

# Tectonic forcing of late Cenozoic climate

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**Global cooling in the Cenozoic, which led to the growth of large continental ice sheets in both hemispheres, may have been caused by the uplift of the Tibetan plateau and the positive feedbacks initiated by this event. In particular, tectonically driven increases in chemical weathering may have resulted in a decrease of atmospheric CO<sub>2</sub> concentration over the past 40 Myr.**

THE past few years have seen increasing interest in the role that tectonic movements may play in controlling global climate, partly stemming from suggestions that the uplift of the Tibetan plateau in the late Cenozoic had a profound effect on atmospheric circulation<sup>1,2</sup>. In addition, it has been proposed that tectonically driven increases in chemical weathering rates over the same interval caused a drawdown of atmospheric carbon dioxide, leading to global cooling<sup>3,4</sup>. These hypotheses have provoked controversy in a wide range of fields. Geophysicists Molnar and England<sup>5</sup> put forth the inverse proposal that climate changes were responsible for the inferred increases in elevation and erosion. Geochemists<sup>6-10</sup> argued whether the ocean strontium isotope record could be used as a proxy for chemical weathering rates in the past. Climate modellers<sup>10-13</sup> discussed whether chemical erosion could increase without a concomitant increase in mantle degassing of CO<sub>2</sub>, for which there is no evidence.

All of these issues are interrelated, with the consequences of one interpretation affecting the others. Here we review the multidisciplinary aspects of the uplift-climate problem, in particular addressing those issues that have provoked the most discussion and controversy over the past few years. We start by presenting evidence for Cenozoic climate trends, and then discuss our uplift-erosion hypotheses, which we believe to be consistent with available late Cenozoic geological data.

## Evidence for Cenozoic climate change

Because sedimentation occurs continuously in the deep sea, the best records of global Cenozoic climate variations come from the ocean. The oxygen isotopic composition of calcite, sensitive to both ice volume and ocean temperature, has been a particularly valuable tool for reconstructing past climate variations. As evaporation-precipitation processes preferentially concentrate the light isotope of oxygen, <sup>16</sup>O, in continental ice sheets, a pronounced enrichment in the oceanic <sup>18</sup>O/<sup>16</sup>O ratio occurs during periods of glaciation. A temperature-dependent <sup>18</sup>O/<sup>16</sup>O fractionation between calcite and water also works in the same direction (colder temperatures cause <sup>18</sup>O/<sup>16</sup>O enrichment); thus, the CaCO<sub>3</sub> shells of marine organisms recovered from deep-sea sediment cores isotopically record information about both ice volume and ocean temperature in the past.

Oxygen isotope data for the past 55 Myr (refs 14, 15) show a long-term increase in  $\delta^{18}\text{O}$  values punctuated by times of relatively rapid change (Fig. 1). The sharp increase in isotopic values in the early Oligocene (~36 Myr BP) may reflect the first major ice-growth event on Antarctica<sup>16</sup>, which followed roughly 20 million years of cooling. Further increases in  $\delta^{18}\text{O}$  in the middle Miocene and late Pliocene represent subsequent increases in ice volume in Antarctica and in the Northern Hemisphere respectively, although the exact portion attributable to ice volume rather than deep ocean cooling remains uncertain. Assuming no ice before the early Oligocene, Miller *et al.*<sup>14</sup> infer deep-water temperatures up to 12 °C warmer than present in the early Eocene, and Shackleton and Kennett<sup>15</sup> report data from the southern Pacific Ocean indicating deep and surface water temperatures up to 15 °C warmer in the early Eocene than at present.

In addition to the long-term cooling inferred from oxygen isotopes, enhanced differentiation of the mid-latitude continents into areas of wetter and drier climates is also indicated by the terrestrial fossil record (see ref. 17 for a summary of this evidence). At higher latitudes, Palaeocene subtropical flora have been identified in Alaska. Middle Oligocene vegetation at this latitude, however, indicates a temperature climate, and by the Pliocene, an essentially modern flora became established<sup>18</sup>. Likewise, progressive cooling of the polar regions is suggested by faunal macrofossil evidence from Ellsmere Island<sup>19</sup>. At high southern latitudes, the presence of Oligocene fossil pollen assemblages in Antarctica and Australia<sup>20</sup> suggests climates much warmer than present.

Overall, a preponderance of oceanic and terrestrial evidence points to a marked, progressive cooling of global climate beginning in the early Eocene. The most striking and convincing evidence of this climatic 'deterioration' is the abundant glacial detritus (tills, moraines, ice-rafted sediments and so on) which now blankets the mid-to-high latitudes of both hemispheres, both in the ocean and on land. Superimposed on this long-term cooling trend are shorter climate oscillations and steps, such as the rapid shifts in  $\delta^{18}\text{O}$  at ~35 Myr and 15 Myr (Fig. 1). These inferred ice-growth events may be in response to abrupt changes in 'forcing', or they may reflect nonlinear (threshold) responses to more gradual changes in climate boundary conditions. In addition to the 'steps', short intervals of climate amelioration are observed in the  $\delta^{18}\text{O}$  record in the early Oligocene, late Oligocene and Miocene. Wolfe<sup>18</sup> inferred similar transient warming phases from floral macrofossil evidence from Alaska and the American northwest.

## Climate forcing mechanisms

Since early last century, scientists have been searching for explanations for the climate cooling of the late Cenozoic<sup>21-23</sup>. The drift of continents over the poles is a popular idea, particularly for the Palaeozoic (for example refs 24-26); but such drift towards the poles has been minimal over the past 100 Myr and cannot explain the observed magnitude of Cenozoic global cooling<sup>27</sup>. Another family of climate-change mechanisms calls on tectonic movements that form sills or ocean gateways in climatically 'critical' locations. For instance, thermal isolation caused by the development of the circum-Antarctic current when Antarctica separated from South America and Australia has often been invoked to explain the rapid cooling and glaciation of Antarctica in the Cenozoic<sup>28,29</sup>. These investigators suggest that greater meridional heat transport before the initiation of circum-Antarctic ocean circulation resulted in higher sea surface temperatures around Antarctica, effectively preventing continental ice sheets from developing. Experiments with general circulation models (GCMs)<sup>30</sup> suggest, however, that rather than decreasing the likelihood of glaciation, greater meridional heat transport would promote glaciation by increasing snowfall in an already cold polar region.

