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# Ice-driven CO<sub>2</sub> feedback on ice volume

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CPD

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## Abstract

The origin of the major ice-sheet variations during the last 2.7 million years remains a mystery. Neither the dominant 41 000-year cycles in  $\delta^{18}\text{O}$  and ice-volume during the late Pliocene and early Pleistocene nor the late-Pleistocene variations near 100 000 years is a linear (“Milankovitch”) response to summer insolation forcing. Both result from non-linear behavior within the climate system. Greenhouse gases (primarily  $\text{CO}_2$ ) are a plausible source of this 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,  $\text{CO}_2$  and ice volume (marine  $\delta^{18}\text{O}$ ) have varied in phase both at the 41 000-year obliquity cycle and within the  $\sim 100\,000$ -year eccentricity band. This timing argues against greenhouse-gas forcing of a slow ice response and instead favors ice control of a fast  $\text{CO}_2$  response. Because the effect of  $\text{CO}_2$  on temperature is logarithmic, the temperature/ $\text{CO}_2$  feedback on ice volume is also logarithmic.

In the schematic model proposed here, ice sheets were forced by insolation changes at the precession and obliquity cycles prior to 0.9 million years ago and responded in a linear way, but  $\text{CO}_2$  feedback amplified (roughly doubled) the ice response at 41 000 years. After 0.9 million years ago, as polar climates continued to cool, ablation weakened.  $\text{CO}_2$  feedback continued to amplify ice-sheet growth at 41 000-year intervals, but weaker ablation permitted ice to survive subsequent insolation maxima of low intensity. These longer-lived ice sheets persisted until peaks in northern summer insolation paced abrupt deglaciations every  $100\,000 \pm 15\,000$  years. Most ice melting during deglaciations was achieved by the same  $\text{CO}_2$ /temperature feedback that had built the ice sheets, but now acting in the opposite direction. Several processes have the northern geographic origin, as well as the requisite orbital tempo and phasing, to have been the mechanisms by which ice sheets controlled  $\text{CO}_2$  and drove 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  $\delta^{18}\text{O}$  as an ice-volume proxy, Hays et al. (1976) confirmed that ice sheets fluctuate at those periods and verified Milankovitch's prediction that the responses lag several thousand years behind orbital forcing (Fig. 1a).

Milankovitch did not anticipate two characteristics of marine  $\delta^{18}\text{O}$  records. One is the strong  $\delta^{18}\text{O}$  (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. The second unexpected feature is the strength of the 41 000-year cycle in  $\delta^{18}\text{O}$  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 2 to 3 times stronger at the period of precession than at the period of tilt.

Milankovitch's insolation hypothesis thus provides a valid starting point for an orbital theory of climate, but not a complete answer. As a result, many scientists have explored the next most important orbital-scale variable in the climate system – changes in concentration of carbon dioxide ( $\text{CO}_2$ ) and methane ( $\text{CH}_4$ ). At this point, however, widely divergent views coexist about the effect of greenhouse gases (particularly  $\text{CO}_2$ ) on ice sheets. Two end-member views are currently in play.

One possibility is that  $\text{CO}_2$  forces ice volume (Pisias and Shackleton, 1986; Genthon et al., 1987; Imbrie et al., 1992, 1993; Broecker and Henderson, 1998; Shackleton, 2000). In this view, changes in  $\text{CO}_2$  “push” the slow-responding ice sheets, which respond with lags of several thousand years (Fig. 1b). These lags are analogous to the forced ice-volume response to changes in insolation (Fig. 1a).

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A different view holds that CO<sub>2</sub> concentrations are controlled by ice volume and act as a positive feedback on ice-sheet mass balance (Ruddiman, 2003; see also Clark et al., 1999). In this case, little or no lag exists between changes in CO<sub>2</sub> and ice volume (Fig. 1c). Each increment of ice growth (whether over a millennium or a century) causes greenhouse-gas concentrations to fall by a certain amount, 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<sub>2</sub> concentrations are parts of the overall response of a highly coupled climate system with complex internal feedbacks, progress in understanding cause and effect at orbital scales has been difficult. Laurent Labeyrie once aptly noted the basic problem: “Everything is correlated to everything”.

One clue to the cause-and-effect problem lies in 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. Relative phasing of CO<sub>2</sub> and ice volume: spectral analysis

The SPECMAP Project (Imbrie et al., 1992, 1993) laid out 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<sub>2</sub> variations, SPECMAP attempted to use geochemical proxies for this purpose. They concluded that CO<sub>2</sub> changes arise as an intermediate-stage internal response within the climate system, but then act as the primary forcing of a slow northern ice response. (SPECMAP did not consider the role of methane.)

At the periods of orbital precession and tilt, 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<sub>2</sub>

variations that in turn act as the ultimate driver of the northern ice sheets (Fig. 2a).

Attention has turned to the CO<sub>2</sub> record preserved in air bubbles at Vostok (Petit et al., 1999; EPICA Community Members, 2004). Shackleton (2000) created a gas time scale for Vostok by tuning the precession component of  $\delta^{18}\text{O}_{\text{air}}$  to an insolation target signal with a September phase. Ruddiman and Raymo (2003) developed an independent gas time scale by tuning the precession component of the CH<sub>4</sub> signal to an insolation target with mid-July (summer monsoon) phasing. Although these two time scales differed at specific levels, they yielded average phases that agreed to within less than 100 years for the greenhouse gases (see also Bender, 2002).

At the 23 000-year precession period, both CO<sub>2</sub> and CH<sub>4</sub> have phases on or close to that of northern mid-summer insolation (Fig. 2c). For methane, this timing is supported by the match between the CH<sub>4</sub> 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). For CO<sub>2</sub>, the phase at the 23 000-year period falls less than 1000 years after that of July insolation (Fig. 2c). These early phases for both methane and CO<sub>2</sub> mean 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<sub>2</sub> vary in phase with  $\delta^{18}\text{O}/\text{ice volume}$  at the 41 000-year period of obliquity (Fig. 2b). This in-phase behavior rules out a slow ice response to greenhouse-gas forcing. Instead, the ice sheets must drive a fast greenhouse-gas response with little or no lag. The gas variations then provide positive feedback to both the growth and melting of ice sheets.

For the 100 000-year period, SPECMAP (Imbrie et al., 19993) proposed that CO<sub>2</sub> changes again occur as an intermediate step within a long chain of responses and force ice volume (Fig. 3a). The initial driver of the CO<sub>2</sub> signal remained unidentified and was referred to as an “internal thermal forcing” (“ITF”). SPECMAP speculated that this initial CO<sub>2</sub> response somehow arose within the climate system when ice sheets began

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to exceed a certain size threshold and create other feedbacks. But the immediate role of CO<sub>2</sub> was to force “sluggish” ice sheets that responded with a lag of ~12 000 years.

Shackleton (2000) later determined that the ~100 000-year CO<sub>2</sub> signal in Vostok ice has a much later phase than SPECMAP had inferred from geochemical proxies, one lying on or very close to the phase of eccentricity (Fig. 3b). He proposed that the 100 000-year CO<sub>2</sub> signal arises independently from processes operating within the carbon system. Shackleton further proposed that the 100 000-year marine δ<sup>18</sup>O signal in benthic foraminifera carries a large temperature overprint and that the actual 100 000-year phase of ice volume is offset some 12 000 years later than the δ<sup>18</sup>O signal. This new interpretation maintained CO<sub>2</sub> as the immediate forcing of ice volume, but it shifted the entire forcing-and-response relationship some 10 000 years later in time compared to the scheme proposed by SPECMAP (Fig. 3b).

