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The Late Cenozoic uplift – climate change paradox

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Abstract The geologic evidence for worldwide uplift of mountain ranges in the Neogene is ambiguous. Estimates of paleoelevation vary, according to whether they are based on the characteristics of fossil floras, on the masses and grain sizes of eroded sediments, or on calculations of increased thickness of the lithosphere as a result of faulting. Detrital erosion rates can be increased both by increased relief in the drainage basin and by a change to more seasonal rainfall patterns. The geologic record provides no clear answer to the question whether uplift caused the climatic deterioration of the Neogene or whether the changing climate affected the erosion system in such a way as to create an illusion of uplift. We suggest that the spread of C_4 plants in the Late Miocene may have altered both the erosion and climate systems. These changes are responsible for the apparent contradictions between data supporting uplift and those supporting high elevations in the past.

Keywords Uplift · Relief · Climate change · Neogene · Erosion

Introduction

The apparent uplift of many mountain ranges and plateaus during the later Cenozoic (Fig. 1) has been considered by some geologists to be either the direct or the indirect cause of the “climatic deterioration” leading to the Late Neogene glaciations. Others (unable to imagine a global tectonic mechanism for uplift of widely separated mountain ranges of very different ages) have suggested that the global cooling of the Late Cenozoic is responsible for the apparent uplift. This has led to two alternative hypotheses to explain the Late Cenozoic history of the planet: (1) the climate change is the direct or indirect result of tectonic uplift, or (2) the climate change has nothing to do with uplift, but has altered earth surface processes in such a way as to simulate widespread uplift. The arguments for the hypothesis that tectonic uplift is directly or indirectly responsible for the climate change have been presented in Ruddiman (1997), summarized by Ruddiman and Prell (1997) and Ruddiman et al. (1997b). The arguments for the hypothesis that the climate change has caused erosive processes to change in such a way as to simulate uplift are more scattered, but are summarized in Molnar and England (1990) and England and Molnar (1990).

The purpose of this study is to explore interrelationships between the apparent worldwide uplift of mountain ranges in the Neogene and the Cenozoic “climatic deterioration”, resulting in the onset of glaciation in Antarctica followed by the oscillating glaciation of the northern continents surrounding the Arctic. We will then speculate on the relative importance of the factors contributing to the climate change, suggesting that evolution of the biosphere may have played a hitherto unsuspected major role.

A brief historical perspective

Lyell (1830) was impressed by the marked contrast between the conditions required for deposition of the Carboniferous coals as opposed to the present climate. Convinced that the climate of the British Isles had been different in the past, he proposed that the change reflected differences in the global latitudinal distribution of land and sea, arguing that with more extensive land areas in the polar regions, the earth would become cooler, and conversely, with more land in the equatorial regions, the earth would be warmer.

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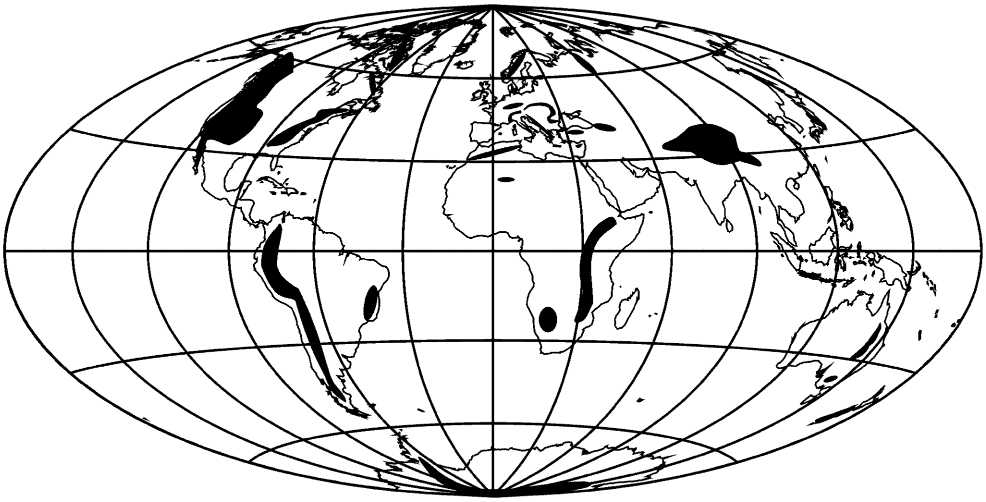

Fig. 1 Areas described as having undergone uplift in the Late Neogene (for sources see text)

With recognition of the existence of more extensive mountain glaciation and of continental ice sheets in the immediate geologic past (e.g., Agassiz 1840), the idea arose that the earth was cooling. In the latter half of the 19th century, this was explained in terms of heat loss as the earth cooled from an originally molten state. Lord Kelvin's (1863, 1864) calculations of the loss of heat by the earth indicated its age to be about 100×10^6 years. He further found that the internal heat flux from the cooling earth exceeded the solar energy flux until the Cenozoic. The earth's meridional temperature gradient developed, and the general, global cooling trend began only after the heat flux from the interior dropped below that of radiation from the sun. Neumayr (1883) demonstrated that climate zones had existed during the Jurassic and Cretaceous, negating the argument that latitudinal climate zones did not exist before the Cenozoic.

The discovery of evidence for extensive glaciation in the southern hemisphere during the Late Paleozoic put the notion of a cooling earth to rest. Recently, evidence for extensive glaciation in the Early Proterozoic and for a possible "snowball earth" in the Neoproterozoic (Hoffman et al. 1998) have placed the Cenozoic cooling trend and glaciation in a new context. Hambrey (1999) presented an excellent review of the history of glaciation of the earth since the Archaean. One peculiarity of this history is that, except for the Precambrian and Pliocene–Quaternary glaciations, only the southern polar region has been ice-covered. This is presumed to be the result of past distributions of land and sea.

The development of the theory of glaciation in the last century was accompanied by the realization that the climate of the earth must have cooled substantially during the Cenozoic, at least in the polar regions. However, the degree of cooling was only expressed in qualitative terms. It was not until late in the latter half of the 20th century that quantitative methods were developed for estimating paleotemperatures. The detailed ideas of cooling during the Cenozoic are based on oxygen isotope ratios in benthic Foraminifera, interpreted as reflecting surface water temperatures at high latitudes (Fig. 2). Compilations by Savin (1977, 1982), Douglas and Woodruff (1981), Moore et al. (1982), and Miller et al. (1987) have become standard references in the interpretation of the Cenozoic temperature history of the planet. The compilations show a trend towards more positive values of $\delta^{18}O$ since the Middle Eocene. This has been

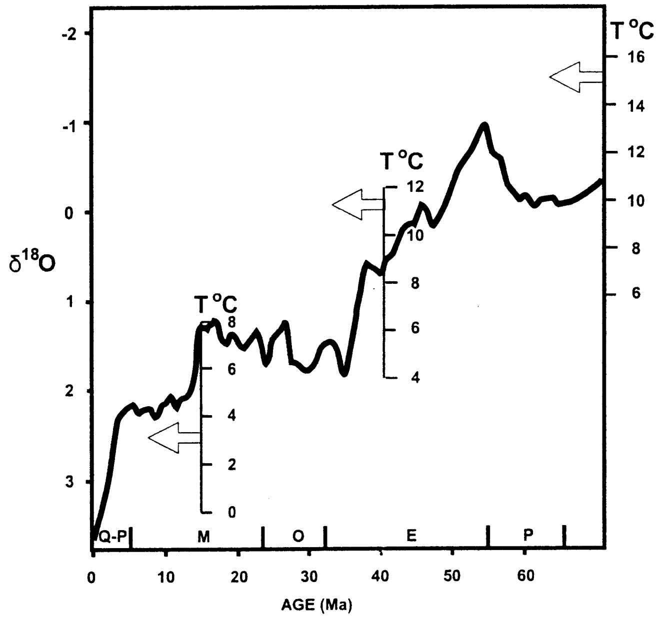

Fig. 2 The decline in ocean bottom water temperatures in the Atlantic indicated by oxygen isotope ratios of the tests of deep-sea benthic Foraminifera (after Miller et al. 1987). Three temperature scales are shown: the scale on the *left* is for the younger part of the diagram (Late Miocene to Recent) and assumes that the $\delta^{18}O$ of seawater is –0.28‰. The scale on the *right* applies to the older part of the diagram (Late Cretaceous through Eocene) and assumes an ice-free earth, with a $\delta^{18}O$ of seawater of –1.2‰. The scale in the *middle* applies to the mid-Tertiary (Oligocene through Middle Miocene) and assumes an intermediate $\delta^{18}O$ of seawater. The interpreted decline in bottom water temperatures is assumed to reflect decline of polar surface temperatures during the Cenozoic

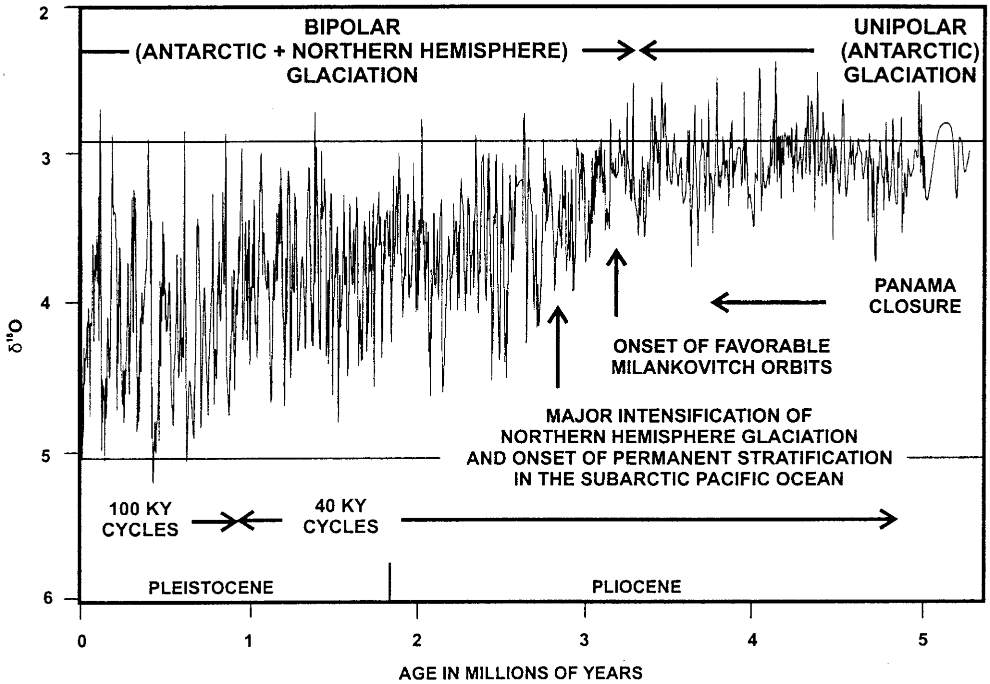

Fig. 3 The Late Neogene climate change reflected in oxygen isotopes of deep-sea benthic Foraminifera (after Tiedemann (1991). The timing of closure of the Atlantic–Pacific passage across Panama and onset of favorable (cool northern-hemisphere summer) orbits is after Haug and Tiedemann (1998). Timing of the onset of permanent stratification in the subarctic Pacific Ocean is after Haug et al. (1999). The upper horizontal *line* is the present-day (Holocene) value for $\delta^{18}O$; the lower horizontal *line* is the value of $\delta^{18}O$ at the Last Glacial Maximum

interpreted as reflecting a general decline in bottom water temperatures and the buildup of ice on Antarctica. It is now thought that ice began to accumulate on Antarctica in the Late Eocene (Brancolini et al. 1995). Ice reached the coast at Prydz Bay in the Indian Ocean sector at the Eocene–Oligocene boundary (34.5 Ma). By the Early Oligocene, ice had reached other parts of the margin of East Antarctica (Hambrey et al. 1991; Exon et al. 2000) and had begun to accumulate on the Antarctic Peninsula (Dingle and Lavelle 1998). Glaciation of the Antarctic continent underwent a major expansion in the Late Miocene (Flower and Kennett 1994; Barrett 1994). There is currently an ongoing discussion (Miller and Mabin 1998) whether Antarctica remained fully glaciated with a cold-dry ice sheet throughout the Pliocene (Stroeven et al. 1998) or whether it became a warm-wet ice sheet with a significant reduction of ice volume in the mid-Pliocene (Kennett and Hodell 1995; Harwood and Webb 1998). Estimating temperatures on land required other methods, some of the most successful being based on plant fossils. Paleobotanical studies (Wolfe 1980, 1985, 1992a; Wolfe and Poore 1982) documented a time of polar warmth with Middle Eocene, broad-leaved deciduous trees in the Arctic being replaced by conifers in the Late Eocene. Younger floras reflect a general cooling since then.

The term “Late Neogene climate change” has become largely synonymous with the climatic changes associated with northern hemisphere glaciation, although this is merely the culmination of the cooling trend which began in the Eocene. Ice rafting of terrigenous material in the northern hemisphere began in the Late Miocene, at about 9.5 Ma (Wolf and Thiede 1991). The glaciation of southern Greenland was underway by 6 Ma (Jansen et al. 1990; Jansen and Sjöholm 1991). Tiedemann (1991) traced the expansion of the northern hemisphere glaciation in oxygen isotope records from DSDP cores taken off northwest Africa (Fig. 3). Significant glaciation did not start until about 3.2 Ma. From the study of oxygen isotopes and the proportion of sand-size foraminiferal tests in carbonate sediments at ODP sites 929 (equatorial west Atlantic) and 999 (Colombian Basin), Haug and Tiedemann (1998) concluded that the separation of the Caribbean and Pacific occurred at about 4.5 Ma but, because of unfavorable orbital conditions, the large-scale northern hemisphere glaciation did not begin until 3.25 Ma. Maslin et al. (1996) have described the major steps in the glaciation of the Arctic, with the Scandinavian and northeast Asian glaciation becoming pronounced at 2.75 Ma, followed by Alaskan glaciation at about 2.65 Ma and initiation of the Laurentian ice sheet at 2.54 Ma.

