

Climate change and trace gases

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Palaeoclimate data show that the Earth's climate is remarkably sensitive to global forcings. Positive feedbacks predominate. This allows the entire planet to be whipsawed between climate states. One feedback, the 'albedo flip' property of ice/water, provides a powerful trigger mechanism. A climate forcing that 'flips' the albedo of a sufficient portion of an ice sheet can spark a cataclysm. Inertia of ice sheet and ocean provides only moderate delay to ice sheet disintegration and a burst of added global warming. Recent greenhouse gas (GHG) emissions place the Earth perilously close to dramatic climate change that could run out of our control, with great dangers for humans and other creatures. Carbon dioxide (CO₂) is the largest human-made climate forcing, but other trace constituents are also important. Only intense simultaneous efforts to slow CO₂ emissions and reduce non-CO₂ forcings can keep climate within or near the range of the past million years. The most important of the non-CO₂ forcings is methane (CH₄), as it causes the second largest human-made GHG climate forcing and is the principal cause of increased tropospheric ozone (O₃), which is the third largest GHG forcing. Nitrous oxide (N₂O) should also be a focus of climate mitigation efforts. Black carbon ('black soot') has a high global warming potential (approx. 2000, 500 and 200 for 20, 100 and 500 years, respectively) and deserves greater attention. Some forcings are especially effective at high latitudes, so concerted efforts to reduce their emissions could preserve Arctic ice, while also having major benefits for human health, agricultural productivity and the global environment.

Keywords: climate change; trace gases; climate feedbacks; black carbon; sea level; Arctic

1. Introduction

Trace atmospheric gases have played a leading role in climate change throughout Earth's history. Thus, empirical data on trace gas histories and climate change provide invaluable information on climate sensitivity. The Earth's climate

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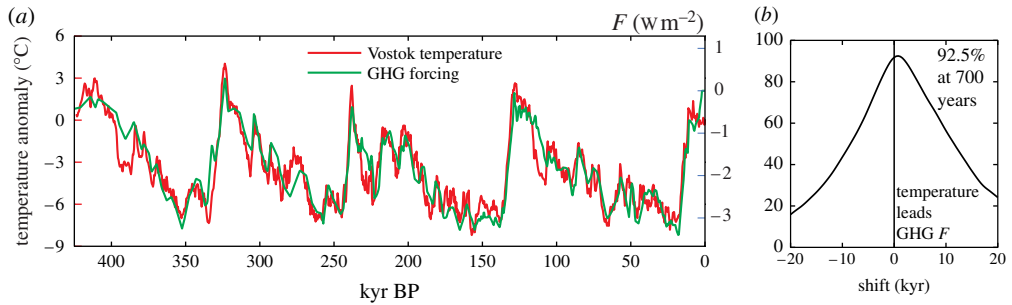


Figure 1. (a) Greenhouse gas forcing and Vostok temperature: Antarctic temperature (left scale) from Vostok ice core (Vimeux *et al.* 2002) and global climate forcing (right scale) due to CO_2 , CH_4 and N_2O . Forcing zero point is for 1850 gas amounts specified by Hansen & Sato (2004). Temperature zero point is ‘present’ value from the Vostok ice core (Vimeux *et al.* 2002). Ratio of temperature and forcing scales ($3.02^\circ\text{C} (\text{W m}^{-2})^{-1}$) is chosen such that their standard deviations in this figure are equal. (b) Correlation (%) diagram showing lead of temperature over GHG forcing in years.

history also provides our best indication of the level of global warming that would constitute ‘dangerous interference’ with climate. The empirical data, abetted by appropriate calculations, imply that control of trace gases must play a critical role in preserving a planet resembling the one on which civilization developed.

2. Climate sensitivity

Our emphasis is on planet Earth as a whole. We must pay attention to dynamical reorganizations of ocean and atmosphere circulation, which can have global effects and cause large regional change. Such reorganizations also make it difficult to assess global change from measurements at a small number of places. Yet global climate forcings evoke a clear global response, which may be of paramount importance.

(a) Antarctic data

Records of climate change over the past several hundred thousand years carry a rich bounty of information about climate sensitivity. Here we use Antarctic temperature data of Vimeux *et al.* (2002) derived from an ice core extracted near Vostok (Petit *et al.* 1999), approximately 1000 km from the South Pole. Although a longer Antarctic record has been obtained (EPICA 2004), the Vimeux *et al.* (2002) temperatures are corrected for climate variation in the water vapour source regions and the record length is sufficient to match the sea-level data of Siddall *et al.* (2003).

The red curve in figure 1 is Antarctic temperature based on the Vostok ice core, time running left to right. The Holocene is the current warm (‘interglacial’) period, now almost 12 000 years in duration. This climate record reveals repeated irregular cooling over periods of *ca* 100 000 years, terminated by rapid warmings of approximately 10°C in Antarctica. The largest temperature swings occurred almost synchronously throughout the planet. The amplitude of these

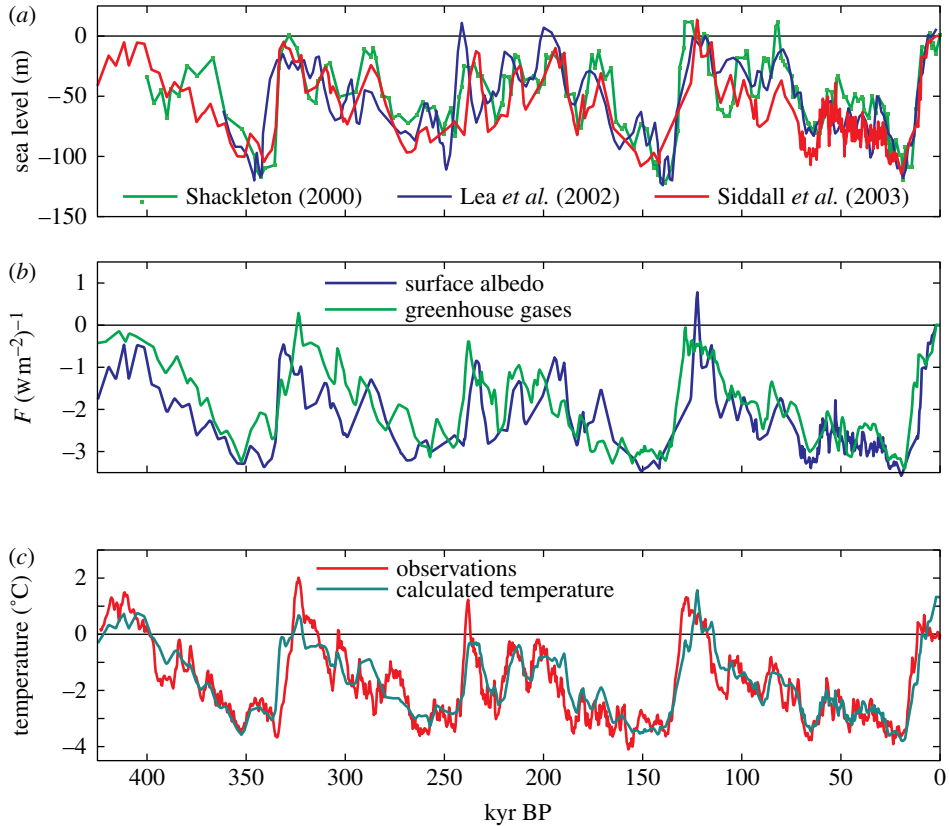


Figure 2. (a) Sea-level records of Shackleton (2000; green), Lea *et al.* (2002; blue) and Siddall *et al.* (2003; red), (b) climate forcings due to GHGs (CO_2 , CH_4 and N_2O) and surface albedo and (c) palaeoclimate temperature change: calculated and observed temperature. Calculated temperature is the product of forcing (b) and climate sensitivity ($3/4^\circ\text{C} (\text{W m}^{-2})^{-1}$). Observed temperature is Vostok temperature (figure 1) divided by 2.

temperature swings is typically $3\text{--}4^\circ\text{C}$ in tropical ocean regions (as revealed by the Mg/Ca composition of microscopic creatures that lived near the ocean surface, whose shells are preserved in ocean sediments; Lea *et al.* 2000), approximately 5°C on global average and 10°C near the poles.

The same Vostok ice core that defines past Antarctic temperature also reveals the history of long-lived atmospheric gases. Bubbles of air are trapped as annual snowfalls pile up and compress gradually into ice. The Vostok records (Petit *et al.* 1999) of the two principal greenhouse gases (GHGs), CO_2 and CH_4 (methane), have been shown many times and are not repeated here. The record of the third major long-lived GHG, N_2O , is not preserved as well owing to reactions with organic matter in dust particles that are also trapped in the ice. However, the amplitude of the glacial–interglacial N_2O change is established from instances when dust amount was small (Spahni *et al.* 2005). Since the N_2O climate forcing is a small fraction of the total GHG forcing, and because N_2O time variations, where available, are similar to those of CO_2 and CH_4 , it is possible to reconstruct accurately the climate forcing caused by the sum of all three long-lived GHGs.

A word of explanation about climate forcings is needed. A forcing is an imposed change of the planet's energy balance with space. The most common technical measure of forcing (Hansen *et al.* 1997; IPCC 2001) is the **adjusted forcing (Fa)**. Fa is the imbalance, caused by the forcing agent, between solar energy absorbed by the planet and thermal emission to space, measured after stratospheric temperature adjusts to presence of the agent. However, it is useful to account for the fact that some forcing agents have greater 'efficacy' than others for changing global temperature, especially when indirect effects of the forcing agent are included (Hansen *et al.* 2005a). Thus, **CH₄** has efficacy of approximately **1.4**, i.e. it causes 40% more temperature change than does a CO₂ forcing of the same magnitude, primarily because increased CH₄ causes an increase in tropospheric O₃ and stratospheric H₂O.

The **effective forcing (Fe)** due to all three long-lived GHGs, shown in figure 1, is

$$Fe = 1.15[Fa(CO_2) + 1.4Fa(CH_4)], \quad (2.1)$$

where Fa for CO₂ and CH₄ is obtained from analytic expressions of Hansen *et al.* (2000). The factor 1.4 accounts for the efficacy of CH₄ and the factor **1.15** accounts approximately for forcing by **N₂O**, as the glacial–interglacial N₂O forcing is approximately **15%** of the sum of CO₂ and CH₄ glacial–interglacial forcings (Hansen *et al.* 2005a; Spahni *et al.* 2005).

Figure 1a reveals remarkable correspondence of Vostok temperature and global GHG climate forcing. The **temperature change appears to usually lead the gas changes by typically several hundred years**, as discussed below and indicated in figure 1b. This suggests that warming climate causes a net release of these GHGs by the ocean, soils and biosphere. GHGs are thus a powerful amplifier of climate change, comparable to the surface **albedo feedback**, as quantified below. The GHGs, because they change almost simultaneously with the climate, are a major 'cause' of glacial-to-interglacial climate change, as shown below, even if, as seems likely, they slightly lag the climate change and thus are not the initial instigator of change.

The temperature–GHG lag is imprecise because the time required for snow to pile high enough (approx. 100 m) to seal off air bubbles is typically a few thousand years in central Antarctica. The estimated age difference between ice and its air bubbles is accounted for in the time-scale of figure 1, which refers to the ice age. Despite multiple careful studies, uncertainties in the ice–gas age differences for the Vostok ice core remain of the order of 1 kyr (Bender *et al.* 2006). Therefore, we can only say with certainty that the temperature and gas changes are nearly synchronous. Data from a different Antarctic (Dome C) ice core with slightly higher snow accumulation rate (Monnin *et al.* 2001) and an independent analysis based on argon isotopes (Caillon *et al.* 2003) support **temperature leading GHGs by ca 600–800 years**. In addition, carbon cycle models yield increases of GHGs in response to warming oceans and receding ice sheets. Ice cores from Maud Land (EPICA 2006), which has very high snow deposition rates, should establish leads and lags accurately, but the present paper has only slight dependence on that result.

(b) Ice sheet and sea-level change

Earth's energy balance is affected by changes on the planetary surface, as well as in the atmosphere. The important surface change is the **albedo** (reflectivity)

for solar radiation. Surface albedo changes as areas of ice, vegetation and exposed land change. Maps of these quantities have been reconstructed in detail for the last ice age (CLIMAP 1981), which peaked *ca* 20 000 years ago. The greatest albedo change, compared to the present interglacial period, was due to the large Laurentide ice sheet that covered Canada and reached into the US.

Hansen *et al.* (1993) calculated the ice age forcing due to surface albedo change to be $3.5 \pm 1 \text{ W m}^{-2}$. The total surface and atmospheric forcings led Hansen *et al.* (1993) to infer an equilibrium global climate sensitivity of $3 \pm 1^\circ\text{C}$ for doubled CO_2 forcing, equivalent to $3/4 \pm 1/4^\circ\text{C W}^{-1} \text{ m}^{-2}$. This empirical climate sensitivity corresponds to the Charney (1979) definition of climate sensitivity, in which ‘fast feedback’ processes are allowed to operate, but long-lived atmospheric gases, ice sheet area, land area and vegetation cover are fixed forcings. Fast feedbacks include changes of water vapour, clouds, climate-driven aerosols¹, sea ice and snow cover. This empirical result for the ‘Charney’ climate sensitivity agrees well with that obtained by climate models (IPCC 2001). However, the empirical ‘error bar’ is smaller and, unlike the model result, the empirical climate sensitivity certainly incorporates all processes operating in the real world.