Ruddiman and Raymo (2003) confirmed that the CO<sub>2</sub> signal at ~100 000 years has a phase close to that of eccentricity, but Ruddiman (2003) concluded that the extremely late phase inferred by Shackleton (2003) for the 100 000-year component of ice volume is not supported by other evidence. For example, it predicts that an ice-volume minimum at that period should have occurred 98 000 years ago, 15 000 years after the preceding eccentricity maximum (Fig. 4). This timing implicitly requires that ice must have been gradually melting at the ~100 000-year period for the entire first half of marine isotopic stage 5. But securely dated coral reefs show that global ice volume was already smaller than today by 125 000 years ago (substage 5.5). How (and where) could ice sheets still have been melting throughout this ice-volume minimum? Oxygen-isotopic and coral-reef evidence also 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 also seems particularly 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 melting at the same time elsewhere on Earth? Similar problems recur in other interglaciations.

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A second inconsistency in proposing such a late ice-volume response is that it would lag thousands of years behind a group of northern responses long regarded as “ice-driven”: North Atlantic sea-surface temperatures, northern hemisphere dust fluxes, and NADW flow ((Ruddiman and McIntyre, 1984; Imbrie et al., 1993). If these signals are indeed driven by ice sheets, how could they lead their “drivers” by ~10 000 years (Fig. 3b)?

Based on these criticisms, Ruddiman (2003) concluded that  $\delta^{18}\text{O}$  is a good proxy for ice volume during the large climatic oscillations of the late Pleistocene, as previously concluded by Hays et al. (1976), SPECMAP (Fig. 3a), and many other studies over several decades. If so, CO<sub>2</sub> and  $\delta^{18}\text{O}$  (ice-volume) have very nearly the same 100 000-year phase. Such a close phasing rules out CO<sub>2</sub> forcing of “sluggish” ice sheets with lags of ~12 000 years (Fig. 3a, b) and indicates that ice volume controls a fast CO<sub>2</sub> response, with CO<sub>2</sub> feeding back positively on ice volume (Fig. 3c).

In summary, time-series analysis suggests that the predominant role of CO<sub>2</sub> over the last ~400 000 years has been as a fast feedback on changes in ice volume, rather than as a forcing of ice sheets. CO<sub>2</sub> acts as an ice-forced in-phase feedback both at the moderately strong 41 000-year period and within the much stronger ~100 000-year band. At the 23 000-year precession period, CO<sub>2</sub> does force ice volume, but the amount of CO<sub>2</sub> power at that period is small (Petit et al., 1999; Shackleton, 2000; Ruddiman and Raymo, 2003).

### 3. Other evidence of a predominant feedback role for CO<sub>2</sub>

Time series analysis is not an optimal way to assess leads and lags between CO<sub>2</sub> and  $\delta^{18}\text{O}$  in the ~100 000-year band (Ruddiman, 2003). Both signals have highly asymmetric shapes that drift gradually toward “glacial” values (low CO<sub>2</sub>, positive  $\delta^{18}\text{O}$ ) but then switch abruptly back to “interglacial” conditions (high CO<sub>2</sub>, more negative  $\delta^{18}\text{O}$ ) during deglacial transitions. In contrast, spectral analysis decomposes climate signals into a series of sine waves characterized by smooth, gradual changes. A symmet-



rical sine wave with a period near 100 000 years cannot capture either the abruptness of deglacial terminations or the fundamental sawtooth asymmetry of the major glacial/interglacial cycles near  $\sim 100\,000$  years (Muller and MacDonald, 2000).

### 3.1. Relationship between $\text{CO}_2$ and $\delta^{18}\text{O}$ (ice volume) during the Last 140 000 years

5 An alternative approach is to examine the relative overall timing of the full  $\text{CO}_2$  and ice-volume signals. For the last sawtooth-shaped climatic oscillation (Fig. 5), numerous well-constrained estimates of sea level from coral reefs exist (Chappell and Shackleton, 1986; Bard et al., 1990). These can be converted directly to ice volume and also to changes in  $\delta^{18}\text{O}$  (assuming 10 m of sea-level change per 0.11‰ of  $\delta^{18}\text{O}$  shift). By this  
10 measure, every major  $\delta^{18}\text{O}$  transition in Fig. 5 is dominated by sea-level (ice volume): 50–70% of terminations II (the stage 6/5 boundary) and I (the stage 2/1 boundary); 50–60% of the substage 5.5/5.4  $\delta^{18}\text{O}$  transition; 100% of the substage 5.4/5.3 boundary,  $\sim 70\%$  of the stage 5/4 transition, and  $\sim 65\%$  of the stage 3/2 boundary. In each case, ice volume accounts for well over half of the  $\delta^{18}\text{O}$  change. For this climatic oscillation,  
15 no major offset between  $\delta^{18}\text{O}$  and ice volume is possible.

The most obvious message from comparing the overall saw-tooth shapes of the  $\text{CO}_2$  and  $\delta^{18}\text{O}$  signals during the last climatic oscillation is one of very close similarity (Fig. 5). This similarity again rules out a 12 000-year  $\text{CO}_2$  lead relative to ice volume ( $\delta^{18}\text{O}$ ) and it further supports the interpretation that ice sheets control a fast  $\text{CO}_2$   
20 response at the 100 000-year period.

### 3.2. $\text{CO}_2$ /ice phasing on terminations

A related approach is to focus on the relative timing of  $\text{CO}_2$  and ice-volume changes across the distinctive and abrupt deglacial terminations. Because terminations are times when changes at all three orbital periods are superimposed, this approach as-  
25 sesses their combined effect. During termination I (the best-dated deglaciation),  $\text{CO}_2$  changes led direct coral-reef (sea-level) indices of ice volume by 1000 years or less

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(Alley et al., 2002). The estimated CO<sub>2</sub> lead relative to δ<sup>18</sup>O on termination II was ~3000 years (Broecker and Henderson, 1998), although sizeable uncertainties exist in dating and in the various “ice–volume” indices (Alley et al., 2002).

The results from spectral analysis (Sect. 2) provide a plausible interpretation for the CO<sub>2</sub>/δ<sup>18</sup>O phasing during these deglaciations. The two signals are nearly synchronous across the terminations because they are dominated by the in-phase timing at ~100 000 and 41 000 years (Fig. 2c, 3c). The small CO<sub>2</sub> lead is explained by a lesser contribution from the 23 000-year period, at which CO<sub>2</sub> leads ice volume by ~5000 years (Fig. 2c). If 15% of the overall signal results from the ~5000-year lead at 23 000 years, and the other 85% from the in-phase timing at 100 000 and 41 000 years, the net lead will be ~750 years (0.15×5000 years + 0.85×0 years).

Other trends in Fig. 5 are also consistent with the results from spectral analysis. The δ<sup>18</sup>O maxima at marine isotopic stages 4 and 2 are clear 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<sub>2</sub> minima at these times indicate that CO<sub>2</sub> acted as an in-phase feedback at the obliquity cycle, amplifying the size of these ice-volume maxima without altering their phase.

