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Abstract. The origin of the major ice-sheet variations during the last 2.7 million years is a long-standing mystery. Neither the dominant 41 000-year cycles in δ^{18}O/ice-volume during the late Pliocene and early Pleistocene nor the late-Pleistocene oscillations near 100 000 years is a linear ("Milankovitch") response to summer insolation forcing. Both responses must result from non-linear behavior within the climate system. Greenhouse gases (primarily CO_2) are a plausible source of the required non-linearity, but confusion has persisted over whether the gases force ice volume or are a positive feedback. During the last several hundred thousand years, CO_2 and ice volume (marine δ^{18}O) have varied in phase at the 41 000-year obliquity cycle and nearly in phase within the ~100 000-year band. This timing rules out greenhouse-gas forcing of a very slow ice response and instead favors ice control of a fast CO_2 response.

In the schematic model proposed here, ice sheets responded linearly to insolation forcing at the precession and obliquity cycles prior to 0.9 million years ago, but CO_2 feedback amplified the ice response at the 41 000-year period by a factor of approximately two. After 0.9 million years ago, with slow polar cooling, ablation weakened. CO_2 feedback continued to amplify ice-sheet growth every 41 000 years, but weaker ablation permitted some ice to survive insolation maxima of low intensity. Step-wise growth of these longer-lived ice sheets continued until peaks in northern summer insolation produced abrupt deglaciations every ~85 000 to ~115 000 years. Most of the deglacial ice melting resulted from the same CO_2/temperature feedback that had built the ice sheets. Several processes have the northern geographic origin, as well as the requisite orbital tempo and phasing, to be candidate mechanisms for ice-sheet control of CO_2 and their own feedback.

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1 Introduction

Milankovitch (1941) proposed that orbitally controlled changes in summer insolation at high northern latitudes drive ice-volume responses at the 23 000-year period of precession and the 41 000-year period of tilt. Using marine δ^{18}O as an ice-volume proxy, Hays et al. (1976) confirmed that ice sheets fluctuate at those periods and also verified Milankovitch's prediction that the responses lag several thousand years behind orbital forcing (Fig. 1a). Fluctuations at non-orbital periods are not addressed here.

Milankovitch did not anticipate two features found in marine δ^{18}O records. One is the strength of the 41 000-year cycle in δ^{18}O and other climate proxies prior to 900 000 years ago (Muller and MacDonald, 2000; Raymo and Nisancioglu, 2003). This dominance is inconsistent with the Milankovitch hypothesis because summer insolation variations at high northern latitudes are considerably stronger at the period of precession than at the period of tilt. The second unexpected feature is the strong δ^{18}O (ice-volume) oscillation centered on a period near 100 000 years during the late Pleistocene (Shackleton and Opdyke, 1976). The small effect of orbital eccentricity on incident solar radiation rules out insolation as the direct cause of these longer-wavelength changes in ice volume.

Milankovitch's insolation hypothesis thus provides a valid starting point for an orbital theory of climate, but not a full explanation. As a result, many scientists have explored the next most important orbital-scale variable in the climate system – changes in concentration of carbon dioxide (CO_2) and methane (CH_4). At this point, however, widely divergent views coexist about the effect of greenhouse gases (particularly CO_2) on ice sheets. Two end-member views are currently in play.

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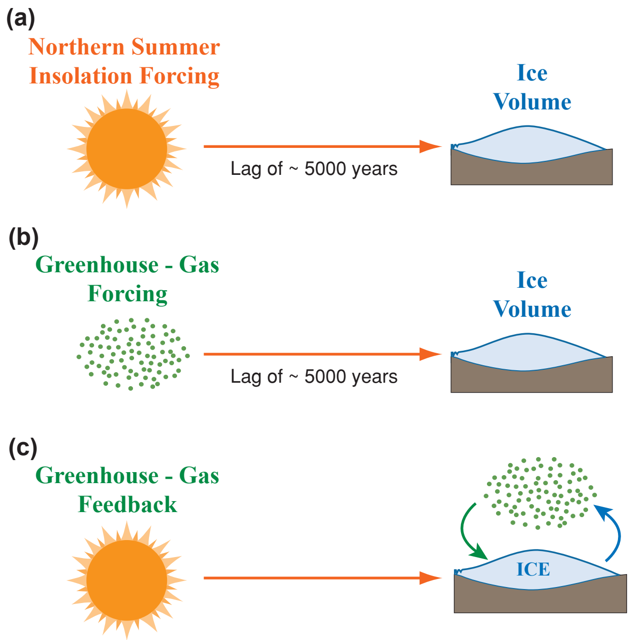

Fig. 1. (a) Northern hemisphere summer insolation forces ice sheets with lags of several thousand years (Milankovitch, 1941; Hays et al., 1976). Greenhouse gases could either force ice sheets with the same lag **(b)**, or be driven by ice-sheet variations and provide positive feedback to the ice **(c)**.

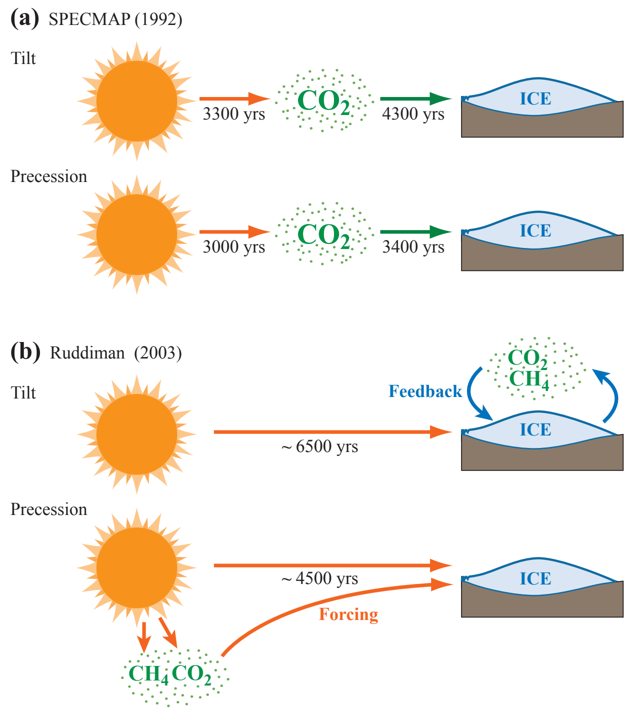

Fig. 2. Phase relationships among insolation, greenhouse gases, and ice-volume at the periods of orbital precession and tilt. **(a)** SPECMAP (Imbrie et al., 1992) inferred that CO_2 forces ice volume as part of a chain of responses to orbital insolation. **(b)** Tuning of gas records in Vostok ice (Ruddiman and Raymo, 2003; Shackleton, 2000) indicates that CO_2 and CH_4 combine with insolation to force ice volume at 23 000 years, but act as ice-driven feedbacks at 41 000 years.

One possibility is that CO_2 forces ice volume (Pisias and Shackleton, 1986; Genthon et al., 1987; Imbrie et al., 1992, 1993; Shackleton, 2000). In this view, changes in CO_2 "push" the slow-responding ice sheets, which respond with lags of 5000 years or more (Fig. 1b). These lags are analogous to the ice-volume response to changes in insolation (Fig. 1a).