Hence, the Late Neogene climate change started during the Middle Pliocene and, by the Late Pliocene, the present system of waxing and waning of large ice sheets was established. A further modification, the change from 40,000- to 100,000-year glacial–interglacial cycles, occurred about 750,000 years ago, probably as the result of a change in the substrate of the ice sheets from soil and residuum to solid rock, altering the characteristics of the basal flow and affecting the dynamics of the ice sheet (Clark and Pollard 1998).

Possible causes of the Cenozoic cooling trend

The list of possible causes of the Cenozoic cooling trend expanded greatly during the 20th century. Hay (1992) discussed the causes proposed to explain the cooling of climate leading to glaciation, relying on those cited in literature reviews (Flint 1971; Crowell 1982). These included (1) variations in solar emissivity; (2) changes in the concentration of dust in space; (3) variations in the earth's orbital motions; (4) continental drift; (5) the effect of vertical movements ("uplift") on the radiation balance and atmospheric circulation; (6) changes in oceanic circulation; (7) changes in sea-ice cover; (8) variations in the concentration of greenhouse gasses in the atmosphere; and (9) increased aerosol content of the atmosphere. Since then, two additional factors have been suggested: (10) changes in occurrence of polar stratospheric clouds; and (11) changes in polar vegetation. Much has happened during the last decade, and a re-evaluation of these factors is appropriate here.

Variations in solar emissivity

Variations in solar emissivity had been discounted by Jerzykiewicz and Serkowski (1968). Stellar evolution models show that the emissivity of stars the size of our sun increases slowly with time. This should lead to warming of the planet, but the changes in emissivity are very small on the time scale of the Cenozoic. However, there has been considerable uncertainty because early observations of solar neutrinos showed a marked deficit over what was expected from theory. This raised the question whether models of stellar evolution and the mechanism of solar emissivity were erroneous, or whether the nature of neutrinos was not fully understood (Ulrich 1975; Newman 1986; Raghavan 1995). The "solar neutrino mystery" has recently been solved by the discovery that neutrinos can oscillate between their three different "flavors" (Bahcall 2001), vindicating the models for solar evolution. However, another aspect of solar activity has recently emerged with the discovery that cosmic rays induce cloud formation, and that enhanced solar activity, with concomitant increase in the solar wind, deflects cosmic rays, thereby reducing the flux reaching the earth (Svensmark and Friis-Christensen 1997; Svensmark 1998, 2000; Marsh and Svensmark 2000). So far, these variations have been suggested to play a role only in short-term climate change, but the implications for the long term have yet to be explored.

Changes in the concentration of dust in space

Changes in the concentration of cosmic dust had been discounted by Pollack (1982), but a variant of this theme has been recently reintroduced by Brownlee (1995) and Muller and MacDonald (1995, 1997a, 1997b). They proposed that the earth's orbit passes through the zodiacal cloud of dust in the solar system's plane of symmetry (responsible for the "zodiacal light") every 100,000 years, and suggested that this, rather than the eccentricity of the earth's orbit, is the mechanism controlling the 100,000-year glacial-interglacial cycle characteristic of the later Quaternary. Kortenkamp and Dermott (1998) reported that there is indeed a 100,000-year periodicity in the rate of accumulation interplanetary dust.

Milankovitch orbital motions

Milankovitch orbital motions are generally accepted as governing the intensity of northern hemisphere glaciation during the Pliocene and Quaternary, but their long-term effects were discounted by Schwarzacher (1987) because they also occur when the earth is in a nonglaciated state. However, Varadi et al. (2001) have suggested that, at about 65 Ma, there was a dramatic shift in the dynamics of the solar system, involving macroscopic changes in the orbit of Mercury but also affecting the other inner planets. This suggests that Milankovitch orbital variations in the Cenozoic may be different from those of earlier times. During the Cenozoic, the onset of glaciation in Antarctica (DeConto and Pollard 2001) and in the northern hemisphere (Haug and Tiedemann 1998) may be related to specific orbital configurations producing minimal summer temperatures. Haug and Tiedemann (1998) argue that, although the closing of the Central American isthmus set the stage for the onset of northern hemisphere glaciation, the actual trigger was a favorable Milankovitch orbital configuration which occurred between 3.1 and 2.5 Ma.

Continental drift

Continental drift in the regions surrounding the Arctic Ocean had been considered by Donn and Shaw (1977) to be the primary cause of the post-Eocene climate deterioration, but the effects were demonstrated by Barron and Washington (1984) to be of minor importance. Except for India and Australia, the movements of the continental blocks since the Cretaceous have been almost zonal.

Vertical motion of the earth's crust ("uplift")

Vertical motion of the earth's crust as a cause of climate change goes back at least to Dana (1856). Flint (1943) related glaciation directly to uplift. He became convinced that Late Cenozoic uplift had occurred in many parts of the world, and compiled lists of mountains and plateau regions which had been affected (Flint 1957). Hamilton (1968) suggested that uplift was the major cause of the Cenozoic cooling trend, but did not explain precisely how the two were related. Flohn (1974) suggested that a primary factor in the initiation of glaciation is the buildup of snow cover which persists throughout the year. The increased albedo during the summer lowers the temperature and produces a positive feedback increasing the area of snow cover and further increasing the albedo. Topography exerts a primary control because snow persists longer in higher areas.

It has long been recognized that there is a close relationship between climate and the topography of the earth's surface. Hay (1996) presented a review of many aspects of this complex topic. Large-scale regional uplifts, such as in Tibet and the western US, also alter the earth's global climate by changing the radiation balance. Mountain ranges produce mostly local and regional effects, such as the intensification of rainfall on the windward side of the range and the rain-shadow effect on the leeward side (Barry 1981; Barry and Chorley 1982). These local effects are part of human experience, observed, if not understood, by everyone who lives in mountainous regions. Larger scale effects of changes in the topography of the earth have been deduced from climate model experiments. In climate simulations, flat continents produce a more zonal climate with a reduced meridional temperature gradient, whereas mountainous continents produce more differentiated global climates and an enhanced meridional temperature gradient (Kasahara and Washington 1969; Kasahara et al. 1973; Manabe and Terpstra 1974; Barron and Washington 1984; Barron 1985; Hay et al. 1990). Kutzbach et al. (1997) explored the effects of uplift on the hydrologic cycle, using a general circulation, climate model. As expected, uplifts are sites of preferential preservation of snow and ice, and induce the ice-albedo feedback. However, precipitation is enhanced on upwind slopes, resulting in an even greater increase in runoff. Large-scale uplifts also act directly as obstacles to atmospheric circulation, forcing increases in wind velocity, interrupting zonal flow, and inducing the formation of eddies.

Renewed interest in uplift and climate change as cause and effect appeared during the 1980s. Birchfield and Weertman (1983), Ruddiman et al. (1986, 1989), Ruddiman and Raymo (1988), Ruddiman and Prell (1997), Ruddiman et al. (1997a), and Rind et al. (1997) have argued that irregularities in atmospheric circulation are mostly the result of uplift, with plateaus and mountains blocking zonal circulation. They proposed that the uplift of the Himalayas–Tibet and western North America fundamentally altered the pattern of atmospheric circulation in the northern hemisphere, and set up conditions promoting the development of the Laurentide and Eurasian ice sheets. They argued that the uplifts promoted stabilization of the north-south oscillations of westerly winds (Rossby waves), which made it possible for warm waters of the Gulf Stream off New England to become the moisture source for the Laurentide and Scandinavian ice sheets. Manabe and Broccoli (1990) and Broccoli and Manabe (1992, 1997) showed that the perturbations of the northern hemisphere circulation by the Tibetan and western North American uplifts are largely responsible for the mid-latitude aridity of eastern Asia and central North America. Rind et al. (1997) argued that uplift has pervasive effects throughout the atmosphere-ocean system, affecting the flow of winds which in turn drive the surface ocean currents, changing the temperature and salinity of the ocean and hence affecting ocean heat transport and the deep ocean circulation, and even resulting in slight global warming. The arguments for plateau uplift as the cause of Late Cenozoic climate change have been further developed by Ruddiman and Kutzbach (1989, 1990, 1991a, 1991b) and Prell and Kutzbach (1992, 1997).

Eyles (1996) developed the arguments put forth by Flint (1957) and Flohn (1974), and argued that rejuvenated Late Neogene uplift of passive margins around the entire North Atlantic region played a critical role in the initiation of Late Cenozoic glaciation.

Changes in oceanic circulation

Changes in oceanic circulation have been cited as a cause for both the glaciation of Antarctica and the northern hemisphere glaciations. The effects cited are (1) changes in surface circulation forced by the opening and closing of interocean passages, and (2) changes in thermohaline circulation resulting from the relocation of sites of deep-water formation.

Berggren (1982) associated major climate changes during the Mesozoic and Cenozoic with the closing and opening of interocean gateways. The most significant changes were the opening of a circum-Antarctic seaway, elimination of the low-latitude Tethyan seaway, and opening of passages into the Arctic. There are two aspects of ocean circulation which might be important, the surface circulation and the thermohaline circulation. In terms of overall heat transport, the surface circulation is by far more important. The thermohaline circulation accounts for less than 20% of the global ocean heat transport. However, a reversal of the thermohaline circulation could have significant climatic consequences (Chamberlin 1906; Brass et al. 1982).

The opening of the circum-Antarctic passages is considered by many to have set the stage for glaciation of the Antarctic continent. According to Barker and Burrell (1977), the Drake Passage opened as a shallow water passage at about 30 Ma, but first became a deep-water passage at about 20 Ma. Lawver et al. (1992) have argued for an earlier, Oligocene opening of a deep Drake Passage. Watkins and Kennett (1971) and Kennett (1977) attributed the thermal isolation of Antarctica to the deepening of the Tasman–Antarctic passage at about 26 Ma. Recent Ocean Drilling Program data place the opening of the deep Tasman–Antarctic passage at about 34 Ma (Exon et al. 2000). The timing of the opening of a full circum-Antarctic passage does not coincide with the onset of Antarctic glaciation. However, a passage between East and West Antarctica, proposed by Leckie and Webb (1985), would have allowed a proto-circum-Antarctic current system to develop, thermally isolating East Antarctica.

The closing of the zonal, northern low-latitude Tethyan passage from the Indian Ocean into the Mediterranean occurred during the Miocene (Dercourt et al.

1986, 1992). This increased the isolation of the northern hemisphere basins, setting the stage for the northern hemisphere glaciation. The separation of the Atlantic from the Pacific took place gradually, from 13 to 2.5 Ma. Duque-Caro (1990) indicated that a sill with a depth of about 1 km began to block the Panama–Costa Rica Strait during the Middle Miocene at 12 Ma. By 6 Ma, during the Late Miocene, the sill depth had shallowed to about 200 m. Although land faunas indicate that final closure occurred at about 2.5 Ma (Marshall 1988), new evidence from ODP sites in the Caribbean and eastern Equatorial Pacific (Haug and Tiedemann 1998) indicates that the Atlantic and Pacific had become separated by about 4.5 Ma.

Maier-Reimer et al. (1990) examined the effects of changes in interocean gateways, using an ocean general circulation model driven by present-day winds. Comparison of closed and open Central American isthmus models suggests that, with an open passage, the sea-surface slope between the Caribbean and the Norwegian–Greenland seas, and hence the strength of the Gulf Stream–North Atlantic Drift, would be much reduced. A strong subsurface North Equatorial Countercurrent would carry 10 Sverdrups of water into the Atlantic whereas only 1 Sverdrup would pass from the Atlantic to the Pacific on the surface. This would reduce the salinity and density contrasts between the Atlantic and Pacific, so that there would be no production of North Atlantic Deep Water, and hence no “global conveyor” in the sense of Broecker (1991). Although most of the global river runoff flows into the Atlantic (or its tributary Arctic Ocean), the Atlantic is saltier than the Pacific. This is because of its peculiar shape, its widest regions being at latitudes where evaporation greatly exceeds precipitation. As long as it was connected at depth to the tropical Pacific, major salinity differentiation could not occur. The effect of the higher salinity is to set the stage so that further density increases due to cooling and sea-ice formation in the Arctic and GIN (Greenland–Iceland–Norwegian) seas causes sinking of surface waters into the interior. This high-latitude sinking induces an arm of the Gulf Stream–North Atlantic Drift to turn northwards across the Iceland–Scotland Ridge. This relatively warm water provided the moisture source for the massive Quaternary ice sheets over northeastern North America and northwestern Europe (Maslin et al. 1996). The sinking of water in the Arctic and GIN seas forces replenishment of Atlantic surface waters from the Indian Ocean, leading to the strange circumstance that overall ocean heat transport is northwards in the Atlantic, even in the southern hemisphere.

The coincidences between the timing of the opening of the Tasman–Antarctic passage and Antarctic glaciers reaching the coast at Prydz Bay, and the closure of the Panama passage and onset of northern hemisphere glaciation are striking. They suggest that the resulting changes in ocean circulation set the stage for major steps in climate change. However, the actual triggers were probably favorable Milankovitch orbits.

Changes in sea-ice occurrence

Changes in sea-ice cover in the polar regions have been cited as a possible cause for climate change. Perennial sea ice in the Arctic Ocean and seasonal sea ice around the Antarctic effectively change the albedo of the earth’s surface in these regions from low to high.

Ewing and Donn (1956) proposed that the alternation of glacials and interglacials might be in response to an alternately ice-covered and ice-free Arctic Ocean. The idea was abandoned because of lack of independent evidence for an ice-free Arctic Ocean during the Late Pleistocene. Clark (1982) suggested that it was the formation of perennial sea ice in the Arctic Ocean which permitted the atmosphere at northern latitudes to cool by 15–20 °C, causing the onset of widespread northern hemisphere glaciation. However, Herman and Osmond (1984) argued that perennial sea-ice cover did not appear until 0.9 Ma, long after the cyclic northern hemisphere glaciation had become established. Raymo et al. (1990) investigated the impact of sea-ice cover of the Arctic Ocean on the regional climate, using an atmospheric general circulation model. They found that the climatic effects were confined mostly to the region from which the ice cover was removed. The neighboring continental areas were much less affected, and Raymo et al. (1990) concluded that the formation of perennial sea-ice cover was not in itself the cause of the widespread northern hemisphere glaciation.