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 (~250 ppb) methane variations led δ<sup>18</sup>O by several thousand years, indicating that methane 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<sub>2</sub> signal. The main message from Fig. 5, however, is the basic similarity in timing of changes in CO<sub>2</sub> and δ<sup>18</sup>O and the fast-feedback role for CO<sub>2</sub>.

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### 3.3. CO<sub>2</sub> feedback at the last glacial maximum

Additional evidence of the feedback role of CO<sub>2</sub> comes from the last glacial maximum, the only pre-modern interval examined in sufficient regional detail to allow global-scale assessment of the processes affecting 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<sub>2</sub> lowering provided a radiative cooling of ~0.67°C, and the ~320-ppb CH<sub>4</sub> decrease added another ~0.14°C, allowing for 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 for a 2.5°C global-mean climate sensitivity to CO<sub>2</sub> doubling. The resulting total greenhouse-gas cooling of 1.7°C accounted for about 40% of the total cooling of ~4.5°C(±0.7°) at the last glacial maximum.

Albedo-temperature feedback accounted 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 increase in Antarctic sea ice has been widely attributed to lower greenhouse-gas levels (Hansen et al., 1984; Broccoli and Manabe, 1987), this effect can be added to the greenhouse-gas ledger, bringing the total gas contribution to more than 50%. In addition, part of the remaining albedo increase that came from a reduction in glacial vegetation cover was caused by lower CO<sub>2</sub> values, further increasing the greenhouse contribution (Levis and Foley, 1999). In summary, greenhouse gases (mainly CO<sub>2</sub>) were the dominant feedback at the last glacial maximum. The bright high-albedo surfaces of the northern ice sheets accounted for much of the rest of the feedback that

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kept glacial climates cold.

### 3.4. Quantifying the links between CO<sub>2</sub> and ice volume

The ice volume and CO<sub>2</sub> trends in Fig. 5 follow a quasi-linear relationship (Fig. 6a). Muddlessee (2001) noted a similar relationship between Vostok CO<sub>2</sub> and the  $\delta^{18}\text{O}$  stack of Bassinot et al. (1994). Several factors complicate this comparison: (1) time-varying temperature overprints on the  $\delta^{18}\text{O}$  records used in the SPECMAP stack; (2) offsets of several thousand years caused by errors in the relative ages of the CO<sub>2</sub> and  $\delta^{18}\text{O}$  signals; and (3) sub-orbital oscillations in the CO<sub>2</sub> record that were suppressed in the SPECMAP  $\delta^{18}\text{O}$  signal by stacking and smoothing.

The effect of CO<sub>2</sub> on global temperature is logarithmic (Oglesby and Saltzman, 1990). The change in global temperature per unit change of CO<sub>2</sub> increases as CO<sub>2</sub> concentrations fall (Fig. 6b). Because of this relationship, the positive temperature feedback on ice volume from changes in CO<sub>2</sub> is also logarithmic. As ice sheets grow and drive CO<sub>2</sub> values lower in a quasi-linear way, the size of the temperature feedback on the ice sheets increases logarithmically.

## 4. Conceptual models of CO<sub>2</sub> feedback on ice-age cycles

This section summarizes conceptual schematic models of the possible contribution of greenhouse-gas feedback to the major nonlinearities in long-term ice-volume behavior: the dominant 41 000-year ice-volume cycles during the late Pliocene and early Pleistocene, and the dominant oscillations near 100 000 years during the late Pleistocene. To simplify the comparison of these two regimes, both schematic examples use the same 60° N insolation trends (those of the last 150 000 years) to force the ice sheets. In both cases, the ice sheets are assumed to react with simple lagged linear responses to the forcing, analogous to the “coherent” orbital signals that SPECMAP extracted by filtering  $\delta^{18}\text{O}$  signals.

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#### 4.1. The “41 K Ice World”: CO<sub>2</sub> feedback at 41 000 years

As noted in the introduction, Milankovitch did not anticipate the dominance of 41 000-year  $\delta^{18}\text{O}$  variations through nearly two million years of ice-age cycles (Pisias and Moore, 1981; Ruddiman et al., 1986; Raymo et al., 1989) compared to the stronger insolation forcing from precession. This mismatch is to some extent reduced by the fact that ice volume has almost twice as long to respond to forcing at a 41 000-year cycle as it does at a 23 000-year cycle because of the greater interval over which the forcing persists (41 K/23 K $\approx$ 1.8). In the schematic example shown in Fig. 7, the small amount of 41 000-year forcing is thus boosted by a factor of 1.8 in the “forced” ice-volume response. Nevertheless, the average response at the precession period still exceeds that at obliquity by an average of almost 40%.

One way to resolve the remaining mismatch between the forcing and the observed  $\delta^{18}\text{O}$  responses is to suppress the 23 000-year precession component of the  $\delta^{18}\text{O}$  signal, for example by interhemispheric cancellation of oppositely-phased ice responses between the northern and southern hemispheres (Raymo, 2005). The alternative explanation (favored here) is that the obliquity signal was enhanced by internal feedback within the climate system.

One proposed feedback at the 41 000-year cycle is a greater northward flux of tropical moisture because of an increased planetary temperature gradient caused by insolation changes at the obliquity cycle (Young and Bradley, 1984; Raymo and Nisancioglu, 2003; Vettoretti and Peltier, 2004). In general, however, glacial geologists and glaciologists view accumulation as a far weaker factor in ice-sheet mass balance than ablation (Alley, 2003; Denton et al., 2005). In addition, most GCM simulations show reduced precipitation in regions where ice actually accumulates because the cooling climate reduces water vapor in the atmosphere, and because growing ice sheets produce a strongly negative moisture feedback at higher elevations.

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

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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. 7, the direct ice-volume response to insolation forcing at the 41 000-year cycle is assumed to have doubled in size because of positive CO<sub>2</sub> (and ice-albedo) feedback. This doubling makes the 41 000-year ice-volume signal the dominant orbital response. Although the level of dominance in this example does not match that in marine benthic  $\delta^{18}\text{O}$  records, isotopic signals were overprinted by a large (and apparently in-phase) temperature response at the 41 000-year tilt cycle (Raymo et al., 1989). That overprint exaggerated the strength of the 41 000-year signal in benthic  $\delta^{18}\text{O}$  records compared to actual changes in ice volume.

In this schematic model of the 41 000-year regime, insolation forcing accounts for about 70% of the amplitude of the orbital-band ice-volume response, and CO<sub>2</sub> feedback provides the other 30%. Even though CO<sub>2</sub> (and ice-albedo) feedback is critical to explaining the prominence of the 41 000-year signal, this was a world in which forced responses dominated ice-volume behavior.

#### 4.2. The “~100 K Ice World”: CO<sub>2</sub> feedback at 41 000 and ~100 000 years

In the schematic model of the 41 K world, an implicit assumption was made that polar warmth during the late Pliocene and early Pleistocene caused northern hemisphere ice to melt during insolation maxima that followed intervals of ice growth every 41 000 years. But marine oxygen-isotopic evidence (Mix et al., 1995) shows that the planet continued to cool and that the ice-sheet response changed during the last one million years. By 400 000 years ago, ice-volume variations had become larger in size, saw-toothed in shape, and centered in the ~100 000-year eccentricity band. As proposed below, one simple change in the climate-system response – substantially reduced ablation – can account for the growth of larger ice sheets in the ~100 000-year band compared to the earlier 41 000-year cycles.