An alternative possibility is that CO_2 concentrations are controlled by ice volume and act as a fast positive feedback on ice-sheet mass balance (Ruddiman, 2003; see also Clark et al., 1999). In this case, little or no lag should exist between changes in CO_2 and in ice volume (Fig. 1c). Each increment of ice growth (whether over a millennium or a century) causes greenhouse-gas concentrations to fall, and the reduced gas levels immediately promote further ice growth during that same millennium or century. Once the ice sheets begin to shrink, the greenhouse gas levels rise, promoting further ice melting.

Because both ice sheets and CO_2 concentrations are parts of the overall response of a highly coupled climate system with complex feedbacks, progress in understanding cause and effect at orbital scales has been difficult. The problem, once summarized by Laurent Labeyrie, is that "Everything is correlated to everything".

One potential clue to the cause-and-effect problem is the relative phasing of the greenhouse gases and ice volume. Do the gas changes precede ice volume and thus force a slow ice response (Fig. 1b)? Or do they respond in phase with the ice and thus act as a "fast feedback" (Fig. 1c)?

2 CO_2 and ice volume at the obliquity and precession cycles

The SPECMAP group (Imbrie et al., 1992, 1993) presented a comprehensive hypothesis on the role of greenhouse gases in orbital climatic change. At a time when Vostok drilling had not recovered enough ice to constrain the timing of long-term CO_2 variations, SPECMAP attempted to use geochemical proxies for this purpose. At the periods of orbital precession and tilt, SPECMAP proposed that changes in summer insolation at high northern latitudes initiate a complex chain of responses that are transferred south via deep-water flow. Subsequent changes in the south-polar region then produce CO_2 variations that ultimately drive the slow-responding northern ice sheets (Fig. 2a) [SPECMAP did not consider the role of methane.].

Once the actual CO_2 record preserved in Vostok ice air bubbles became available (Petit et al., 1999), Shackleton

(2000) created a gas time by tuning the precession component of $\delta^{18}O_{air}$ to an insolation target signal with a September phase. Subsequently, Ruddiman and Raymo (2003) developed an independent gas time scale by tuning the precession component of the CH_4 signal to an insolation target with mid-July (summer monsoon) phasing. These two time scales yielded average phases for the obliquity and precession components of the CO_2 variations that agreed to within less than 100 years (see also Bender, 2002).

At the 23 000-year precession period, both CH_4 and CO_2 have phases on or close to that of northern mid-summer insolation (Fig. 2b). For methane, this timing is supported by the match between the CH_4 peak 10 500 years ago in annually layered Greenland ice (Blunier et al., 1995) and the age of the most recent July insolation maximum. It is also consistent with mid-summer (July) forcing of monsoon maxima that create methane-generating wetlands in southern Asia (Kutzbach, 1981; Prell and Kutzbach, 1992; Yuan et al., 2004). The phase of CO_2 at the 23 000-year period falls less than 1000 years after that of July insolation. These early phases for both methane and CO_2 indicate that the two greenhouse gases (along with summer insolation) act as sources of forcing of ice volume at the 23 000-year period (Ruddiman, 2003).

In contrast, both methane and CO_2 vary in phase with $\delta^{18}O$/ice volume at the 41 000-year period of obliquity (Fig. 2b). This in-phase behavior rules out greenhouse-gas forcing of a slow (lagged) ice response. Instead, the ice sheets must drive fast gas responses with little or no lag. These ice-driven gas variations then provide positive feedback to both the growth and melting of ice sheets at 41 000 years. The reason for the different behavior of CO_2 and CH_4 at the precession and obliquity cycles is beyond the scope of this paper.

For the most recent glacial-interglacial oscillation, $\delta^{18}O$ changes (Fig. 3) are linked to changes in ice volume by sea-level constraints from coral reefs (Chappell and Shackleton, 1986; Bard et al., 1990). Sea level can be converted to ice volume and to changes in $\delta^{18}O$ by assuming 10 m of change per 0.11‰ of $\delta^{18}O$ shift (Fairbanks and Mathews, 1978). By this measure, ice volume accounts for well over half of the $\delta^{18}O$ change on the major $\delta^{18}O$ transitions: 50–70% of terminations II (the stage 6/5 $\delta^{18}O$ boundary) and I (the stage 2/1 boundary); 50–60% of the substage 5.5/5.4 $\delta^{18}O$ transition; 100% of the substage 5.4/5.3 boundary, ~70% of the stage 5/4 transition, and ~65% of the stage 3/2 boundary.

The $\delta^{18}O$ trends in Fig. 3 are consistent with the results from spectral analysis. The maxima at marine isotopic stages 4 and 2 are manifestations of the 41 000-year cycle. Both occur several thousand years after insolation minima at the obliquity cycle, consistent with a forced (and lagged) ice-volume response (Imbrie et al., 1992). The coincident CO_2 minima at these times indicate that CO_2 acted as an in-phase feedback at the obliquity cycle, amplifying the size of these ice-volume maxima without altering their phase.

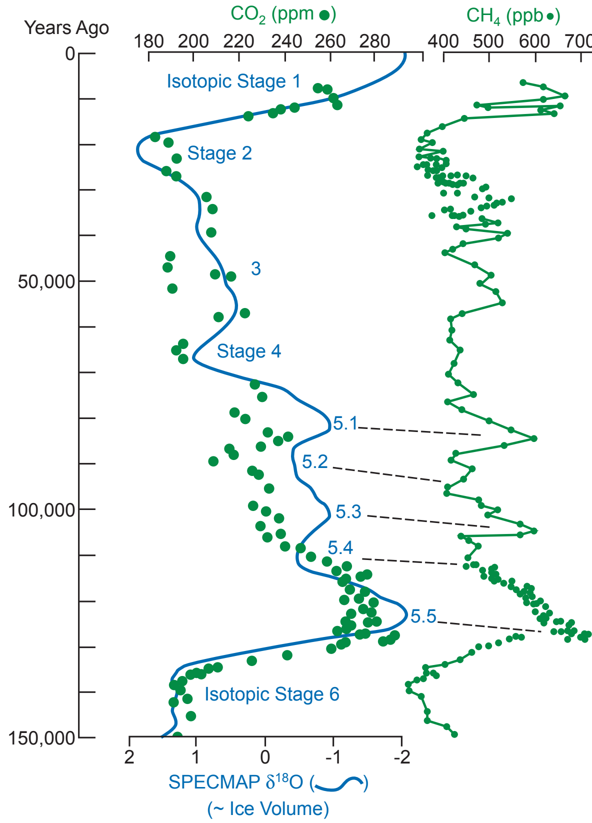

Fig. 3. Comparison of normalized SPECMAP $\delta^{18}O$ record (Imbrie et al., 1984) against Vostok CO_2 and CH_4 signals according to the GT4 time scale of Petit et al. (1999). The SPECMAP time scale is shifted to older ages by 2000 years.

In contrast, evidence of greenhouse-gas forcing of ice volume is present during isotopic stage 5, when insolation changes at the 23 000-year precession cycle were largest. At that time, large methane variations (of ~250 ppb) clearly led $\delta^{18}O$ by several thousand years, indicating that methane changes forced a lagged ice-volume response. Hints of a similar lead appear near isotopic substages 5.5 and 5.1 in the noisier, lower-resolution CO_2 signal.