Hay and Wold (1997) and Hay et al. (1998) discussed the changing salinity of the ocean with time, and concluded that sea-ice formation in the open ocean would have been more difficult before the Late Miocene salt extractions in the Mediterranean, Red Sea, Carpathians and Persian Gulf. They estimated that the mean salinity of the Miocene ocean was about 39‰, significantly higher than the present 34.7‰. Flögel et al. (2000) indicate that salinities in the Triassic and Paleozoic were in the high 40s to low 50s (‰). This suggests that, except for the “snowball earth” conditions of the Precambrian, extensive sea-ice formation in the earth’s oceans may be a Late Cenozoic phenomenon.

Variations in atmospheric greenhouse gas concentrations

Variations in the concentration of greenhouse gasses in the atmosphere is currently a topic of great public interest. At present most of the greenhouse effect of the earth’s atmosphere is produced by water vapor, but its concentration is a function of the temperature of the ocean surface. In effect, water vapor acts as a positive-feedback amplifier of temperature changes induced by other greenhouse gasses. The other major, naturally occurring greenhouse gasses are, in order of their impact on the pre-industrial atmosphere, carbon dioxide (~80% relative contribution), methane (~20%), and nitrous oxide (minor). Relative to CO_2, equal moles of CH_4 are about 25 times as effective in terms of their greenhouse capability (Rodhe 1990). The lifetime of methane in the atmosphere is short (decay time ~10 years vs. 120 years for CO_2), before it is oxidized to become the less effective CO_2.

Climate change as a result of change in the composition of the atmosphere was suggested in the 19th century, and the discussion revolved around CO_2 as a greenhouse gas. Chamberlin (1899) proposed that the glaciations might be brought about by reduction of atmospheric CO_2 during prolonged periods of weathering and erosion as a result of uplift. The subsequent history of the CO_2–climate change hypothesis has been reviewed by Fleming (1998). He noted that Chamberlin later changed his mind about the importance of CO_2, and became convinced it was water vapor which was the major factor.

Raymo (1991) noted that Chamberlin's original hypothesis deserves careful consideration in the light of modern knowledge of geochemical cycles. There are three primary natural sources for CO_2 in the atmosphere: outgassing from the mantle, thermal metamorphism of carbonate rocks, and decomposition or burning of organic matter. The rate of outgassing from the mantle is thought to be a function of the rate of sea-floor spreading (Berner 1991, 1994). The release of CO_2 from metamorphism occurs when carbonates are subducted and CO_2 returned as volcanic exhalations. CO_2 is removed from the atmosphere by the burial of organic carbon and by the weathering of silicate rocks to form carbonates. In brief, increases in seafloor spreading rates enhance volcanism and mountain building which expose silicate volcanic and basement rocks to weathering which removes CO_2 from the atmosphere and causes cooling (Raymo et al. 1988; Raymo 1991; Raymo and Ruddiman 1992; Kutzbach et al. 1997; Blum 1997; Edmond and Huh 1997). In these cases the authors assumed that Late Cenozoic mountain building, specifically the rise of the Himalayas, exposed silicate rocks to weathering. However, McCauley and DePaolo (1997) argued that the Himalayas account for only 5%, rather than 50% of the global silicate weathering flux. They concluded that a 5% imbalance in the CO_2 fluxes would be adequate to drive the Cenozoic cooling trend. Alternatively, removal of CO_2 from the atmosphere may have been the result of burial of organic carbon in rapidly accumulating deltaic and fan deposits, such as those of the Bay of Bengal and Arabian Sea (France-Lanord and Derry 1997). In these instances it was also the Himalayas which provided the detritus promoting burial of organic carbon. Berner and Berner (1997) noted that it is the balance between the supply of CO_2 to the atmosphere and its consumption in weathering which is important in determining whether atmospheric CO_2 levels rise or fall. Raymo (1997) emphasized the uncertainties in our knowledge of the carbon cycle during the Cenozoic. Recent studies show that the erosion of mountainous regions results in an increase in the rate of burial of organic carbon, which may be more important than silicate weathering as a sink for CO_2 (France-Lanord and Derry 1997; Derry and France-Lanord 1997). Turekian and Pegram (1997) suggested that increased weathering of black shales in the Himalayas since 7 Ma may have increased the delivery of phosphorus to the ocean, increasing productivity, and thus lowering atmospheric CO_2. Kump and Arthur (1997) argued that during the Late Neogene, chemical weathering increased in the Himalayas but declined elsewhere, providing for a slow, even decline in atmospheric CO_2. In any case, glaciation would have been the indirect result of uplift via reduction of atmospheric CO_2.

Weathering of $CaCO_3$ does not serve as a long-term sink for atmospheric CO_2. In the dissolution of limestone, CO_2 acts as a catalyst allowing the transport of $CaCO_3$ as Ca^{2+} and HCO_3^- ions to a new site of deposition as $CaCO_3$, whereupon the CO_2 is returned to the atmosphere. Nevertheless, temporary storage of $CaCO_3$ in solution in the ocean may be adequate to account for changes of atmospheric CO_2 of the magnitude of the observed glacial-interglacial changes (Petit et al. 1999). The CO_2 rise during interglacials may be in response to the rapid deposition of $CaCO_3$ on shelves and platforms flooded by sea-level rise (Hay and Southam 1977; Berger et al. 1996).

Because the GEOCARB model (Berner 1994; Berner and Kothavala 2001) does not have adequate temporal resolution to predict the structure of the Cenozoic decline in atmospheric CO_2 concentrations, many geologists tacitly assumed that atmospheric CO_2 decreased in parallel with the $\delta^{18}O$ curve for deep-sea benthic Foraminifera. Although isotopic data from Mesozoic pedogenic carbonates suggest much higher levels of atmospheric CO_2, the younger record is ambiguous (Ekart et al. 1999). Cenozoic paleosols suggest both higher (900–1,000 ppmv) and lower (270–210 ppmv) atmospheric CO_2 concentrations. This view of a general decline in atmospheric CO_2 throughout the Cenozoic has been challenged by recent studies. Pagani et al. (1999a, 1999b) estimated Miocene atmospheric CO_2 concentrations from g_p (magnitude of the carbon isotope discrimination during photosynthesis) values based on $\delta^{13}C$ in diunsaturated alkenones and the shells of shallow-dwelling planktonic Foraminifera from DSDP and ODP sites in the Atlantic, Indian and Pacific oceans. They concluded that atmospheric pCO_2 levels were below 280 ppmv during most of the Miocene. They also found no feature comparable to the sharp Middle Miocene increase in $\delta^{18}O$ interpreted as a major cooling step in the Antarctic. Similar results have been reported for the earlier Cenozoic by Pearson and Palmer (1999, 2000a, 2000b), based on interpretations of atmospheric CO_2 concentrations from estimates of oceanic pH using $\delta^{11}B$ of foraminiferal calcite. On the basis of leaf stomatal indices in *Ginko* and *Metasequoia*, Royer et al. (2001) have concluded that atmospheric CO_2 levels were between 300 and 450 ppmv during the Paleocene, Eocene and Middle Miocene, except for a brief excursion near the Paleocene–Eocene boundary. Veizer et al. (2000) found no direct relationship between the Phanerozoic $\delta^{18}O$ record and the occurrence of glacial episodes documented by geological data, suggesting that the two phenomena are not coupled. Kump (2000) noted that this calls into question the currently accepted relationship between atmospheric CO_2 levels and climate.

Methane has been cited as contributing to climate change in the Cenozoic, but the effects are mostly indirect because it is rapidly oxidized to CO_2. However, sudden large releases of methane from dissociation of gas hydrates have been proposed as an explanation for the sharp, brief warming of climate near the end of the Paleogene (Dickens et al. 1997). Increased tropospheric methane has also been suggested as a source of water for polar stratospheric clouds (see below).

Increased aerosol content of the atmosphere

Increased aerosol content of the atmosphere involves ash and SO_2 from volcanic activity, and windblown dust. Kennett (1981) speculated that the general increase in volcanic activity during the Cenozoic played a major role in the climate deterioration. He noted that the times of increased volcanic activity (as interpreted from the frequency and thickness of ash layers in deep-sea deposits) in the Middle Miocene, Late Miocene, Middle to Late Pliocene, and Quaternary correspond to times of increase in the size of the Antarctic ice sheet and to the onset and intensification of northern hemisphere glaciation. Straub and Schmincke (1998) also found that there has been a marked increase in the amount of tephra in Pacific Ocean sediments during the past 10×10^6 years. Ash has a relatively short life in the atmosphere unless it is injected into the stratosphere. SO_2 serves to nucleate water droplets, producing clouds which may increase the earth's albedo, but it has a relatively short life in the atmosphere unless it is injected into the stratosphere.

Another possible source of SO_2 may be from the dimethylsulfoxide (DMS) directly or indirectly produced by marine phytoplankton. Upon release into the environment, DMS is broken down into simpler compounds and molecular sulfur. Much of the sulfur escapes from the sea surface into the atmosphere where it is oxidized to SO_2. It may be responsible for natural variations in acid rain (Robinson 1995). It has been argued that an increase in the phytoplankton producing DMS might increase global cloudiness and increase the planetary albedo, causing global cooling (Bryant 1997).

Atmospheric dust concentrations are known to have been much higher during times of glaciation than at present (de Angelis et al. 1987; Petit et al. 1990; Sirocko and Lange 1991; McTainsh and Lynch 1996; Yung et al. 1996; Basile et al. 1997), but this is thought to be a result of changes in vegetative cover rather than a cause of glaciation. In general it is thought that, because of their relatively short residence times, changes in aerosol concentration are unlikely to be a cause of climatic deterioration and glaciation.

Polar stratospheric clouds

Polar stratospheric clouds offer a polar-warming specific alternative to increased atmospheric CO_2 (Sloan et al. 1992; Sloan and Pollard 1998). Stratospheric clouds result from frozen water vapor at levels where the temperature lies below –80 °C. Their effect is to produce polar warming by radiative warming in the lower stratosphere. To produce extensive polar stratospheric clouds, it is necessary to introduce more water into the lower stratosphere. Sloan et al. (1992) proposed that this resulted from higher tropospheric methane levels. The methane in the troposphere diffuses into the stratosphere where oxidation produces CO_2 and H_2O. Sloan and Pollard (1998) concluded that, although the effect would be important, it is alone insufficient to account for all of the polar warming of the Eocene.

Polar vegetation

Changes in polar vegetation from evergreen forests to tundra have been cited by Otto-Bliesner and Upchurch (1997) and DeConto et al. (1998) as making a significant contribution to polar cooling. The forests have a low albedo because snow falls from the trees onto the ground, but the tundra has a high albedo reflecting radiation. Although this enhances global cooling, it is not considered a primary cause.

Relative importance of the proposed causes

The causes proposed to explain the Cenozoic cooling trend can be grouped in terms of the present perception of their relative importance. Educated opinion has changed over the past few decades, and will undoubtedly change in the future.

As potential causes of the long-term Cenozoic cooling trend, the major contenders are vertical motions of the crust ("uplift") and concentration of greenhouse gasses in the atmosphere. The most important potential causes of short-term (glacial-interglacial) climate change are variations in the earth's orbital motions (Milankovitch parameters and passage through the zodiacal cloud). The favorite trigger mechanisms for inducing climate steps include changes in surface ocean circulation resulting from the opening and closing of interocean gateways and rare, extended, "cool summer" Milankovitch configurations. Mechanisms which enhance polar cooling include changes in sea-ice cover, reduction of occurrence of polar stratospheric clouds, and change of polar vegetation from evergreen forest to tundra. Possible causes of climate change which are currently "out of favor" include variations in solar emissivity (except for decadal–centennial time scales), changes in the concentration of cosmic dust, and increased aerosol content of the atmosphere.

In the 1980s, declining levels of atmospheric CO_2 were the favored mechanism for the long-term trend.

During the 1990s, uplift became the most widely accepted, causative factor.

The paradigm of Late Cenozoic uplift

The notion of widespread Late Cenozoic uplift became a paradigm of 20th century geology. Dana (1856), on the basis of geomorphologic arguments, believed that the Appalachians had been rejuvenated during the Late Cenozoic. Stille (1936) argued that younger sediments are more abundant than older ones because the rate of orogeny increases with time. Newell (1949) proposed that the central Andes have undergone rapid uplift since the beginning of the Pliocene, an idea recently reaffirmed by Jordan et al. (1997). De Sitter (1952) stated that the Atlas, Pyrenees, and Alps had all been uplifted in the Pliocene and Quaternary. Büdel (1955) described young uplift of the Hoggar Range in the Sahara. Axelrod (1957) argued for very young uplift of the Sierra Nevada of California. Klüpfel (1957) suggested that the Paleozoic mountains of central Europe had been rejuvenated in the Late Neogene. Barbier (1957) proposed that the "sugar loafs" of Brazil were the result of Quaternary removal of a thick weathered mantle. Trümpy (1960) believed that the Alps, which initially formed at the end of the Oligocene, had been reduced to low elevations by the end of the Miocene and were rejuvenated during the Pliocene and Quaternary. Subsequently, many authors proposed that young uplift affected the Sierra Nevada of California (Axelrod 1962; Christensen 1966; Huber 1981; Chase and Wallace 1986; Unruh 1991), the Southern Alps of New Zealand (Suggate 1963; Wellman 1974; Koons 1989), the Himalayas (Curray and Moore 1971; Derbyshire 1996; Fort 1996; Einsele et al. 1996; Copeland 1997), the East African Rift (Baker and Wohlenberg 1971; Mahaney 1987; Coetzee 1987), the Southern California Borderland (Doyle and Bandy 1972), the polar Urals (Maksimov 1973), Indonesia (Tija et al. 1974), the mountains of Scandinavia (Mörner 1977; Cloetingh and Kooi 1992; Hjelstuen et al. 1999), Tibet (Hsu 1978; Xu 1981; Powell 1986; Copeland et al. 1987; Harrison et al. 1992; Molnar et al. 1993; Li et al. 1997; Copeland 1997; Zheng et al. 2000), parts of the Andes (Jordan et al. 1983; Benjamin et al. 1987; Strecker et al. 1989), the Rocky Mountains and High Plains (Epis and Chapin 1975; Gable and Hatton 1983; Sahagian 1987; Hay et al. 1989), the Transantarctic Mountains (Tingey 1985; Webb et al. 1986; Wilch et al. 1989; Behrendt and Cooper 1991), the Altai and Tien Shan (Cunningham et al. 1996; Métevier and Gaudemer 1997), much of southeastern Asia (Métivier et al. 1999), Japan (Harayama 1992; Momohara 1994), and the Apennines (Coltori et al. 1996). A number of other regions, including such diverse areas as the Paleozoic fold belts forming the Australian Alps and Great Dividing Range (Kemp 1981; Galloway and Kemp 1981), and the Kalahari Plateau (Partridge 1997), have all been described as having undergone uplift in the Pliocene and Quaternary (Fig. 1).

