removal of mass from mountain valleys can actually raise the surrounding peaks through isostasy (Molnar and England, 1990). In the Alps and Himalayas, loss of mass in deep fluvial and glacial valleys has been estimated to account for up to 25% of the elevation of the highest mountains (Hay et al., 2002; Montgomery, 1994). Because chemical and physical erosion rates are ultimately controlled by the combined effects of surface relief and climatic parameters such as seasonality, precipitation, and glaciation, observed increases in the delivery of sediment to basins and continental margins can be attributed to climate change, mountain uplift, or both. The paradigm of Cenozoic uplift causing global cooling and Northern Hemisphere glaciation was challenged by Molnar and England (1990), who argued that the appearance of recent uplift in a number of disparate mountain ranges may be an artifact of Cenozoic climate change rather than the cause of the cooling (see also *Mountain uplift and climate change*).

Can the atmosphere drive tectonic processes?

The potential for the atmosphere to be a driver of tectonic processes has only recently begun to be appreciated. In addition to the processes mentioned above, whereby erosion in mountain valleys can raise mountain peaks, differential erosion rates on the windward (wet) versus leeward (dry) flanks of mountains and latitudinal climate gradients can cause structural asymmetries that influence the morphological evolution of entire mountain ranges. This effect has been proposed for the Andes (Montgomery et al., 2001), where structural asymmetries and latitudinal changes in crustal thickness correlate with climatically-controlled erosion patterns. Furthermore, progressive Cenozoic cooling and associated increases in wind stress, upwelling intensity, and cooler sea surface temperatures along the South American margin may be enhancing rain shadow aridity along the Chilean and Peruvian margin. The resulting decrease in precipitation and runoff is thought to be starving the adjacent convergent plate boundary (subduction zone) of lubricating sediment, thus contributing to the stresses supporting the high Andes (Lamb and Davis, 2003).

Plate tectonics and the global carbon cycle

While not the focus of this entry, the effects of tectonic processes on geochemical cycling and atmospheric greenhouse gas concentrations should be considered in any discussion of plate tectonics and climate change. Primary inputs of CO_2 include mantle outgassing at sea floor spreading centers and volcanoes, metamorphism of carbonate rocks along subduction zones, and respiration and burning of organic matter. Long-term sinks for CO_2 are controlled by the burial of organic matter and by the weathering of silicate rocks to form carbonates (Berner and Kothavala, 2001). Cenozoic mountain building, particularly the uplift of the Himalayan-Tibetan Plateau, has been related to CO_2 drawdown, cooling, and glaciation via increased weathering rates and the effects of increased nutrient delivery on ocean productivity and organic carbon burial (Chamberlin, 1899; Filipelli, 1997; Raymo et al., 1988). While the actual mechanisms responsible for CO_2 drawdown on tectonic timescales are complex and continue to be debated, geochemical reconstructions of Cenozoic CO_2 levels do show a dramatic decrease in atmospheric mixing ratios from more than three times present day levels during the warmth of the Eocene, to near modern levels by the early Miocene (Pagani et al., 2005).

Volcanism and large igneous provinces

Volcanism, often associated with tectonic processes, can also play an important role in climatic change. In addition to outgassing CO_2, an important greenhouse gas with the potential to raise global temperatures, magma can be rich in SO_2, which can lead to the production of sulfuric acid (H_2SO_4) aerosols in the atmosphere. When injected into the stratosphere via energetic volcanism, sulfur aerosols are highly reflective to incoming short wave radiation from the Sun, providing a cooling effect on the troposphere and Earth's surface. In 1991, the SO_2-rich eruption of Mt. Pinutubo in the Philippines cooled the Northern Hemisphere by 0.5–0.6 °C (Hansen et al., 1996) and the recent frequency of volcanic events may account for some of the variability recognized in highly resolved climate reconstructions of the Holocene (Mann et al., 1999). Global cooling associated with each injection of stratospheric aerosols is generally limited to a few years. On longer timescales, the atmospheric CO_2 loading associated with pervasive volcanism could have a net warming effect, complicating estimates of the net global environmental impacts of the larger and most long-lived volcanic events in Earth history.