Likewise, the relationship between the emergence of the Isthmus of Panama and the initiation of Northern Hemisphere glaciation is unclear. The formation of the isthmus seems to pre-date widespread Northern Hemisphere glaciation by 0.5 to

2 Myr (ref. 31), and GCM simulations suggest that the deflection of warm Atlantic waters through the Panamanian passage would actually result in cooling in the high-latitude North Atlantic<sup>32</sup>. Thus, although local tectonic changes undoubtedly influence ocean circulation and heat transport, it is unclear how important these changes are in explaining either the stepwise coolings inferred from the Cenozoic  $\delta^{18}\text{O}$  record or the long-term cooling trend observed for the Tertiary.

Many early 'ice age' theories focused on possible effects that uplift and erosion of mountain ranges and plateaux could have on climate. Here we review new variations on these old themes; in particular, we propose that uplift of the Tibetan plateau has been the main driving force behind Cenozoic climate change. Although uplift during the Cenozoic has been inferred for many regions, including the Alps, East Africa, and parts of the Cordillera of North and South America (see ref. 17 for a summary), we focus on the Tibetan plateau because there is unequivocal evidence that it has been created during the Cenozoic, and because its unique size and location make it central to our hypotheses.

With a mean elevation of almost five kilometres and an area half the size of the United States, the Tibetan plateau is the most imposing topographic feature on the Earth's surface. It formed as a result of the collision of the Indo-Australian plate with the Asian plate; hard collision of these continents probably began in earnest in the middle Eocene (44–52 Myr ago; refs 33–35) and continues today. Little information is available on the elevation history of the plateau. In fact, we have only two firm data points: the modern elevation, and the presence of Cretaceous to lower Eocene marine limestones which show that much of the plateau was below sea level around 70–100 Myr ago with some areas still under water until ~52 Myr (refs 33–35). Although we know that the plateau got from point "a" to point "b", so to speak, a more detailed knowledge of its elevation history will clearly be essential both for understanding plate tectonic and geophysical processes<sup>33–35</sup> and for evaluating hypotheses that propose a link between climate change and plateau elevation.

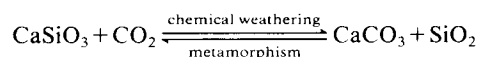
One such hypothesis<sup>1,2,36</sup> notes that the modern Tibetan plateau is so high and so broad that it not only drives a regionally intense monsoon circulation, but also perturbs atmospheric circulation on a hemispheric scale. Ruddiman and colleagues<sup>1,2,36</sup> propose that the absence of this plateau before the Eocene would have resulted in significant climatological differences around the Northern Hemisphere. To test this idea, Ruddiman and Kutzbach<sup>2</sup> initiated a series of experiments using the National Center for Atmospheric Research (NCAR) GCM of the Earth's atmosphere in which they changed only the topography of the present landmasses. Their results (see also ref. 36) showed that many of the global changes in precipitation and temperature inferred by geologists and palaeobotanists to have occurred over the past 40 Myr were consistent with differences in the climates calculated by the model. For instance, relative to the zero-topography experiment, the GCM run with modern topography shows a decrease in summer precipitation around the Mediterranean caused by increased subsidence of dry air masses and stronger northeasterly winds. Similarly, the spread of summer-dry vegetation and increased dustiness in the Mediterranean region since the Miocene has been inferred from geobotanical and sedimentological evidence.

The palaeobotanical record reflects a trend from equable, moist temperate climates in the early and mid-Cenozoic to cold/warm and dry/wet regional patterns in the Northern Hemisphere today. The climate differences in the GCM experiments with and without topography are consistent with this trend toward increased seasonality and regional differentiation of climate. But as discussed by Ruddiman and Kutzbach<sup>2</sup>, the GCM experiments do not show a pronounced drop in high-latitude summer temperatures as elevations are increased, suggesting that development of high topography alone is not

sufficient to initiate the growth of large terrestrial ice sheets in both hemispheres. Experiments using GCM models incorporating important feedback processes (snow albedo, sea ice, mixed-layer ocean temperatures) produce coolings that are larger by almost half than earlier versions (J. Kutzbach, personal communication). For the winter season, many regions cool by amounts that represent a considerable portion of the temperature changes inferred from geological records. For the crucial summer ablation season which controls ice volume, however, the amount of cooling remains far short of that required to explain long-term Cenozoic cooling. Additional factors, such as changes in atmospheric composition, seem to be required to explain global Cenozoic cooling.

### The Cenozoic 'icehouse effect'

It is widely accepted that changes in the concentration of radiatively important trace gases in the atmosphere (particularly  $\text{CO}_2$ ) can alter the Earth's climate. A direct link between atmospheric temperature and  $\text{CO}_2$  concentration is observed throughout the 140,000-year Vostok ice-core record of the last glacial-interglacial cycle<sup>37</sup>, and many investigators have proposed a link between the evolution of global climate over the Cenozoic and changes in the composition of the Earth's atmosphere (for example refs 3, 23, 38–41). On timescales longer than a million years, atmospheric  $\text{CO}_2$  levels are primarily controlled by the balance between the rate of volcanic input from the Earth's interior, and the rate of output through chemical weathering of rocks at the Earth's surface. These reactions can be simplified as



Two types of climate hypotheses have emerged: those in which variations in the atmospheric  $\text{CO}_2$  levels are driven primarily by changes in volcanic outgassing, the main input term<sup>38,40–42</sup>;