Many of the arguments for Late Neogene and Quaternary uplift were based on geomorphologic (e.g. peneplains, river terraces) or tectonic evidence, including raised shorelines far from the regions occupied by Quaternary ice sheets. Others, such as those of Curray and Moore (1971) concerning the Himalayas, Goerler et al. (1988) for the High Atlas, Hay et al. (1989) for the eastern Rocky Mountains and High Plains, and Métevier and Gaudemer (1997) for the Tien Shan, are based on the masses of sediment eroded during the Late Neogene and Quaternary. Hay et al. (1989) noted the coincidence of the development of the Laurentide ice sheet and the uplift of western North America, indicated by the large volumes of sediment delivered to the Gulf of Mexico.

In a longer perspective of earth history, Hay and Wold (1990) proposed that the maxima of sediment deposition seen in the mass-age distribution of Phanerozoic sediments mainly reflect mountain building, but they also correspond to times of climate diversification and increase of the meridional temperature gradient. The correlation between increased erosion and sediment deposition rates and "mountain building" is supported by the Phanerozoic strontium isotope curve (McArthur et al. 2001; Hay et al. 2001). Eyles (1993) argued that there is a recurring relationship between the opening of ocean basins, passive margin uplift, and glaciation. Moore and Worsley (1994) found that there is a close relationship between orogenic activity and glaciation extending throughout the history of the planet.

Independent evidence suggesting increased tectonic activity lies in the record of tephra deposits in the deep sea. A significant Late Cenozoic increase in tephra frequency was noted by Kennett and Thunell (1975, 1977). Kennett (1981), on the basis of further analyses, found that the number and thickness of tephra layers in oceanic sediments increases significantly in the Pliocene and Quaternary, and he related this to increased volcanism. Cambray and Cadet (1994) studied the peri-Pacific arc volcanism and also concluded that there was a sharp increase in Pliocene and Quaternary volcanism. In a recent, more detailed study, Straub and Schmincke (1998) presented detailed results of analysis of tephra occurrences in the Pacific. They concluded that overall 23% of the sediment in the Pacific basin is volcanic. Again, they found that the frequency of tephra increased markedly in the Pliocene and Quaternary, with a less pronounced peak in the Middle Miocene. Increased volcanism is not a certain indicator of increased mountain building, but a relationship seems likely. Ronov (1959) had already noted the relationship between volumes of volcanic rocks and volumes of sediment, and concluded that volcanism and tectonic uplift were intimately linked, a conclusion borne out by his later compilations (Ronov 1980, 1982, 1993).

Many geologists have considered the local evidence for Pliocene–Quaternary uplift in various parts of the world to be compelling. Although it is difficult to imagine a single mechanism which could affect so many different areas, intra-plate compression (Cloetingh and

Kooi 1992) as a result of global plate reorganization is a possibility.

Definitions of "uplift"

One aspect of the uplift paradigm is semantic. A number of related terms are used in discussion of vertical motions of the surface of the solid earth in geology (England and Molnar 1990). The term "uplift" is used ambiguously by geologists to refer either to the motion of the surface of the solid earth or to the motion of rocks. Motion of the surface of the solid earth should be described with respect to the geoid. England and Molnar (1990) defined the "surface" as the average elevation of an area on the scale of the square of the thickness of the continental crust or lithosphere. They used the expression "uplift of rocks" to refer to the displacement of rocks relative to the geoid. Because of erosion, the uplift of rocks may be greater than that of the surface. England and Molnar (1990) defined the upward motion of rocks with respect to the surface as "exhumation", although this term is often used by geologists in the more restricted sense of uncovering an ancient landscape represented by the topography of rocks beneath an unconformity (Jackson 1997). In descriptions of erosion, the term "denudation" is used to describe the rate at which a surface is lowered by removal of soil and rock, not taking into account isostatic adjustment. In the following discussions we use "denudation" as being synonymous with "exhumation" sensu England and Molnar (1990).

Because elevations are usually expressed with reference to sea level, either present or past, motions of the sea surface play a special role in discussions of uplift and climate change. Eustatic (=global) sea-level changes imply relocation of water from the ocean to ice sheets, groundwater, lakes and marginal seas, or changes in the volume of the ocean basins. The relocation of mass changes the shape of the geoid, and the large ice sheets of the northern hemisphere have formed and decayed more rapidly than the earth can respond. To add to the complexity of the problem, ocean water was attracted towards the ephemeral massive ice sheets. Changes of eustatic sea level are accompanied by isostatic adjustments of both the seafloor and the land. Because the geoid, the surface of the solid earth and sea level all move vertically, it would be best to view motions with respect to distance from the center of the earth. This is presently not possible, and hence there are always ambiguities associated with eustatic sea level and elevations on land and beneath the sea (Kendall and Lerche 1988; Dott 1992).

The hypothesis of the illusion of Late Cenozoic uplift

Molnar and England (1990) have challenged the Late Cenozoic uplift paradigm, proposing that it is illusory. They questioned the evidence for Pliocene–Quaternary uplift in many parts of the globe, suggesting that the uplift was only apparent, not real. England and Molnar (1990) argued that many of the estimates of rapid uplift were mistakes in interpretation. It should be noted that they were discussing uplift due to processes other than isostatic adjustment, and hence excluded the rise of a surface in response to removal of load from their definition of uplift. Isostatic adjustment follows the average removal of load over an area, having dimensions of the order of the thickness of the lithosphere, e.g., 10^4 km^2 or more (ca. 1° latitude by 1° longitude). England and Molnar (1990) contended that geodetic determinations of uplift, being based on points, are determinations of the rate of upward motion of discrete rock masses, not of the upward motion of the areally averaged surface. They cited Schaer and Jeanrichard's (1974) estimates of Late Neogene uplift rates of 1 km per 10^6 years in the Central Alps as an example of this kind of error in interpretation. As another example, these authors cited De Sitter's (1952) projection of a Miocene peneplain surface to the crest of the Pyrenees, to conclude that 2 km of uplift had occurred. England and Molnar (1990) argued that most of the apparent uplift could be due to isostatic compensation in response to erosion and dissection of the mountain range.

Upward motion of rocks relative to the land surface can be deduced from fission-track analysis of apatites, which allow the time when a rock presently at the surface passed through a geotherm to be determined. Knowing the geothermal gradient, the geotherm can be related to a geobarometric surface, and its depth determined. The best resolution which can be obtained is when the rock passed through the 50–100 °C isotherm (Faure 1986). However, there are a number of uncertainties associated with the history of the geothermal gradient which could affect the estimate of depth. One of these is climate change. A cooler climate lowers the temperature of the land surface. The seasonal variations reach only a depth of about one to two meters, where the temperature reflects the annual average. Hence, long-term climate change alters the geothermal gradient. Finally, unless there is a direct relationship between the rate of erosion and the rate of uplift, and unless erosion is the only means by which the surface can be lowered (e.g., tectonic unroofing can be ruled out), it is not possible to determine the motion of the land surface from the geobarometric measurement. England and Molnar (1990) concluded that rates of "exhumation" cannot be used to determine either the magnitude or direction of motion of the land surface. They note that there is a relationship between elevation and erosion rates, but that this varies over an order of magnitude and also depends on rainfall. Hence, they argued that elevation cannot be estimated from erosion rates. Finally, they concluded that "We are aware of no reliable, quantitative estimates of rates of surface uplift in mountain ranges that place useful constraints on tectonic processes" (England and Molnar 1990, p. 1176).

Molnar and England (1990) noted that many areas cited as having undergone Plio–Quaternary uplift have had

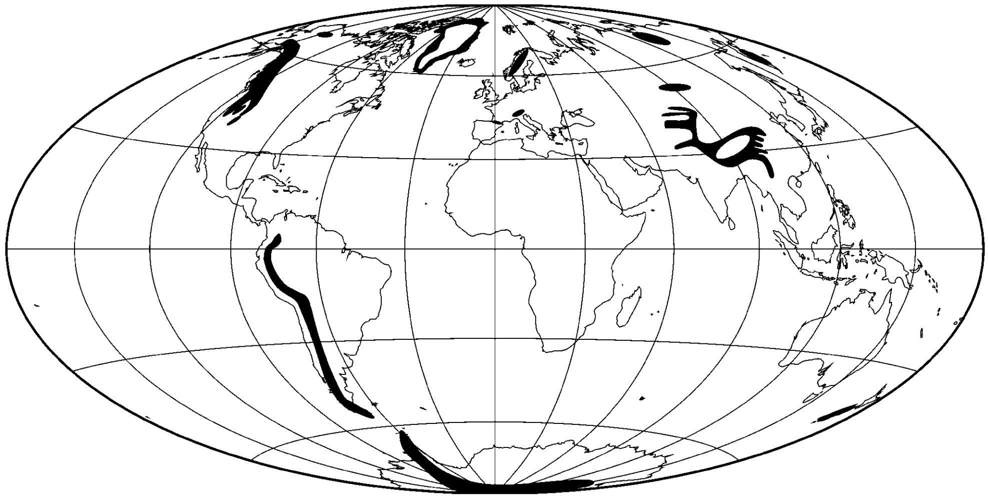

Fig. 4 Regions of Late Neogene–Quaternary mountain glaciation (after Gerasimov 1964)

very different geologic histories, and could find no underlying cause why so many different areas should be affected. They concluded that the erosion system operated differently before the onset of the northern hemisphere glaciations. They argued that erosion rates increased as a result of the climatic changes, and that most of the uplift which has taken place is isostatic response to the erosion of valleys. They suggested that part of the effect might be that mountain glaciers widened valleys from a "V'' to a "U'' shape, providing sediment and causing isostatic uplift of the mountain peaks by making the entire mountain range lighter.

Wold et al. (1989) made a calculation to evaluate the process of valley formation in the erosion and elevation history of the Alps. They found that enough sediment was removed in producing the valleys in the Alps to cause 800-m uplift of the "Gipfelflur" (level of concordance of summits) in response to the lightened load of the mountain range. Thus, about 25% of the present elevation of the peaks in the Alps can be attributed to formation of the valleys.

Montgomery (1994) made a comparison of the effects of valley incision on the isostatic uplift of mountain peaks in the Sierra Nevada of California, the Tibetan Plateau, and the Himalayas. He concluded that isostatically compensated valley incision can only account for 5–10% of the present elevation of peaks in the Sierra Nevada and on the Tibetan Plateau. However, he found that as much as 20–30% of the elevation of the Himalayan peaks could be due to incision of the deep valleys in the region.

On a global scale, the glacial reshaping of valleys may be an important process in causing the uplift of mountain ranges, but not all of the areas in which uplift is thought to have occurred have been sculpted by mountain glaciers (Fig. 4).

The paleoaltitude problem

The problem of distinguishing real and illusory uplift could be settled if there were some means of measuring the altitude of a surface in the past. Six indirect methods have been proposed for estimation of elevation, based on (1) the air pressure at the time of a volcanic eruption, (2) mean annual temperature determined from analysis of plant communities, (3) the amounts of sediment delivered to the surrounding basins, (4) the grain size of sediments delivered to the surrounding basins, (5) lithospheric thickness, and (6) O isotopic composition of minerals formed by weathering.

Estimating paleoaltitude based on air pressure

A nearly direct means of measurement of paleoaltitude has been proposed by Sahagian and Maus (1994) who found that vesicle size distribution in basalts, being a function of the atmospheric pressure plus hydrostatic pressure of the overlying lava, can be used to infer atmospheric pressure and hence assess paleoelevation. Investigations on flows on Mauna Loa (Hawaii) showed that the technique is accurate to about 0.1 bar (=100 hectapascals), and hence has a resolution of about 1,400 m.

The method has not yet been applied to the problem of Cenozoic uplift.

Estimating paleoaltitude based on paleofloras

The most commonly used measure of paleoaltitude is through the analysis of floras, although this has only been used in a few regions (for the southern Rocky Mountains: MacGinitie 1953; Axelrod and Bailey 1976; Meyer 1986; Gregory and Chase 1992, 1994; Wolfe 1992b; Gregory and McIntosh 1996; Wolfe et al. 1997; for Tibet: Hsu 1978; Xu 1981; Powell 1986; Spicer et al. 2001; for Japan: Momohara 1994; for the Andes: Wijninga 1996a, 1996b). The use of fossil plant assemblages has been criticized because the localities where the fossils occur may represent special environments in upland areas, such as valley plains where sediment occurred. Oligocene plant assemblages from Florissant,

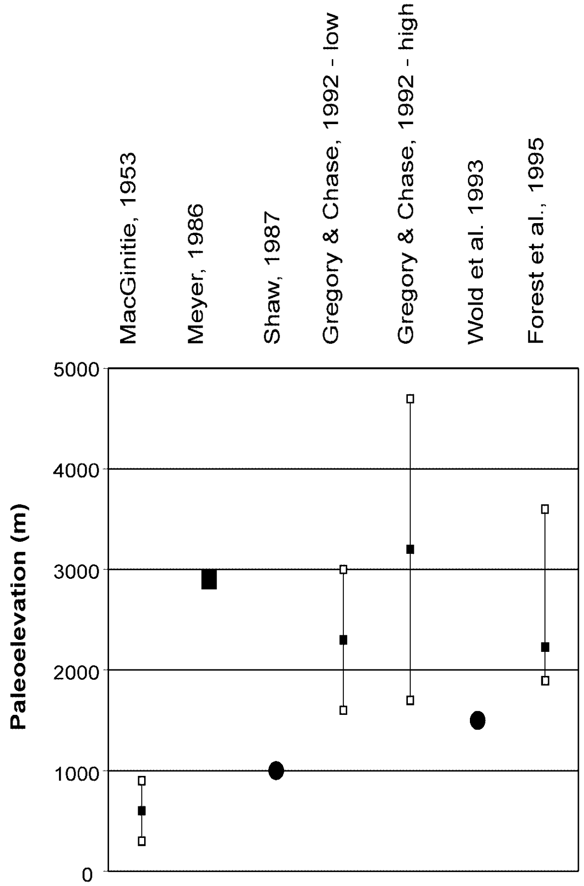

Fig. 5 Estimates of the paleoelevation of the Oligocene Florissant (Colorado, USA) fossil site at the time of its deposition (for discussion, see text)

Colorado, represent a special case because it can be demonstrated that the locality is related to a large erosion surface with little relief.