The prolonged, effusive emplacement of mafic magmas to form large igneous provinces (LIPs) is usually associated with plate tectonic processes including uplift and continental rifting. LIPs can be continental or oceanic, and large oceanic LIPs, such as the Ontong Java Plateau in the western Pacific, can contain more than 50 million km^3 of volcanic and plutonic rocks. The episodic emplacement of LIPs has occurred throughout Earth history; although most surviving examples of continental and oceanic flood basalts are Mesozoic and Cenozoic in age (see *Flood basalts, climatic implications*). While the timing of some emplacement events has been directly associated with climate change (Mahoney and Coffin, 1997), LIP-climate connections are highly complex. In addition to their effect on sea level, LIP and mid-ocean ridge volcanism can release CO_2, SO_2 Cl, F, and H_2O into the atmosphere with significant climatic consequences. For example, the overall warmth of the Cretaceous period has been attributed, at least in part, to a combination of high atmospheric CO_2 and high sea level, with times of peak warmth and ocean anoxia associated with LIP emplacement (Larson and Erba, 1999; Leckie et al., 2002). While a number of studies have related LIP emplacement to extinction events (Rampino and Caldeira, 1993), the specific mechanisms causing environmental stress are difficult to tease apart and may be due to the direct effects of LIPs on climate, the biotic effects of toxic atmospheric fall-out (i.e., acid rain), or some combination of the two.

The Neoproterozoic *Snowball Earth* provides another example of indirect tectonic-climate linkages through volcanism. Between $\sim$750 and 580 Ma, the Earth is thought to have been episodically completely covered by ice, with some glaciations lasting for millions to tens of millions of years (Hoffman et al., 1998). While the evidence of significant ice at the equator remains controversial, tectonic processes (largely through volcanism's contribution to atmospheric CO_2) play a central role in most theoretical models of both the onset and demise of the snowball. First, a prolonged period of tectonic quiescence and minimal outgassing is thought to have allowed CO_2 concentrations to drop below some critical threshold, initiating a powerful ice-albedo cooling feedback (Schrag et al., 2002). The continental breakup of Rodinia (a Proterozoic super-continent), and associated changes in runoff and weathering have also been linked with a decline in CO_2 around this time (Donnadieu et al., 2004). As the Earth cooled, expanding ice sheets and sea ice would have increased planetary albedo. As shown in simple energy balance model (EBM) experiments, the ice-albedo feedback becomes unstable once snow and ice reach $\sim30°$ North and South, resulting in the sudden expansion of ice to the equator (Budyko, 1968). Once the Earth was completely covered by ice, continental weathering of silicate rocks, a long-term sink for CO_2, would have been greatly reduced. This would have allowed atmospheric CO_2 to accumulate to levels sufficient to cause significant warming, although additional warming mechanisms including the presence of other greenhouse gases and/or other feedbacks may have been necessary to trigger deglaciation (Pierrehumbert, 2004).

Summary

A number of tectonic processes are linked both directly and indirectly to a wide range of climate forcings, so the direct role of plate tectonics in specific ancient climate events is often difficult to decipher. In addition to the direct climatic effects of horizontal and vertical tectonics discussed here, paleogeography strongly influences the climate system's sensitivity to external forcing and provides the basic framework for studying ancient climate change over geological timescales.

While the relative contributions of greenhouse gas concentrations and plate tectonics in the evolution of the Earth's climate continues to be debated, tectonic processes influence the composition of the atmosphere, so the two cannot be entirely separated. Tectonically forced changes in ocean and atmospheric circulation have greatly influenced the evolution of specific regional climates; however, their potential to produce the full range of global climatic variability recognized in the geologic record may not be as great as once suspected.

Lastly, while tectonic processes are usually associated with slowly evolving environmental change, there are important exceptions. Volcanism's essentially instantaneous effect on the atmosphere provides one example. Less obvious is the climate system's potential to respond non-linearly to the most gradual tectonic forcing, with sometimes sudden and extreme consequences. Thus, plate tectonic processes should be considered a potential climate forcing mechanism on all timescales.

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Cross references

Albedo Feedbacks
Astronomical Theory of Climate Change
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