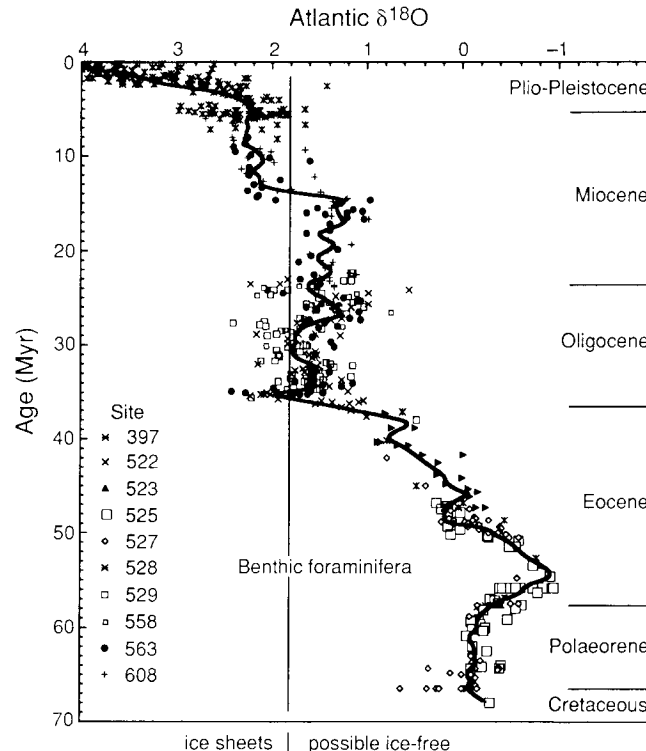


FIG. 1 Compilation of benthic  $\delta^{18}\text{O}$  measurements from Deep Sea Drilling Program sites spanning the past 70 Myr. The long-term increase in  $\delta^{18}\text{O}$  values reflects cooling of the deep ocean and growth of ice sheets at high latitudes (after ref. 14).  $\delta^{18}\text{O} = [({}^{18}\text{O}/{}^{16}\text{O})_{\text{sample}}/({}^{18}\text{O}/{}^{16}\text{O})_{\text{standard}} - 1]$ , where standard is PDB.

and those in which CO<sub>2</sub> variations are driven by changes in chemical erosion rates, the main output term<sup>3,4,23</sup>. The first class of models predicts an association between times of high CO<sub>2</sub> levels and high rates of chemical weathering; in contrast, the erosion models predict an association between low CO<sub>2</sub> levels and high chemical weathering rates. We evaluate these two predictions later.

Berner, Lasaga and Garrels<sup>38</sup> (hereafter referred to as BLAG) developed a numerical model of CO<sub>2</sub> inputs (through metamorphism and decarbonation reactions) and outputs (through chemical weathering) to the atmosphere over the past 100 Myr. They related the flux of CO<sub>2</sub> into the atmosphere linearly to the rate of seafloor production and hence subduction, and the output flux of CO<sub>2</sub> from the atmosphere to the area of continents available for chemical weathering, with atmospheric temperatures acting as a strong negative feedback (assuming that higher temperatures cause more chemical weathering). As formulated, their model produces higher CO<sub>2</sub> levels in the Cretaceous because of (1) more rapid seafloor spreading and volcanic outgassing of CO<sub>2</sub> at that time, and (2) a decrease in the land area available to chemical erosion, because of globally higher sea levels. The higher sea levels are in large part related to faster seafloor spreading rates and increased ridge volume.

Although the BLAG model and subsequent modifications<sup>41,42</sup> correctly yield higher global temperatures in the Cretaceous, the timing of the principal temperature change indicated by these models does not match the timing of the major climate cooling inferred from the geological record. In the BLAG models, the largest decline in atmospheric CO<sub>2</sub> and global temperatures occurs between 100 and 50 Myr ago (pre-Eocene), driven primarily by decreases in global seafloor spreading rates which control mantle outgassing and, indirectly, sea level. But geological evidence<sup>43</sup> suggests that the greater climate cooling occurred primarily between 50 Myr and the present (post-Eocene).

A possible explanation for this mismatch is suggested by erosion-driven CO<sub>2</sub> models<sup>3,4,23</sup>. On the basis of modern river studies<sup>44,45</sup>, Raymo *et al.*<sup>3</sup> suggested that the long-term removal rate of CO<sub>2</sub> from the atmosphere by chemical erosion is a strong function of continental relief rather than area, and that atmospheric temperature exerts a relatively weak control on global chemical erosion rates. The logic is as follows: today, much of the Earth's land surface is characterized by low-lying, deeply weathered shield areas which contribute little to the global river flux of solutes. Even though chemical weathering is pervasive in these regions, the absolute rates are low. In contrast, mountainous regions are dominated by rapid mechanical erosion which, in turn, increases the surface area of fresh minerals available for chemical attack. Three further factors enhance chemical breakdown of detrital grains in mountainous regions: the abundance of easily weathered sedimentary silicates derived from uplifted passive margin sequences (weathering of carbonates does not ultimately alter atmospheric CO<sub>2</sub>), the orographic concentration of rainfall on mountain slopes, and steep slopes which flush away mechanical and chemical erosion products, constantly exposing fresh minerals.

These effects are most marked in the Himalaya. The collision of India with Asia resulted not only in the uplift, deformation and erosion of an antecedent passive margin, but also in the formation of the Tibetan plateau where incident solar heating in summer drives the strong atmospheric convection and rainfall associated with the Asian monsoon. (Numerical models suggest that the high topography of the Indian subcontinent is largely responsible for the intensity of the Asian monsoon<sup>2,36,46-48</sup>.) Data from the eight largest rivers draining the Himalayan-Tibetan region (Table 1) show that almost 25% of the total dissolved load reaching the ocean today comes from a watershed area that represents only 5% of the Earth's land surface<sup>49,50</sup>. Thus, because of geologically unusual circumstances (the pres-

TABLE 1 Total river dissolved fluxes from Himalayan-Tibetan region

River	Dissolved solids (10 <sup>6</sup> tons yr <sup>-1</sup> )	Drainage area (10 <sup>6</sup> km <sup>2</sup> )
Ganges/Brahmaputra	323.5	1.669
Yangtze (Chiang)	166	1.827
Si Kiang	132	0.464
Irrawaddy	91	0.414
Indus	62	0.916
Mekong	60	0.849
Huangho	33.5	0.814
Percentage of global total	25%	4.2%

Fluxes are from refs 49, 50.

ence of a large plateau near a warm ocean), a disproportionate fraction of the Earth's chemical weathering occurs in this small region in Asia. The monsoon precipitation that we invoke is not the 'normal' orographic rainfall found in areas where prevailing westerlies or easterlies impinge on meridional mountain ranges, but is caused by a thermally driven northward flux of moisture from the equatorial Indian Ocean. The Tibetan plateau lies in the latitude of the westerlies, but far from the main Atlantic moisture source, and 'normal' orographic precipitation (on the western slopes) is small.