As summarized in Sect. 3, the most recent ~100 000-year ice-growth oscillation is the best-dated and also the one in which  $\delta^{18}\text{O}$  increases can be unambiguously linked

to ice-volume increases (Fig. 5). In the benthic marine  $\delta^{18}\text{O}$  stack of Lisecki and Raymo (2005), the net  $\delta^{18}\text{O}$  shift to glacial-maximum values during this interval was achieved in three distinct steps across prominent  $\delta^{18}\text{O}$  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  $\delta^{18}\text{O}$  signal (and global ice volume) reach values higher than at any time earlier in this ice-growth oscillation (as is also evident in Fig. 5).

All three 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. Two other precession minima produced smaller  $\delta^{18}\text{O}$  increases (one at the substage 5.3/5.2 boundary and the other within stage 3), but neither of these increases shifted the  $\delta^{18}\text{O}$  signal to values more positive than before. The three ice-growth transitions thus appear to be uniquely linked to insolation minima at the 41 000-year cycle. All three  $\delta^{18}\text{O}$  transitions were also times when  $\text{CO}_2$  concentrations fell to prominent minima (Fig. 5).

These shifts toward more positive  $\delta^{18}\text{O}$  values in the late Pleistocene are similar in magnitude to those during the earlier 41 K world. In the benthic marine  $\delta^{18}\text{O}$  stack of Lisecki and Raymo (2005), the average  $\delta^{18}\text{O}$  increase during all 41 000-year cycles prior to 0.9 million years ago was 0.7‰ (range of 0.4 to 1.1‰). The three  $\delta^{18}\text{O}$  transitions during the last climatic oscillation average 0.83‰, or ~20% larger than the mean for the 41 K world.

The primary difference from the earlier 41 K regime is that  $\delta^{18}\text{O}$  values did not return to their original levels after reaching these prominent new maxima. The 1.0‰ increase across the substage 5.5/5.4 boundary was followed by a 0.35‰  $\delta^{18}\text{O}$  decrease across the substage 5.4/5.3 transition, resulting in a net isotopic shift of +0.65‰. The 0.8‰ increase across the stage 5/4 boundary was followed by a 0.3‰  $\delta^{18}\text{O}$  decrease across the stage 4/3 transition, producing a net isotopic shift of +0.5‰. The stage 3/2 increase of 0.7‰ preceded termination I.

These observations suggest a simple schematic model of how the ~100 000-year oscillations in ice volume of the late Pleistocene developed (Fig. 8). As in the 41 K world,

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insolation changes at the periods of both precession and tilt drove lagged ice-volume responses assumed to have been linear. CO<sub>2</sub> feedback again selectively amplified the ice-volume response at 41 000 years, but now by an amount ~20% larger than the doubling previously assumed for the 41 K world. The one major difference from the earlier 41 K world is that ablation is now assumed to have been much lower during the insolation maxima that followed ice-growth episodes every 41 000 years: roughly 65% of the new ice that had grown on these major transitions survived and formed a new baseline for further growth. These simple assumptions (linear insolation forcing, CO<sub>2</sub> feedback at 41 000 years, and reduced ablation) produce a sawtooth-shaped ice buildup over this ~100 000-year interval (Fig. 8). Ice-albedo feedback would also have joined CO<sub>2</sub> in promoting the growth of ice sheets.

Why would ablation have decreased so markedly between the 41 K world and the ~100 K world? Because marine  $\delta^{18}\text{O}$  trends show ongoing polar cooling for the last several million years, basic glaciology points to a plausible explanation (Fig. 9). Because ice ablation is an exponential function of warm-season temperature, a relatively small polar cooling could have produced a large decrease in summer ablation and a substantial positive effect on ice-sheet mass balance. But because winter-season accumulation of snow (ice) is much less sensitive to changes in temperature, polar cooling would have caused little increase in snow accumulation. In a colder world, larger ice sheets would have grown in a step-wise fashion because of reduced ablation.

An obvious trend toward reduced ablation is implicit in the longer-term history of northern ice sheets. Prior to 2.8 million years ago, strong ablation in a warmer world kept northern ice sheets of significant size from forming even during favorable orbital configurations. From 2.8 to 0.9 million years, cooling and reduced ablation permitted the growth of ice sheets every 41 000 years (as described above), but subsequent insolation maxima (whether weak or strong) melted the ice. After 0.9 million years ago, further cooling and reduction in ablation permitted ice sheets to survive weak insolation maxima and persist for longer intervals of ~85 000 to ~115 000 years (Sect. 6). The Antarctic ice sheet represents the next step in this ice-age progression: ablation rates

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in Antarctica have fallen so low that ice survives even the largest insolation maxima.

Non-linear CO<sub>2</sub>/temperature feedback would also have helped to reduce rates of ablation. The growth of very large ice sheets after the stage 5/4 isotopic boundary should have driven CO<sub>2</sub> concentrations lower than those attained during the earlier 41 K world. These lower CO<sub>2</sub> values would have further cooled temperatures via the logarithmic relationship shown in Fig. 6b and thereby further reduced ablation of the large ice sheets.

In the previous schematic model of the 41 K world (Fig. 7), insolation forcing of a linear ice-volume response accounted for ~70% of the amplitude of ice-sheet variations. In the larger oscillations of the ~100 K world (Figs. 5, 8), Milankovitch forcing accounts for well under half the amplitude (Imbrie et al., 1993). Most of the ~100 K signal results from the transformation of CO<sub>2</sub> feedback at the 41 000-year cycle into asymmetric, saw-tooth cycles in the ~100 K band.

## 5. How did ice sheets control CO<sub>2</sub>?

What mechanisms are responsible for ice-sheet control of atmospheric CO<sub>2</sub> concentrations with little or no lag? The processes linking the ice and CO<sub>2</sub> must have been directly driven by the ice and must also have been capable of altering atmospheric CO<sub>2</sub> concentrations. They must also have varied primarily at the 41 000-year period prior to 0.9 million years ago, and at both ~100 000 and 41 000 years since that time, with phases near that of both ice volume and CO<sub>2</sub>.

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, 1987; Raymo et al., 1989). Variations in 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<sub>3</sub> dissolution on the Atlantic sea floor. Because southward flow in the deep Atlantic is rapid, these changes would alter the

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mixed-layer chemistry (alkalinity) of the Southern Ocean a few hundred years later when deep Atlantic waters later reach the surface, and thereby affect atmospheric CO<sub>2</sub>. During the northern hemisphere ice-age cycles, the δ<sup>13</sup>C proxy for “NADW” flow (Raymo et al., 1997) varied at 41 000 years until 0.9 million years ago and subsequently within the ~100 000-year band. Both variations were phased with δ<sup>18</sup>O (ice volume).

Other potential ways of altering atmospheric CO<sub>2</sub> may be 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 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 million years 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 are in phase with, or lag only slightly behind, changes in δ<sup>18</sup>O/ice volume (Kohfeld et al., 2005).

A second potential mechanism is increased stratification of surface waters or increases in sea-ice cover that might have limited the release of CO<sub>2</sub> 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<sub>2</sub> fluxes to the atmosphere. These changes occurred at a 41 000-year tempo prior to 0.9 million years ago, and later within the ~100 000 year band (Morley and Dworetsky, 1991).