The empirical climate sensitivity based on the last ice age can be tested for longer periods using sea-level data. Figure 2a shows general agreement among several records on the magnitude of glacial–interglacial change. For illustrative purposes, in figure 2b,c we use the Siddall *et al.* (2003) record, which has the highest temporal resolution. The impact of differences among the three records on results in figure 2b,c is readily envisaged, as effects are linear. We cannot rely on timing of sea-level changes to better than several thousand years because it includes ‘orbital tuning’, i.e. slight time-scale adjustments to make major features line up with Earth’s orbital changes. Thus, although relative timing of GHG and Antarctic temperatures, from the same ice core, are good within *ca* 1000 years or less, dating inconsistency of sea-level change with respect to these other two quantities is as much as several thousand years. Temporal resolution in the sea-level data is also coarser than in the ice core data.

Sea-level change yields an estimate of ice sheet area change. As ice sheets grow they become thicker, as well as larger in both horizontal dimensions. Thus, we take ice sheet horizontal area as proportional to the two-third power of the amount of water locked in the ice sheet. Normalization of surface albedo forcing is 3.5 W m^{-2} (equivalent to approximately 1.5% reduction of solar irradiance, as the Earth absorbs approximately 240 W m^{-2} of solar energy) at the time of the

¹We include climate-driven aerosol changes and their cloud effects as a ‘fast feedback’ because aerosols respond rapidly to climate change. This choice yields a more precise empirical climate sensitivity because aerosol forcing depends sensitively on uncertain aerosol absorption. Our inferred climate sensitivity, 3°C for doubled CO_2 , is the same as estimated by Hansen *et al.* (1993), who did not classify aerosols as a fast feedback, because our present omission of the small net aerosol forcing is compensated by larger effective GHG forcings, especially the high efficacy (140%) of CH_4 . Ice core data show that aerosols decrease as the climate warms, probably because increased water vapour and rainfall wash out aerosols. Aerosol amount in the Earth’s atmosphere seems to have decreased in the past two decades (Mishchenko *et al.* 2007), while human-made aerosol sources were believed to be increasing. We suggest that the aerosol decrease may be due to rapid global warming, approximately 0.2°C per decade (Hansen *et al.* 2006a), and resulting moistening of the atmosphere.

last ice age, when sea level was approximately 110 m lower than today. Smaller albedo effects due to continental shelf exposure and vegetation migration are included within this empirical evaluation. The resulting surface albedo climate forcing is shown, along with the GHG forcing, in figure 2b.

When the surface albedo and GHG forcings of figure 2b are added and multiplied by the climate sensitivity ($3/4^{\circ}\text{C} (\text{W m}^{-2})^{-1}$), the calculated temperature shown by the blue curve in figure 2c is obtained. This calculated temperature is compared to the Vostok temperature change divided by 2, which we take as an approximation of global temperature change². The remarkable coincidence of calculated and observed temperatures cannot be accidental. The close agreement has dramatic implications for interpretation of past climate change and for expectation of future climate change due to human-made climate forcings.

(c) Causes of palaeoclimate fluctuations

Figure 2 shows that, with surface albedo and long-lived GHG amounts specified, the magnitude of Pleistocene climate variations is accounted for by fast feedback processes (climate-driven changes of water vapour, aerosols, clouds, sea ice and snow)³. However, implications of the large palaeoclimate swings in figure 2 reach far beyond confirmation that the Charney (fast feedback) climate sensitivity is approximately 3°C for doubled CO_2 .

² Antarctic temperature change divided by 2 serves as a crude 'global thermometer' for large global climate change on time-scales of several thousand years or longer. Limitations of a local thermometer are obvious on time-scales of 1–2 kyr or less, when Antarctic and Greenland temperature fluctuations are often on a 'see-saw', i.e. out of phase (EPICA 2006). Leads and lags of temperature changes at different locations are crucial for understanding the mechanisms of climate change, and these short-term variations can involve complex dynamical processes, including possible 'reorganizations' of ocean and atmospheric circulation. However, global temperature changes must be coherent in the two hemispheres for any climate forcings large enough to change tropical ocean temperature, because the tropics, via ocean and atmosphere, export heat to both hemispheres. Indeed, a coherent global response occurs even for forcings predominately located in one hemisphere, such as anthropogenic aerosols or change of ice sheet area, although the response is larger in the hemisphere with greater forcing (Hansen et al. 2005a).

³ The last glacial cycle, the most accurately dated, has two notable discrepancies between observed and calculated temperature. The first calculation discrepancy is failure to obtain a deep minimum temperature at Marine Isotope Stage (MIS) 5d (ca 110 kyr BP). We suggest that the sea-level curve of Siddall et al. (2003) may understate sea-level fall at that time. Several other sea-level records, summarized in fig. 6 of Potter & Lambeck (2003), show sea level much lower in MIS 5d than in MIS 5c (ca 105 kyr BP). The second discrepancy occurs in the last 8000 years, with calculated temperature rising rapidly while observed temperature fell. Calculated warming is due to increase of CO_2 from approximately 260 to 275 ppm and CH_4 from approximately 600 to 675 ppb. Ruddiman (2003) suggests that the GHG increases are due to deforestation beginning ca 8 kyr BP and rice agriculture beginning ca 5 kyr BP. Indeed, much of the observed GHG increase is plausibly anthropogenic, but we would expect early negative anthropogenic forcings from the same agricultural and deforestation activities, due to aerosols and surface albedo change (not included in figure 2), to exceed positive GHG forcings. The aerosol forcing, especially indirect effects on clouds, is strongly nonlinear, with human-made aerosols in a nearly pristine atmosphere being much more effective than those added to the present atmosphere. Thus, although Ruddiman's basic thesis is probably correct, his conclusion that humans saved the Earth from an ice age is probably not right.

Surface albedo and GHG amounts are themselves feedbacks that respond to climate change, implying that actual climate sensitivity is much greater than that due to fast feedbacks. Realization that climate sensitivity is larger on longer time-scales is not new, but larger sensitivities are usually thought to apply to millennial time-scales. We will argue that ‘slow’ feedbacks (ice sheet, vegetation and GHG) substantially influence century, and perhaps shorter time-scales.

Empirical analysis depends upon accurate knowledge of time-dependent forcings. One forcing mechanism is well known (Berger 1978): changes in the seasonal distribution of solar radiation impinging on the planet due to slow Earth’s orbital changes (inclination of the spin axis, eccentricity of the orbit and season of closest approach to the Sun, i.e. precession of the equinoxes). The global mean forcing due to orbital variations is small: with a fixed surface albedo distribution, the maximum global mean forcing due to orbital variations is approximately 0.25 W m^{-2} . This small forcing leaves an easily discernable impact on the spectrum of climate variability (Hays *et al.* 1976), even though a greater portion of variability has the character of red noise (Wunsch 2003). It appears that global climate is remarkably sensitive to even small forcings, and thus also to unforced climate fluctuations (chaos).

Timing of insolation changes is known with great precision. Unfortunately, dating of past climate change is often influenced by orbital data (orbital tuning) using preconceived ideas about orbital effects on climate. This limits the degree to which the climate records in figures 1 and 2 can be used to infer mechanisms of climate change.

Analysis must begin with the predominant feature, the asymmetry of the ice ages, defined by global warmings that terminated the major ice ages. The warmings at *ca* 15, 130, 240 and 330 kyr BP are named Terminations I, II, III and IV, respectively. Such huge rapid climate change had to involve large positive feedbacks. Indeed, as figure 2 shows, those feedbacks were surface albedo and GHGs. What we want to know is how those feedbacks worked.

Note first that ‘minor’ mismatches in timing of observed and calculated temperatures in figure 2c are due to dating errors and, to a lesser degree, limitations of a local thermometer¹. Proof is obtained by considering the contrary: ice sheet forcing approximately 3 W m^{-2} and a 5 kyr timing gap between forcing and response, as appears to be the case at Termination IV (figure 2c), is $15\,000 \text{ W yr m}^{-2}$, enough to warm the upper kilometre of the ocean by approximately 160°C (see table S1 in Hansen *et al.* (2005b)). Obviously, no such warming occurred, nor did warming more than approximately 1/100th of that amount. Forcing and temperature change had to be synchronous within a few centuries, at most, for the large global climate change at terminations.

Rapid warming at terminations, we assert, must be due to the fact that ice sheet disintegration is a wet process that, spurred by multiple thermodynamical and dynamical feedback processes (Hansen 2005), can proceed rapidly. Chief among these feedbacks is the large change in absorbed solar energy that occurs with the ‘albedo flip’ when snow and ice become wet. This process determines the season at which insolation anomalies are most important.

The Milankovitch (1941) theory of the ice ages assumes that summer insolation anomalies at high latitudes in the Northern Hemisphere (NH) drive the ice ages: minimum summer insolation allows snow and ice accumulated in the cold season to survive, while maximum summer insolation tends to melt the ice sheets.

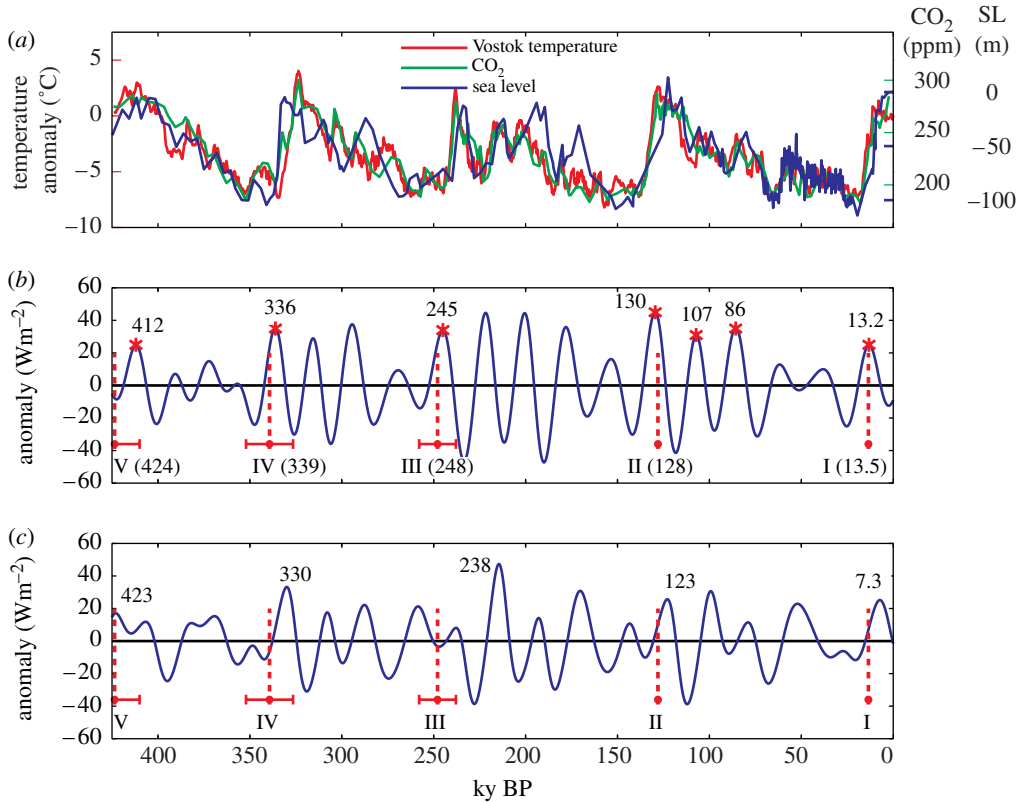


Figure 3. (a) Temperature (Vimeux *et al.* 2002), CO₂ (Petit *et al.* 1999) and sea level (SL; Siddall *et al.* 2003), (b) late spring (April–May–June) insolation at 60° N and (c) late spring (October–November–December) insolation at 75° S.

We suggest, however, that **spring is the critical season for terminations**, because the albedo feedback works via the **large change in absorbed sunlight that begins once the ice/snow surface becomes wet**, after which the surface albedo remains low until thick fresh snow accumulates. **A spring maximum of insolation anomaly pushes the first melt earlier in the year**, without comparable shortening of autumn melt, thus **abetting ice sheet disintegration**. **And an increase of GHGs stretches the melt season both earlier and later**, while also **increasing midsummer melt**. Thus, it is **not surprising that Terminations I, II, III and IV all had strong maxima in GHG forcing**, as well as, we presume, favourable insolation.

Let us test the **‘spring melt’** proposition and examine consequences. Figure 3a shows sea level, CO₂ and Antarctic temperature, while figure 3b,c shows insolation anomalies for **late spring at 60° N** and late spring at 75° S, respectively. Only the insolation curves are dated precisely; others include orbital tuning to summer or June insolation. Estimated termination dates are from Raymo (1997).