On the basis of these and similar observations, we propose that late Cenozoic uplift of the Himalayan region and Tibetan plateau would have resulted in regionally, and hence globally, higher chemical erosion rates, causing a drawdown of atmospheric CO<sub>2</sub> and global cooling. The timing of this tectonically driven CO<sub>2</sub> decrease should be post-Eocene, coincident with the formation of the Tibetan plateau and in agreement with geological evidence for when global cooling was most rapid.

### Proxies for chemical weathering rates

One important difference between the two types of hypotheses described above is the prediction of when chemical weathering rates are greatest: in BLAG-type models, weathering rates increase as global temperatures rise, whereas in our model, weathering rate increases are correlated with falling global temperatures. To test these predictions, we need a proxy for global chemical erosion rates. Such a proxy may be the <sup>87</sup>Sr/<sup>86</sup>Sr isotope record preserved in marine carbonates<sup>3,6-9,51-55</sup>.

The seawater <sup>87</sup>Sr/<sup>86</sup>Sr ratio recorded by marine carbonates (Fig. 2a) reflects a balance between input of radiogenic material, relatively high in <sup>87</sup>Sr/<sup>86</sup>Sr, weathered from continents (average value in present-day runoff is 0.7119; ref. 51) and nonradiogenic material of low <sup>87</sup>Sr/<sup>86</sup>Sr introduced by hydrothermal activity (average value 0.7035). The isotopic composition of the river strontium flux is not invariant but depends on the lithology of the rocks being weathered<sup>51-53</sup>. Typically, carbonate weathering results in high river fluxes of strontium with low isotope ratios (<0.709), which would tend to buffer the oceanic value, whereas silicate weathering releases a low flux of relatively radiogenic strontium (>0.710; refs 8, 51, 53). An exception to this relationship has been proposed by Palmer and Edmond<sup>8</sup> who speculate that metamorphic repartitioning of elements in the orogenic zone of the Himalayas has resulted in the formation of extremely radiogenic (>0.720) easily weathered carbonates. Measurements of the isotopic composition of Himalayan limestones are needed to confirm this hypothesis. A third strontium input to the ocean, from redissolution of marine carbonates (average value 0.7084; ref. 51), buffers the isotope ratio of the oceanic strontium reservoir towards the integrated mean value of the preceding 10-15 Myr.

The pronounced post-Eocene rise in <sup>87</sup>Sr/<sup>86</sup>Sr ratios (Fig. 2a) could be due to an increase in the delivery of radiogenic strontium from land, or to a decrease in seafloor hydrothermal activity. The widely held assumption that hydrothermal activity

is a function of seafloor spreading rates, which are estimated from marine magnetic anomalies, can be used to estimate the hydrothermal flux of nonradiogenic strontium through time. Seafloor spreading rates seem to have changed little over the past 30–40 Myr (ref. 56), and it is widely agreed that the pronounced late Cenozoic increase in the oceanic  $^{87}\text{Sr}/^{86}\text{Sr}$  ratio can therefore only be explained by a considerable increase in the delivery of radiogenic strontium from land. This could have been caused by increases in global chemical weathering rates<sup>3,6,9,12</sup> and/or the mean river isotopic composition of strontium<sup>8–10</sup>.

Given the association between silicate weathering and high  $^{87}\text{Sr}/^{86}\text{Sr}$  values, there are only two ways that rivers could become more radiogenic over the past 40 Myr without increasing silicate weathering rates. One is to decrease dramatically the delivery of low  $^{87}\text{Sr}/^{86}\text{Sr}$  ratio strontium from carbonate weathering, requiring an unlikely reduction in the weathering rates of these types of rocks. The second, mentioned above, is to repartition radiogenic strontium by metamorphism, from resistant granites and gneisses into easily weathered carbonates<sup>8</sup>. Palmer and Edmond<sup>8</sup> attribute the unusually high strontium isotopic composition of the Ganges–Bramaputra river system to such ‘orogenic’ repartitioning. But a related study<sup>53</sup> of the Ganges–Bramaputra headwaters proposes that the high  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios in these rivers are in fact due to intense chemical weathering of granites and gneisses (silicates).

Many studies of the Cenozoic strontium record<sup>3,6,7,9,12,53</sup> have concluded that the pronounced Cenozoic increase in oceanic  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios was due to increased rates of chemical weathering over this interval. Richter *et al.*<sup>9</sup> suggest that the strontium flux to the ocean has increased by ~40% over the past 40 Myr, an estimate which conservatively assumes that the strontium isotopic composition of Himalayan rivers increased from a ‘normal’ global value of 0.7105 in the Oligocene to its present anomalously high value of 0.7127, and that hydrothermal exchange was proportional to the rate of seafloor generation. In addition, this study and others<sup>6,53</sup> concur with the suggestion<sup>3</sup> that erosion in the Himalaya has been responsible for a significant fraction of the observed increase in  $^{87}\text{Sr}/^{86}\text{Sr}$ . But the fraction attributable to erosion in low-latitude mountainous regions such as the Himalayas, as opposed to erosion by high-latitude continental ice sheets or alpine glaciers elsewhere, is poorly constrained. By assuming a link to more easily assessed rates of mechanical erosion, Hodell *et al.*<sup>6</sup> estimate that up to 50% of the increase in oceanic  $^{87}\text{Sr}/^{86}\text{Sr}$  values over the past five million years may have been due to chemical erosion of the Canadian shield by the Laurentide ice sheet. Likewise, Miller *et al.*<sup>57</sup> proposed that glacial erosion in Antarctica was responsible for much of the increase in  $^{87}\text{Sr}/^{86}\text{Sr}$  since the Eocene.

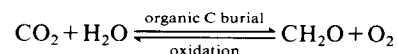
Although the evolution of the seawater strontium curve is consistent with a pronounced increase in chemical weathering since the Eocene, this interpretation is clearly not unique; a considerable fraction of the  $^{87}\text{Sr}/^{86}\text{Sr}$  change could be due to variations in the isotope ratios in source rocks. The interpretation is further complicated by the fact that some strontium (albeit at generally low isotopic values) derives from the weathering of carbonates, a process which does not result in any net removal of  $\text{CO}_2$  from the atmosphere. Ideally, we need a proxy for just the silicate component of weathering. Only with a number of weathering proxies, possibly including carbonate accumulation rates<sup>3,58,59</sup> or Ge/Si ratios<sup>60</sup>, will we be able accurately to define global weathering rates in the geological past.