Glacial increases in southern hemisphere dust fluxes (Ridgeway and Watson, 2002) and in Southern Ocean sea-ice cover (Stephens and Keeling, 2000) and surface-water stratification (Francois et al., 1997; Sigman and Boyle, 2000) have considerable potential to have reduced atmospheric CO<sub>2</sub> levels, but convincing links to the driving changes in northern ice sheets have been difficult to demonstrate. One way to project changes from the north to the south is via greenhouse-gas changes. “First-stage” changes

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in CO<sub>2</sub> tied directly to the northern ice sheets and proximal northern responses could cause “second-stage” changes in distal southern hemisphere dust fluxes and Southern ocean stratification that could further amplify the CO<sub>2</sub> response.

Large uncertainties persist about how large a CO<sub>2</sub> response each of these processes could explain (Sigman and Boyle, 2000). The explanation may well lie in a combination of processes.

## 6. Terminations

The case has been made here that CO<sub>2</sub> feedback was the key non-linearity in the climate system that built large ice sheets during the last 0.9 million years. The question remains why abrupt deglacial terminations, the most obvious feature of large late Pleistocene ice oscillations, occurred within a band centered near 100 000 years.

An initial clue that these deglaciations are linked to modulation of precession came from the discovery that the ~100 000-year component of  $\delta^{18}\text{O}$  is phase-locked to eccentricity (Hays et al., 1976). SPECMAP (Imbrie et al., 1992, 1993) noted that the last several terminations correlate with near-coincident insolation maxima at tilt and precession. The combined insolation effects from these two periods were closely aligned at terminations I and IV, and offset by only 4000 years on termination II. Tilt and precession forcing were not closely aligned during Termination III (Imbrie et al., 1984), and, probably as a result, the deglaciation at that time was incomplete.

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

On the other hand, Huybers and Wunsch (2004) found that terminations occur at or near multiples of the 41 000-year tilt cycle: either after two cycles (~82 000 years) or

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three (~123 000 years). One problem with invoking only tilt 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 (Figs. 5, 8), but less than half of the ice present at that time melted (~30 m of sea-level equivalent).

Both precession and tilt, acting in reasonably close alignment, probably play a role in pacing (determining the timing of) terminations. But insolation changes alone cannot explain some aspects of deglaciations, especially the “the 400 K problem” (Imbrie and Imbrie, 1980). This dilemma is best exemplified by termination V, a very large deglaciation that occurred at a time of particularly weak insolation forcing. A small 65° N insolation maxima at the precession cycle occurred near the beginning of termination V (425 000 years ago), and another at the end (408 000 years ago), but the mid-point of this deglaciation coincided with a weak insolation minimum near 416 000 years ago. The largest mystery about termination V is why ice would have continued to melt through the insolation minimum 416 000 years ago.

### 6.1. Greenhouse-gas roles during terminations

If greenhouse gases primarily act as a fast feedback on ice volume, they could contribute to terminations in three ways. First, many studies have noted that large deglaciations seem connected to the size of the ice sheets, as if the operative mechanism is stored up as a kind of “potential energy” that is unleashed during ice melting. This view is consistent with the observation by Imbrie et al. (1993) that direct insolation forcing explains well under half of the amplitude of terminations and that most of these deglaciations results from processes within the climate system. If CO<sub>2</sub> feedback is the primary internal process that causes non-linear growth of ice sheets during each interglacial-glacial oscillation (Fig. 8), this same feedback will be available to enhance rapid and non-linear melting during times of favorable orbital forcing.

A second contribution from CO<sub>2</sub> feedback arises from the fact that it amplifies the forced ice response at 41 000 years. In so doing, CO<sub>2</sub> feedback in effect adds to the forcing side of the ledger. The net result is the same as if the insolation forcing at

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41 000 years was strengthened (for the models shown in Figs. 7 and 8), by a factor of two or more.

This “boost” to the 41 000-year insolation forcing may explain another enigma in late-Pleistocene  $\delta^{18}\text{O}$ /ice-volume responses. A typical example is the relative strength of the 65° N insolation maxima that drove the ice melting on isotopic substages 5.3 and 5.1 compared to the insolation maximum that occurred on termination I. The two stage 5 maxima were larger than the one on Termination I (Fig. 8), yet they failed to melt the small amount of stage 5 ice that was present, while the weaker insolation maximum 11 000 years ago somehow melted all of a large volume of ice.

A plausible explanation for this mismatch may be found in  $\text{CO}_2$  feedback amplification of the forced response at 41 000 years. In the latter part of stage 5, tilt was out of alignment with the two insolation maxima at the precession cycle, and thus its feedback-amplified contribution to ice melting was unavailable on the substage 5.4/5.3 and 5.2/5.1 transitions. In contrast, the tilt maximum was almost perfectly aligned with the precession maximum on termination 1, and its  $\text{CO}_2$ -amplified contribution at the 41 000-year period added to the effects of precession.

$\text{CO}_2$  feedback also affects the timing of terminations in a third, more indirect, way. Because major (net) ice build-up only occurs every 41 000 years, major deglaciations can only fall after multiples of that period. Given that terminations are defined as times when a large volume of ice melts, a single insolation minimum at the tilt cycle will not allow enough ice build-up to permit a true termination. Two or three intervals of ice growth at the 41 000-year period (that is, either  $\sim 82\,000$  or  $\sim 123\,000$  years) are needed to accumulate enough ice for a termination to occur. This timing constraint, combined with the one imposed by maximum ablation at multiples of the precession cycle (Raymo, 1997), will tend 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 (three tilt cycles and five precession cycles).

Finally, greenhouse-gas changes at the 23 000-year precession cycle add directly to insolation forcing of ice volume. At this cycle, both  $\text{CO}_2$  and  $\text{CH}_4$  have the same “early”

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phase as summer insolation and thus they also force ice volume (Fig. 2b). Because of the modulation of precession by eccentricity, this forcing should be strongest during interglacial isotopic substages, including those that end terminations. CO<sub>2</sub> and CH<sub>4</sub> maxima occurred 11 000 to 10 000 years ago near the end of termination I, at roughly 128 000 years ago near the end of termination II, and at similar stages of the two previous terminations (Petit et al., 1999; Shackleton, 2000; Ruddiman and Raymo, 2003). In all four cases, the greenhouse-gas maxima were approximately coincident with insolation maxima at the precession cycle and thereby provided additional forcing to melt ice.

## 6.2. Greenhouse gases and termination V

These greenhouse-gas roles also add a new perspective to the sequence of events on termination V. First, the fact that an unusually large volume of ice had accumulated in isotopic stage 12 means that an unusually large amount of “stored” feedback (from CO<sub>2</sub> and ice-albedo) was available to accelerate ice melting. Because insolation forcing across this deglaciation was unusually weak, these positive feedbacks would have played a relatively larger role than they did on other terminations.

Second, termination V took almost twice as long as other terminations. Marine isotopic  $\delta^{18}\text{O}$  records from planktic foraminifera (Imbre et al., 1984) and benthic foraminifera (Lisecki and Raymo, 2005) indicate a duration of about 20 000 years, compared to ~10 000 years for terminations I through IV. This long 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. As noted earlier, the key question is why ice melting persisted through the small 65° N insolation minimum at 416 000 years ago.