The original method used floristics, the direct comparison of the species and genera present in the fossil assemblage with modern assemblages and assuming that the fossils lived under the same climatic conditions as those in the modern assemblage. The floristics method assumes that the climatic tolerances of taxa have not changed with time. It was used by MacGinitie (1953) to estimate that, after the formation of the Late Eocene erosion surface, the Oligocene plant and insect-rich deposits at Florissant, Colorado, presently 2,500 m above sea level, formed at elevations of 300 to 900 m (Fig. 5).

Paleoclimate interpretation from floristics has been replaced by the foliar physiognomic method which assumes that leaf characters such as size, shape, thickness, and margin type reflect adaptation towards maximizing efficiency under a particular suite of climatic conditions (Givnish 1987). Foliar physiognomic analysis uses leaf and wood characteristics to assign an assemblage to a vegetation type. A mean annual temperature (MAT) for the vegetation type can be determined by analogy with the temperature parameters characteristic of modern vegetation types. Wolfe (1978) developed the physiognomic method in his description of the development of Tertiary climates in North America. Foliar physiognomic analysis of Late Cretaceous floras from North America was used by Wolfe and Upchurch (1987) to estimate both paleotemperature and precipitation. This method of interpretation of paleoclimate from fossil leaf assemblages assumes that genera and species may have adapted to different conditions, but that the physiologic response remains constant (Chaloner and Creber 1990).

Axelrod and Bailey (1976) presented a detailed discussion of the methodology involved in interpreting paleoaltitude from the paleotemperatures indicated by the fossil flora. Meyer (1986) made a new estimate of the original elevation of the Oligocene Florissant site in Colorado (Fig. 5). He proposed determining paleoaltitude by subtracting the MAT for the vegetation type indicated by physiognomic analysis of fossil leaf assemblages from an estimate of what the MAT at sea level would be for the same age and place, and then dividing this difference by the “terrestrial lapse rate” (change in temperature with elevation measured on the surface). He noted that the determination of the MAT at sea level for a given site is complicated by three factors: continentality, elevated base level, and latitude. The continentality effect is a result of distance from the moderating effect of the sea. The elevated base-level effect reflects the greater heating of an elevated surface due to greater receipt of insolation via the shorter path through the atmosphere. Projected down to sea level, the surface temperatures are warmer than that of the sea. In Colorado the combined effect of continentality and elevated base level is estimated to increase the sea-level MAT by about 6 °C. Using the global mean lapse rate of 5.98 °C km^{-1} assumed by Meyer (1986), the correction for continentality and elevated base level corresponds to an elevation difference of 1 km. The latitude effect describes the meridional change in temperature. Meyer (1986) used the change along the west coast of North America of 0.33 °C degree^{-1} latitude at sea level, which closely approximates the modern average meridional temperature gradient for the northern hemisphere. His final estimate of the paleoelevation of the Florissant site in the Oligocene was 2,900 m, 400 m above its present level.

Wolfe (1990) noted that earlier studies had used only one leaf variable at a time to relate to a climate parameter; the results suggested that the ratio of species with toothed versus smooth leaf margins was most closely related to MAT. In his study of changes across the Cretaceous-Tertiary boundary, he introduced a new climate–leaf analysis multivariate program (CLAMP) which attempts to interrelate many variables with climate parameters simultaneously. The results were presented in terms of a nonlinear two-dimensional temperature–wetness space. Earlier estimates of the temperature change based on the ratio of the number of species with toothed margins had indicated no temperature difference across the boundary (Wolfe and Upchurch 1986). According to CLAMP, the temperature change across the boundary was a 10 °C increase.

Gregory and Chase (1992) made a reassessment of the paleoaltitude of the Florissant flora. They used multiple regression analysis of five leaf characters and temperature to develop a predictive equation for MAT. They obtained an MAT of 10.7±1.5 °C for the Florissant flora. They considered the greatest uncertainty in Meyer's (1986) estimate of the paleoelevation to lie in the projection of the mean annual temperature at sea level beneath the site, and proposed two different methods to refine it. They followed Meyer in assuming that the projected sea-level mean annual temperature increases inland due to the continentality effect but noted that it is difficult to evaluate a continentality-elevated base level effect. For one method (Fig. 5: "Gregory and Chase 1992 low") they chose the coeval Oligocene La Porte flora from northeastern California as representative of the sea-level temperature in Colorado because the present MAT of the La Porte site approximates the projected sea-level temperature in northern Colorado today. They determined the MAT of the La Porte flora, corrected for 1° latitude difference since the Eocene, to be 22.8±1.5 °C. Using the global lapse rate of 5.89 °C km^{-1} of Meyer (1986), they determined a paleoaltitude of 2,300±700 m. For the second method (Fig. 5: "Gregory and Chase 1992 high"), the inland temperatures were compared directly with those of the coast (Wolfe 1992b), and no adjustment was made for latitude effect. They chose the coeval Goshen flora of Oregon to represent the coastal temperature, determining its MAT to be 20.3±1.5 °C. They used Wolfe's (1992b) mean regional lapse rate of 3.0 °C km^{-1} to determine a paleoelevation of 3,200±1,500 m. The two estimates are 200±700 m lower, and 700±1,500 m higher than the present elevation, giving a total possible elevation range of 1,600 to 4,700 m. They arrive at the conclusion that the Late Eocene erosion surface at Florissant and the High Plains of Colorado have been at about their present elevations for 35×10^{-6} years.

Gregory and Chase (1992) stated that from 35 Ma until the last few million years, when deep incision has taken place, this region had a denudation rate less than 0.003 mm year^{-1}. This is comparable to that reported by Lisitzin (1972) for the Lena River basin of Siberia and has not been appreciably revised by more recent studies (Stein 2000). This is one of the lowest denudation rates for a major river basin, but characteristic of the rivers of the Siberian Arctic. Gregory and Chase (1992) conclude that this is strong evidence in support of the Molnar and England (1990) hypothesis that the global erosion system operated differently before the onset of northern hemisphere glaciation. However, the Lena is an Arctic river which flows only a few months each year. The mean annual temperature of the Lena River basin is the lowest of any large river, –8.6 °C (Ludwig et al. 1996), and like the other Siberian Arctic rivers, much of its basin is underlain by permafrost and hence is not easily eroded.

It is evident that the major factor in determining paleoelevation using the floristic methods discussed above is the choice of a "terrestrial lapse rate". This should not be confused with the "lapse rate" (or "free-air lapse rate") used by meteorologists. The free-air lapse rate is the pressure-induced, vapor-moderated decline in temperature of air with elevation measured by weather balloons. Lapse rates are discussed in their ecological context by Hay (1996). The lapse rate for dry air is uniformly about 10 °C km^{-1}. The lapse rate for air saturated with vapor is less because of the conversion of vapor to water and release of latent heat; the "wet" lapse rate is a nonlinear function of temperature and varies from nearly 10 °C km^{-1} for very cold air to 4 °C km^{-1} for tropical temperatures. "Global" lapse rates have been established by integrating the records of many weather balloons; the most commonly cited global lapse rates are 6.5 °C km^{-1} (Barry and Chorley 1982) and 6 °C km^{-1} (Peixoto and Oort 1992). The "topographic lapse rate" is the rate of decrease of temperature measured a couple of meters above the surface as one ascends a mountain range. It is less than the lapse rate measured by the vertically ascending balloon because of the transfer of sensible heat (that is, the form of heat representing the average rate of molecular motion and measured by a thermometer) from the ground. The "terrestrial lapse rate" used in the floristic studies is a regional version of the topographic lapse rate. However, because of its regional nature it may be influenced by factors other than elevation. In the Gregory and Chase (1992) study, the two lapse rates used differed by a factor of two. Choice of one or the other would double or halve the paleoelevation estimates to 4,000±1,500 and 1,600±700 m for a total range of 900 to 5,500 m.

To avoid the problem of choice of lapse rate, Forest et al. (1995) proposed to use energy conservation principles to estimate paleoaltitude from fossil leaf assemblages. Instead of correlating the leaf assemblages with MAT, they make a correlation with the total specific energy content of the air, h, defined as

$$h = c_p T + L_v q + gZ = H + gZ$$

where c_p is the specific heat capacity at constant pressure of moist air, T is temperature (K), the term c_pT is the "sensible heat" of the air, L_v is the latent heat of vaporization, q is the specific humidity, $L_v q$ is the "latent heat" of the air, g is gravitational acceleration, Z is altitude, and gZ is the "potential energy" of the air. H is the moist enthalpy, the sum of sensible and latent heat contents of the air. It should be noted that in meteorology, the term enthalpy is synonymous with the sensible heat content of the air (Barry and Chorley 1982), so Forest et al.'s (1995) distinction of H as moist enthalpy is appropriate. Knowing the moist enthalpy at sea level and at the site where the floral assemblage occurs, the elevation, Z, is

$$Z = \frac{H_{sealevel} - H_{site}}{g}$$

Forest et al. (1995) plotted the spatial distribution of mean annual static moist energy for North America and found the lines are nearly zonal, arcing northwards only about 5° latitude across the Rocky Mountains. Making the assumptions that the surface moist static energy is zonally symmetrical and that the estimates of enthalpy for the regional paleoclimate (derived from 29 foliar physiognomic characteristics) are applicable, they determined the paleoelevation at Florissant to be 2.9±0.7 km with respect to the present geoid (Fig. 5).

All of the attempts to use floristics to determine paleoelevation suffer from the problem that *the vegetation, through evapotranspiration, modifies the climate*. In the continental interior the specific humidity is largely a function of the transpiration by the vegetation. Furthermore, vegetation has evolved rapidly during the Cenozoic, with water-conserving C_4 plants replacing more freely transpiring C_3 plants over large areas about 8×10^6 years ago (Cerling 1997). Recent climate model simulations (DeConto 1996; DeConto et al. 1999a, 1999b). show that the effect of vegetation on the climate of the Cretaceous was much larger than had been anticipated. The general spread of water-conserving C_4 plants in the Late Cenozoic should have had the effect of reducing the specific humidity in the continental interiors. This should have increased the vertical lapse rate in the interior of the continents. It should also have increased regional lapse rates and reduced the moist enthalpy in the continental interiors. If the lapse rates over land were less before the spread of C_4 plants, then paleoelevation estimates based on floristics and on moist enthalpy would need to be revised upwards.

Estimating paleoaltitude based on eroded sediment

Most sediment is detrital, and the most obvious factor affecting the load of detrital sediment carried by rivers is the relief of the drainage basin (Pinet and Souriau 1988; Summerfield and Hulton 1994; Hay 1998). It is also widely believed that steeper slopes promote the transport of coarser sediment. Accordingly, the sediment masses eroded from upland areas and the coarseness of the resulting deposits have been used to infer the elevation of the source areas.

Estimating paleoaltitude based on masses of sediment

Shaw (1987) used mass-balanced replacement of sediment presently in the Gulf of Mexico to its site of origin in the interior of North America to reconstruct paleotopography back to 65 Ma. Shaw made the assumption that all of the sediment derived from the present Mississippi and western Gulf Coast drainage had been deposited in the Gulf of Mexico. His maps indicate an elevation for Florissant at 35 Ma of about 1,000 m. The method of making the topographic reconstructions was described in detail in Hay et al. (1989). The sediment replacement involves a formula for denudation rate proportional to elevation:

$$\frac{\delta T d_e}{\delta t} = (H - Ed_b) Ktd_e$$

where $\delta Td_e/\delta t$ is the rate of mechanical denudation expressed in terms of the thickness of detrital sediment eroded each year per meter of elevation above the detrital erosional base level, H is the elevation of the grid point, Ed_b is the elevation of the detrital erosional base level, and Ktd_e is a constant expressing the rate of denudation in terms of the thickness of the solid phase of material eroded each year per meter of elevation.

Shaw (1987), Hay et al. (1989), and Shaw and Hay (1990) used Ed_b=200 m and Ktd_e=0.113×10^{-6} m m^{-1}_{elev} $year^{-1}$; these values were based on a global average of large rivers, but coincidentally correspond to the values for the Mississippi River basin. Wold et al. (1993) noted that not all of the sediment eroded from the present-day Mississippi drainage in the continental interior was necessarily deposited in the Gulf of Mexico. Comparison of their Figs. 1 and 2 (mass-age distributions for the Gulf of Mexico and Atlantic margins of the United States, respectively) suggests that as much as one third of the Oligocene sediment eroded from the continental interior might have been deposited on the Atlantic margin, probably via the ancient St. Lawrence drainage system. Adding this sediment back into the continental interior would raise the estimate of the elevation at Florissant in the Oligocene to about 1,500 m (Fig. 5).