Because the 41 000-year signal is roughly twice as strong as that at 23 000 years in the signals of both CO_2 (Petit et al., 1999) and $\delta^{18}O$ (Imbrie et al., 1984, 1992), the feedback role for CO_2 outweighs the ice-forcing role in the combined effects of these two periods.

3 Conceptual model: 41 000-year cycles of ice volume from 2.7 to 0.9 million years ago

The above observations can be applied to the early regime of Northern Hemisphere glaciation. Near 2.75 million years ago, north polar latitudes cooled to the point that reduced summer insolation periodically allowed moderately large ice

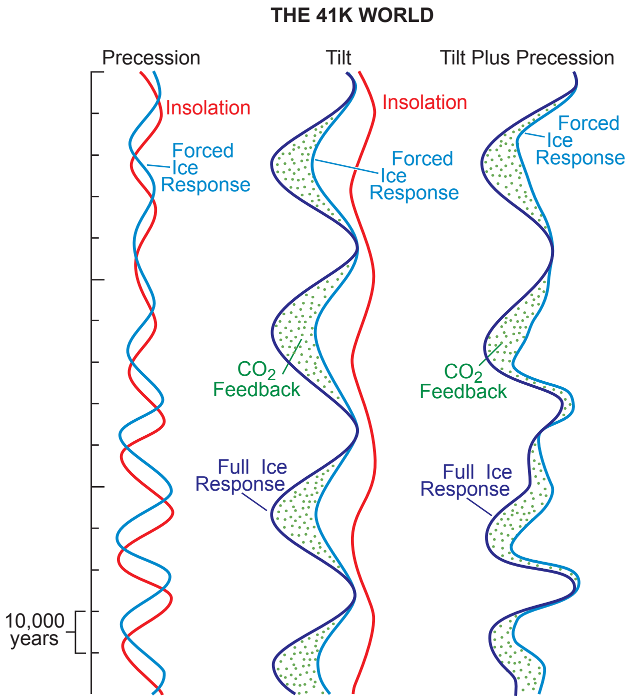

Fig. 4. Schematic model of CO_2 feedback on ice volume in the 41 K world prior to 0.9 million years ago. Summer insolation at 65° N forces ice volume at the precession cycle. Summer insolation forcing of ice volume at the tilt cycle is 40% of that at the precession cycle. The greater length of the tilt cycle increases the forced ice response compared to that at precession. Positive CO_2 feedback amplified the 41 000-year ice response so that the combined ice-volume signal was dominated by tilt. (Insolation trends shown are those of the last 150 000 years).

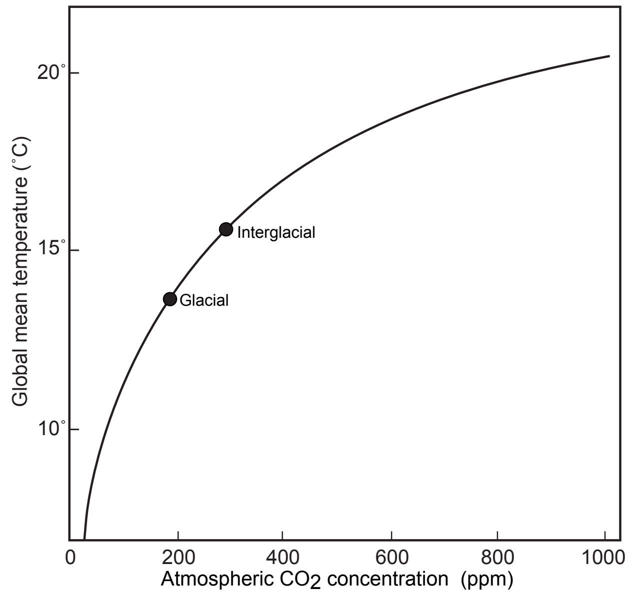

Fig. 5. Logarithmic relationship between CO_2 concentration and global temperature for the model examined by Oglesby and Saltzman (1990).

sheets to form, but all or most of the ice melted during each subsequent interval of increased insolation. For nearly the first two million years of the northern hemisphere ice ages, $\delta^{18}O$ variations were dominated by 41 000-year variations (Pisias and Moore, 1981; Ruddiman et al., 1986; Raymo et al., 1989), despite the fact that changes in summer insolation forcing were stronger at the 23 000-year precession cycle, both on a monthly basis and for the entire caloric summer half-year.

This mismatch between forcing and response is reduced to some extent by the fact that ice volume has almost twice as long to respond to forcing at the 41 000-year cycle as it does at the 23 000-year cycle because of the greater interval over which the forcing persists (41 000/23 000=1.8. For an e-folding ice response, the ratio would be somewhat smaller, ∼1.5. Even with allowance for this longer time of integration of the effect of tilt forcing, the average ice-volume response at the precession period still exceeds that at obliquity by almost 50%.

One way to resolve the remaining mismatch between the summer insolation forcing and the observed $\delta^{18}O$ responses is to suppress the 23 000-year precession component of the $\delta^{18}O$ signal by interhemispheric cancellation of oppositely phased ice responses between the northern and southern hemispheres (Raymo et al., 2006). Another proposed resolution is the fact that the large amplitude of precession peaks is offset by changes in length of the summer season (P. Huybers, personal communication, 2006).

The suggestion has also been made that insolation changes at the obliquity cycle enhanced the planetary temperature gradient and drove a greater northward flux of tropical moisture (Young and Bradley, 1984; Raymo and Nisancioglu, 2003; Vettoretti and Peltier, 2004). Glacial geologists and glaciologists, however, generally view accumulation as a far weaker factor in ice-sheet mass balance than ablation (Alley, 2003; Denton et al., 2005). Most simulations with general circulation models show reduced precipitation in regions where ice accumulates because local cooling reduces the amount of water vapor in the atmosphere.

The explanation favored here is greenhouse-gas feedback at the 41 000-year period. If greenhouse-gas changes at this cycle acted as a positive feedback (as they have done for the last 400 000 years; Fig. 2b), they would have amplified the 41 000-year ice-volume response to direct insolation forcing and caused additional (non-linear) ice growth at that cycle. In the example shown in Fig. 4, the direct ice-volume response to insolation forcing at the 41 000-year cycle is arbitrarily assumed to have doubled in size because of positive CO_2 feedback. This proposed doubling would make the 41 000-year ice-volume signal the dominant orbital response.