Greenhouse-gas feedback provides a plausible answer. The insolation minimum 416 000 years ago was produced by precession, which dominates 65° N insolation changes. But the tilt cycle reached a maximum just as precession reached this minimum. If, as concluded above, CO<sub>2</sub> feedback amplified the effect of insolation forcing at

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the tilt cycle, the CO<sub>2</sub>-enhanced effect 416 000 years ago would have outweighed the weak precession minimum. This amplified signal could have bridged the gap between the two precession insolation maxima and provided a continuous source of forcing across the entire termination.

5 A final observation about termination V is that it marks the start of a new ice-age regime. All peak interglaciations since stage 11 have differed from those that preceded in having more negative  $\delta^{18}\text{O}$  values (indicating warmer temperatures or smaller ice volume), and lower CO<sub>2</sub> concentrations (Siegenthaler et al., 2005). Part of the mystery of termination V may be the unique conditions associated with the transition from one  
10 regime to another.

## 7. 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 a linear (“Milankovitch”) ice-volume response at the tilt and precession cycles; and (2) amplification of the forced  
15 41 000-year ice response by CO<sub>2</sub> feedback. The growth of 41 000-year ice sheets in the 41 K world can be explained by CO<sub>2</sub>-feedback amplification of the forced ice response to changes in tilt. After 0.9 million years ago, similar episodes of CO<sub>2</sub>-amplified ice growth continued at 41 000-year intervals, but polar cooling suppressed ice ablation in the intervals that followed. The net result was step-like transitions toward greater ice  
20 volume that created the asymmetric sawtooth-shaped ice oscillations of the ~100 K world. The same positive CO<sub>2</sub> feedback that caused non-linear growth of ice sheets in this new regime was then available to enhance the amplitude of subsequent ice melting during times when insolation forcing became favorable. Although precession dominates insolation cycles at 65° N and helped to pace terminations, amplification  
25 of forced ice-volume changes by CO<sub>2</sub> feedback at the 41 000-year cycle made tilt a comparably important factor in pacing deglacial terminations.

CO<sub>2</sub> feedback also helps explain why the northern and southern hemispheres re-

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spond 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, many climatic responses are driven by a CO<sub>2</sub> signal that is directly controlled by (and in phase with) the northern ice sheets. As a result, most global climatic signals are ice-driven and thus nearly synchronous. An exception is the tropics, where summer insolation forcing produces strong monsoon responses that are largely independent of northern ice.

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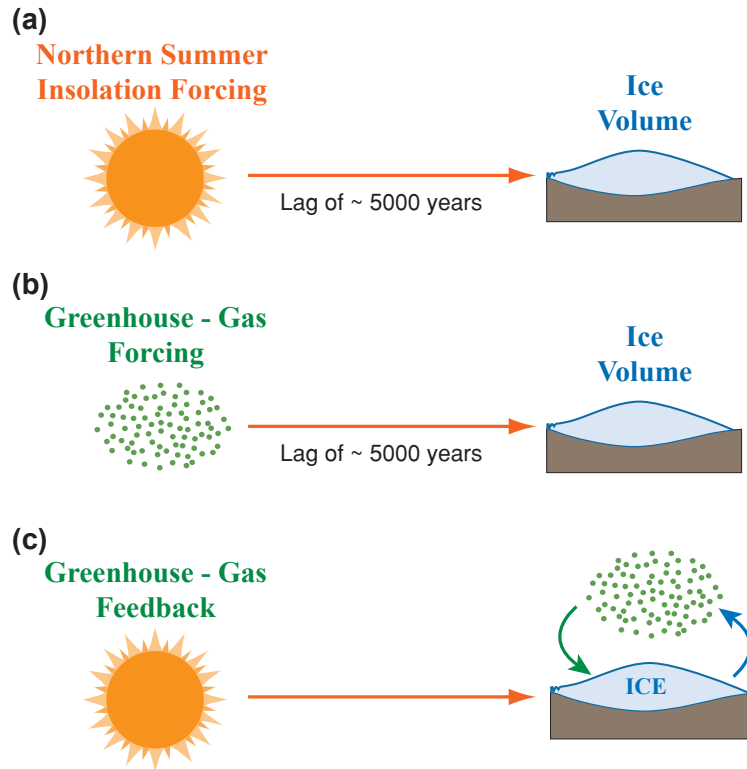
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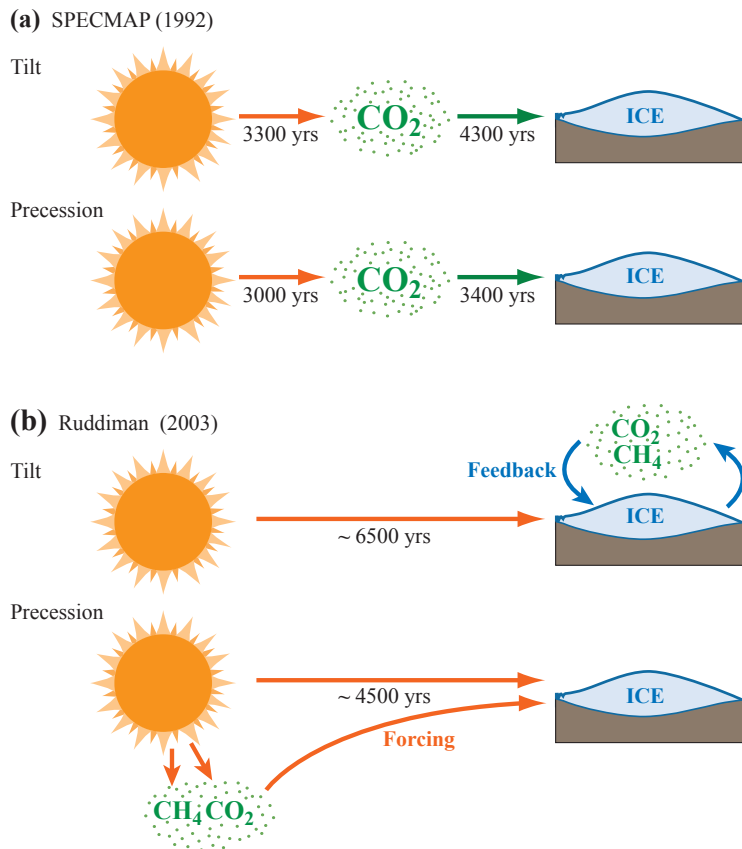
**Fig. 1.** (a) Northern hemisphere summer insolation forces ice sheets with lags of several thousand years (Milankovitch, 1941; Hays et al., 1976). Greenhouse gases could either force ice sheets with the same lag (b), or be driven by ice-sheet variations and provide positive feedback to the ice (c).

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**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<sub>2</sub> 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<sub>2</sub> and CH<sub>4</sub> combine with insolation to force ice volume at 23 000 years, but are ice-driven feedbacks at 41 000 years.