The sea-level quantity most important to our discussion, and to society, is the **rate of sea-level change** (Roe 2006). We expect the rate of sea-level **rise** to be maximum when **NH spring insolation peaks**. This expectation can be checked and refined via accurately dated Termination I. Temperature increase of **Termination I occurred between 18 kyr BP and the Younger Dryas–Preboreal**

Table 1. Dates (kyr BP) of maximum insolation at 60° N and 75° S. (MAM, March–April–May; AMJ, April–May–June; JJA, June–July–August; OND, October–November–December.)

Northern Hemisphere			Southern Hemisphere		
MAM	AMJ	JJA	OND	terminations (Raymo 1997)	
14.5	13.2	8.5	7.3	I	13.5
38.7	38.1	30.8	27.3		
63.8	61.8	54.3	52.1		
86.8	85.7	80.4	79		
108.7	107.2	101.5	99.3		
130.8	129.5	124.4	122.9	II	128.0
155.0	153.5	146.3	143.4		
179.6	178.1	172.2	170.3		
201.7	200.5	195.0	192.9		
223.3	221.7	216.5	214.5		
246.1	245.0	239.6	237.9	III	247.9 ± 10.0
271.6	269.3	260.9	258.5		
295.7	294.4	289.1	287.7		
316.9	315.6	310.2	307.9		
337.4	336.1	331.4	329.9	IV	339.3 ± 12.8
357.7	356.8	351.1	348.5		
373.1	372.2	369.4	369.2		
392.1	390.6	384.4	381.8		
412.9	412.0	407.5	406.7		
431.5	430.0	424.7	423.0	V	423.6 ± 13.6
449.3	448.7	445.3	445.6		
467.8	466.4	461.3	459.6		
488.4	487.2	482.1	480.3		
509.1	507.7	502.5	500.6		
533.7	532.6	526.3	525.2	VI?	534.5 ± 23.6
560.7	559.4	552.8	549.1		
582.1	580.7	575.5	573.9		
603.3	602.0	596.7	594.4		
624.5	623.2	618.2	616.5	VII	621.6 ± 20.6

transition at 11.7 kyr BP (EPICA 2006), dated by the new Greenland ice core chronology, GICC05 (Rasmussen *et al.* 2006). The most rapid sea-level rise during Termination I, when sea level rose on average 3–5 m per century for several centuries (Fairbanks 1989), is called Meltwater Pulse 1a (MWP 1a). Stanford *et al.* (2006) place MWP 1a at 14.15–13.7 kyr BP, with almost half of the glacial–interglacial sea-level rise complete by 13 kyr BP.

Figure 3b shows that maximum summer (June–July–August; JJA) insolation occurred at 8.5 kyr BP. Maximum spring insolation was at 14.5 kyr BP. Using Termination I for empirical guidance, April–May–June (AMJ) insolation, peaking at 13.2 kyr BP (table 1), provides optimum fit to peak ice sheet disintegration and sea-level rise. AMJ (‘late spring’) also is optimum from the albedo flip perspective: insolation at the latitude of ice sheets is changing most rapidly at the spring equinox, but mean March insolation is weak, so surface

melt is unlikely. Since half-width of positive insolation anomalies is *ca* 6 kyr, the late-spring albedo-flip forcing implies that the earliest positive forcing of ice sheets (i.e. towards melting) would be expected at *ca* 18 kyr BP.

Confirmation of this interpretation of terminations requires additional accurately dated cases. Termination II, long an enigma owing to suggestions that the climate change preceded presumed orbital forcing, provides a stern test. Figure 3 and table 1 show summer, late spring (AMJ) and spring insolation peaking at 125, 129.5 and 131 kyr BP, respectively. Radiometric dating of a marine oxygen isotope record at one Bahamian site yields an age 135 ± 2.5 kyr BP for Termination II (Henderson & Slowey 2000). A high resolution study of sediments in the Santa Barbara Basin (Cannariato & Kennett 2005), via correlation with a ^{230}Th -dated stalagmite in China (Yuan *et al.* 2004), places the strongest warming rate at *ca* 131 kyr BP and the beginning of Termination II at *ca* 134 kyr BP. ^{230}Th dating of high resolution stalagmite data from southern Europe (Drysdale *et al.* 2005) suggests that deglaciation was essentially complete by 129 ± 1 kyr BP. Although better definition and dating of Termination II is needed, available data are inconsistent with summer forcing of termination. Late spring forcing, considering the 6 kyr half-width of the insolation anomaly, is reasonably consistent with available data.

Table 1 provides timing of insolation maxima at 60°N and 75°S , which can be compared with sea-level records. The albedo-flip mechanism for ice sheet disintegration should work in the Southern Hemisphere late spring (October–November–December), as well as in the Northern Hemisphere. Resulting sea-level high stands due to Antarctic shrinkage would be less than those produced by the Laurentide ice sheet, but they might account for some sea-level anomalies described by Thompson & Goldstein (2005) as ‘sub-orbital’, which are more frequent than Northern Hemisphere insolation anomalies.

Note that terminology for seasons varies in the palaeoclimate literature. We use ‘summer’ for meteorological summer, JJA in the Northern Hemisphere, the season of highest temperature at middle latitudes. Some others, e.g. Huybers (2006) and Roe (2006), take the summer solstice, approximately 21 June, as the midpoint of summer. The midpoint of late spring, 16 May, is just over five weeks earlier than 21 June, so we do not expect fundamental inconsistencies between our conclusion that late spring insolation drives ice sheet decay and the analyses of Huybers (2006) and Roe (2006).

Precise dating is needed for additional terminations and sea-level high stands that might be associated with other insolation peaks (figure 3). However, available data for the two terminations with near-absolute dating do not provide evidence for multi-millennial lag between insolation forcing and ice sheet response. If our interpretation of near synchronicity of forcing and ice sheet response is correct, implications for humanity are profound.

(d) Implications of albedo-flip mechanism

Our primary interest in palaeoclimate rapid global warmings is their implication for twenty-first century climate change. For this purpose, additional comments about palaeoclimate are appropriate.

The high fast-feedback climate sensitivity (approx. 3°C for doubled CO_2) implies that moderate additional positive feedback can produce large climate change, because climate ‘gain’ is already not far from unity (Hansen *et al.* 1984).

Thick ice sheets provide not only a positive feedback, but also the potential for cataclysmic collapse, and thus an explanation for the asymmetry of the ice ages. The albedo flip property of ice/water provides a trigger mechanism. If the trigger mechanism is engaged long enough, multiple dynamical feedbacks will cause ice sheet collapse (Hansen 2005). We argue that the required persistence for this trigger mechanism is at most a century, probably less. Global warming necessarily accompanies ice sheet loss and decreased surface albedo. Global warming, based on both palaeoclimate data and carbon cycle models, is accompanied by increased GHGs. The result is large global warming at terminations.

What determines the magnitude of ice melt and thus associated global warming? Ice sheet albedo change is not a 'runaway' feedback. Continual unforced (chaotic) climate variability initiates ice loss well before global climate gain reaches unity. The magnitude of global warming after melting is initiated, whether by insolation anomaly or otherwise, is limited by ice sheet size. Thus, a colder climate with larger ice sheets should have the possibility of a greater sudden warming. Data for the past several million years (Lisiecki & Raymo 2005), during which the planet has been cooling, confirm this characteristic. Any given warming depends upon details, including the degree to which GHG positive feedback is brought into play. Chaotic behaviour is expected and abundant, but so too is increased amplitude of terminations for cooler climates.

Climate has been unusually stable during the warm Holocene. This may be, at least in part, because the planet is warm enough for the Laurentide and Fennoscandian ice sheets to be absent, but not warm enough for much reduction in size of Antarctica or Greenland. The question is how much human-made climate forcing is needed to cause the albedo-flip mechanism on West Antarctica and/or Greenland on a scale large enough to initiate multiple feedbacks and nonlinear ice sheet collapse? Our best guide, again, may be palaeoclimate data, along with evidence of current ice sheet change.

3. Dangerous climate change

Emergence of human-caused global warming raises the question: what level of further warming will be 'dangerous' for humanity? As discussed elsewhere (Hansen *et al.* 2006a,b), it may be useful in considering this issue to contrast today's climate with the warmest interglacial periods and with the middle Pliocene, when global temperature was 2–3°C warmer than today.

Antarctic temperature was a few degrees warmer in the warmest interglacial periods, but temperature there is magnified by high latitude feedbacks and dependent upon the altitude of the ice surface. The tropical Pacific and Indian Oceans are especially relevant: the Pacific is a driver of global climate, and the Indian Ocean has the highest correlation with global temperature in the period of instrumental data (Hansen *et al.* 2006a). Figure 4 compares instrumental temperatures and palaeo-proxy temperatures in those two regions.

There is an uncertainty of approximately 1°C in the calibration of palaeo-proxy temperature with modern data. However, ocean surface temperature at the beginning of modern measurements (late nineteenth century) must have been within the Holocene temperature range, so the error in matching up the two scales in figure 4 should not exceed several tenths of a degree Celsius.

5-2 millones de años atrás

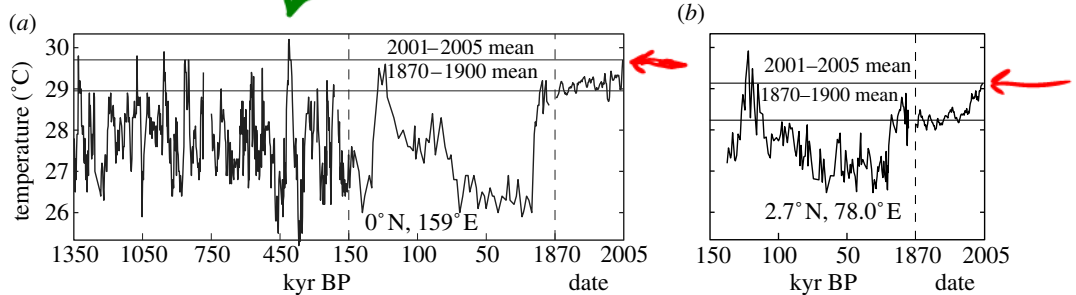


Figure 4. (a) Western Equatorial Pacific (Medina-Elizade & Lea 2005) and (b) Indian Ocean (Saraswat *et al.* 2005) sea surface temperatures (SSTs) based on palaeoclimate proxy data and modern surface temperature measurements, as described in the text.

We conclude that the warming of the past several decades has brought today's temperature to or near the Holocene maximum and within approximately 1°C of the warmest interglacial periods.

Sea level following Termination II may have reached 4 ± 2 m higher than today (Overpeck *et al.* 2006), which would already qualify as dangerous change. It is possible, but uncertain, that such a sea-level rise would occur with additional warming less than 1°C today. But what is clear is that global warming to the level of the middle Pliocene, when sea level was 25 ± 10 m higher, would be exceedingly dangerous.

Global warming of approximately 3°C is predicted by practically all climate models for 'business-as-usual' (BAU) growth of GHGs (IPCC 2001, 2007). Yet IPCC (2001, 2007) foresees twenty-first century sea-level rise of only a fraction of a metre with BAU global warming. Their analysis assumes an inertia for ice sheets that, we argue, is incompatible with palaeoclimate data and inconsistent with observations of current ice sheet behaviour.

BAU global warming (approx. 3°C) would be magnified on the ice sheets, based on general high latitude amplifications found in palaeo records and in climate models, as well as local ice sheet warming due to albedo flip. As a result, large portions of West Antarctica and Greenland would be bathed in melt water. Already areas of summer melt have increased rapidly on Greenland (Steffen *et al.* 2004), the melt season is beginning earlier and lasting longer, and summer melt is being observed on parts of West Antarctica.

There is little doubt that projected warmings under BAU would initiate albedo-flip changes as great as those that occurred at earlier times in the Earth's history. The West Antarctic ice sheet today is at least as vulnerable as any of the earlier ice sheets. The processes that give rise to nonlinear ice sheet response (almost universal retreat of ice shelves buttressing the West Antarctic ice sheet and portions of Greenland, increased surface melt and basal lubrication, speed-up of the flux of icebergs from ice streams to the ocean, ice sheet thinning and thus lowering of its surface in the critical coastal regions, and an increase in the number of 'icequakes' that signify lurching motions by portions of the ice sheets) are observed to be increasing (see §8).

Despite these early warnings about likely future nonlinear rapid response, IPCC continues, at least implicitly, to assume a linear response to BAU forcings. Yet BAU forcings exceed by far any forcings in recent palaeoclimate history.