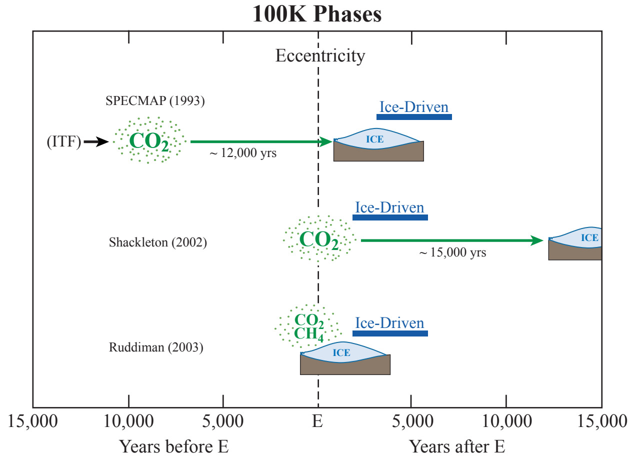

Fig. 6. Phase relationship between CO_2 and ice volume in the ~100 000-year band. SPECMAP (Imbrie et al., 1993) inferred that CO_2 forces ice volume as part of a chain of responses to orbital insolation. Shackleton (2000) also inferred CO_2 forcing of ice volume, but with the entire process offset ~10 000 years later in time. Ruddiman (2003) inferred that CO_2 is primarily a fast feedback on ice volume, based on the similar phasing of CO_2 and $\delta^{18}O$. Phases of "ice-driven responses" (North Atlantic sea-surface temperature, dust, deep-water circulation) are indicated.

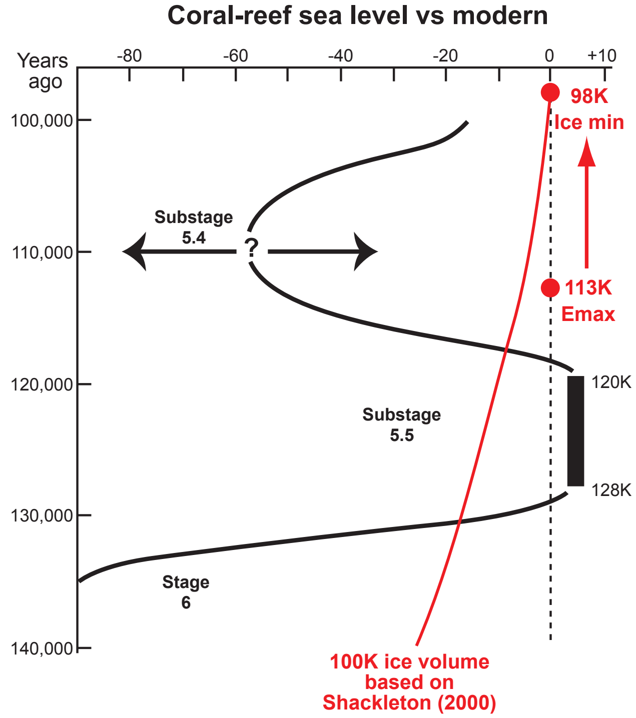

Fig. 7. Estimated sea-level changes during marine isotopic substages 5.5 and 5.4 from coral reefs and $\delta^{18}O$ signals (Chappell and Shackleton, 1986; Bard et al., 1990) compared with filtered 100 000-year ice-volume signal based on Shackleton (2000).

The effect of CO_2 on global temperature is weakly logarithmic: the change in global temperature per unit change of CO_2 increases with lower concentrations (Fig. 5). As a result, the positive temperature feedback on ice volume arising from changes in CO_2 at the 41 000-year cycle would also have been weakly logarithmic.

Although the level of dominance in the conceptual example shown in Fig. 4 does not match that in marine benthic $\delta^{18}O$ records, the isotopic signals are overprinted by a large (and apparently in-phase) temperature response at the 41 000-year cycle (Raymo et al., 1989). That overprint exaggerates the strength of the 41 000-year signal in benthic $\delta^{18}O$ records compared to actual changes in ice volume, and the amount of amplification is not precisely known. The schematic example shown may thus be in reasonable accord with the real behavior of ice sheets.

In this schematic model of the 41 000-year regime, the combined insolation forcing at the precession and tilt cycles accounts for about 70% of the amplitude of the orbital-band ice-volume response, while CO_2 feedback provides the other 30%. Even though CO_2 feedback is critical to explaining the unexpected prominence of the 41 000-year signal in this earlier climatic regime, forced responses still dominated ice-sheet behavior.

4 Relative phasing of CO_2 and ice volume in the ~100 000-year eccentricity band

4.1 Spectral analysis

Late Pleistocene climatic records of many kinds have been dominated by oscillations in a band centered on 100 000 years, especially during the last 400 000 years. SPECMAP (Imbrie et al., 1993) proposed that CO_2 changes in this band occurred as an intermediate step within a chain of responses (Fig. 6). The initial driver of the CO_2 signal was referred to as an (unidentified) "internal thermal forcing" or ITF. The CO_2 response arose within the climate system when ice sheets began to exceed a certain size threshold and produce other feedbacks. But once again the direct role proposed for CO_2 in the SPECMAP scheme was to force ice sheets, which in this case responded with a much longer lag of ~12 000 years.

Shackleton (2000) later determined that the ~100 000-year CO_2 signal in Vostok ice has a much later phase than Imbrie et al. (1993) had inferred from geochemical proxies, one on or very close to the phase of eccentricity. He suggested that the 100 000-year CO_2 signal arose from processes operating within the carbon system independently of the ice sheets. Shackleton further proposed that the 100 000-year marine $\delta^{18}O$ signal in benthic foraminifera carries a very large temperature overprint and that the actual 100 000-year phase of ice volume is offset some 12 000 years later than the δ^{18}O signal. This revised interpretation maintained CO_2 as the immediate source of forcing of ice volume (Fig. 6), but it shifted the entire forcing-and-response relationship ~10 000 years later in time compared to the scheme of Imbrie et al. (1993).

Ruddiman (2003) concluded that the extremely late phase inferred by Shackleton (2000) for the 100 000-year component of ice volume is not consistent with other evidence. For example, it would mean that an ice-volume minimum at the 100 000-year period should have occurred 98 000 years ago, 15 000 years after the preceding eccentricity maximum (Fig. 7). This timing implicitly requires that ice was gradually melting at the ~100 000-year period through the entire first half of marine isotopic stage 5. But oxygen-isotopic and coral-reef evidence indicate that renewed ice growth during marine isotopic substage 5.4 culminated in an ice-sheet maximum of substantial size near 110 000 years ago (Imbrie et al., 1984; Chappell and Shackleton, 1986). This scenario is highly implausible: if large ice sheets were rapidly growing during substage 5.4 in the only known centers of northern hemisphere glaciation, how could "100 000-year" ice sheets have been slowly melting at this same time elsewhere on Earth?

A second problem with such a late ice-volume response is that it lags thousands of years behind a group of northern responses widely regarded as "ice-driven": North Atlantic sea-surface temperatures, northern hemisphere dust fluxes, and NADW flow (Ruddiman and McIntyre, 1984; Raymo et al., 1989; Imbrie et al., 1993). If these signals were driven by the ice sheets, they could not have led their "drivers" by ~10 000 years. Based on these criticisms, Ruddiman (2003) concluded that δ^{18}O is a good first-order proxy for ice volume during the large climatic oscillations of the late Pleistocene, in agreement with the traditional view of CLIMAP (Hays et al., 1976) and SPECMAP (Imbrie et al., 1984), and many other studies over several decades.