## The need for negative feedbacks

The erosion hypothesis described above predicts that chemical weathering rates can increase without a concomitant increase in mantle degassing rates. Increasing the erosional drawdown of  $\text{CO}_2$  while holding the volcanic input constant would, however, deplete the atmosphere of all its  $\text{CO}_2$  within a few million years<sup>10</sup>. By invoking tectonics as the main control on

weathering intensity, we implicitly negate the importance of the temperature–weathering feedback which prevents a runaway ‘icehouse’ (or ‘greenhouse’) in BLAG-type carbon cycle models. Hence, a different negative feedback is needed, one that adds  $\text{CO}_2$  back to the ocean–atmosphere reservoir without adding alkalinity (for instance,  $\text{Ca}^{2+}$ ) which removes  $\text{CO}_2$ . Three processes that could provide such a feedback are discussed: imbalances in the organic carbon subcycle, precipitation of silicate minerals in the deep sea and seafloor weathering of basalt.

In addition to the carbonate–silicate cycle, the organic carbon subcycle, described schematically as



has a small but significant impact on the cycling of atmospheric  $\text{CO}_2$ . On average over the Phanerozoic, roughly 80% of the carbon removed from the ocean–atmosphere system was deposited as carbonates, whereas ~20% was buried as organic matter<sup>61</sup>. Because organic matter is extremely enriched in  $^{12}\text{C}$  during photosynthesis, any change in the relative size of this reservoir through time would be reflected in a change in the  $^{13}\text{C}/^{12}\text{C}$  ratio of the world’s ocean and atmosphere (the surface carbon reservoir from which the organic matter is derived, or added to). If, for instance, organic carbon burial decreased over the past 40 Myr, the ocean–atmosphere reservoirs would become enriched in  $^{12}\text{C}$ , the  $^{13}\text{C}/^{12}\text{C}$  ratio of marine calcite would decrease, and a net addition of  $\text{CO}_2$  to the atmosphere would occur. Such a change in the relative fraction of organic carbon buried would affect atmospheric  $\text{CO}_2$  levels independently of global silicate weathering and mantle degassing rates.

The most complete compilation of bulk carbonate  $\delta^{13}\text{C}$  values during the Cenozoic (Fig. 2b; ref. 61) shows just such a decrease. Using these data, Shackleton<sup>61</sup> estimated that the fraction of carbon buried as organic matter rather than carbonate dropped from 20% to 10% over the course of the Neogene, causing a net addition of  $\text{CO}_2$  to the atmosphere (alternatively, the data are consistent with an increase in kerogen weathering relative to organic carbon burial, also leading to a net addition of  $\text{CO}_2$  to the atmosphere). Such a change in the organic carbon subcycle is of the correct magnitude and direction to prevent a tectonically sustained increase in silicate weathering (of ~40%) from completely stripping the atmosphere of  $\text{CO}_2$ . Estimates depend on assumptions of how photosynthetic fractionation factors changed, when to define ‘steady state’ (or the present long-term mean) and what carbon isotope data set is used. Similar studies using different assumptions and data bases have come to more conservative conclusions<sup>12,41,42</sup>.

Why would the long-term burial rate of organic carbon have decreased over the past 40 Myr? Most organic carbon burial today takes place in shallow marine environments such as estuaries, continental shelves and deltas. In these well oxygenated environments, the burial rate of organic carbon is a positive function of the total sediment burial rate<sup>62</sup>. But because carbon isotope data suggest that less organic carbon was being buried in the late Cenozoic, while other evidence suggests that worldwide sedimentation rates have been increasing<sup>63–65</sup>, it seems that this was not the main factor controlling organic carbon burial in recent geological history. On the other hand, increased erosion of continental shelves, estuaries and deltas as sea level dropped through the Cenozoic could have resulted in a decrease in organic carbon burial in the shallow marine realm<sup>61</sup> and an increase in the erosion of previously deposited organic-rich sediments.

A second possible control on organic carbon burial is nutrient supply: a decrease in organic carbon burial may have been caused by a decline in the amount of phosphorus and nitrogen available to form organic matter. This does not seem to be the case for the past 40 Myr, as the nutrient supply from rock weathering has, if anything, increased (as discussed above), and

because much dissolved phosphorus currently entering the ocean is removed by inorganic precipitation<sup>66</sup>.

A third possible control is the oxygen concentration in the deep ocean. Today many areas of the deep ocean are nearly anoxic<sup>67</sup>. Even lower dissolved oxygen concentrations could result in a considerable increase in marine organic carbon burial<sup>68-70</sup>, although this point is controversial<sup>71,72</sup>. The oxygen content of the deep ocean is controlled by the temperature of the water as it leaves the surface, with O<sub>2</sub> concentrations increasing as sea surface temperatures decrease. Today deep waters form in polar regions with a dissolved O<sub>2</sub> concentration of  $\sim 350 \mu\text{mol kg}^{-1}$ . If high-latitude sea surface temperatures were  $\sim 10^\circ\text{C}$  warmer in the mid-Cenozoic, a level consistent with planktonic isotopes<sup>14,73</sup>, then O<sub>2</sub> concentrations would have been only  $\sim 275 \mu\text{mol kg}^{-1}$ . The inferred increase through time in ocean oxygen concentrations could have resulted in decreased organic carbon burial.

Deep-ocean oxygen concentrations would decrease and sea level would fall as CO<sub>2</sub> concentrations in the atmosphere began to fall: as the climate gets colder, oxygen solubility in sea water increases and continental ice growth causes sea level to drop. Each of these changes, in turn, could lead to decreased organic carbon burial and addition of CO<sub>2</sub> to the atmosphere, providing a negative feedback to climate change. Increased continentality and aridity, associated with cooling global temperatures, could also have resulted in the shrinkage and drainage of freshwater marshes and swamps, again leading to a decrease in the burial of organic-rich sediments.