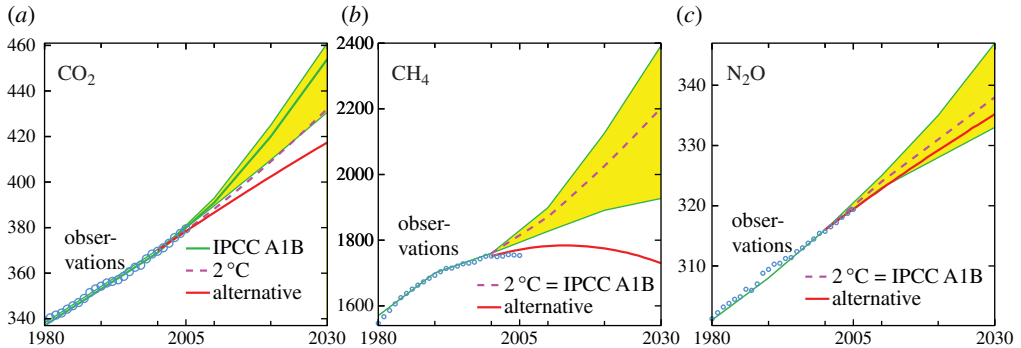


Figure 5. Observed amounts and twenty-first century scenarios for the three principal GHGs. Well-mixed greenhouse gas mixing ratios: (a) CO₂ (ppm), (b) CH₄ (ppb) and (c) N₂O (ppb). Coloured area denotes extreme IPCC (2001) scenarios. Alternative and 2°C scenarios are from Hansen & Sato (2004).

Part of the explanation for the inconsistency between palaeoclimate data and IPCC projections lies in the fact that existing ice sheet models are missing realistic (if any) representation of the physics of ice streams and icequakes, processes that are needed to obtain realistic nonlinear behaviour. In the absence of realistic models, it is better to rely on information from the Earth's history.

That history reveals large changes of sea level on century and shorter time-scales. All, or at least most, of glacial-to-interglacial sea-level rise is completed during the *ca* 6 kyr quarter cycle of increasing insolation forcing as additional portions of the ice sheet experience albedo flip. There is no evidence in the accurately dated terminations (I and II) of multi-millennia lag in ice sheet response. We infer that it would be not only dangerous, but also foolhardy to follow a BAU path for future GHG emissions.

4. Climate forcing scenarios

The IPCC BAU scenarios continue to be used as standard forcings for climate simulations, but there is no inherent reason that the world must follow BAU GHG growth rates. Almost a decade has elapsed since the IPCC scenarios were defined, so it seems worthwhile to compare measurements of atmospheric gases with the IPCC BAU scenarios.

We also compare with the 'alternative' scenario of Hansen *et al.* (2000), which was defined with the objective of keeping added human-made climate forcing in the twenty-first century no larger than 1.5 W m^{-2} . This limit keeps further global warming (after 2000) less than 1°C , and thus within the range of previous interglacial periods, assuming that the fast feedback climate sensitivity is approximately 3°C for doubled CO₂. This 1°C limit requires that CO₂ should not exceed 450–475 ppm, the exact CO₂ limit depending on the level of non-CO₂ forcings, as discussed below.

Figure 5 compares scenarios and observations for the three principal long-lived GHGs. It is difficult to discriminate among CO₂ scenarios, because they diverge gradually. However, emissions of fossil fuel CO₂ increased rapidly in the past

decade, consistent with IPCC BAU and more rapid than the alternative scenario. If CO₂ emissions continue to follow BAU for another decade, with annual emission increases averaging 2% per year, the emissions in 10 years will be 40% above those in the alternative scenario. In this case, it would be difficult, probably expensive and implausible, to get back on the path of the alternative scenario this half-century.

Non-CO₂ climate forcings are important, despite the fact that CO₂ is the largest human-made climate forcing. Indeed, expected difficulties in slowing the growth rate of CO₂ and eventually stabilizing atmospheric CO₂ amount make the non-CO₂ forcings all the more important. It now appears that only if reduction of the non-CO₂ forcings is achieved, and CO₂ growth is slowed, will it be possible to keep global temperature within or near the range of the warmest interglacial periods.

Fortunately, observed growth paths of non-CO₂ forcings exhibit promise. CH₄ (figure 5b) is increasing slower than even the alternative scenario, much slower than IPCC scenarios. This may be partly due to reduced losses from fossil fuels (reduced CH₄ loss from leaky pipelines and venting at oil wells, and capture at coal mines), as well as efforts to capture CH₄ at landfills and waste management facilities.

There is potential for greater reduction of CH₄ emissions. Such reduction could also reduce tropospheric ozone (O₃), an important GHG and a pollutant contributing to asthma and other respiratory diseases. The 'global warming potential' (GWP) assigned to CH₄ in the Kyoto Protocol understates its effect on climate because it excludes indirect effects.

Growth of N₂O is also falling below most scenarios, but only slightly. N₂O is especially important owing to its long atmospheric lifetime, of the order of a century. There is substantial potential for reducing its growth rate, which is due in part to excessive use of nitrogen in fertilization practices. There are potential multiple benefits in reducing N₂O emissions, but better understanding of nitrogen cycle is needed. It deserves greater attention and emphasis in climate mitigation efforts.

It would be better if all climate forcings were not packaged together and made interchangeable with CO₂ in mitigation strategies. Sources of different gases are usually independent and greater progress is likely from complementary focused programmes. However, in regulations of a specific activity or industry, the rules should be based on information about the effect of the activity on all climate forcings.

5. Carbon cycle and climate change

About one-quarter of fossil fuel CO₂ emissions will stay in the air 'forever', i.e. more than 500 years. This carbon cycle fact is well established (Archer 2005). However, implications of this fact have not penetrated the consciousness of the public and policy makers. We take 500 years as a practical definition of forever because it is long enough for large responses from both the ocean and ice sheets. Resulting climate changes would be, from humanity's perspective, irreversible.

Figure 6a shows the fate of a small pulse of CO₂ added to the atmosphere, for example via the burning of fossil fuels, as calculated from the indicated five-term fit to the Bern carbon cycle model (Joos et al. 1996; Shine et al. 2005). One-third of the emitted CO₂ is still in the air after a century, and almost one-quarter is still there

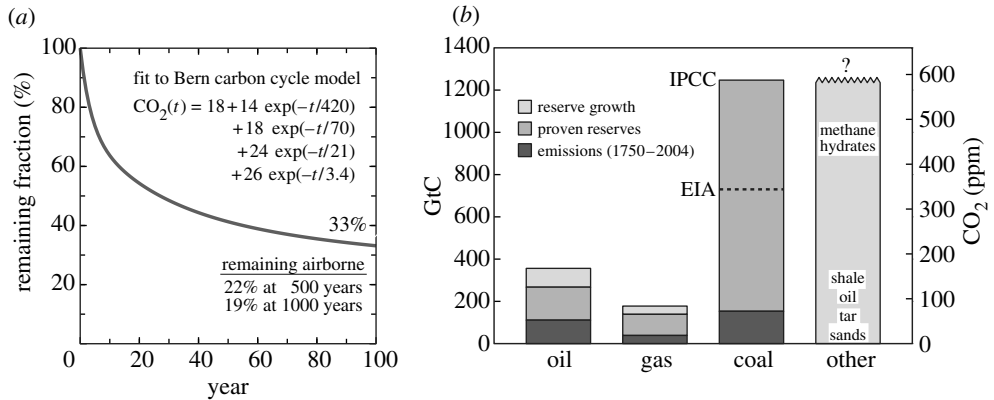


Figure 6. Carbon cycle constraints. (a) Decay of pulse CO_2 emission: decay of a small pulse of CO_2 added to today's atmosphere, based on the indicated analytic approximation to the Bern carbon cycle model and (b) fossil fuel reservoirs: fossil fuel emissions to date, proven reserves and estimated economically recoverable reserve growth based on EIA (2006) and, in the case of coal, IPCC (2001). Reservoirs of 'other' fossil fuels are uncertain but large, most estimates for hydrates being in the range 5000–10 000 GtC (Archer 2005).

after five centuries. A substantial part of the CO_2 is taken up by the ocean quickly. But the CO_2 added to the ocean, in effect, exerts a back pressure on the atmosphere that prevents all of the added CO_2 from being absorbed by the ocean until some CO_2 is buried in ocean sediments, which requires a very long time.

The pulse response function (figure 6a) is optimistic, being valid only for moderate CO_2 emission rates. It may be accurate for the alternative scenario, with slowing CO_2 emissions, but with BAU emissions the ocean chemistry becomes notably nonlinear (an increasing 'Revelle factor') and the biosphere is expected to become less effective at CO_2 uptake (Cox *et al.* 2000; Fung *et al.* 2005; Jones *et al.* 2006).

Given the estimated size of fossil fuel reservoirs (figure 6b), the chief implication is that we, humanity, cannot release to the atmosphere all, or even most, fossil fuel CO_2 . To do so would guarantee dramatic climate change, yielding a different planet than the one on which civilization developed and for which extensive physical infrastructure has been built.

Estimated oil and gas reservoirs (figure 6b), with only modest further use of coal, are sufficient to bring atmospheric CO_2 to approximately 450–475 ppm limit of the alternative scenario (Kharecha & Hansen 2007). Given the convenience of liquid and gas fuels, it seems likely that readily available oil and gas reservoirs will be exploited. Thus, attainment of the alternative scenario implies the need to phase out coal use, except where the CO_2 is captured and sequestered, and to impose the same constraint on development of unconventional fossil fuels. In practice, achievement of these goals surely requires a price (tax) on CO_2 emissions sufficient to discourage extraction of remote oil and gas resources as well as unconventional fossil fuels. Furthermore, the time required to develop fossil-free energy sources implies a need to stretch supplies of conventional oil and gas. In turn, this implies a need for near-term emphasis on energy efficiency.

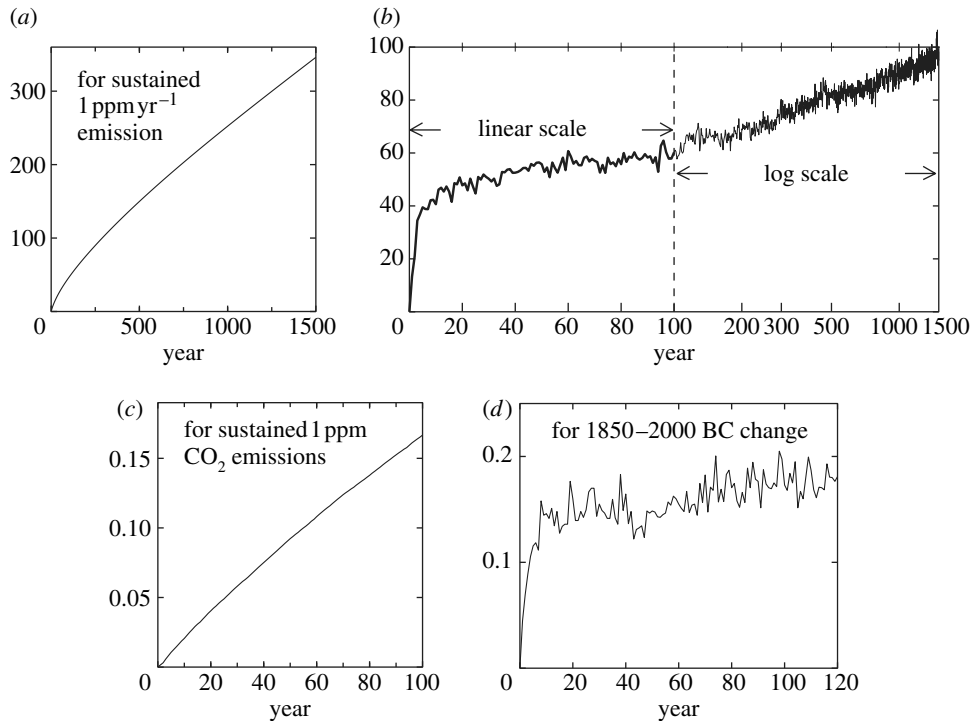


Figure 7. (a) Accumulated atmospheric CO₂ (ppm) for sustained emission of 1 ppm yr⁻¹, based on analytic fit (figure 6) to Bern carbon cycle model, (b) climate response function (%) based on global surface air warming of GISS modelE coupled to Russell ocean model, and (c,d) surface air temperature (°C) responses to sustained changes in (c) CO₂ emissions (1 ppm yr⁻¹) and (d) fossil fuel BC emissions (50% of 1850–2000 BC change).

Figure 7 provides additional fundamental data concerning the carbon cycle and climate. These data aid quantitative analyses of the contributions of CO₂ and non-CO₂ forcings to climate change.

Figure 7a shows accumulated atmospheric CO₂ for sustained 1 ppm yr⁻¹ emission of CO₂, as calculated with the five-term carbon cycle model in figure 6. Note that after 1500 years, the amount of emitted CO₂ remaining in the air is still nearly one-quarter of emissions during the 1500 years.

Figure 7b shows the climate response function defined as the fraction of equilibrium global warming that is obtained as a function of time. This function is based on a 3000-year simulation after instantaneous doubling of CO₂, using GISS modelE coupled to the Russell ocean. Note that only 60% of the equilibrium response is achieved after a century, and 90% after a millennium. This long response time is caused by slow uptake of heat by the deep ocean, which occurs primarily in the Southern Ocean.