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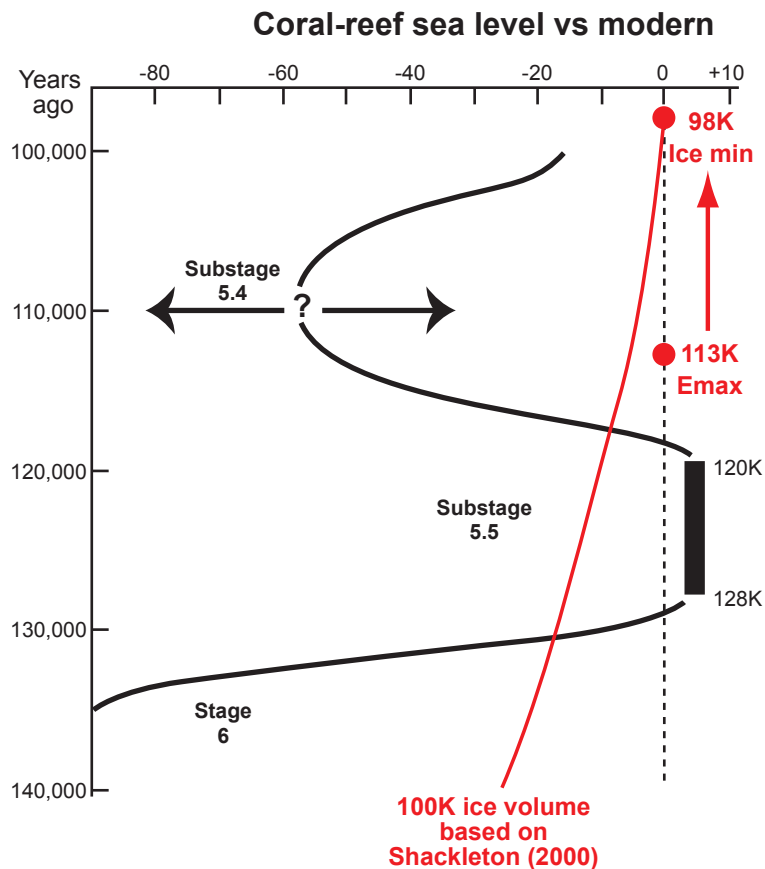
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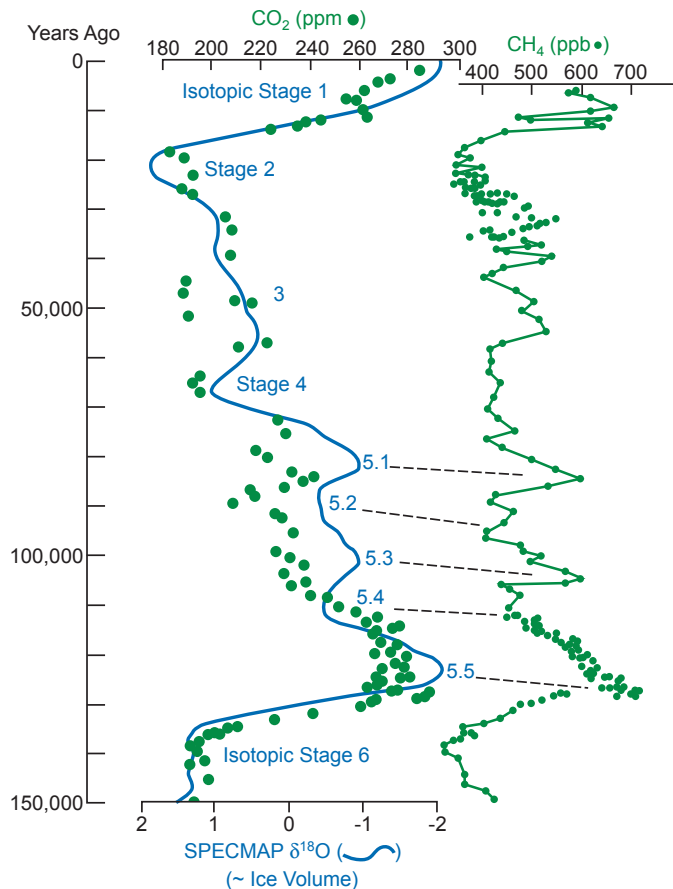
**Fig. 4.** Comparison of sea-level changes during marine isotopic substages 5.5 and 5.4 from coral reefs and  $\delta^{18}\text{O}$  signals (Chappell and Shackleton, 1986; Bard et al., 1990) with filtered 100 000-year ice-volume signal based on Shackleton (2000).

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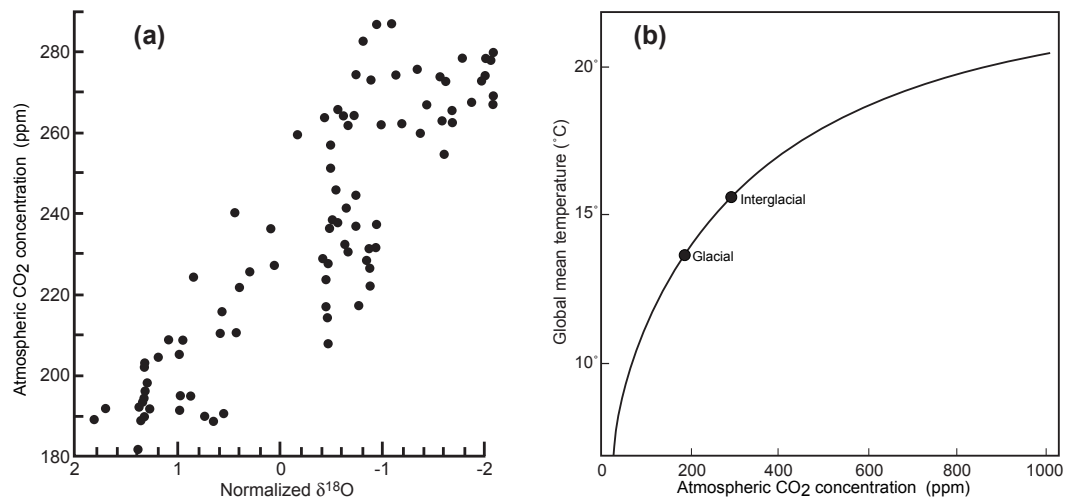


**Fig. 5.** Comparison of normalized SPECMAP  $\delta^{18}\text{O}$  record (Imbrie et al., 1984) against Vostok  $\text{CO}_2$  and  $\text{CH}_4$  signals (Petit et al., 1999).  $\text{CO}_2$  and  $\delta^{18}\text{O}$  changes are closely correlated and nearly synchronous.  $\text{CH}_4$  leads  $\delta^{18}\text{O}$  by several thousand years during isotopic stage 5 (dashed lines).