As originally pointed out by Shackleton<sup>61</sup>, the bulk carbonate  $\delta^{13}\text{C}$  data provide convincing evidence that there has been a net addition of light carbon (and CO<sub>2</sub>) to the ocean and atmosphere since the Oligocene. We propose that this carbon flux

would have partially counteracted a drawdown of atmospheric CO<sub>2</sub> by enhanced erosion of silicates. Before invoking organic carbon burial as an effective negative feedback to the global carbon cycle, however, one also has to consider the relative residence times of CO<sub>2</sub> ( $<1$  Myr) and O<sub>2</sub> ( $\sim 10$  Myr) in the atmosphere. Any decrease in organic matter burial will result in net removal of O<sub>2</sub> from the atmosphere. Eventually the continued drawdown of O<sub>2</sub> will cause the partial pressure of oxygen in the atmosphere to fall, probably leading to organic carbon burial rates increasing again. Thus, the negative feedback to the carbonate-silicate cycle provided by the organic carbon subcycle would gradually become less effective five to ten million years after the 'event' that initiated the change, a problem which may not be critical in the case discussed here; namely, gradually more intense forcing as the Tibetan plateau increases in elevation.

Two other processes have been proposed as negative feedbacks to atmospheric CO<sub>2</sub> decreases driven by enhanced silicate weathering. One is precipitation of silicate minerals on the sea floor (ref. 67, and W. Broecker, personal communication). Although this process is poorly understood (see summaries in refs 67, 74), much of the sodium and potassium entering the sea must be precipitated as silicate minerals. On the other hand, calcium derived from silicate weathering is mostly precipitated as carbonate, resulting in the net removal of CO<sub>2</sub> from the atmosphere. As atmospheric CO<sub>2</sub> falls, however, ocean pH rises (OH<sup>-</sup> rises), creating conditions more favourable to the precipitation of Ca as deep-sea silicates rather than carbonates. In this way, one could again decouple silicate weathering rates from atmospheric CO<sub>2</sub> levels. Seafloor weathering of basalt, described in the carbon-cycle model of François and Walker<sup>12</sup>, provides a third mechanism to decouple silicate weathering, alkalinity and atmospheric CO<sub>2</sub> levels. In this process, hydrogen ions in ocean water react with seafloor basalts, releasing calcium ions which then combine with ocean carbon to form calcium carbonate. The net result is the removal of CO<sub>2</sub> from the ocean-atmosphere reservoir. But as atmospheric CO<sub>2</sub> falls, ocean pH and OH<sup>-</sup> rise, creating conditions less favourable to the removal of CO<sub>2</sub> by this mechanism, again providing a negative feedback to a runaway 'icehouse'.

Given the suggestions above, it is certainly possible, if not probable, that increases in silicate weathering rates could be sustained for millions to tens of millions of years without an increase in mantle degassing rates. In particular, the assumption that rates of volcanism and silicate weathering need to be balanced on timescales longer than a million years<sup>10</sup> may not be necessary. One last possibility is that seafloor spreading rates are not a good proxy for mantle degassing rates. Caldeira<sup>13</sup> suggests that increased subduction of pelagic carbonates since the Jurassic has resulted in an increased metamorphic CO<sub>2</sub> flux from the mantle which is responsible for the higher rates of chemical erosion inferred for the late Cenozoic. This could provide the 'extra' CO<sub>2</sub> needed to sustain high silicate weathering rates at a time when seafloor spreading rates are constant.

### The role of positive feedbacks

Molnar and England<sup>5</sup> proposed an interaction between climate and tectonics that on first reading seems the exact inverse of our uplift hypothesis, but in fact may provide a fully complementary source of positive feedback. They argue that late Cenozoic global cooling was responsible for increased mechanical erosion, which through isostasy would create mountain ranges with high peaks and deep valleys, rather than uplift causing climate changes and erosion. They suggested two mechanisms by which climate change intensified erosion: expansion of mountain glaciers around the world; and changes in atmospheric circulation, and thus in precipitation patterns. At the limit, this hypothesis might be interpreted as suggesting that high mountain ranges existed throughout the Cenozoic, and that the only change during the Cenozoic has been the effect of climate change

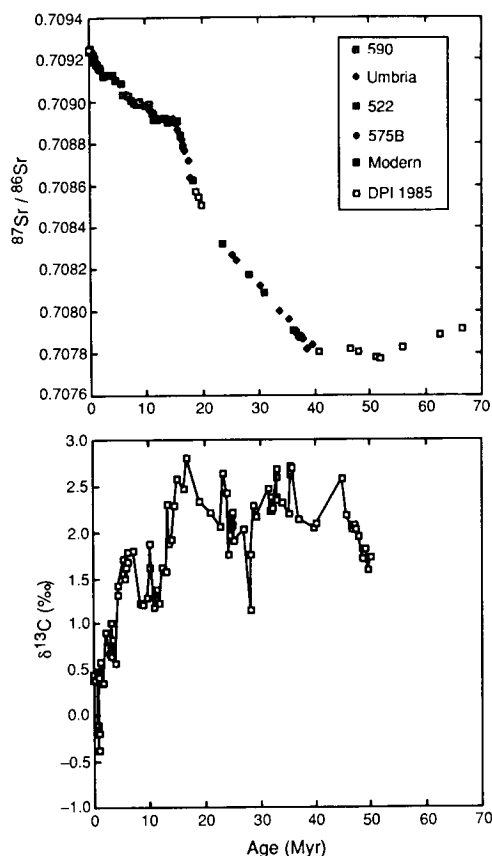


FIG. 2 a, Strontium isotopic composition of sea water for the past 70 Myr, based on analysis of marine carbonates (from ref. 9). b,  $\delta^{13}\text{C}$  of bulk carbonate over the Cenozoic (from ref. 61).  $\delta^{13}\text{C} = [({}^{13}\text{C}/{}^{12}\text{C})_{\text{sample}}/({}^{13}\text{C}/{}^{12}\text{C})_{\text{standard}} - 1]$ , where standard is PDB.

on mountain relief. Such an interpretation would, however, leave the cause of Cenozoic cooling and circulation changes unresolved. In fact, Molnar and England<sup>5</sup> noted the need for a first cause of Cenozoic cooling, and acknowledged Tibetan uplift as a possible cause of climate change. Their basic disagreement is with the inference of accelerating late-Cenozoic uplift throughout the world.

It is consistent with both hypotheses to argue that uplift of Tibet (and possibly elsewhere) is a plausible first cause of Cenozoic climate changes (through circulation changes and weathering), but that climate change (in particular, glacier activity) then caused additional erosion, exhumation and isostatic uplift in other regions, including some mountain belts regarded as orogenically 'dead' from the standpoint of horizontal plate tectonic motions. Thus, the arguments of Molnar and England can be viewed as a positive feedback mechanism, whereby uplift-induced erosion initiates global cooling, which then causes further glacial erosion and cooling worldwide.