If δ^{18}O is a valid proxy for ice volume, the phase lag between CO_2 and ice volume (δ^{18}O) at the 100 000-year cycle cannot be more than a few thousand years. In the absence of significant insolation forcing at this period, eccentricity is used here a convenient reference point for comparing the phases of CO_2 and δ^{18}O. The SPECMAP group (Imbrie et al., 1984, 1989) estimated that the δ^{18}O signal lags eccentricity by ~3300 years, but subsequent U-series dating of coral reefs and sea level (Edwards et al, 1987; Bard et al., 1990) led to revisions of this estimate. Several studies concluded that the SPECMAP time scale is on average too young by 1500 to 2000 years (Pisias et al., 1990; Shackleton, 2000; Ruddiman, 2003). If the phase of the δ^{18}O signal at the ~100 000-year period is adjusted by 2000 years, its lag behind eccentricity would be only ~1300 years. As for the CO_2 signal, tuned Vostok time scales indicate that it could either be in phase with eccentricity (Shackleton, 2000) or lead it by as much as 3500 years (Ruddiman and Raymo, 2003; Ruddiman, 2003). Consequently, the overall CO_2 lead versus δ^{18}O (ice volume) in the ~100 000-year band would be ~1300 to ~4800 years (Fig. 6).

4.2 Cross-correlation analysis

Fourier analysis is not an optimal way to assess leads and lags between CO_2 and δ^{18}O in the ~100 000-year band. Both signals have highly asymmetric shapes that drift gradually toward "glacial" values (low CO_2 and positive δ^{18}O) and then shift abruptly to "interglacial" conditions (high CO_2 and negative δ^{18}O) across deglacial transitions. A symmetrical sine wave with a period near 100 000 years cannot capture either the abruptness of the terminations or the fundamental saw tooth asymmetry of the major glacial/interglacial cycles near ~100 000 years (Muller and MacDonald, 2000).

An alternative approach is to compare the relative timing of the full CO_2 and ice-volume signals. Using cross-correlation analysis, Mudelsee (2001) found that the CO_2 signal at Vostok (Petit et al., 1999) lagged SPECMAP δ^{18}O by 3700±2800 years between 420 000 and 200 000 years ago, but led it by 3200±4000 years since that time. Comparison with a benthic δ^{18}O record from the Indian Ocean indicated a CO_2 lead of 2700±1300 years over the last 420 000 years. Comparison to the $\delta^{18}O_{air}$ record at Vostok indicated a CO_2 lead of 3900±500 years.

This analysis shows that CO_2 and δ^{18}O are in phase to within a few thousand years, with a general tendency toward a small CO_2 lead. Because both the CO_2 and δ^{18}O signals are dominated by power in the ~100 000-year band, this result further confirms the conclusion that the phase separation between CO_2 and ice volume at that period is small.

4.3 Leads and lags on terminations

Several studies have focused on the relative timing of CO_2 and ice-volume changes across deglacial terminations. During termination I, CO_2 changes lead coral-reef (sea-level) indices of ice volume by 1000 years or less (Broecker and Henderson, 1998; Alley et al., 2002). The estimated CO_2 lead relative to δ^{18}O on termination II was ~3000 years (Broecker and Henderson, 1998), but sizeable uncertainties exist in both the dating of both signals and in the accuracy of the ice-volume proxies (Alley et al., 2002).

In summary, three methods of comparing the timing of changes in CO_2 and δ^{18}O (ice volume) converge on a small CO_2 lead in the range of 1000 to 4000 years, with uncertainties of a few thousand years.

5 CO_2/ice phasing at ~100 000 years: CO_2 forcing or feedback?

Spectral analysis, cross-correlation analysis, and analysis of leads and lags on terminations all rule out the possibility that

CO_2 leads ice volume by 12 000 years at the 100 000-year period, as proposed by Imbrie et al. (1993) and Shackleton (2000). That conclusion also eliminates the interpretation that CO_2 forces ice sheets with a large time constant of ∼15 000 years at this period. The observed phasing between CO_2 and δ^{18}O leaves open two interpretations: either (1) CO_2 forces fast-responding ice sheets that lag by only a few thousand years, or (2) CO_2 is in phase with ice volume and acts as a positive feedback.

In order for CO_2 to lead and force ice sheets in the 100 000-year band, the ice response time would have to be very rapid. The response time of ice sheets at a given period can be calculated from the phase lag of δ^{18}O behind CO_2 from the arctangent relationship of Imbrie et al. (1984): $\Phi = \arctan 2\pi f T$. where Φ is the phase lag of the ice sheets (in degrees out of 360°) relative to the forcing, f is the frequency of the forcing in years (here, 1/100 000), and T is the ice time constant in years. For a phase lag of 1000–4000 years at the 100 000-year period, the time constant required for the ice sheets would lie in the same range (1000–4100 years).

Such a small time constant does not seem unreasonable for the rapid disintegration of marine portions of ice sheets that calve icebergs to the ocean, but it is far more problematic for the bulk of the ice in continental interiors. The strongest evidence against so short an ice response is the lag of 5000 to 8000 years of marine δ^{18}O signals behind summer insolation at the forced cycles of 23 000 and 41 000 years. These lags formed the quantitative basis for the SPECMAP marine time scale and they require ice-sheet response times of at least 5000 years (Imbrie et al., 1993), longer than the values of 1000 to 4100 years calculated above.

The other plausible interpretation is that CO_2 acts primarily as an ice-driven feedback that has the same phase as the ice, or very nearly so. Given the large uncertainties in most of the above estimates, this interpretation is permitted by the very small phase difference between CO_2 and δ^{18}O in the band near ∼100 000 years.

The interpretation favored here is that the phasing of CO_2 and ice volume in the ∼10 000-year band is the result of both feedback and forcing processes, but that the in-phase feedback contribution is the larger of the two. This conclusion is consistent with the strong in-phase CO_2 feedback at the 41 000-year period and the weaker CO_2 forcing at the 23 000-year period during the last 400 000 years (Fig. 2b).

5.1 CO_2 feedback at the last glacial maximum

The evidence summarized above suggests a prominent feedback role for CO_2 in orbital-scale variations. Evidence that CO_2 may in fact be the dominant orbital-scale feedback in the climate system comes from the last glacial maximum, the only pre-modern interval examined in sufficient regional detail to allow a global-scale assessment of processes that affect climate (Hansen et al, 1984; Raynaud et al, 1988; Hoffert and Covey, 1992). Because summer and winter solar radiation values 21 000 years ago were similar to those today, insolation differences are not regarded as a major explanation of the colder glacial-maximum temperatures. Instead, these studies suggest that the primary feedbacks in operation were greenhouse gases and albedo.

The ∼90-ppm CO_2 lowering at the last glacial maximum provided a radiative cooling of ∼0.67°C, and the ∼320-ppb CH_4 decrease added another ∼0.14°C, including the chemical effects of methane on stratospheric ozone (Raynaud et al, 1988). The combined radiative cooling of ∼0.81°C would have been amplified by a factor of 2.1 by other factors (primarily water vapor) for a 2.5°C global-mean sensitivity to CO_2 doubling. The resulting total greenhouse-gas cooling of 1.7°C would then account for about 40% of the total global cooling of ∼4.5°C (±0.7°) at the last glacial maximum.