Hay et al. (1992) used the sediment mass balance to estimate the elevation history of the Alps since the Oligocene. Their reconstructions generally agreed with the estimates of the elevation of the young Alps which had been made by Hantke (1984, 1985). They also noted that the sediment data suggest that the Alps were low during the Late Miocene and Pliocene, as had been suggested by Trümpy (1960) and recently supported by Wagner (1996). The idea that the Alps had been reduced to low elevations by the end of the Miocene rests on interpretation of the Augensteinschotter, quartzitic gravel which rests on the calcareous plateaus of the Eastern Alps, as the deposits of braided streams, but other interpretations are possible (Kuhlemann 2000), and the age of the Augensteinschotter has been revised to Early Oligocene–Early Miocene (Frisch et al. 2001). However, the recent and thorough account of the erosion history of the Alps by Kuhlemann (2000) also suggests lower elevations in the Late Miocene–Early Pliocene, *assuming that erosion rates are directly related to elevation*.

The value used for the thickness of the solid phase of material eroded each year per meter of elevation (Ktd_e of Hay et al. 1992) is critical to the reconstruction of elevation. Doubling the value will halve the elevations. In discussing the erosion history of the Appalachians, Wold et al. (1993) used the local Ktd_e value of 0.024×10^{-6} m m^{-1}_{elev} $year^{-1}$, determined from Curtis et al. (1973). From the data given by Allen (1997), the global average value of Ktd_e is 0.115×10^{-6} meters per meter of elevation per year (m m^{-1}_{elev} $year^{-1}$), with the maximum being 0.905×10^{-6} m m^{-1}_{elev} $year^{-1}$ for the Red River of Vietnam, and the minimum being 0.003×10^{-6} m m^{-1}_{elev} $year^{-1}$ for the Lena and Yenisei rivers of the Siberian Arctic. Because of the marked regional variations, evaluation of the local erosion rate parameters is the most uncertain aspect of mass-balanced paleotopographic reconstruction.

Gregory and Chase (1992) noted that the elevation they proposed for the Florissant site in the Oligocene would require that the post-Eocene to Pleistocene erosion rate be equal to that of the modern Lena River basin, i.e., the lowest on earth. We find this difficult to accept because the Late Eocene erosion surface over the Colorado Front Range formed a continuous surface with the Great Plains to the east. Prior to its erosion, the surface over the Great Plains was underlain by unconsolidated, easily erodible Cretaceous and Early Cenozoic shales and sands. The slope between Colorado and the Gulf of Mexico would have been greater than 1 m per km. The only way to protect the unconsolidated sediment from erosion would be to have either no runoff or a vegetation cover so dense that the detrital material was firmly held in place by the plants at all seasons of the year. Yet this is the area which is supposed to have developed from forest to savanna during the later Cenozoic (Wolfe 1978), forcing the evolution of ungulate taxa, such as horses and camels (Webb 1983). Furthermore, savannas should, according to the studies of Langbein and Schumm (1958), have maximal erosion rates.

Estimating paleoaltitude based on the grain size of sediments

In their atlases of Paleozoic and Mesozoic–Cenozoic paleogeographic maps, Ronov et al. (1984, 1989) distinguished between lowland, upland and highland areas. In the text they stated that this is based on the volume and grain size of the sediments delivered to the surrounding basins, but the specifics were not given. Coarse sediments, such as conglomerates, are often taken to be evidence for steep gradients in streams and hence an indication of the elevation of the hinterland.

Gravel and conglomerate have often been cited as evidence for upland elevation in the source area. Potter (1955) inferred Late Cenozoic uplift of the southern Appalachians from the Lafayette gravel. Hantke (1984, 1985) cited the coarse conglomerates of the Molasse as evidence for high elevations in the Alps at the end of the Oligocene and in the early Miocene. However, Frostick and Reid (1989) found that infrequent large storms can cause abrupt coarsening of fanglomerate material, mimicking the effect of tectonic uplift.

Garner (1959), in a study of sediments produced by erosion in different areas of the Andes, had noted that the grain size of material produced from the same lithologies differed markedly according to the climate of the region. In dry areas the sediments were coarse, in wet regions, fine-grained. It has become evident that sediment grain size can be a function of both elevation and climate.

Estimating paleoaltitude from the thickness of the lithosphere

Assuming that a region has always been isostatically adjusted, its paleoaltitude at times in the past could be determined by knowing the history of change of thickness of the lithosphere. Jordan et al. (1997) used the crustal shortening in the Andes to estimate lithospheric thickening and uplift of the central Andes. From analysis of the structural geology of the region, they concluded that parts of the central Andean region were at or near sea level in the Cretaceous. The initial shortening occurred in the Eocene (Incaic deformation) but was regionally restricted; it accounted for one-quarter to one-half of the total Cenozoic thickening in the Western Cordillera. The major phase of thickening occurred since the mid-Oligocene, accounting for most of the uplift to the present elevation in excess of 3 km.

This method is limited to regions where the structural deformation permits reconstruction of the undeformed lithosphere, and sedimentary deposits give an indication of sea level prior to the deformation.

The inverse of this method, reconstructing the history of the lithosphere from paleoelevation estimates (based on fossil floras), has been described by Gregory and Chase (1994). They conclude that about 24 km of crustal thickening occurred beneath Colorado during the Laramide Orogeny, and that the elevation of the region has been declining slowly ever since in response to erosion and isostatic adjustment.

Estimating paleoaltitude of mountain belts from the isotopic composition of authigenic minerals

Chamberlain and Poage (1999) proposed that the paleotopography of mountain belts can be estimated by the rain-shadow depletions of ^{18}O and D in rainwater which are incorporated into authigenic minerals in the lee of the mountains. From early studies they propose an ~2-km uplift of the Southern Alps of New Zealand since the Early Pliocene, but no change in the mean elevation of the Sierra Nevada of North America during the past 16 Ma.

What controls detrital sediment yields?

The argument over whether there has been Late Neogene uplift hinges on whether erosional processes have accelerated, creating the impression of tectonic uplift. Two changes which may have accelerated erosion are climate change and the spread of C_4 plants.

The detrital sediment yield from a drainage basin is the mass eroded per unit time; denudation rate is the thickness eroded per unit time. Although in the short term detritus may be stored temporarily in flood plains, on geologic time scales erosion in the source area equals deposition in the surrounding basins. Hay (1998) reviewed data and ideas on the delivery of detrital sediment to the sea. He concluded that major controls on detrital yields are geology, elevation, and climate. In large basins, the geology is varied and the relative susceptibility of different lithologies to erosion are thought to average out; hence, the details of the geology are usually neglected. Five particularly relevant works have appeared since Hay (1998) was sent to press: Ludwig et al. (1996), Allen (1997), Hovius (1998), Harrison (2000), and Hooke (2000).

Elevation

Hay (1998) noted that a number of studies indicate that the relief of the drainage basin is the dominant factor affecting the sediment yield and denudation rate.

Milliman and Syvitski (1992, 1994) divided data on 280 rivers into five elevation categories according to the maximum elevation in the drainage basin: high mountain (>3,000 m), mountain (1,000–3,000 m, further subdivided into three regional categories), upland (500–1,000 m), lowland (100–500 m), and coastal plain (<100 m). They concluded that mountainous rivers have detrital sediment yields between 0.140 and 1.700 kg m^{-2} $year^{-1}$, upland rivers have yields between 0.060 and 0.250 kg m^{-2} $year^{-1}$, and lowland rivers have yields between 0.020 and 0.060 kg m^{-2} $year^{-1}$. Milliman and Syvitski (1992, 1994) and Milliman (1997) concluded that small montane rivers are disproportionately important in delivering sediment to the sea.

Summerfield and Hulton (1994) made a detailed analysis of 31 rivers. They found that basin relief, the relief length ratio (maximum elevation/greatest length of the basin), local relief, and modal elevation all correlate with the detrital sediment yield.

Allen (1997) published new information from the extensive Oxford Global Sediment Flux Database. These data cover 97 drainage basins, but not all data are available for all river basins. Figure 6A, B show the relationship between the basin relief and relief/length ratios and the detrital sediment yields for the 97 rivers, and Fig. 6C shows the relationship between mean elevation and detrital sediment yield for 53 rivers There is always much scatter in the data on detrital sediment yields, and the correlation coefficients are always lower for larger data sets. However, the trends indicating that elevation affects the detrital sediment yield are unmistakable.

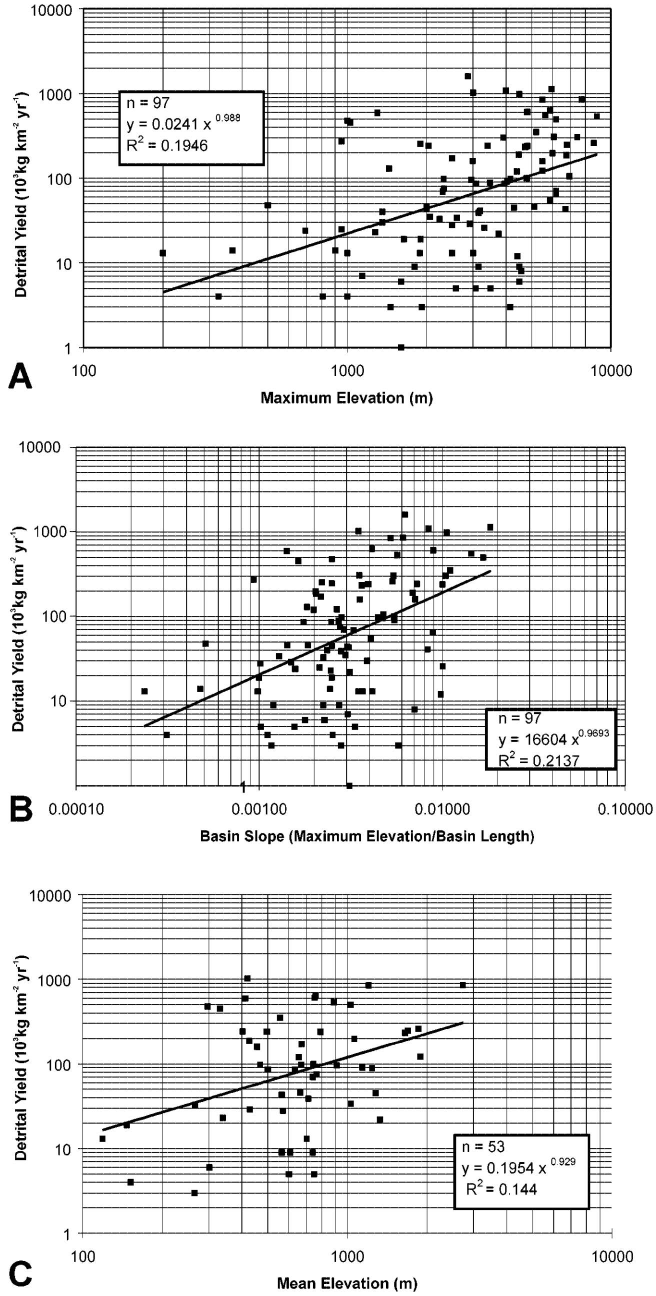

Fig. 6A–C Detrital yield as a function of three elevation parameters in the larger data set on rivers in the Oxford Global Sediment Flux Database published by Allen (1997). **A** Maximum elevation (=basin relief). **B** Basin slope (=relief/length ratio). **C** Mean elevation

Climate

Although it is often stated that there is a link between erosion rates and climate, quantitative relationships between sediment loads and climatic factors have generally been elusive, the clearest one having been documented for Arctic rivers. Emptying into the sea in regions where the mean annual temperature is well below 0 °C, these have anomalously low sediment loads attributed to their short periods of flow during the summer and to the permafrost which underlies their drainage basins. Berner and Berner (1987, 1996) noted that seasonal rainfall, particularly the monsoons which affect southern Asia, is especially important in producing a large suspended sediment load. They also noted that although rivers in desert regions may have high concentrations of detrital sediment, their intermittent flow and low water volumes preclude them from transporting large loads to the sea, so that the sediment yields of Australia and Africa are low.

Summerfield and Hulton (1994) found that precipitation and runoff were correlated with the detrital sediment yield, but did not detect a clear relationship between runoff variability and detrital yield. They concluded that there are two regimes in which the hydrologic cycle plays a major role in denudation: (1) in semiarid regions (annual precipitation between 200 and 400 mm), where the vegetation cover is not continuous, and (2) in areas with very high mean annual precipitation (>1,000 mm), mostly on the borders of orogenic belts.

Ludwig et al. (1996) reexamined the relationships between topography, climatic factors and detrital sediment yield. They found that the sediment flux is best modeled by the equation

$$F_{TSS} = 0.0176(Q * Slope * I_{Fournier}) + 45$$

where F_{TSS} is the detrital sediment yield expressed in t km^{-2} year^{-1}, Q is the water discharge expressed in terms of mm, *Slope* is the basin slope (after Moore and Mark 1986) expressed in 10^{-3} radians, and $I_{Fournier}$ is a modified Fournier index:

$$I_{Fournier} = \sum_{i=1}^{12} PP_i^2 / APPT$$

where PP_i is the precipitation in month *i* and APPT is the total annual precipitation. This modified Fournier index is significantly different from Fournier's (1960) original, which was the square of the precipitation in the maximum month over the total annual precipitation. By taking all 12 months into account, this is more sensitive to the overall distribution of rainfall throughout the year. It should be noted that the values for *Slope* given in Table 1 of Ludwig et al. (1996) are cited as being in radians but are actually expressed in terms of 10^{-3} radians. The Ludwig et al. equation indicates that the detrital sediment yield varies linearly with the runoff, slope, and their modified Fournier index. The increase with runoff can be explained as related to the capacity of the river to transport detritus. The linear increase with increasing slope corresponds to the relationship between elevation and detrital yield. The increase with the modified Fournier index is best explained in terms of the effects of seasonal rainfall on the vegetative cover, and of sudden heavy rains on erosion of loose sedimentary material.