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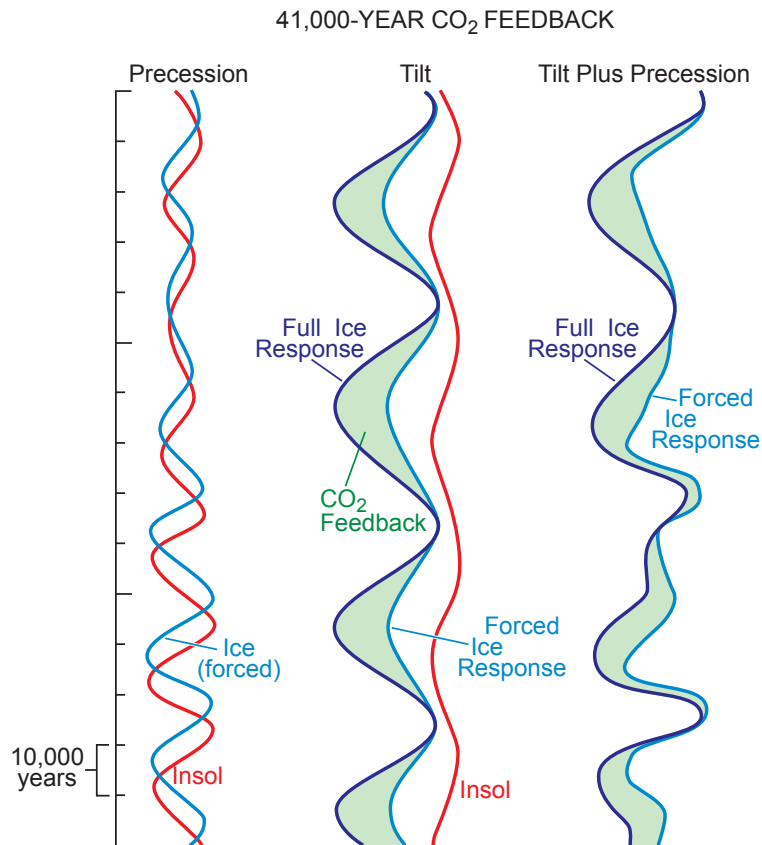
**Fig. 6.** (a) Correlation between SPECMAP  $\delta^{18}\text{O}$  and Vostok CO<sub>2</sub> records shown in Fig. 5. CO<sub>2</sub> analyses projected onto smoothed  $\delta^{18}\text{O}$  trend. (b) Logarithmic relationship between CO<sub>2</sub> concentration and global temperature (Oglesby and Saltzman, 1990).

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**Fig. 7.** Schematic model of CO<sub>2</sub> feedback on ice volume in the 41 K world prior to 0.9 million years ago. Insolation (and greenhouse gases) forced ice volume at the precession cycle. Insolation also forced ice volume at the tilt cycle, but CO<sub>2</sub> feedback amplified the forced ice response at 41 000 years so that the combined ice-volume signal was dominated by tilt. Insolation trends are those of the last 150 000.

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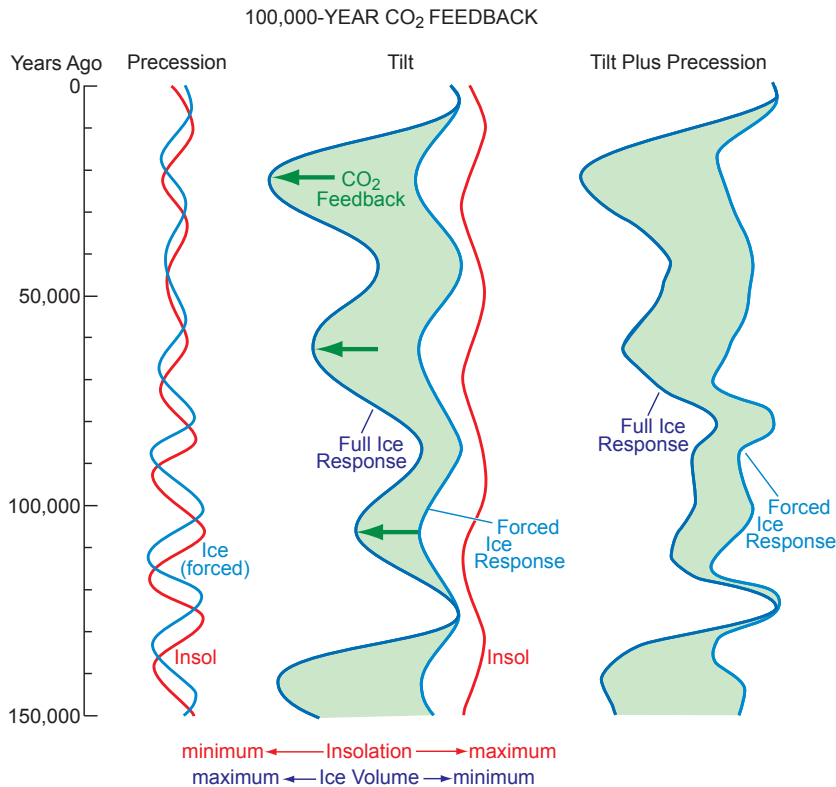
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**Fig. 8.** Schematic model of CO<sub>2</sub> feedback on ice volume in the ~100 K world since 0.9 million years ago. Insolation (and greenhouse gases) forced ice volume at the precession cycle. Insolation also forced ice volume at the tilt cycle, but CO<sub>2</sub> feedback amplified ice growth every 41 000 years (green arrows). Following these ice-growth episodes, weaker ablation in a colder world allowed much of the new ice to survive weak insolation minima, and the net ice-volume increases at 41 000-year intervals were transformed into a longer-period response.

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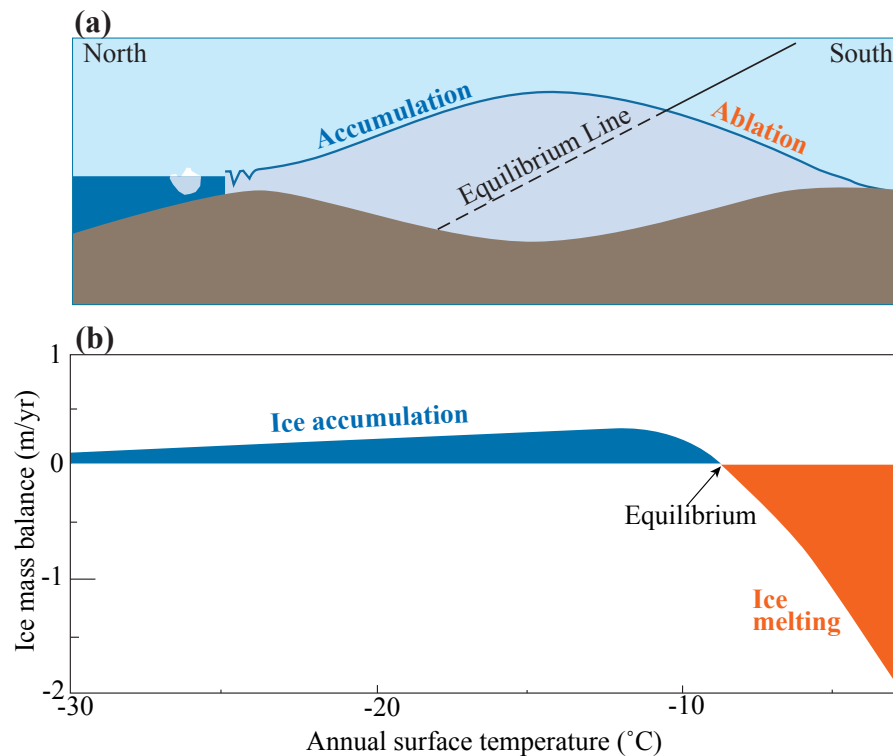
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**Fig. 9.** Geographic and climatic constraints on ice-sheet mass balance. **(a)** Zonal cross section of northern ice sheets. Equilibrium line separates areas of net accumulation and ablation. **(b)** Ice mass balance as a function of mean annual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

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