The long response time of the climate system reduces the peak global warming due to human forcings, if these forcings eventually decline. Thus in the alternative scenario, despite the fact that there is approximately 0.5°C more global warming ‘in the pipeline’ due to gases already in the air, it is possible to keep further global warming (beyond that in 2000) less than 1°C despite assumed

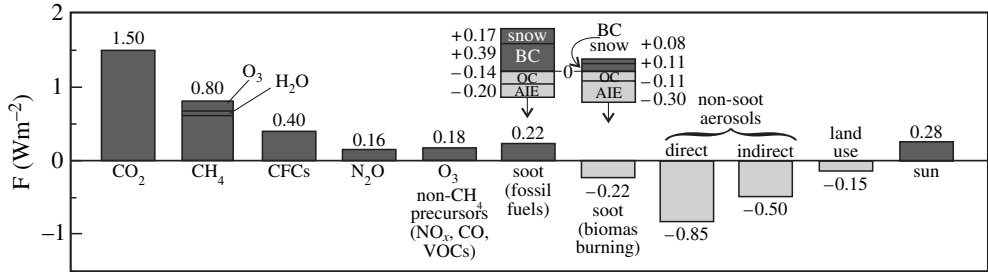


Figure 8. Estimated effective climate forcings for the industrial era, 1750–2000, with primary indirect effects grouped with the sources of direct forcing (Hansen *et al.* 2005a). BC, black carbon; OC, organic carbon; AIE, aerosol indirect effect; CFCs, chlorofluorocarbons; VOCs, volatile organic compounds.

additional 1.5 W m^{-2} GHG forcing this century. This is possible owing to the ocean's slow response and the assumption that there will be a slow long-term decrease in GHGs after 2100.

However, the climate system's long response time and slow mixing of heat into the ocean are a mixed benefit to society. Warming of the deep ocean may have at least two long-range detrimental effects (Hansen *et al.* 2006b): erosion of ice shelves around Antarctica and Greenland (Rignot & Jacobs 2002), and destabilization of methane hydrates on continental shelves (Harvey & Huang 1995; Archer 2007). The ocean's slow response delays such effects, but there is the danger of setting in motion a warming of the deep ocean that will lock in disastrous impacts which will unfold for future generations.

6. Non-CO₂ forcings

If fossil fuel CO₂ emissions continue to increase unabated, other climate forcings are relatively unimportant. However, this scenario is unlikely. Global warming is becoming apparent. Efforts to slow GHG emissions and stabilize global climate may increase. In this case, especially if the warming that constitutes 'danger' to the planet is as small as we estimate, non-CO₂ forcings become very important. Indeed, because some further increase of CO₂ is inevitable, it is probably implausible to keep additional global warming less than 1°C, unless non-CO₂ forcings are addressed aggressively.

Figure 8 summarizes known global climate forcings. Units are effective forcing in W m^{-2} in 2000 relative to pre-industrial times. Well-established indirect effects are grouped with the primary forcing. The bases for these estimates are given by Hansen *et al.* (2005a).

Methane is the largest climate forcing other than CO₂. Indeed, including indirect effects on tropospheric O₃ and stratospheric H₂O, forcing by CH₄ is half as large as that by CO₂. It is assumed in the alternative scenario of Hansen *et al.* (2000) that aggressive efforts to reduce human-made CH₄ emissions will be undertaken, such that CH₄ abundance decreases to 1300 ppb in 2100. Attainment of this decrease may require reduction of anthropogenic CH₄ sources by about one-third.

Methane release from permafrost (Zimov *et al.* 2006; Walter *et al.* 2007), should it accelerate with global warming, could spoil the efforts to reduce CH₄. Hansen & Sato (2004) argue that large release from methane hydrates is unlikely

if additional global warming is kept under 1°C, based on the fact that CH₄ increase was moderate during previous interglacial periods that were warmer than at present by up to 1°C. However, warming greater than 1°C raises the likelihood of a large positive feedback from methane hydrates. There seems to be a dichotomy of possible futures: either achieve strong reductions of both CO₂ and CH₄ emissions or both gases are likely to increase substantially.

Soot from fossil fuel burning, i.e. highly absorbing aerosols that contain black carbon (BC) and organic carbon (OC), are estimated to cause a global climate forcing of 0.22 W m⁻². This value accounts for the low efficacy of BC as a climate forcing. This (0.22 W m⁻²) is a conservative estimate for fossil fuel BC forcing, as discussed by Hansen *et al.* (2005a), because it assumes a high OC/BC ratio for fossil fuel emissions. In addition, it assigns 50% of the aerosol indirect effect (which causes cooling) to soot (BC/OC).

We calculate a GWP for BC relative to CO₂, per unit mass of emission cuts, as follows. We compare a sustained cut of fossil fuel BC emissions of 50% and a sustained CO₂ emission change of 1 ppm yr⁻¹. Global temperature change due to the direct BC forcing is calculated from a single run with GISS modelE coupled to the Russell ocean (figure 7d), and the result is reduced by the factor 0.22/0.39 to account for the indirect effects of BC (figure 8). CO₂ in the air (figure 7a) for sustained 1 ppm yr⁻¹ emission is based on the analytic fit (figure 6a) to the Bern carbon cycle model. Global temperature change for this CO₂ scenario (figure 7c) is obtained by integrating (to 20, 100 and 500 years) the product of the CO₂ forcing and the climate response function of the GISS model (figure 7b). The resulting GWP might be called a 'global temperature potential' (Shine *et al.* 2005), because it differs from the IPCC (2001) definition. However, our assumption of sustained emission cuts is appropriate, as there is no expectation that a country reducing BC emissions would later increase them, given the many benefits of reduced atmospheric soot.

Sustained 1 ppm yr⁻¹ reduction of CO₂ emissions yields $\Delta T \sim 0.04, 0.17$ and 0.67°C after 20, 100 and 500 years, respectively. The 50% change in fossil fuel BC yields $\Delta T \sim 0.15, 0.18$ and 0.24°C after 20, 100 and 500 years. To convert these results to GWPs, we first reduce the ΔT calculated for BC by the ratio 0.22/0.39 to account for indirect effects, which are predominately negative (figure 8). We then divide each temperature change by the mass of emission change (1 ppm yr⁻¹ CO₂ is approximately 7560 Tg yr⁻¹). Atmospheric BC amount is based on the Sato *et al.* empirical derivation of a global mean 'effective externally mixed optical depth' of 0.01 for total atmospheric BC. If the BC were externally mixed spheres of density 1 g cm⁻³, the implied BC mass would be 0.43 mg m⁻² or 0.22 Tg globally. Assuming that realistic BC particle shapes and internal mixing increase the optical absorption effectiveness of BC by a factor of 2 yields actual BC mass of 0.11 Tg globally. An assumed lifetime of 5.5 days yields total BC emissions of approximately 7.5 Tg yr⁻¹. We assume that fossil fuels and biomass are each half of atmospheric BC.

The resulting GWP for soot (BC) is approximately 2000 for 20 years, approximately 500 for 100 years and approximately 200 for 500 years. These GWPs are an average for the current distribution of BC sources. But emission reductions in some regions may be more effective than others. In particular, BC sources that contribute to snow albedo changes may be particularly effective. Moreover, if BC emissions contribute to ice melt, for example on Greenland,

they warrant special attention outside the value implied by their GWP. It is highly desirable to minimize emission of BC in the Arctic by ships or other sources.

Anthropogenic N₂O is now causing a climate forcing approximately 10% as large as that of CO₂ (figure 8). N₂O, like CO₂, has a long atmospheric lifetime. Current increase of N₂O is near the low end of IPCC scenarios (figure 5). There is potential to reduce this growth, for example via reduction in over-application of nitrogen fertilizers and treatment of wastewater (Barton & Atwater 2002). With better understanding of the nitrogen cycle, it may be possible to define practices that would yield an N₂O scenario well below both the IPCC and alternative scenarios (figure 5), with ancillary benefits in reduced pollution of water and air (Giles 2005).

7. Arctic

Recent warming in the Arctic is having notable effects on regional ecology, wildlife and indigenous peoples (ACIA 2004). BAU growth of climate forcings would be expected to yield an ice-free Arctic Ocean in the summer, as was the case during the middle Pliocene when global temperature was only 2–3°C warmer than today (Crowley 1996). It has been argued that the Greenland ice sheet would not probably survive with an ice-free Arctic Ocean (Hansen 2005). This raises the question: are there realistic scenarios that could avoid large Arctic warming and an ice-free Arctic Ocean?

Figure 9 summarizes some experiments with the GISS climate model aimed at investigating the contributions to Arctic climate change in the past century due to CO₂ and non-CO₂ forcings. The ‘all forcings’ simulations included both positive and negative forcings, the latter being mainly due to aerosols. CO₂ alone yields a global warming about three-fourths as large as observed global warming. However, the global response to the sum of the ‘air pollutants’ (tropospheric O₃, CH₄, BC, OC and the aerosol indirect effect of BC and OC) is as large as that for CO₂. Indeed, the Arctic warming due to these pollutants exceeds the Arctic warming due to CO₂.

Therefore, we suggest that it may still be possible to save the Arctic from complete loss of ice. If an absolute reduction of air pollutant forcings is achieved, along with reduction of the CO₂ growth rate as in the alternative scenario, there may be little additional loss of sea ice. Confirmation of this suggestion requires better measurements of the non-CO₂ forcings and their climate effects.

The Arctic epitomizes the global climate situation. The most rapid feasible slowdown of CO₂ emissions, coupled with focused reductions of other forcings, may just have a chance of avoiding disastrous climate change.

8. Discussion

Earth’s climate is remarkably sensitive to forcings, i.e. imposed changes of the planet’s energy balance. Both fast and slow feedbacks turn out to be predominately positive. As a result, our climate has the potential for large rapid fluctuations. Indeed, the Earth, and the creatures struggling to exist on the planet, have been repeatedly whipsawed between climate states. No doubt this rough ride has driven progression of life via changing stresses, extinctions and

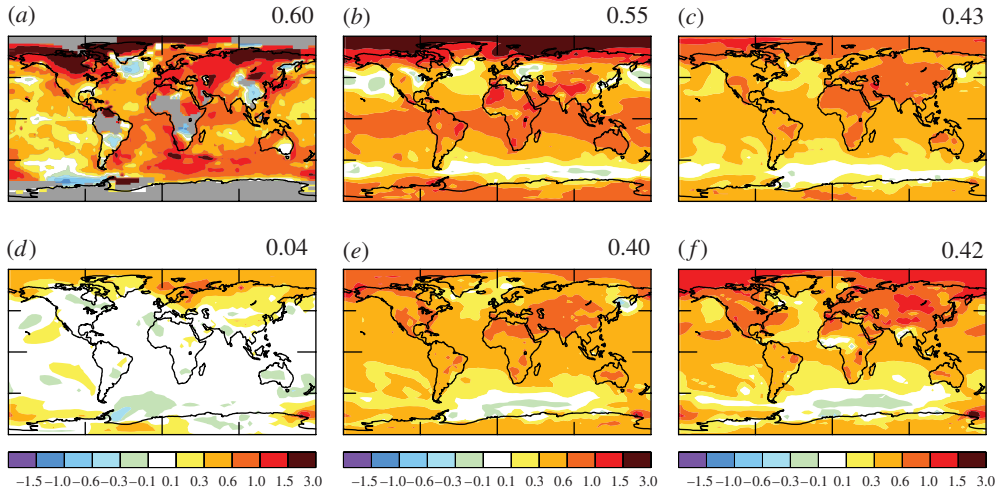


Figure 9. Surface temperature change ($^{\circ}\text{C}$) (based on local linear trends) for (a) observations and (b–f) simulations employing various combinations of transient 1880–2003 forcings: (b) all forcings, (c) CO_2 , (d) BC snow albedo effect, (e) tropospheric $\text{O}_3 + \text{CH}_4$ and (f) (e) + BC + OC + AIE. CH_4 forcing in (e) includes indirect effect on stratospheric H_2O . Snow albedo effect has 1880–2003 $\text{Fa} \sim 0.05 \text{ W m}^{-2}$. Results in (f) include forcings of (e) plus direct effects of black carbon (BC) and organic carbon (OC) aerosols from fossil fuels and biomass burning, their aerosol indirect effects (AIE) and the snow albedo effect.

species evolution. But civilization developed, and constructed extensive infrastructure, during a period of unusual climate stability, the Holocene, now almost 12 000 years in duration. That period is about to end.

(a) Climate sensitivity: the whipsaw

‘Fast-feedbacks’, including changing water vapour, clouds, sea ice, aerosols (dust, airborne organic particles, etc.) and effects of aerosols on clouds, determine climate response on decadal time-scales. Earth’s history yields a fast-feedback equilibrium climate sensitivity of approximately 3°C for doubled CO_2 forcing, i.e. approximately $3/4^{\circ}\text{C} (\text{W m}^{-2})^{-1}$ of forcing. Climate models concur. This sensitivity characterizes fast-feedback processes in the analyses of climate change.⁴

Real world climate response differs from this idealized case in two ways. First, response on decadal time-scales is much less than the fast-feedback equilibrium response. Half of the equilibrium response is obtained in 30 years, but, as the climate response function (figure 7b) shows, the other half requires a millennium. Second, assumption of fixed surface properties (vegetation cover and ice sheet area) becomes invalid long before equilibrium is achieved.