Data from the Oxford Sedimentary Flux Database published by Allen (1997) can be used to compare the relationship between variables thought to affect erosion rates. Hovius (1998) proposed that the mechanical erosion rate is dependent on five variables:

$$\ln(e_m) = -0.416 \ln A + 4.26 * 10^{-4} h_m + 0.15T + 0.095T_r + 0.0015q + 3.585$$

where e_m is the mechanical erosion rate (g m^{-2} year^{-1}), h_m is the maximum elevation in the drainage basin (m), A is the area (km^2), T is the mean annual temperature (°C), T*r* is the annual temperature range (°C), and *q* is runoff (mm year^{-1}). Harrison (2000) used multicorrelation analysis to develop an improved equation which explains more of the variance in the Oxford data:

$$\ln(e_m) = \ln s_4^{1.20} + \ln L_b^{0.50} + 7.52 * 10^{-4} q_m + 0.151T + 0.085T_r - 5.66$$

where the terms are as above, but with $s_4 = \sqrt{A}$, L_b being the basin length, and q_m being the maximum monthly runoff. Hooke (2000) has developed a complete expression from first principles, but numerical values for some of the terms in his equation are only poorly known or unknown. His analysis makes it clear why there is such great variability in the data. In all of these expressions, the elevation term dominates except for extreme values of other terms.

The Late Neogene increase in sediment deposition rates

Analysis of Cenozoic sediment recovered by the Deep Sea Drilling Project (DSDP) indicated that there has been a three-fold increase in delivery of sediment to the ocean basins since the Middle Miocene (Davies et al. 1977). They suggested that the increase may have resulted from the rapidly changing climates of the Late Cenozoic. Rapid climate change could have accelerated weathering by preventing soil profiles from coming to equilibrium. In addition to changing the rate of weathering, climate change would alter the amount of runoff from land. Changing the amounts of water flowing in rivers will alter their competence and overall ability to transport sediment. Without changing their gradient, rivers with more water can transport larger amounts of detrital sediment in suspension and by bottom traction. The discussion by Davies et al. (1977) was very brief, and they did not present any quantitative argument, but suggested that the evidence of increased sedimentation rates was global, and hence must be related to a global phenomenon.

Donnelly (1982) calculated the accumulation rates of alumina at selected DSDP sites in the Atlantic, Pacific and Indian oceans. He found that there has been a six-fold increase in accumulation rates in the northern hemisphere during the last 15×10^6 years, but the signal was more ambiguous in the southern oceans. He argued that tectonic elevation could be discounted as a possible cause because the signal is similar in the "tectonic" Pacific and "nontectonic" Atlantic. He also discounted glaciation because the increase is observed at both low and high latitudes. He concluded that the effect is most likely related to changes in plant cover in response to a change from more equable, earlier Cenozoic climates to Late Tertiary climates "dominated by irregularities in precipitation and in the incidence of catastrophic wet and dry periods" (Donnelly 1982, p. 454).

Perlmutter and Matthews (1990, 1991) discussed the effect of changes in the weathering regimes which should be expected on Milankovitch time scales. They presented a diagram (reproduced in modified form here as Fig. 7) showing the changes which occur when Milankovitch forcing results in glaciation and deglaciat-

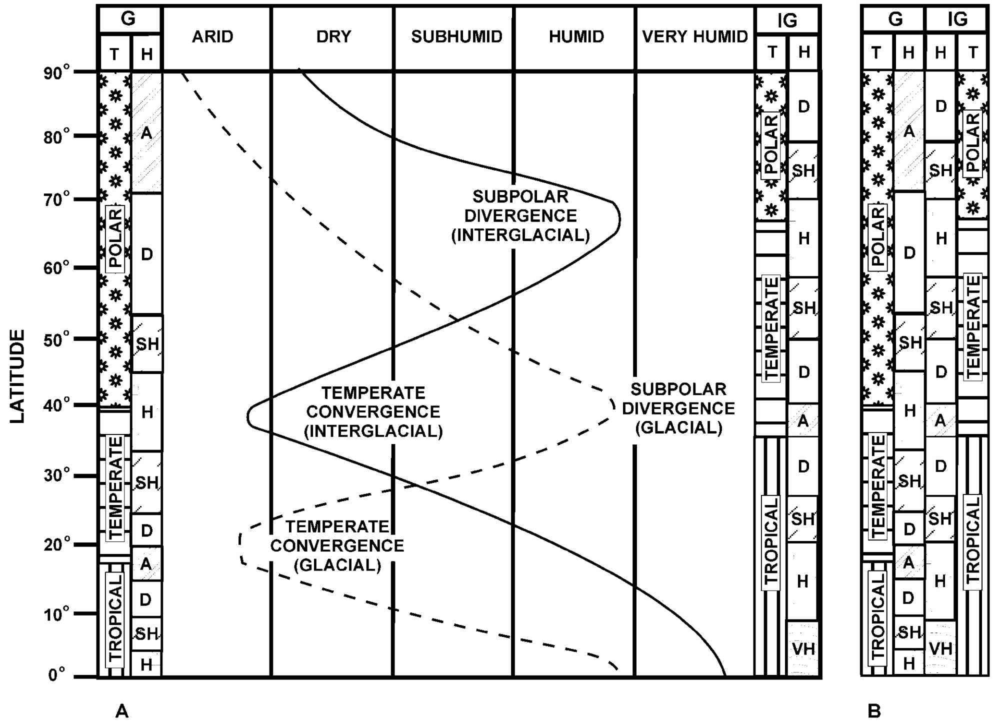

Fig. 7A, B The effect of Milankovitch cycles on global climatic zones (after Perlmutter and Matthews (1990). **A** The effects shown in terms of perturbation of the atmospheric convergences and boundaries between temperature and humidity belts. *G* “Glacial” (Milankovitch insolation minimum), *iG* “interglacial” (Milankovitch insolation maximum), *t* temperature zone, *h* humidity zone. *a* Arid, *d* dry, *sH* semi-humid, *h* humid, *vH* very humid. **B** “Glacial” and “interglacial” belts compared. Note particularly the expected changes in hydrologic conditions at different latitudes, indicating that in the extreme case there can be alternations between humid and arid conditions

ion. They indicated that in some areas the climate might alternate between that of a polar desert and temperate humid region, or between a subtropical desert and tropical humid belt. Perlmutter and Matthews speculated that, over Milankovitch time scales, most of the earth is affected by alternations of climate, influencing weathering rates and transport of material to the sea almost everywhere. The magnitude of the effects would be directly proportional to the magnitude of the meridional temperature gradient, with severe climate changes occurring during times when the poles are glaciated, and less pronounced climate changes when the global climates are more equable.

The Neogene erosion rate increase is more impressive when the global mass-age distribution of sediment is considered (Hay 1994). Compilation of global volumes and masses began with the sediment on the continental blocks (Ronov 1980, 1982). However, about one fourth of the Cenozoic sediment resides on ocean crust (Gregor 1985; Ronov et al. 1986; Hay et al. 1988). Some of the apparent Neogene increase is due to subduction of older ocean crust. For the Cenozoic, the area/age relationship of ocean crust existing today is almost linear, declining from a present area of 282×10^6 km^2 to 156×10^6 km^2 (55% of the present area) at 65 Ma. The correction of the global mass required to take loss of ocean floor and sediment to subduction is about $3/4+1/4\times100/55=1.20$ for the Paleocene, and less for each of the younger epochs. Some of the greater mass of young sediment is due to erosion and recycling of older sediments to make younger sediments (Wold and Hay 1990, 1993), a process which occurs even in the deep sea (Moore and Heath 1977; Moore et al. 1978; Aubry 1995). For the Cenozoic this effect is small, with the correction being only about 12% for the Paleocene, and less for each of the younger epochs. Thus, even when the recycling of older sediment and loss of older deep-sea sediment to subduction are taken into account, the amount of Late Neogene and Quaternary sediment is still very large compared with the amount of older Cenozoic and Mesozoic sediment, as shown in Fig. 8. The masses of sediment deposited in the Paleogene (corrected for subduction and erosion) correspond to a flux rate of about 7×10^{18} kg per 10^6 years. This increases to about 8×10^{18} kg per 10^6 years in the Miocene, 15×10^{18} kg per 10^6 years in the Pliocene, and 26×10^{18} kg per 10^6 years in the Quaternary. Hay (1994) found that two-thirds of the mass of Quaternary sediment, essentially the entire excess, resides in the continental rises and deep-sea abyssal plains.

If the approximately three-fold increase in masses of Late Neogene–Quaternary sediment is interpreted as reflecting uplift of mountains and plateau regions, it implies an increase in the average elevation of land from about 190 m during the Miocene to its present average elevation (excluding Antarctica) of about 565 m (Harrison et al. 1983). Only a minor part of this increase, about 65 m, is attributable to the growth of the Antarctic and Greenland ice caps.

The Cenozoic climate change has involved a lowering of global temperatures, with a strong decrease in polar temperatures, and a reduction in precipitation, also greatest in the polar regions (Hay et al. 1997). The lowering of temperatures during the Cenozoic would have acted to reduce chemical weathering rates, although greater seasonal temperature differences would enhance physical

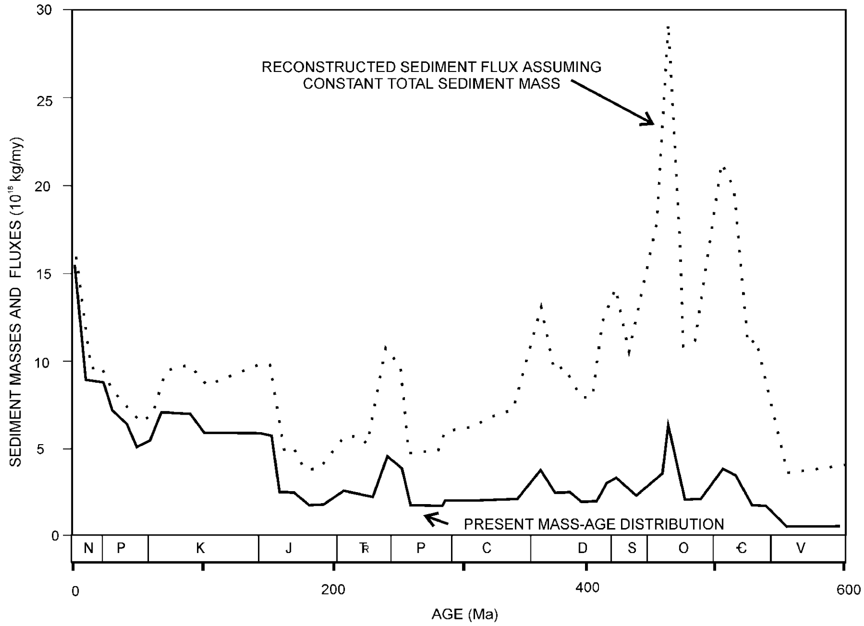

Fig. 8 The fluctuations in the global mass-age distribution of Phanerozoic sediment, based on data from Ronov (1993). Fluxes were reconstructed using the method of Wold and Hay (1993)

weathering rates. If the lowering of temperature were the result of a decline in atmospheric CO_2, this would further reduce chemical weathering rates. Physical weathering is only effective where parent rock at or close to the surface is exposed to temperature extremes, such as in desert and periglacial regions. Soils act as thermal insulators, reducing seasonal temperature variations in the parent rock to null at a depth of 1 to 2 m, so that with soil present the reduction of chemical weathering rates should be the dominant effect. This means that less parent rock will be broken down into particles small enough to be eroded. Because of the nonlinear relationship between temperature and saturation vapor pressure, lower temperatures imply a decrease in precipitation (Hay et al. 1987, 1998). At the same time, lower temperatures imply that evaporation and evapotranspiration rates will also decrease, so that land areas can be expected to show a greater decline than oceanic regions. Precipitation is limited by temperature, the availability of moisture sources and the occurrence of conditions which produce saturation of vapor, whereas evaporation is strictly a function of temperature and availability of moisture. As the temperature decreases, precipitation will also decrease but the proportion of precipitation which becomes runoff will increase. The net effect of cooling of the climate should be a decrease of both precipitation and runoff. The ability of precipitation to erode detritus and the capacity of streams to transport detritus should both decrease. The conclusion is that, other factors remaining constant, the reduced temperatures and precipitation cause by the cooling trend of the Cenozoic should have reduced erosion rates, not increased them.

A wholly different alternative for explaining the increase in sedimentation rates, one which does not involve changing the rates of erosion of upland areas and which corresponds to the fact that most of the Quaternary sediment is in the continental rises and abyssal plains, was suggested by Hay and Southam (1977). They proposed that Late Cenozoic sea-level changes in response to glaciation and deglaciation may have caused large amounts of sediment to be offloaded from the continental shelves into the deep sea. Hay and Southam (1977) assumed that, after a long period with a stable sea level, the continental shelf would be "at grade", that is, it would have developed an equilibrium profile from the shore to the shelf break. They then calculated the effect of lowering sea level by 100 m. To achieve the same equilibrium profile at the lower sea level requires erosion of 400–600 m sediment because of isostatic adjustment. Using an estimate of the shelf area based on a global hypsographic curve, and assuming complete isostatic adjustment, they concluded that a lowering of sea level of 100 m would result in offloading of about 5×10^{18} kg sediment. This is 5% of the total deep-sea sediment mass of Pliocene–Quaternary age (100×10^{18} kg). If this process could be repeated every 200,000 years, it alone could account for the entire apparent increase in young sediment by selective recycling of sediment from the continental shelves into the deep sea. However, the offloading of sediment from the shelves cannot be repeated indefinitely. The continental shelf must be resupplied with sediment during sea-level high stands. During the Quaternary the high sea-level stands have been too short to resupply the shelves with sediment. Hay and Southam (1977) also noted that the global hypsographic curve indicates that the slope on the continental shelves (0 to –200 m) is greater than the slope on the coastal plains (+200 to 0 m). This is counterintuitive – one would expect the slope of sediments beneath the water to be less than the slope of sediments above water. From this and other evidence they concluded that the present continental shelves are not "at grade", but have profiles in equilibrium with Pleistocene sea-level low stands (–130 m). Nevertheless, their estimate of the mass of material which would be mobilized is of the order of magnitude to account for a substantial part of the Late Neogene–Quaternary sediment mass without requiring increased rates of erosion in the continental interiors.