An important implication of the plateau uplift hypothesis is that even steady-state plate motions can lead to non-steady-state effects on climate and, hence, possibly global relief. Continent-continent collisions which result in plateaux of the magnitude of Tibet are infrequent and episodic. Consequently, the accompanying effects of uplift (perturbations of atmospheric flow, large monsoons, and intense erosion and weathering), should be comparably infrequent. Over the past 700 Myr, only two other time periods were characterized by Tibetan-size plateaux, the late Precambrian and the late Palaeozoic<sup>75</sup>; these were also intervals of widespread continental glaciation<sup>4</sup>. In summary, despite the continuous presence throughout geological history of high mountain terrain along the convergent margins of the world, it may be the rarer occurrence of plateaux that can drive climate away from steady state and decouple rates of horizontal and vertical tectonic movement.

## Evaluating the effects on climate

We propose that over the past 40 Myr, uplift of the Tibetan plateau has resulted in stronger deflections of the atmospheric jet stream, more intense monsoonal circulation, increased rainfall on the front slopes of Himalayas, greater rates of chemical weathering and, ultimately, lower atmospheric CO<sub>2</sub> concentrations. These changes in climate may initiate strong positive feedbacks to global cooling through glacier-driven erosion<sup>5</sup>. A negative feedback, acting either through the organic carbon subcycle or elsewhere, must also occur to prevent the atmosphere from being completely stripped of CO<sub>2</sub>. This model implies that atmospheric CO<sub>2</sub> levels are not maintained at steady state by the temperature-weathering feedback during times of intense tectonism (which may last many millions of years).

To evaluate the linkages between uplift, weathering and climate proposed here, proxy data for global chemical weathering rates over the Cenozoic and a better understanding of the elevation history of the Tibetan plateau are needed. We also need a quantitative understanding of the long-term carbon cycle, in particular the magnitude and source of variations in the organic subcycle and seafloor weathering. Ultimately, direct determination of atmospheric CO<sub>2</sub> levels over the past 100 Myr will be essential if we are to evaluate the uplift-erosion model<sup>3,4</sup> and other carbon-cycle climate models<sup>38,41</sup>. The most promising technique for this is the inference of atmospheric CO<sub>2</sub> levels from measurements of  $\delta^{13}\text{C}$  in marine organic matter<sup>76-78</sup>. □