Climate sensitivity with surface properties free to change (but with GHG specified as a forcing, a choice relevant to the twenty-first century) is defined in figure 1, which reveals Antarctic temperature increase of $3^{\circ}\text{C} (\text{W m}^{-2})^{-1}$. Global temperature change is about half that in Antarctica, so this equilibrium global climate sensitivity is $1.5^{\circ}\text{C} (\text{W m}^{-2})^{-1}$, double the fast-feedback (Charney) sensitivity.

⁴Climate sensitivity, including the fast-feedback sensitivity, changes with the mean state of the Earth’s climate (Hansen et al. 2005a). However, climate sensitivity is practically independent of time over the past several hundred thousand years (figure 2).

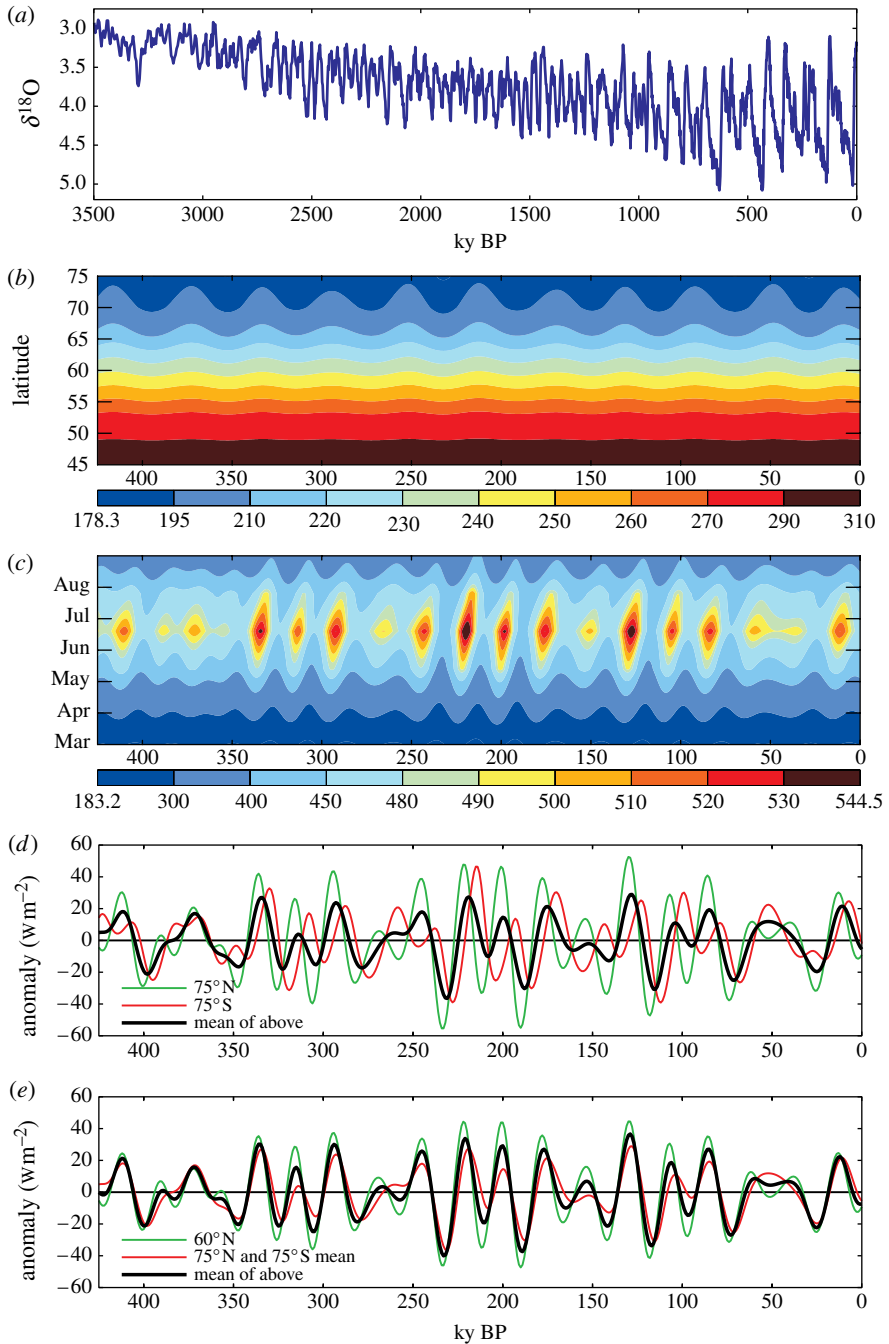


Figure 10. (a) $\delta^{18}\text{O}$ global sea-level/temperature proxy: global benthic oxygen isotope change, a proxy record of Plio-Pleistocene temperature and ice volume. (b–e) Late Pleistocene insolation as a function of latitude and season: (b) annual mean insolation (W m^{-2}), (c) 60° N insolation (W m^{-2}) and (d, e) late spring insolation: (d) 75° latitude and (e) 60° and 75° latitudes.

Is this $1.5^{\circ}\text{C} (\text{W m}^{-2})^{-1}$ sensitivity, rather than $0.75^{\circ}\text{C} (\text{W m}^{-2})^{-1}$, relevant to human-made forcings? Yes, for many purposes, in view of our conclusion that vegetation and ice sheets can change within the ocean response time (figure 7*b*). It might be argued that $1.5^{\circ}\text{C} (\text{W m}^{-2})^{-1}$ was derived from a (palaeoclimate) situation with large vulnerable ice sheets. True, but today further warming of even half the 5°C warming since the last ice age will make West Antarctica and at least the South Dome of Greenland vulnerable. Moreover, the $1.5^{\circ}\text{C} (\text{W m}^{-2})^{-1}$ sensitivity does not include GHG feedback. CO_2 , CH_4 and N_2O all seem to be positive feedbacks in palaeo records, with a lag of at most several hundred years, perhaps related to ocean mixing time.

Unlike albedo feedback, which may be decreasing as ice sheet area declines, GHG (carbon cycle and nitrogen cycle) feedback seems likely to grow as Earth warms. As tundra melts and the ocean floor warms the potential for methane hydrate release increases (Harvey & Huang 1995; Archer 2007). Global warming of at least 6°C at the Palaeocene–Eocene thermal maximum, *ca* 55 Myr BP, involved catastrophic carbon release to the atmosphere and mass extinctions (Bowen *et al.* 2006), one of at least several such large rapid warmings in Earth's history. Some portion of that carbon release and warming was probably a climate feedback.

The ultimate climate whipsaw occurred with snowball Earth events, most recently at the end of the pre-Cambrian, *ca* 540 Myr BP, when the oceans froze (Kirschvink *et al.* 2000). Whether the planet was a hard 'iceball' or a 'slushball', weathering seems to have been reduced to such a slow rate that volcanic CO_2 accumulated in the air until a super greenhouse melted tropical ice, and the albedo feedback whipped the planet to hellish hothouse conditions (Hoffman & Schrag 2002).

There is no inherent reason for both fast and slow feedbacks to be positive and strong, but they are on Earth. And given the origin of these feedbacks, in water, carbon and nitrogen cycles, it is probable that other terrestrial planets, circling our Sun and other stars, also have unstable climates. Evidence of past running water on Mars, albeit subject to alternative interpretation, suggests that Mars has undergone dramatic climate change. Venus probably experienced a runaway greenhouse effect, a state from which there is no return.

(*b*) Plio-Pleistocene whipsaw

The whipsaw between cold and warm climates grew dramatically from the middle Pliocene, 3.5 Myr BP, through the Pleistocene epoch (past 1.8 Myr). Earth cooled during this period, and the area of ice grew. When the planet became cold enough to have a large (Laurentide) ice sheet in North America, the albedo forcing was larger in the Northern Hemisphere than in the Southern Hemisphere, and global albedo forcing had magnitude similar to GHG forcing (figure 2*b*). This large Northern Hemisphere ice sheet at relatively low latitude caused transition from 41 kyr whipsaw periodicity to Late Pleistocene variations that give more prominence to 100 and 23 kyr periodicities.

Figure 10*a* shows growth of the climate whipsaw during the Plio-Pleistocene. $\delta^{18}\text{O}$ preserved in shells of benthic (deep ocean dwelling) foraminifera depends on both ocean temperature and the mass of water locked in ice sheets. The $\delta^{18}\text{O}$ record in figure 10*a* is the combination of 57 globally distributed ocean sediment cores (Lisiecki & Raymo 2005). In addition to a growing whipsaw amplitude, the

salient feature in the $\delta^{18}\text{O}$ curve is the change from 41 to 100 kyr periodicity *ca* 1 Myr ago. The higher temporal resolution in figures 2 and 3 reveals also 23 kyr periodicity, especially in sea level (figure 2*a*). Indeed, climate variations in the past million years contain strong 23, 41 and 100 kyr signals (Hays *et al.* 1976; Huybers 2006).

During the past 3 Myr, the Earth's surface cooled on average approximately 2–3°C (Dowsett *et al.* 1994; Raymo *et al.* 1996) with little change in the tropics. Sea level 3 Myr ago was 25 ± 10 m higher than today (Dowsett *et al.* 1994). The cooling appears to be related to continental drift and mountain building (Ruddiman & Kutzbach 1989), which increased weathering and drawdown of atmospheric CO_2 . CO_2 amount *ca* 3 Myr ago has been estimated as 380–425 ppm (Raymo *et al.* 1996), but the proxy measures of CO_2 make the uncertainty difficult to quantify. Change of meridional heat transport (Rind & Chandler 1991), perhaps associated with orographic and ocean bottom topography changes, could have contributed to mean temperature change.

Cooling increases the area of ice and the amplitude of climate fluctuations. Ice area is a powerful feedback that, together with associated GHG changes, fully accounts for Pleistocene global temperature change (figure 2*c*). Change from 41 to 100 kyr variability necessarily occurred once global cooling permitted growth of the Laurentide ice sheet.

Changing tilt of the Earth's spin axis, with its predominant 41 kyr frequency, principally determines annual insolation variation versus latitude (figure 10*b*). The 41 kyr variations are increasingly large, in absolute and relative insolation, towards the pole. Hemispheric 41 kyr forcings are in phase, increased tilt melting ice in both hemispheres, so global 41 kyr forcing is always present for high latitude ice sheets. Antarctic and Greenland ice sheets are at high latitudes where the 41 kyr tilt forcing rules.

When the planet is cold enough to harbour an ice sheet at lower latitude, precession (thus the season when the Earth is closest to the Sun) becomes more important, with its predominant 23 kyr frequency. The magnitude of precession forcing depends on the eccentricity of the Earth's orbit, disappearing with a circular orbit. Thus, *ca* 100 kyr eccentricity variations come into play along with precession. Figure 10*c* shows insolation forcing at the latitude of the Laurentide ice sheet (60° N), exhibiting both the 23 and 100 kyr frequencies.

It remains to connect insolation variations in figure 10*b,c* back to figures 2 and 3. Figure 10*d* is the late spring (April–May–June in NH, October–November–December in SH) insolation at 75° latitude. Precession contributions are out of phase in the two hemispheres, and thus the average of the two hemispheres yields a 41 kyr frequency.

The result differs when the Laurentide ice sheet is added (figure 10*e*). Nearly identical curves are obtained for absorbed energy using present day satellite-observed albedo. Albedo forcing by the Laurentide ice sheet is comparable to that for Antarctica plus Greenland/Arctic. Thus the combined forcing for the Laurentide and polar ice sheets (thick line in figure 10*e*) exhibits all of the periodicities. The lower latitude damps amplitude of 41 kyr variations, but the primary effect is the absence of a mid-latitude ice sheet in the Southern Hemisphere. The Laurentide ice sheet, when it is present, is globally dominant.

Thus, insolation anomalies drive temperature and sea-level change. Terminations occur after a period in which precession maxima miss a beat or two (figure 3). Weak precession maxima permit ice amount to grow especially large and sea level to achieve an extreme minimum.

This ice albedo feedback is not a runaway effect. As insolation (or other) forcing increases, the area of ice vulnerable to melting increases. Ice sheet demise may occur in pulses as additional ice sheets or portions of ice sheets (e.g. West Antarctica or the South Dome of Greenland) become vulnerable. As long as there is ice on the planet, the response time to insolation forcings can be no shorter than the shortest insolation period (*ca* 6 kyr half-width of precession anomalies), even if ice sheets have no inertia (instantaneous response).

This analysis of the Plio-Pleistocene whipsaw has two important implications. First, the multi-millennial time-scale for ice sheet disintegration probably reflects the forcing time-scale, not an inertial time-scale for ice sheets. Second, climate sensitivity that includes the effect of slow feedbacks implies an ominously low level for the amount of human-made GHGs which will constitute ‘dangerous’ change.

(c) *Albedo flip: rapid climate change*

A salient feature of terrestrial climate change is its asymmetry. Warmings are rapid, usually followed by slower descent into colder climate. Given the symmetry of orbital forcings (figures 3 and 10), the cause of rapid warming at glacial ‘terminations’ must lie in a climate feedback. Clearly, the asymmetric feedback is the albedo flip of ice and snow that occurs when they become warm enough to begin melting.