Peizhen et al. (2001) have made a detailed analysis of the increase in sedimentation rates and sediment grain sizes which began 2–4×10^6 years ago. They demonstrate that the increase in sedimentation rates and corresponding required increase in erosion rates occur in many different areas of the world, including continental margins and interiors from tropical to high latitudes. They conclude that this can only be due to climate change, the only globally synchronous process. They argue that the erosion rate increase was the result of the transition which took place 3–4×10^6 years ago, from a "period of climate stability, in which landscapes had attained equilibrium configurations, to a time of frequent and abrupt changes in temperature, precipitation and vegetation, which prevented fluvial and glacial systems from establishing equilibrium states".

Discussion

The question as to whether Late Cenozoic climate change is the direct or indirect result of tectonic uplift in critical regions, or whether the climate change increased erosion rates and the associated isostatic adjustments in such a way as to simulate tectonic uplift cannot be answered in a definitive way. There are interrelationships between uplift, climate, erosion and vegetation such that the effect of a change in one process cannot easily be disentangled from the related changes in the others.

There is no direct or indirect method of unambiguously determining paleoelevation, so the role of Late Neogene uplift in climate change cannot be objectively assessed. Estimates based on determining air pressure from the sizes of bubbles in flow basalts have a potential error of 1.5 km, and in any case have never been attempted for determining paleoelevation. The interpretation of paleofloras as indicators of elevation is based on comparison of foliar physiognomy of fossil with that of modern plants. Changing ideas about the interpretation of paleofloras have increased estimates of the paleoelevation at the best studied fossil site, Florissant, Colorado (Oligocene) from 900 to 3,200 m (MacGinitie 1953; Meyer 1986; Gregory and Chase 1992). However, it may be that the spread of evapotranspiration-regulating C_4 plants has significantly altered the atmospheric energy transport system and, along with it, relative and absolute humidities and lapse rates. The apparently more sophisticated method of basing paleoelevation on moist enthalpy (Forest et al. 1995) still suffers from the problem that the boundary conditions may have changed 8×10^6 years ago with the spread of the C_4 plants.

The argument that the increase in Late Neogene sedimentation rates reflects uplift rests on the present-day observation that the prime factor controlling erosion rates is slope within the drainage basin. It can also be argued that much of the Neogene increase in erosion rates may be related to changes in global vegetation patterns, as suggested by Donnelly (1982). Specifically, it could reflect the spread of C_4 plants. Many C_4 plants are considered to be adapted to arid conditions but, because of their ability to control evapotranspiration, it may be that C_4 plants are responsible for the spread of arid conditions. Part of the sedimentation on the lower continental slopes and abyssal plains can be the result of offloading of sediment from the continental shelves, as proposed by Hay and Southam (1977). Finally, the increase in erosion rates may reflect the inability of the weathering–erosion system to achieve equilibrium conditions because of the rapid oscillations of climate in the Late Neogene, as suggested by Davies et al. (1977) and described in detail by Peizhen et al. (2001).

With atmospheric CO_2 levels apparently having declined much earlier, and uplift being at least in part an illusion caused by isostatic adjustment to sculpturing of the landscape by rivers and glaciers, what was the cause of the Cenozoic climate change?

We suggest it had no single cause, but we can set up a hypothetical sequence which can serve as a model for testing. The decline in CO_2 from high levels in the Late Mesozoic and Early Cenozoic was a major factor in the Eocene–Oligocene decline. The development of circum-Antarctic passages allowing thermal isolation of the continent set the stage for glaciation of Antarctica. The trigger for the Antarctic ice-sheet buildup was a series of southern hemisphere "cool summer" orbits. The closure of the Atlantic–Pacific passage across Central America allowed the development of the Gulf Stream and delivery of warm water to the northern North Atlantic where it could serve as a moisture source for northern hemisphere ice sheets. Again, the trigger for ice-sheet buildup was a series of "cool summer" orbits, this time in the northern hemisphere.

But what caused the step-wise decline in temperature suggested by the Middle Miocene (14–12 Ma) increase in the $\delta^{18}O$ of benthic Foraminifera, and the sea-level fall which occurred a few million years later? Today it can be argued that this has no counterpart in the CO_2 record, corresponds to no particular uplift event, and occurred well after the opening of the Drake Passage. What caused the spread of C_4 plants at around 8 Ma? Why are estimates of paleoelevation based on different methods in conflict? What could enhance the uneven distribution of rainfall during the year? Part of the effect is due to the general cooling but this affects only middle and higher latitudes. Was the effect extended to lower latitudes through the closing of the Tethys and the increased continentality of the Eurasian–African region? Or was it dependent on the development of the monsoon systems which depend on the elevations of Tibet and the East African Rift? Or was it the spread of water-conserving C_4 plants which inhibit water transport into the interiors of the continents and promote aridity?

An outrageous hypothesis

The evolution and spread of land plants had a great impact on weathering, erosion, and climate in the Middle Paleozoic (Berner 1997). We believe that the earth's weathering, erosion and climate system may have undergone another major change with the advent and spread of C_4 plants. They may have radically altered the hydrologic cycle, changed atmospheric heat transport mechanisms, and restructured weathering and erosion processes.

What are C_4 plants and how could they have such a large impact on the earth? $C_3$and C_4 plants differ in the way they carry out photosynthesis (Zeltich 1987). The two different forms of metabolism occur in both monocotyledons and dicotlyedons, but C_4 metabolism is much more abundant in the former (Ehleringer et al. 1997).

Most trees and shrubs and forbs (herbaceous plants other than grasses) are C_3 plants. They utilize the Calvin cycle directly for photosynthesis (Shopes and Govindjee 1987). The enzyme RUBISCO (ribulose 1,5-bisphosphate carboxylase oxygenase), the most abundant protein on earth, is the catalyst for photosynthesis. RUBISCO catalyzes the reaction of CO_2 and ribulose bisphosphate (RuBP) to produce two molecules of phosphoglyceric acid (PGA) and release O_2. The PGA molecules produced by this carboxylation reaction contain three carbons; therefore, the plants using this pathway are termed C_3 plants. The PGA can then be reduced by adenosine triphosphate (ATP) and reduced nicotinamide adenine dinucleotide (NADPH) to sugar which, upon being converted to starch, regenerates RUBISCO. Unfortunately, RUBISCO also catalyzes the oxygenation of RuBP, consuming O_2 and releasing CO_2, and producing one molecule of PGA and one molecule of phosphoglycolic acid. The relative reaction rates of carboxylation to oxygenation depend on the atmospheric ratio of CO_2 to O_2, the temperature, and the brightness of the light. C_3 plants have their maximum efficiency at high levels of CO_2, low levels of O_2, temperatures between 15 and 25 °C, and medium illumination. Their efficiency decreases with decreasing levels of atmospheric CO_2, temperatures above 25 °C, and in bright light. The decreasing levels of atmospheric CO_2 during the Cenozoic have resulted in decreased efficiency of CO_2 fixation by the C_3 pathway. The limiting value for C_3 plants, at which the carboxylation and oxygenation reactions are equal and there is no net fixation of CO_2, is probably between 150 and 50 ppm atmospheric CO_2 (Cerling 1997). C_3 plants typically have deep roots, and can tap the moisture in the deeper layers of soil. Because the stomata must remain open during the day to take in CO_2, they transpire readily and return much of the moisture from rainfall to the atmosphere. They play an important role in recycling water over land areas. Because of their moisture release, they are also largely responsible for the difference in lapse rate measured on the ground versus that measured by an ascending balloon.

Many grasses and sedges, and some other plants typical of grasslands, savannas, and semiarid regions utilize a photosynthetic pathway which is markedly different from that of the C_3 plants. These plants have a distinctive structure (Kranz anatomy), with the vascular tissue surrounded by a dense layer of bundle sheath cells with a high concentration of chloroplasts. Layers of mesophyll surround the bundle sheath cells. Different parts of the photosynthetic process take place in these different parts of the plant. The initial fixation of CO_2 takes place in mesophyll cells, where the enzyme phosphoenolpyruvate carboxylase causes CO_2 to be added to a three-carbon acid phosphoenolpyruvate (PEPA) to form a four-carbon acid, oxaloacetic acid which is then reduced to malic acid, another four-carbon compound. These acids give the name C_4 plants to those utilizing this pathway. The malate is then transported to bundle sheath cells. There it enters the chloroplasts, where it is oxidized (decarboxylated) by malic enzyme to release CO_2. Then photosynthesis involving RUBISCO proceeds as in C_3 plants. The extra step in the C_4 pathway reduces its efficiency, but this loss is more than compensated for by an increased efficiency of carboxylation over oxygenation. The increased efficiency of carboxylation is achieved by enrichment of the CO_2 concentration in the bundle sheath cells until it is almost an order of magnitude higher than the atmospheric concentration (Cerling 1997). The spread of C_4 plants is regarded as an adaptation to the lowering of atmospheric CO_2 levels during the Cenozoic (Ehrelinger and Monson 1993), but they are also capable of much more rapid fixation rates than C_3 plants. The most rapidly growing plants, such as maize and sugar cane, are C_4 plants. Because of their overall greater efficiency in fixing CO_2, C_4 plants lose less water through transpiration per unit C fixed. Their maximum efficiency occurs at temperatures between 30 and 40 °C and under bright light. Hence, they are also adapted to warmer, drier, and brighter conditions. They typically have shallow roots and remove moisture only from the upper layers of the soil. Because of their efficient use of water and inability to use water in the deeper soil layers, they have an important effect on the hydrologic cycle. They restrict the return of water to the atmosphere to as little as 25% that of C_3 plants, promoting drier conditions downwind. Grasses, which largely use C_4 metabolism, are probably the most abundant plant group on earth. Grasses have several special adaptations – their leaves are produced from the central stem at ground level rather than from the tips of branches. Hence, they can quickly recover from grazing. They are also capable of becoming brown and dormant during dry periods; again because of the way in which leaves are produced, they can produce green leaves in a matter of days after rainfall. Their ability to become dormant and recover rapidly would enhance unevenness of rainfall. The grasses are also especially adapted to metabolize and use silica as a skeletal material (opal phytoliths). Although little is known of their effect on weathering processes, their ability to utilize silica suggests that they have had a major impact on the chemical weathering of silicate. It is probably not coincidental that silica deposition in the ocean seems to have increased markedly in the Late Miocene.

Because of their greater photosynthetic efficiency, it has been widely assumed that the development and spread of C_4 plants was a response to declining levels of atmospheric CO_2. In the light of new estimates of Miocene atmospheric CO_2 levels at or below pre-industrial levels, this seems unlikely. It may be that the spread of C_4 plants is in response to their own aridification of the environment. It is easy to visualize that, by restricting the return of moisture to the atmosphere, they could set in motion a positive feedback mechanism to cause local, then regional, and eventually continental drying.

Another group of plants is even more extreme in restricting transpiration. The Crassulaceae, cacti, euphorbias, and other succulents, use a temporal separation of CO_2 uptake and photosynthesis (Ricklefs 1997) within the same cells. This modification of the photosynthetic pathway is termed crassulacean acid metabolism (CAM). The CAM plants open their stomata and take up CO_2 at night, when the temperatures are lower. Like the C_4 plants, they use phosphoenolpyruvate carboxylase as catalyst, to combine CO_2 with phosphoenolpyruvate (PEP) to make oxaloacetic acid which is then converted to malic acid. These acids are then decarboxylated during the day so that photosynthesis can be carried out. During the hot day the stomata remain closed, conserving water. They are not as efficient at CO_2 fixation as either C_3 or other C_4 plants, but they can live under inhospitable conditions. CAM plants are especially adapted to semiarid and arid conditions, and include agaves, cacti, and other succulents. Their maximum efficiency occurs at temperatures around 35 °C and under bright light. Their roots are usually restricted to shallow layers of the soil. Although they are a very minor component of the global biomass and have very low growth rates, their capabilities of water retention suggests that they may play a role in enhancing arid conditions.

DeConto (1996) and Hay et al. (1997) noted that the atmospheric energy transfer system made much greater use of latent heat transport before the C_4 and CAM plants appeared. The higher rates of evapotranspiration associated with C_3 plants appear to have been the factor responsible for the less pronounced temperature gradients between the coasts and continental interiors. We propose that the spread of C_4 plants during the Miocene, climaxing around 8 Ma, played a major role in modifying the earth's climate. It could be the underlying cause of the "Late Neogene climatic deterioration".

Summary and conclusions

Until very recently the favored mechanisms cited to explain the cooling trend during the Cenozoic have been declining levels of atmospheric CO_2 and widespread uplift. Both of these have recently been called into question. It now appears that atmospheric CO_2 levels declined in the Paleogene, and have remained at about pre-industrial levels since the beginning of the Neogene. Analyses of paleobotanical evidence and sedimentary mass balance provide contradictory answers to the question whether there has been widespread Late Neogene uplift which has contributed to climate change or whether the Late Neogene climate change has created the illusion of uplift.

Sediment mass balance indicates that erosion rates increased globally during the Late Neogene. From what is known today, this could be due either to widespread uplift or to global enhancement of seasonality of rainfall. Alternatively, the increase may be the result of the rapid climatic oscillations of the Late Cenozoic which prevented the weathering and erosion systems from reaching equilibrium.

We contend that the decline in CO_2 from high levels in the Late Mesozoic and Early Cenozoic was a major factor in the Eocene–Oligocene decline. The development of circum-Antarctic passages allowing thermal isolation of the continent set the stage for the glaciation of Antarctica, but the trigger which started Antarctic ice-sheet buildup was a series of "cool summer" orbits. The closure of the Atlantic–Pacific passage across Central America allowed the development of the Gulf Stream and delivery of warm water to the northern North Atlantic where it could serve as a moisture source for northern hemisphere ice sheets, and again the trigger for ice-sheet buildup was a series of "cool summer" orbits.

However, if atmospheric CO_2 decline and widespread uplift are ruled out, what caused the climate change during the Late Neogene? We believe that the earth's climate system may have undergone a major change with the advent and spread of C_4 plants. They have radically altered the hydrologic cycle, changed atmospheric heat transport mechanisms, and restructured weathering and erosion processes.

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