Albedo-temperature feedback accounts for most of the remaining glacial-maximum cooling (Hansen et al, 1984; Hoffert and Covey, 1992). About half of this albedo increase came from the bright surfaces of the northern hemisphere ice sheets, but a substantial part resulted from the increased extent of Southern Ocean sea ice. Because the glacial increase in Antarctic sea ice has been attributed to lower greenhouse-gas levels (Broccoli and Manabe, 1987), this effect should probably be added to the greenhouse-gas side of the ledger, bringing the total gas contribution to more than 50%. In addition, part of the remaining albedo increase came from a reduction in glacial vegetation cover caused by lower CO_2 values, further increasing the indirect greenhouse contribution (Levis and Foley, 1999). Greenhouse gases (mainly CO_2) thus appear to have been the dominant feedback that kept the world cold at the last glacial maximum. The bright high-albedo surfaces of the northern ice sheets accounted for much of the rest of the feedback.

In summary, CO_2 acts primarily as an in-phase feedback on ice volume, while CO_2 forcing of ice sheets plays a smaller role. The next two sections summarize a conceptual model of ice-volume oscillations in the 100 000-year band that builds on these conclusions.

6 Conceptual model of ∼100 000-year ice-volume oscillations

6.1 Ice growth every 41 000 years

In the schematic model of the 41 K world (Sect. 3), northern hemisphere ice sheets grew at intervals of 41 000 years, but then melted entirely during the subsequent insolation maxima. Gradually, however, temperatures were cooling during the late Pliocene and early Pleistocene, causing a slow drift to more positive marine δ^{18}O values (Mix et al., 1995). In the late Pleistocene, variations in ice volume became larger in size and longer in wavelength. By 400 000 years ago, sawtooth-shaped δ^{18}O oscillations centered on the

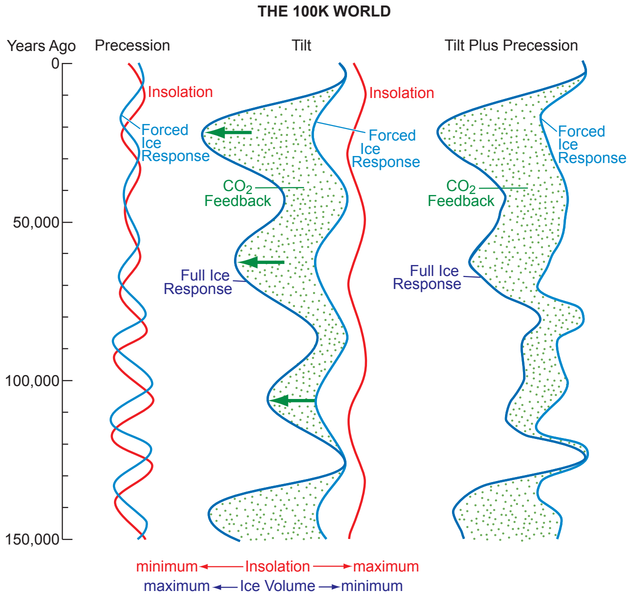

Fig. 8. Schematic model of CO_2 feedback on ice volume in the ~100 K world since 0.9 million years ago. As in the 41 K world, insolation forced ice volume at the precession and tilt cycles, the greater length of the tilt cycle increased the forced ice response compared to precession, and CO_2 feedback amplified ice growth every 41 000 years. Following ice-growth episodes, reduced ablation in a colder world allowed much of the new ice to survive weak insolation minima, and the ice-volume steps at 41 000-year intervals (green arrows) were transformed into a longer-period (~100 K) response.

~100 000-year band had become dominant. The analysis in this section suggests that one simple change in the climate-system response — a substantial reduction in ice ablation — could account for the growth of larger ice sheets centered in the ~100 000-year band rather than 41 000-year cycles.

The focus here is again on the most recent ~100 000-year ice-growth oscillation (Fig. 3), because it is both the best dated and the one with the clearest evidence that δ^{18}O increases represent ice volume. In the benthic marine δ^{18}O stack of Lisiecki and Raymo (2005), the net δ^{18}O shift to glacial-maximum values during this interval was achieved in three distinct steps at prominent δ^{18}O boundaries: substage 5.5/5.4 (+1.0‰), stage 5/4 (+0.8‰), and stage 3/2 (+0.7‰). Only at these three times did the δ^{18}O signal (and global ice volume) reach values higher than at any time earlier in this oscillation. In addition, all three δ^{18}O transitions were times when CO_2 concentrations were falling to prominent minima.

All three of these transitions occurred during times of near-alignment of northern summer insolation minima at the tilt and precession cycles, and thus at times of coincident forcing of ice volume by both cycles. A critical question is whether these new increments of ice growth were related to both insolation cycles or just to one of them.

Two other precession insolation minima during this interval produced small δ^{18}O increases (at the substage 5.3/5.2 boundary and within stage 3), but neither of these increases shifted the δ^{18}O signal to values more positive than those reached previously. This observation suggests that the three ice-growth transitions are critically linked to processes tied to the 41 000-year cycle.

The three shifts toward more positive δ^{18}O values are similar to those during the earlier 41 K world. In the benthic marine δ^{18}O stack of Lisiecki and Raymo (2006), the average δ^{18}O increase during 41 000-year cycles prior to 0.9 million years ago was 0.7‰ (the range was 0.4–1.1‰). The increases across the three δ^{18}O transitions within the last climatic oscillation averaged 0.83‰, or ~20% larger than the mean for the 41 K world.

The primary difference from the earlier 41 K regime is that δ^{18}O values failed to return to their original levels after each new maximum. The 1.0‰ increase across the substage 5.5/5.4 boundary was followed by a 0.35‰ δ^{18}O decrease across the substage 5.4/5.3 transition, leaving a net isotopic shift of +0.65‰. The 0.8‰ increase across the stage 5/4 boundary was followed by a 0.3‰ δ^{18}O decrease across the stage 4/3 transition, leaving a net isotopic shift of +0.5‰ (The stage 3/2 increase of 0.6‰ was erased by termination I.).

These observations suggest a simple schematic model of how the larger ice-volume oscillations of the late Pleistocene developed (Fig. 8). Just as in the 41 K world, insolation changes at the periods of precession and tilt drove lagged ice-volume responses (again assumed to be linear). And again, CO_2 feedback selectively amplified the ice-volume response at 41 000 years, although now by ~20% more than the doubling assumed for the 41 K world (Fig. 4).

The major difference compared to the 41 K world is that ablation was now much lower during the insolation maxima that followed ice-growth episodes every 41 000 years. Roughly 65% of the new ice that grew on these major transitions survived and formed a new base level for further growth. These three assumptions (linear insolation forcing, CO_2 feedback at 41 000 years, and reduced ablation) produce a sawtooth-shaped buildup of ice over intervals that last longer than a single 41 000-year cycle (Fig. 8).

Why would ablation have decreased so markedly between the 41K world and the ~100 K world? The explanation probably lies in the ice mass-balance relationship shown in Fig. 9. Because ice ablation is an exponential function of warm-season temperature, a relatively small polar cooling could have greatly diminished summer ablation. But since ice mass balance in winter is much less sensitive to temperature, polar cooling would have caused little change in net snow accumulation. Consequently, in a world with colder north-polar regions, larger ice sheets would have grown in a step-wise fashion primarily because of reduced ablation.