The albedo-flip feedback helps explain the rapidity of deglaciations and their early beginnings relative to Milankovitch’s summer insolation maxima. A positive perturbation to insolation is most effective in spring because it lengthens the melt season. Once the albedo is flipped to dark, it usually stays dark until the cold season returns. Increased absorption of sunlight caused by albedo flip provides the energy for rapid ice melt. When the insolation forcing reverses, ice sheet regrowth can be slower, as it is limited by the rate of snowfall in cold regions.

Except for snowball Earth conditions, albedo flip is not a runaway feedback. However, the magnitude of the potential global climate response increases as ice sheet size increases. Thus, as the Earth cooled from the Pliocene through the Pleistocene, the amplitude of global temperature fluctuations increased.

Sea-level increases (figure 2a) associated with insolation anomalies have characteristic response time similar to the time-scale of the forcing (minimum half-width *ca* 6 kyr). This is consistent with a persistence time of *ca* 7 kyr found by Mudelsee & Raymo (2005) for ice volume changes reflected in marine oxygen isotope records. If these long time-scales are interpreted as an inherent time-scale for ice sheet disintegration and built into ice sheet models, then they provide a false sense of security about sea level.

The unusual stability of the Earth’s climate during the Holocene is probably due to the fact that the Earth has been warm enough to keep ice sheets off North America and Asia, but not warm enough to cause disintegration of the Greenland or Antarctic ice sheets.

An ice sheet in equilibrium may have summer melt on its fringes, balanced by interior ice sheet growth. Large climate change will occur only if a forcing is sufficient to initiate rapid dynamical feedbacks and disintegration of a substantial portion of the ice sheet. Rapidly rising temperatures in the past three decades (figure 4), evidence that the Earth is now substantially out of energy balance (Hansen *et al.* 2005*b*), and indications of accelerating change on West Antarctica and Greenland (see below) indicate that the period of stability is over.

(*d*) Planet Earth today: imminent peril

The imminent peril is initiation of dynamical and thermodynamical processes on the West Antarctic and Greenland ice sheets that produce a situation out of humanity's control, such that devastating sea-level rise will inevitably occur. Climate forcing of this century under BAU would dwarf natural forcings of the past million years, indeed it would probably exceed climate forcing of the middle Pliocene, when the planet was not more than 2–3°C warmer and sea level 25 ± 10 m higher (Dowsett *et al.* 1994). The climate sensitivities we have inferred from palaeoclimate data ensure that a BAU GHG emission scenario would produce global warming of several degrees Celsius this century, with amplification at high latitudes.

Such warming would assuredly activate the albedo-flip trigger mechanism over large portions of these ice sheets. In combination with warming of the nearby ocean and atmosphere, the increased surface melt would bring into play multiple positive feedbacks leading to eventual nonlinear ice sheet disintegration, as discussed by Hansen (2005). It is difficult to predict time of collapse in such a nonlinear problem, but we find no evidence of millennial lags between forcing and ice sheet response in palaeoclimate data. An ice sheet response time of centuries seems probable, and we cannot rule out large changes on decadal time-scales once wide-scale surface melt is underway. With GHGs continuing to increase, the planetary energy imbalance provides ample energy to melt ice corresponding to several metres of sea level per century (Hansen *et al.* 2005*b*).

With this danger in mind, it is appropriate to closely monitor ice sheet conditions. Area of summer melt on Greenland increased from approximately 450 000 km² in the first few years after satellite observations began in 1979 to more than 600 000 km² in recent years (Steffen *et al.* 2004). Iceberg discharge from Greenland increased markedly over the past 15 years. Mass loss increased from 4–50 km³ yr⁻¹ in 1993–1998 to 57–105 km³ yr⁻¹ in 1999–2004, based on radar altimeters, with probable losses at the higher ends of those ranges (Thomas *et al.* 2006). Recent analyses of satellite gravity field data yield a net annual loss of 101 ± 16 km³ yr⁻¹ during 2003–2005 (Luthcke *et al.* 2006).

The gravest threat we foresee starts with surface melt on West Antarctica and interaction among positive feedbacks leading to catastrophic ice loss. Warming in West Antarctica in recent decades has been limited by effects of stratospheric ozone depletion (Shindell & Schmidt 2004). However, climate projections (Hansen *et al.* 2006*b*) find surface warming in West Antarctica and warming of nearby ocean at depths that may attack buttressing ice shelves. Loss of ice shelves allows more rapid discharge from ice streams, in turn a lowering and warming of the ice sheet surface, and increased surface melt. Rising sea level helps unhinge the ice from pinning points.

West Antarctica seems to be moving into a mode of significant mass loss (Thomas *et al.* 2004). Gravity data yielded mass loss of approximately $150 \text{ km}^3 \text{ yr}^{-1}$ in 2002–2005 (Velicogna & Wahr 2006). A warming ocean has eroded ice shelves by more than 5 m yr^{-1} over the past decade (Rignot & Jacobs 2002; Shepherd *et al.* 2004). Satellite QuickSCAT radiometer observations (Nghiem *et al.* 2007), initiated in 1999, reveal an increasing area of summer melt on West Antarctica and an increasing melt season over the period of record. Attention has focused on Greenland, but the most recent gravity data indicate comparable mass loss from West Antarctica. We find it implausible that BAU scenarios, with climate forcing and global warming exceeding those of the Pliocene, would permit a West Antarctic ice sheet of present size to survive even for a century.

Our concern that BAU GHG scenarios would cause large sea-level rise this century (Hansen 2005) differs from estimates of IPCC (2001, 2007), which foresees little or no contribution to twenty-first century sea-level rise from Greenland and Antarctica. However, the IPCC analyses and projections do not well account for the nonlinear physics of wet ice sheet disintegration, ice streams and eroding ice shelves, nor are they consistent with the palaeoclimate evidence we have presented for the absence of discernable lag between ice sheet forcing and sea-level rise.

The best chance for averting ice sheet disintegration seems to be intense simultaneous efforts to reduce both CO_2 emissions and non- CO_2 climate forcings. As mentioned above, there are multiple benefits from such actions. However, even with such actions, it is probable that the dangerous level of atmospheric GHGs will be passed, at least temporarily. We have presented evidence (Hansen *et al.* 2006b) that the dangerous level of CO_2 can be no more than approximately 450 ppm. Our present discussion, including the conclusion that slow feedbacks (ice, vegetation and GHG) can come into play on century time-scales or sooner, makes it probable that the dangerous level is even lower.

Present knowledge does not permit accurate specification of the dangerous level of human-made GHGs. However, it is much lower than has commonly been assumed. If we have not already passed the dangerous level, the energy infrastructure in place ensures that we will pass it within several decades.

We conclude that a feasible strategy for planetary rescue almost surely requires a means of extracting GHGs from the air. Development of CO_2 capture at power plants, with below-ground CO_2 sequestration, may be a critical element. Injection of the CO_2 well beneath the ocean floor assures its stability (House *et al.* 2006). If the power plant fuel is derived from biomass, such as cellulosic fibres⁵ grown without excessive fertilization that produces N_2O or other offsetting GHG emissions, it will provide continuing drawdown of atmospheric CO_2 .

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⁵The potential of these ‘amber waves of grain’ and coastal facilities for permanent underground storage ‘from sea to shining sea’ to help restore America’s technical prowess, moral authority and prestige, for the sake of our children and grandchildren, in the course of helping to solve the climate problem, has not escaped our attention.

References

- Archer, D. 2005 Fate of fossil fuel CO₂ in geologic time. *J. Geophys. Res.* **110**, C09 S05. (doi:10.1029/2004JC002625)
- Archer, D. 2007 Methane hydrates and anthropogenic climate change. *Biosci. Discuss.* **4**, 993–1057.
- Arctic Climate Impact Assessment (ACIA) 2004 *Impacts of a warming Arctic*. Cambridge, UK: Cambridge University Press.
- Barton, P. K. & Atwater, J. W. 2002 Nitrous oxide emissions and the anthropogenic nitrogen in wastewater and solid waste. *J. Environ. Eng.* **128**, 137–150. (doi:10.1061/(ASCE)0733-9372(2002)128:2(137))
- Bender, M. L., Floch, G., Chappellaz, J., Suwa, M., Barnola, J. M., Blunier, T., Dreyfus, G., Jouzel, J. & Parrenin, F. 2006 Gas age—ice age differences and the chronology of the Vostok ice core, 0–100 ka. *J. Geophys. Res.* **111**, D21 115. (doi:10.1029/2005JD006488)
- Berger, A. L. 1978 Long term variations of daily insolation and quaternary climate changes. *J. Atmos. Sci.* **35**, 2362–2367. (doi:10.1175/1520-0469(1978)035<2362:LTVODI>2.0.CO;2)
- Bowen, G. J. *et al.* 2006 Eocene hyperthermal event offers insight into greenhouse warming. *EOS Trans. Am. Geophys. Union* **87**, 165–169.
- Caillon, N., Severinghaus, J. P., Jouzel, J., Barnola, J. M., Kang, J. & Lipenkov, V. Y. 2003 Timing of atmospheric CO₂ and Antarctic temperature changes across Termination III. *Science* **299**, 1728–1731. (doi:10.1126/science.1078758)
- Cannariato, K. G. & Kennett, J. P. 2005 Structure of the penultimate deglaciation along the California margin and implications for Milankovitch theory. *Geology* **33**, 157–160. (doi:10.1130/G21065.1)
- Charney, J. 1979 *Carbon dioxide and climate: a scientific assessment*. Washington, DC: National Academy of Sciences Press.
- CLIMAP 1981 Seasonal reconstructions of the Earth's surface at the last glacial maximum in Map Series. Technical Report MC-36. Boulder, CO: Geological Society of America.
- Cox, P. M., Betts, R. A., Jones, C. D., Spall, S. A. & Totterdel, I. J. 2000 Acceleration of global warming due to carbon cycle feedbacks in a coupled climate model. *Nature* **408**, 184–187. (doi:10.1038/35041539)
- Crowley, T. J. 1996 Pliocene climates: the nature of the problem. *Mar. Micropaleontol.* **27**, 3–12. (doi:10.1016/0377-8398(95)00049-6)
- Dowsett, H. J., Thompson, R., Barron, J., Cronin, T., Fleming, F., Ishman, R., Poore, D., Willard, D. & Holtz, T. 1994 Joint investigations of the Middle Pliocene climate. *Global Planet. Change* **9**, 169–195. (doi:10.1016/0921-8181(94)90015-9)
- Drysdale, R. N., Zanchetta, G., Hellstrom, J. C., Fallick, A. E. & Zhao, J. 2005 Stalagmite evidence for the onset of the Last Interglacial in southern Europe at 129 ± 1 ka. *Geophys. Res. Lett.* **32**, L24 708. (doi:10.1029/2005GL024658)
- Energy Information Administration (EIA), US Dept. Energy 2006 *International energy outlook*. See <http://www.eia.doe.gov/oiaf/ieo/>.
- EPICA community members 2004 Eight glacial cycles from an Antarctic ice core. *Nature* **429**, 623–628. (doi:10.1038/nature02599)
- EPICA community members 2006 One-to-one coupling of glacial climate variability in Greenland and Antarctica. *Nature* **444**, 195–198. (doi:10.1038/nature05301)
- Fairbanks, R. G. 1989 A 17,000-year glacio-eustatic sea level record—influence of glacial melting rates on the Younger Dryas event and deep-sea circulation. *Nature* **342**, 637–642. (doi:10.1038/342637a0)
- Fung, I. Y., Doney, S. C., Lindsay, K. & John, J. 2005 Evolution of carbon sinks in a changing climate. *Proc. Natl Acad. Sci. USA* **102**, 11 201–11 206. (doi:10.1073/pnas.0504949102)
- Giles, J. 2005 Nitrogen study fertilizes fears of pollution. *Nature* **433**, 791. (doi:10.1038/433791a)
- Hansen, J. 2005 A slippery slope: how much global warming constitutes “dangerous anthropogenic interference”? *Clim. Change* **68**, 269–279. (doi:10.1007/s10584-005-4135-0)