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- Ruddiman, W. F. & Raymo, M. E. *The Past Three Million Years: Evolution of Climatic Variability in the North Atlantic Region* (eds Shackleton, N. J., West, R. G. & Bowen, D. Q.) 227-234 (Cambridge Univ. Press, Cambridge, 1988).
- Ruddiman, W. F. & Kutzbach, J. E. *J. geophys. Res.* **94**, 18409-18427 (1989).
- Raymo, M. E., Ruddiman, W. F. & Froelich, P. N. *Geology* **16**, 649-653 (1988).
- Raymo, M. E. *Geology* **19**, 344-347 (1991).
- Molnar, P. & England, P. *Nature* **346**, 29-34 (1990).
- Hodell, D. A., Mead, G. A. & Mueller, P. A. *Chem. Geol. (Isotop. Geosci. Sec.)* **80**, 291-307 (1990).
- Capo, R. C. & DePaolo, D. J. *Science* **249**, 51-55 (1990).
- Palmer, M. R. & Edmond, J. M. *Geochim. cosmochim. Acta* **56**, 2099-2111 (1992).
- Richter, F. M., Rowley, D. B. & DePaolo, D. J. *Earth planet. Sci. Lett.* **109**, 11-23 (1992).
- Berner, R. A. & Rye, D. M. *Am. J. Sci.* **292**, 136-148 (1992).
- Sundquist, E. T. *Quat. Sci. Rev.* **10**, 283-296 (1991).
- François, L. M. & Walker, J. C. G. *Am. J. Sci.* **292**, 81-135 (1992).
- Caldeira, K. *Nature* **357**, 578-581 (1992).
- Miller, K. G., Fairbanks, R. G. & Mountain, G. S. *Paleoceanography* **2**, 1-19 (1987).
- Shackleton, N. J. & Kennett, J. P. *Init. Rep. Deep Sea Drilling Proj.* **29** (ed. Kennett, J. P.) 801-807 (U.S. Government Printing Office, Washington DC, 1975).
- Barrett, P. J., Elston, D. P., Harwood, D. M., McKelvey, B. C. & Webb, P. N. *Geology* **15**, 634-637 (1987).
- Ruddiman, W. F., Prell, W. L. & Raymo, M. E. *J. geophys. Res.* **94**, 18379-18391 (1989).
- Wolfe, J. A. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* **9**, 25-57 (1971).
- McKenna, M. C. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* **30**, 349-362 (1980).
- Kemp, E. M. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* **24**, 169-208 (1978).
- Lyell, C. *Principles of geology* (Murray, London, 1875).
- Dana, J. D. *Am. J. Sci.* **22**, 305-334 (1856).
- Chamberlain, T. C. *J. Geol.* **7**, 545-584; 667-685; 751-787 (1899).
- Crowell, J. C. & Frakes, L. A. *Am. J. Sci.* **268**, 193-224 (1970).
- Caputo, M. V. & Crowell, J. C. *Geol. Soc. Am. Bull.* **96**, 1020-1036 (1985).
- Crowley, T. J., Mengel, J. G. & Short, D. A. *Nature* **329**, 803-807 (1987).
- Barron, E. J. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* **50**, 45-61 (1985).
- Kennett, J. P. *J. geophys. Res.* **82**, 3843-3860 (1977); *Mesozoic and Cenozoic Oceans* (ed. Hsu, K. J.) 119-122 (Am. geophys. Un., Washington DC, 1986).
- Barker, P. F. & Burrell, J. *Antarctic Geoscience* (ed. Craddock, C.) 377-385 (Univ. of Wisconsin, Madison, 1982).
- Oglesby, R. J. *Clim. Dynam.* **3**, 135-156 (1989).
- Keigwin, L. D. *Science* **217**, 350-353 (1982).
- Maier-Reimer, E., Mikolajewicz, U. & Crowley, T. *Paleoceanography* **5**, 349-366 (1990).
- Mercier, J.-L., Armijo, R., Tapponnier, P., Carey-Gailhardis, E. & Lin, H. T. *Tectonics* **6**, 275-304 (1987).
- Molnar, P. *Phil. Trans. R. Soc. Lond. A326*, 33-88 (1988); *Am. Scientist* **77**, 350-360 (1989).
- Harrison, T. M., Copeland, P., Kidd, W. S. F. & Yin, A. *Science* **255**, 1663-1670 (1992).
- Kutzbach, J. E., Guetter, P. J., Ruddiman, W. F. & Prell, W. L. *J. geophys. Res.* **94**, 18393-18407 (1989).
- Barnola, J. M., Raynaud, D., Korotkevich, Y. S. & Lorius, C. *Nature* **329**, 408-414 (1987).
- Berner, R. A., Lasaga, A. C. & Garrels, R. M. *Am. J. Sci.* **283**, 641-683 (1983).
- Barron, E. J. & Washington, W. M. *The Carbon Cycle and Atmospheric CO<sub>2</sub>: Natural Variations Archaean to Present* (eds Sundquist, E. T. & Broecker, W. S.) 546-553 (Am. geophys. Un., Washington DC, 1985).
- Walker, J. C. G., Hays, P. B. & Kasting, J. F. *J. geophys. Res.* **86**, 9976-9782 (1981).
- Lasaga, A. C., Berner, R. A. & Garrels, R. M. *The Carbon Cycle and Atmospheric CO<sub>2</sub>: Natural Variations Archaean to Present* (eds Sundquist, E. T. & Broecker, W. S.) 397-410 (Am. geophys. Un., Washington DC, 1985).
- Berner, R. A. *Science* **249**, 1382-1386 (1990); *Am. J. Sci.* **291**, 339-376 (1991).
- Crowley, T. J. & North, G. N. *Paleoclimatology*, 339 (Oxford Univ. Press, New York, 1991).
- Gibbs, R. J. *Science* **156**, 1734-1737 (1967).
- Stallard, R. F. & Edmond, J. M. *J. geophys. Res.* **88**, 9671-9688 (1983); Stallard, R. F. *Physical and Chemical Weathering in Geochemical Cycles* (eds Lerman, A. & Meybeck, M.) 225-246 (Kluwer, Dordrecht, Holland, 1988).
- Manabe, S. & Terpstra, T. B. *J. Atmos. Sci.* **31**, 3-42 (1974).
- Hahn, D. G. & Manabe, S. *J. Atmos. Sci.* **32**, 1515-1541 (1975).
- Druyan, L. M. *J. Clim.* **2**, 127-139; 347-355 (1982).
- Meybeck, M. *Hydrol. Sci. Bull.* **21**, 265-284 (1976).
- Pinet, P. & Souriau, M. *Tectonics* **7**, 563-582 (1988).
- Palmer, M. R. & Edmond, J. M. *Earth planet. Sci. Lett.* **92**, 11-26 (1989).
- Palmer, M. R. & Edlerfield, H. *Nature* **314**, 526-528 (1985).
- Krishnaswami, S., Trivedi, J. R., Sarin, M. M., Ramesh, R. & Sharma, K. K. *Earth planet. Sci. Lett.* **109**, 243-253 (1992).
- Brass, G. W. *Geochim. cosmochim. Acta* **40**, 721-730 (1976).
- Burke, W. H. *et al. Geology* **10**, 516-519 (1982).
- Komintz, M. A. *Am. J. Petrol. Geol. Mem.* **46**, 109-127 (1984).
- Miller, K. G., Feigenson, M. D., Wright, J. D. & Clement, B. M. *Paleoceanography* **6**, 33-52 (1991).
- Davies, T. A., Hay, W. W., Southam, J. R. & Worsley, T. R. *Science* **197**, 53-55 (1977).
- Opdyke, B. N. & Wilkinson, B. H. *Paleoceanography* **3**, 685-703 (1988).
- Shemesh, A., Mortlock, R. A. & Froelich, P. N. *Paleoceanography* **4**, 221-234 (1989).
- Shackleton, N. J. *Marine Petroleum Source Rocks* (eds Brooks, J. & Fleet, A. J.) 423-434 (Geological Society of London, 1987).
- Berner, R. A. & Canfield, D. E. *Am. J. Sci.* **289**, 333-361 (1989).
- Donnelly, T. W. *Geology* **10**, 451-454 (1982).
- Hay, W. W. *Proc. int. Geol. Congr. 27th* **6**, 15-38 (1984).
- Hay, W. W., Shaw, C. A. & Wold, C. N. *Geol. Rundsch.* **78**, 207-242 (1989).
- Froelich, P. N., Bender, M. L., Luedtke, N. A., Heath, G. R. & DeVries, T. *Am. J. Sci.* **282**, 474-511 (1982).
- Broecker, W. S. & Peng, T.-H. *Tracers in the Sea* (Eldigio, Palisades, New York, 1982).
- Demailson, G. J. & Moore, G. T. *Bull. Am. Assoc. Petrol. Geol.* **64**, 1179-1209 (1980).
- Emerson, S., Fischer, K., Reimers, C. & Heggie, D. *Deep Sea Res.* **32**, 1-21 (1985).
- Emerson, S. *The Carbon Cycle and Atmospheric CO<sub>2</sub>: Natural Variations Archaean to Present* (eds Sundquist, E. T. & Broecker, W. S.) 78-87 (Am. geophys. Un., Washington DC, 1985).
- Pedersen, T. F. & Calvert, S. E. *Am. Assoc. Petrol. Geol. Bull.* **74**, 454-466 (1990).
- Betts, J. N. & Holland, H. D. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* **97**, 5-18 (1991).
- Keigwin, L. D. & Corliss, B. H. *Geol. Soc. Am. Bull.* **97**, 335-345 (1986).
- Drever, J. I. *The Geochemistry of Natural Waters* (Prentice-Hall, Englewood Cliffs, New Jersey, 1988).
- Dewey, J. F. & Burke, K. C. A. *J. Geol.* **81**, 683-692 (1973).
- Arthur, M. A., Dean, W. E. & Claypool, G. E. *Nature* **315**, 216-218 (1985).
- Jasper, J. P. & Hayes, J. M. *Nature* **347**, 462-464 (1990).
- Rau, G. H., Froelich, P. N., Takahashi, T. & DesMarais, D. J. *Paleoceanography* **6**, 335-347 (1991).

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