Such a trend toward reduced ablation is implicit in the longer-term history of northern ice sheets. Prior to

2.75 million years ago, strong ablation in a warmer world kept northern ice sheets of significant size from forming even during the most favorable orbital configurations. From 2.75 to 0.9 million years ago, reduced ablation in a cooler world allowed growth of ice sheets every 41 000 years, but the next insolation maximum (whether weak or strong) melted most or all of the ice. After 0.9 million years ago, further cooling and additional reduction in ablation permitted ice sheets to survive weaker insolation maxima and persist for longer intervals. In this progression toward a colder world, the Antarctic ice sheet is the next step: ablation rates there have fallen so low that ice survives even the largest insolation maxima.

Non-linear CO_2/temperature feedback also helped to reduce rates of ablation. The growth of very large ("100 K") ice sheets presumably drove CO_2 concentrations lower than those that had been attained during the earlier 41 K world. The lower CO_2 values would have further cooled temperatures because of the logarithmic relationship shown in Fig. 5. The cooler temperatures would have further reduced ablation of very large ice sheets. Ice-albedo feedback would also have aided CO_2 in promoting the growth of larger ice sheets.

In this conceptual model (Fig. 8), the growth of larger ice sheets in the ~100 K world primarily results from CO_2 feedback. Insolation forcing of a linear (Milankovitch) ice response accounts for ~25% of the net increase in ice volume between the marine isotopic substage 5.5 interglaciation and the last glacial maximum at isotopic stage 2. The other 75% of the amplitude of the ice build-up results from the transformation of CO_2 feedback at the 41 000-year cycle into asymmetric, saw-tooth oscillations in the ~100 K band. Unlike the 41 K world, the ~100 K world was dominated by feedbacks (primarily CO_2, but also ice albedo).

6.2 Deglaciation every ~100 000 years

A major question remains. Why did abrupt deglacial terminations occur within a band centered near 100 000 years? The discovery that the ~100 000-year component of $\delta^{18}O$ is phase-locked to eccentricity (Hays et al., 1976) suggested that these deglaciations might be linked in some way to modulation of precession by eccentricity at periods averaging near 100 000 years.

Raymo (1997) noted that the time span between successive terminations tends to fall on or near multiples of the 23 000-year precession period, either after four cycles (~92 000 years) or five (~115 000 years). She concluded that the ~100 000-year "cycle" is actually quantum in nature, rather than periodic, and that it is paced mainly by eccentricity modulation of precession peaks in insolation, with insolation maxima at the tilt cycle playing a lesser role.

In contrast, Huybers and Wunsch (2004) concluded that terminations occur at or near multiples of the 41 000-year tilt cycle: after either two cycles (~82 000 years) or three (~123 000 years). One problem with invoking only tilt as the key forcing is that a large amount of ice had accumulated by isotopic stage 4, and a strong 41 000-year insolation maximum followed in stage 3, but less than half of the ice present at that time melted (~30 m of sea-level equivalent).

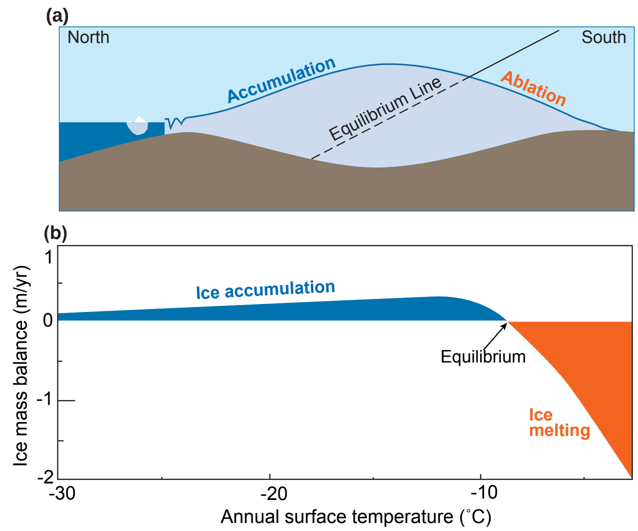

Fig. 9. Geographic and climatic constraints on ice-sheet mass balance. **(a)** Zonal cross section of northern ice sheets. Equilibrium line separates areas of net accumulation and ablation. **(b)** Ice mass balance as a function of mean annual temperature. Long-term cooling reduces strong warm-season ablation.

Insolation forcing at the periods of both precession and tilt probably plays a role in determining the timing of terminations (Imbrie et al., 1992, 1993). The individual contributions from these two periods were closely aligned at terminations I and IV and were offset by only ~4000 years on termination II. Tilt and precession forcing were not closely aligned during Termination III, and probably as a result, deglaciation at that time was incomplete.

But only a small fraction of the amount of ice melting at terminations can be explained by a linear ice response to insolation forcing. Most of the melting resulted from feedback processes operating within the climate system (Imbrie et al., 1993).

CO_2 is the primary internal feedback that maintained cold temperatures during the glacial maximum (Sect. 5), and CO_2 was proposed earlier in this section as the key positive feedback that causes non-linear growth of large ice sheets (Fig. 8). When favorable orbital configurations provided sufficient forcing to initiate deglacial melting, this same positive feedback was readily available to amplify ice melting. As the large glacial-maximum ice sheets began to melt, their shrinkage allowed CO_2 levels to rise, and the rising CO_2 concentrations enhanced melting of the remaining ice. In effect, the positive CO_2 feedback embedded in the existence of large glacial-maximum ice sheets holds much of the impetus of the non-linear destruction of the ice.

CO_2 feedback also has a direct impact on the timing of deglaciations because of its amplification of the forced ice response at 41 000 years. This amplification makes tilt a more important factor during terminations than its strength in 65° N insolation trends would suggest. As a result, close alignment of tilt and precession insolation forcing is critical.

CO_2 feedback also affects the timing of terminations in an indirect way. Accumulation of "extra" ice volume occurs every 41 000 years, but a single tilt cycle does not build enough ice to allow a major termination (see also Huybers and Wunsch, 2004). Two or three intervals of ice growth are needed for enough ice to accumulate (∼82 000 or ∼123 000 years). This ice-growth constraint, combined with the one imposed by maximum ablation at multiples of the precession cycle (Raymo, 1997), tends to limit terminations to one of two intervals: either every 82 000–92 000 years (two tilt cycles and four precession cycles) or every 115 000–123 000 years (five precession cycles and three tilt cycles).

As noted in Sect. 2, greenhouse-gas changes at the 23 000-year precession cycle also add directly to insolation forcing of ice volume. At this cycle, both CO_2 and CH_4 have the same "early" phase as summer insolation and thus force ice volume (Fig. 2b). Because of the modulation of precession by eccentricity, this forcing should be strongest during interglacial isotopic substages, beginning with large peaks on terminations. Coincident CO_2 and CH_4 maxima that occurred late in the last four deglaciations approximately coincident with insolation maxima at the precession cycle provided additional forcing to melt ice.