- Hansen, J. & Sato, M. 2004 Greenhouse gas growth rates. *Proc. Natl Acad. Sci. USA* **101**, 16 109–16 114. (doi:10.1073/pnas.0406982101)
- Hansen, J., Lacis, A., Rind, D., Russell, G., Stone, P., Fung, I., Ruedy, R. & Lerner, J. 1984 Climate sensitivity: analysis of feedback mechanisms. In *Climate processes and climate sensitivity* (eds J. E. Hansen & T. Takahashi). Geophysics Monograph Series 29, pp. 130–163. Washington, DC: American Geophysical Union.
- Hansen, J., Lacis, A., Ruedy, R., Sato, M. & Wilson, H. 1993 How sensitive is the world's climate? *Natl Geogr. Res. Explor.* **9**, 142–158.
- Hansen, J., Sato, M. & Ruedy, R. 1997 Radiative forcing and climate response. *J. Geophys. Res.* **102**, 6831–6864. (doi:10.1029/96JD03436)
- Hansen, J., Sato, M., Ruedy, R., Lacis, A. & Oinas, V. 2000 Global warming in the twenty-first century: an alternative scenario. *Proc. Natl Acad. Sci. USA* **97**, 9875–9880. (doi:10.1073/pnas.170278997)
- Hansen, J. et al. 2005a Efficacy of climate forcings. *J. Geophys. Res.* **110**, D18 104. (doi:10.1029/2005JD005776)
- Hansen, J. et al. 2005b Earth's energy imbalance: confirmation and implications. *Science* **308**, 1431–1435. (doi:10.1126/science.1110252)
- Hansen, J., Sato, M., Ruedy, R., Lo, K., Lea, D. W. & Medina-Elizade, M. 2006a Global temperature change. *Proc. Natl Acad. Sci. USA* **101**, 16 109–16 114. (doi:10.1073/pnas.0406982101)
- Hansen, J. et al. 2006b Dangerous human-made interference with climate: a GISS model E study. *Atmos. Chem. Phys. Discuss.* **6**, 12 549–12 610.
- Harvey, L. D. D. & Huang, Z. 1995 Evaluation of the potential impact of methane clathrate destabilization on future global warming. *J. Geophys. Res.* **100**, 2905–2926. (doi:10.1029/94JD02829)
- Hays, J. D., Imbrie, J. & Shackleton, N. J. 1976 Variations in the Earth's orbit: pacemaker of the ice ages. *Science* **194**, 1121–1132. (doi:10.1126/science.194.4270.1121)
- Henderson, G. M. & Slowey, N. C. 2000 Evidence from U–Th dating against Northern Hemisphere forcing of the penultimate deglaciation. *Nature* **404**, 61–66. (doi:10.1038/35003541)
- Hoffman, P. F. & Schrag, D. P. 2002 The snowball Earth hypothesis: testing the limits of global change. *Terra Nova* **14**, 129–155. (doi:10.1046/j.1365-3121.2002.00408.x)
- House, K. Z., Schrag, D. P., Harvey, C. F. & Lackner, K. S. 2006 Permanent carbon dioxide storage in deep-sea sediments. *Proc. Natl Acad. Sci. USA* **103**, 12 291–12 296. (doi:10.1073/pnas.0605318103)
- Huybers, P. 2006 Early Pleistocene glacial cycles and the integrated summer insolation forcing. *Science* **313**, 508–511. (doi:10.1126/science.1125249)
- Intergovernmental Panel on Climate Change (IPCC) 2001 In *Climate change 2001: the scientific basis* (eds J. T. Houghton, Y. Ding, D. J. Griggs, M. Noguer, P. J. van der Linden, X. Dai, K. Maskell & C. A. Johnson). Cambridge, UK: Cambridge University Press.
- IPCC 2007 *Climate change 2007: the physical basis—summary for policymakers*. See <http://www.ipcc.ch/SPM2feb07.pdf>.
- Jones, C. D., Cox, P. M. & Huntingford, C. 2006 Climate–carbon cycle feedbacks under stabilization: uncertainty and observational constraints. *Tellus B* **58**, 603–613. (doi:10.1111/j.1600-0889.2006.00215.x)
- Joos, F., Bruno, M., Fink, R., Stocker, T. F., Siegenthaler, U., Le Quere, C. & Sarmiento, J. L. 1996 An efficient and accurate representation of complex oceanic and biospheric models of anthropogenic carbon uptake. *Tellus B* **48**, 397–417. (doi:10.1034/j.1600-0889.1996.t01-2-00006.x)
- Kharecha, P. A. & Hansen, J. E. 2007 Implications of 'peak oil' for atmospheric CO₂ and climate. arXiv:0704.2782v1.
- Kirschvink, J. L., Gaidos, E. J., Bertani, L. E., Beukes, N. J., Gutzmer, J., Maepa, L. N. & Steinberger, R. E. 2000 Paleoproterozoic snowball Earth: extreme climatic and geochemical global change and its biological consequences. *Proc. Natl Acad. Sci. USA* **97**, 1400–1405. (doi:10.1073/pnas.97.4.1400)

- Lea, D. W., Pak, D. K. & Spero, H. J. 2000 Climate impact of late Quaternary equatorial Pacific sea surface temperature variations. *Science* **289**, 1719–1724. (doi:10.1126/science.289.5485.1719)
- Lea, D. W., Martin, P. A., Pak, D. K. & Spero, H. J. 2002 Reconstructing a 350 ky history of sea level using planktonic Mg/Ca and oxygen isotope records from a Cocos Ridge core. *Q. Sci. Rev.* **21**, 283–293. (doi:10.1016/S0277-3791(01)00081-6)
- Lisiecki, L. E. & Raymo, M. E. 2005 A Pliocene–Pleistocene stack of 57 globally distributed benthic $\delta^{18}\text{O}$ records. *Paleoceanography* **20**, PA1 003. (doi:10.1029/2004PA001071)
- Luthcke, S. B., Zwally, H. J., Abdalati, W., Rowlands, D. D., Ray, R. D., Nerem, R. S., Lemoine, F. G., McCarthy, J. J. & Chinn, D. S. 2006 Recent Greenland ice mass loss by drainage system from satellite gravity observations. *Science* **314**, 1286–1289. (doi:10.1126/science.1130776)
- Medina-Elizade, M. & Lea, D. W. 2005 The mid-Pleistocene transition in the Tropical Pacific. *Science* **310**, 1009–1012. (doi:10.1126/science.1115933)
- Milankovitch, M. 1941 *Kanon der Erdbestrahlung und seine Anwendung auf das Eiszeitenproblem*. Belgrade, Serbia: Royal Serbian Academy.
- Mishchenko, M. I., Geogdzhayev, I. V., Rossow, W. B., Cairns, B., Carlson, B. E., Lacis, A. A., Liu, L. & Travis, L. D. 2007 Long-term satellite record reveals likely recent aerosol trend. *Science* **315**, 1543. (doi:10.1126/science.1136709)
- Monnin, E., Indermühle, A., Dällenbach, A., Flückiger, J., Stauffer, B., Stocker, T. F., Raynaud, D. & Barnola, J.-M. 2001 Atmospheric CO_2 concentrations over the last glacial termination. *Science* **291**, 112–114. (doi:10.1126/science.291.5501.112)
- Mudelsee, M. & Raymo, M. E. 2005 Slow dynamics of the Northern Hemisphere glaciation. *Paleoceanography* **20**, PA4 022. (doi:10.1029/2005PA001153)
- Nghiem, S. V., Steffen, K., Neumann, G. & Huff, R. 2007 Snow accumulation and snowmelt monitoring in Greenland and Antarctica. In *Dynamic planet: monitoring and understanding a dynamic planet with geodetic and oceanographic tools*, vol. 130 (eds C. Rizos & P. Tregoning), pp. 31–38. New York, NY: Springer.
- Overpeck, J. T., Otto-Bliesner, B. L., Miller, G. H., Muhs, D. R., Alley, R. B. & Kiehl, J. T. 2006 Paleoclimate evidence for future ice-sheet instability and rapid sea level rise. *Science* **311**, 1747–1750. (doi:10.1126/science.1115159)
- Petit, J. R. *et al.* 1999 420,000 years of climate and atmospheric history revealed by the Vostok deep Antarctic ice core. *Nature* **399**, 429–436. (doi:10.1038/20859)
- Potter, E. K. & Lambeck, K. 2003 Reconciliation of sea-level observations in the Western North Atlantic during the last glacial cycle. *Earth Planet. Sci. Lett.* **217**, 171–181. (doi:10.1016/S0012-821X(03)00587-9)
- Rasmussen, S. O. *et al.* 2006 A new Greenland ice core chronology for the last glacial termination. *J. Geophys. Res.* **111**, D06 102. (doi:10.1029/2005JD006079)
- Raymo, M. E. 1997 The timing of major climate terminations. *Paleoceanography* **12**, 577–585. (doi:10.1029/97PA01169)
- Raymo, M. E., Grant, B., Horowitz, M. & Rau, G. H. 1996 Mid-Pliocene warmth: stronger greenhouse and stronger conveyor. *Mar. Micropaleontol.* **27**, 313–326. (doi:10.1016/0377-8398(95)00048-8)
- Rignot, E. & Jacobs, S. S. 2002 Rapid bottom melting widespread near Antarctic ice sheet grounding lines. *Science* **296**, 2020–2023. (doi:10.1126/science.1070942)
- Rind, D. & Chandler, M. 1991 Increased ocean heat transports and warmer climate. *J. Geophys. Res.* **96**, 7437–7461.
- Roe, G. 2006 In defense of Milankovitch. *Geophys. Res. Lett.* **33**, L24 703. (doi:10.1029/2006GL027817)
- Ruddiman, W. F. 2003 The anthropogenic greenhouse era began thousands of years ago. *Clim. Change* **61**, 261–293. (doi:10.1023/B:CLIM.0000004577.17928.f0)
- Ruddiman, W. F. & Kutzbach, J. E. 1989 Forcing of late Cenozoic Northern Hemisphere climate by plateau uplift in southern Asia and the American west. *J. Geophys. Res.* **94**, 18 409–18 427.

- Saraswat, R., Nigam, R., Weldeab, S., Mackensen, A. & Naidu, P. D. 2005 A first look at past sea surface temperatures in the equatorial Indian Ocean from Mg/Ca in foraminifera. *Geophys. Res. Lett.* **32**, L24 605. (doi:10.1029/2005GL024093)
- Shackleton, N. J. 2000 The 100,000-year ice-age cycle identified and found to lag temperature, carbon dioxide, and orbital eccentricity. *Science* **289**, 1897–1902. (doi:10.1126/science.289.5486.1897)
- Shepherd, A., Wingham, D. & Rignot, E. 2004 Warm ocean is eroding West Antarctic ice sheet. *Geophys. Res. Lett.* **31**, L23 402. (doi:10.1029/2004GL021106)
- Shindell, D. T. & Schmidt, G. A. 2004 Southern Hemisphere climate response to ozone changes and greenhouse gas increases. *Geophys. Res. Lett.* **31**, L18 209. (doi:10.1029/2004GL020724)
- Shine, K. P., Fuglestedt, J. S., Hailemariam, K. & Stuber, N. 2005 Alternatives to the global warming potential for comparing climate impacts of emissions of greenhouse gases. *Clim. Change* **68**, 281–302. (doi:10.1007/s10584-005-1146-9)
- Siddall, M., Rohling, E. J., Almogi-Labin, A., Hemleben, Ch., Meischner, D., Schmelzer, I. & Smeed, D. A. 2003 Sea-level fluctuations during the last glacial cycle. *Nature* **423**, 853–858. (doi:10.1038/nature01690)
- Spahni, R. et al. 2005 Atmospheric methane and nitrous oxide of the late Pleistocene from Antarctic ice cores. *Science* **310**, 317–321. (doi:10.1126/science.1120132)
- Stanford, J. D., Rohling, E. J., Hunter, S. E., Roberts, A. P., Rasmussen, S. O., Bard, E., McManus, J. & Fairbanks, R. G. 2006 Timing of meltwater pulse 1a and climate responses to meltwater injections. *Paleoceanography* **21**, PA4 103. (doi:10.1029/2006PA001340)
- Steffen, K., Nghiem, S. V., Huff, R. & Neumann, G. 2004 The melt anomaly of 2002 on the Greenland Ice Sheet from active and passive microwave satellite observations. *Geophys. Res. Lett.* **31**, L20 402. (doi:10.1029/2004GL020444)
- Thomas, R. et al. 2004 Accelerated sea-level rise from West Antarctica. *Science* **306**, 255–258. (doi:10.1126/science.1099650)
- Thomas, R., Frederick, E., Krabill, W., Manizade, S. & Martin, C. 2006 Progressive increase in ice loss from Greenland. *Geophys. Res. Lett.* **33**, L10 503. (doi:10.1029/2006GL026075)
- Thompson, W. G. & Goldstein, S. L. 2005 Open-system coral ages reveal persistent suborbital sea-level cycles. *Science* **308**, 401–404. (doi:10.1126/science.1104035)
- Velicogna, I. & Wahr, J. 2006 Measurements of time-variable gravity show mass loss in Antarctica. *Science* **311**, 1754–1756. (doi:10.1126/science.1123785)
- Vimeux, F., Cuffey, K. M. & Jouzel, J. 2002 New insights into Southern Hemisphere temperature changes from Vostok ice cores using deuterium excess correction. *Earth Planet. Sci. Lett.* **203**, 829–843. (doi:10.1016/S0012-821X(02)00950-0)
- Walter, K. M., Smith, L. C. & Chapin, F. S. 2007 Methane bubbling from northern lakes: present and future contributions to the methane budget. *Phil. Trans. R. Soc. A* **365**, 1657–1676. (doi:10.1098/rsta.2007.2036)
- Wunsch, C. 2003 The spectral description of climate change including the 100 ky energy. *Clim. Dyn.* **20**, 353–363.
- Yuan, D. et al. 2004 Timing, duration, and transitions of the last interglacial Asian monsoon. *Science* **304**, 575–578. (doi:10.1126/science.1091220)
- Zimov, S. A., Schuur, E. A. G. & Chapin, F. S. 2006 Permafrost and the global carbon budget. *Science* **312**, 1612–1613. (doi:10.1126/science.1128908)