Amplification of the tilt cycle by CO_2 feedback may help to resolve two other shortcomings in the Milankovitch insolation theory. One problem is that the 65° N insolation maxima that drove ice melting on isotopic substages 5.3 and 5.1 were larger than the one on Termination I (Fig. 8), yet they failed to melt the small amount of stage 5.4 and stage 5.2 ice, whereas the weaker insolation maximum 11 000 years ago melted all of the stage 2 glacial-maximum ice. This inconsistency in the Milankovitch hypothesis may be explained by amplification of the forced response at 41 000 years by CO_2 feedback. In the latter part of stage 5, tilt was out of alignment with the two precession-cycle insolation maxima and thus made no feedback-amplified contribution to ice melting. In contrast, the insolation maximum at the tilt cycle late in termination 1 was in almost perfect alignment with the insolation maximum at precession, and the CO_2-amplified contribution at the 41 000-year period added to the precession forcing.

The most perplexing enigma for the Milankovitch theory is Termination V, a large deglaciation that occurred during a time when 65° N insolation forcing was weak and when the tilt and precession trends were totally out of alignment. Greenhouse-gas feedback may help to explain this anomalous response in two ways. First, because an unusually large volume of ice had accumulated in isotopic stage 12, an unusually large amount of "stored" feedback (from CO_2 and ice-albedo) was available to amplify any impetus toward ice melting. These feedbacks would have played a relatively larger role during termination V than on other terminations.

Second, termination V took ∼20 000 years (Imbrie et al., 1984; Lisiecki and Raymo, 2006), while terminations I through IV only lasted ∼10 000 years. This longer duration requires slow but steady forcing that extended from the weak insolation maximum at the precession cycle 425 000 years ago to the second weak insolation maximum at the precession cycle 408 000 years ago. The key question is why ice melting persisted through a weak 65° N insolation minimum 416 000 years ago.

Although that insolation minimum was produced by the precession cycle, the tilt cycle reached a maximum at the same time. If CO_2 feedback amplified the effect of insolation forcing at the tilt cycle, it could have offset this weak precession minimum, bridged the gap, and produced continuous melting across the entire termination.

7 How did ice sheets control CO_2?

What mechanisms were responsible for ice-sheet control of atmospheric CO_2 concentrations? Obviously, the linkage would have to have occurred through processes with the same behavior in time as ice volume and CO_2: prominent variations at the 41 000-year cycle prior to 0.9 million years ago, large variations at both 41 000 years and ∼100 000 years since that time, and phases near those of both ice volume and CO_2. Processes with these time characteristics could link CO_2 and ice volume during both ice growth and decay.

One possible link is a fast polar-alkalinity response (Broecker and Peng, 1989). Changes in atmospheric circulation driven by northern ice sheets affect deep-water circulation in the Atlantic (Boyle and Keigwin, 1985; Raymo et al., 1989). Variations in the depth of penetration of northern-source deep waters alter the relative areas of Atlantic sea floor bathed by corrosive southern-source waters and less corrosive northern-source waters, with resulting effects on $CaCO_3$ dissolution on the Atlantic sea floor. With rapid southward flow in the deep Atlantic, these changes alter the mixed-layer chemistry (alkalinity) of the Southern Ocean a few hundred years later when deep Atlantic waters later reaches the surface, and thereby affect atmospheric CO_2. During the northern hemisphere ice-age cycles, the $\delta^{13}C$ proxy for 'NADW' flow (Raymo et al., 1997) varied mainly at 41 000 years until 0.9 million years ago and subsequently within the ∼100 000-year band. Both variations were phased with $\delta^{18}O$ (ice volume).

One way of potentially altering atmospheric CO_2 is tied to glacial strengthening of the Asian winter monsoon and resulting effects on the relatively carbon-rich surface waters of the North Pacific. Fertilization of Pacific surface waters by monsoon-generated dust could cause increased production and sinking of planktic algae, with greater sequestration of carbon out of contact with the atmosphere (Martin, 1990). Asian loesses accumulated at a cycle near 41 000 years before 0.9 Myr ago and within the ~100 000-year band since that time (Kukla et al., 1990). Glacial maxima also produced increased Eurasian dust fluxes to the western North Pacific (Hovan et al., 1989) and to Greenland ice (Mayewski et al., 1996). The Eurasian dust fluxes were in phase with, or lagged slightly behind, changes in δ^{18}O/ice volume.

Another mechanism is increased stratification of surface waters or increases in sea-ice cover that might have limited the release of CO_2 to the atmosphere (Francois et al., 1997; Sigman and Boyle, 2000; Stephens and Keeling, 2000). In the western subpolar North Pacific, frigid winter monsoon winds from Asia caused glacial-maximum increases in both sea ice and surface-water stratification (Morley and Hays, 1983; Jaccard et al., 2005) that could have reduced CO_2 fluxes to the atmosphere. These changes occurred at a 41 000-year tempo prior to 0.9 million years, and later within the ~100 000 year band (Morley and Dworetsky, 1991).

Glacial increases in southern hemisphere dust fluxes (Ridgwell and Watson, 2002), in Southern Ocean sea-ice cover (Stephens and Keeling, 2000), and in surface-water stratification (Francois et al., 1997; Sigman and Boyle, 2000) also have considerable potential to alter atmospheric CO_2 levels, but plausible links to northern ice sheets have proven elusive. One way to project changes from north to south is via the greenhouse-gas changes themselves. "First-stage" variations in CO_2 tied directly to the northern ice sheets via proximal northern responses could cause "second-stage" changes in distal southern hemisphere dust fluxes and Southern ocean sea ice or surface stratification that could further amplify the CO_2 response.

Very large uncertainties persist as to how large a CO_2 response each of these processes might explain (Sigman and Boyle, 2000), and processes not included here may also have provided ice-driven CO_2 feedback. Whatever the mechanism or mechanisms that operated, the evidence in this paper requires that the CO_2 feedback links must have occurred at the 41 000-year cycle and within the ~100 000-year band, but not at the 23 000-year cycle.

8 Summary

In the hypothesis presented here, intervals of ice-sheet growth during the last 2.7 million years share two characteristics: (1) insolation forcing of linear ("Milankovitch") ice-volume responses at the tilt and precession cycles; and (2) amplification of the forced 41 000-year ice response by CO_2 feedback. The growth of 41 000-year ice sheets prior to 0.9 million years ago can be explained by CO_2-feedback amplification of the forced ice response to changes in tilt. After 0.9 million years ago, similar episodes of CO_2-amplified ice growth continued at 41 000-year intervals, but polar cooling suppressed ice ablation during subsequent intervals. The net result was a series of step-like transitions toward greater ice volume that produced the asymmetric sawtooth-shaped ice oscillations of the ~100 K world. The same positive CO_2 feedback that caused non-linear growth of ice sheets in this new regime was then available to enhance the amplitude of ice melting during times when insolation forcing became favorable. Although precession dominated insolation cycles at 65° N and constrained the pacing of terminations to the ~100 000-year band, amplification of ice-growth at 41 000-year cycles by CO_2 feedback was also important in the timing of deglaciations.

The CO_2 feedback hypothesis can explain why the northern and southern hemispheres responded nearly in phase on terminations (Broecker and Denton, 1989). Near the ice sheets, changes in ice-sheet size set the climatic tempo. Far from the ice sheets, most climatic responses were strongly affected by an atmospheric CO_2 signal that was largely controlled by (and in phase with) the northern ice sheets. As a result, most global climatic signals were ice-driven and nearly synchronous. An exception is the tropics, where summer insolation forcing produced very strong monsoon responses that were largely independent of northern ice (Kutzbach, 1981).

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