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Group Report: How Can We Use Paleodata to Evaluate the Internal Variability and Feedbacks in the Climate System?

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INTRODUCTION

Our knowledge of the modern climate system is encoded in conventional atmosphere–ocean general circulation models (AOGCMs), and we expect them to be useful as aids in predicting the way in which global climate will evolve in response to changes in internal or external forcing. In spite of their considerable numerical complexity, however, the finite resolution at which we can afford to integrate them ensures that many crucial physical processes can be represented only in a heavily parameterized form (e.g., tropical cumulous convection, clouds, boundary layer processes, radiative transfer, etc.). Although we may adjust the variables that control the impact of these parameterized processes to obtain a high quality fit to modern climate, we have no reason to believe that the model "tuned" in this way will behave realistically when differently forced. There is, therefore, good reason to attempt to test these models further by comparing their predictions to observed states of the climate system that differ significantly from that which obtains at present. If the AOGCM is able to pass such additional tests, then we will be more justified in placing credence in its prediction of the response to the ongoing anthropogenic increase of the radiatively active trace gases carbon dioxide (CO_2) and methane (CH_4). Paleodata provide the only source of the information required to perform such additional tests.

Global Changes in the Perspective of the Past
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In the following discussion, we will focus especially on the way in which such data on past states of the climate system may be invoked to constrain variations through time of the behavior of several important subcomponents of the system. Such information is vital in helping us to appreciate the nature of internal climate system variability and the physical–chemical–biological feedbacks that underlie this variability. Since the response to external forcing associated with variations of the Earth's orbit around the Sun is apparently a crucial factor in triggering such variability throughout late Quaternary time, and since the dating of significant climatic events is far better in this most recent epoch of Earth history than for earlier times, we will focus most of the ensuing discussion on this period. Within this period, the dominant climatic signal is that associated with the 100,000-year cycle of the ice ages, which began around 900,000 years ago. This cycle is now known to involve dramatic changes in the thermohaline circulation of the oceans (Shackleton et al. 1983; Boyle and Keigwin 1985), in the terrestrial and marine biosphere, and in atmospheric trace gases as well as in continental ice cover. We will explore the interactions between these subelements of the Earth system that are involved in the glacial cycle. As we will suggest, the inferences that follow from this exploration provide useful insights into climate system function and sensitivity.

GREENHOUSE GASES AND CLIMATIC DETERMINISM

Before entering our discussion of climate system variability and specific feedbacks, it will be useful to review a number of general concepts concerning the behaviour of complex nonlinear systems. One of the more important concepts is connected to the question of the existence of multiple-equilibria. If a system is "transitive," for example, it is one which has a unique (perhaps statistical) equilibrium state for each set of boundary conditions, regardless of the initial conditions from which this end state is approached. To be specific, if the Earth's climate system were transitive and the atmospheric concentration of CO_2 were 300 ppm, then the same steady-state climate would obtain, whether this state was reached by reducing CO_2 from 450 ppm to 300 ppm or by increasing CO_2 from 150 ppm to 300 ppm. On the other hand, if the system is intransitive (or "nonergodic"), then it may posses more than a single equilibrium state for the same set of boundary conditions. In some climate models, for example, if the Earth is initially covered by ice, it will remain ice-covered even with the incident solar insolation fixed at its present-day value. On the other hand, if the Earth is initially ice-free and the solar intensity is similarly fixed, then the same model will reproduce the present-day climate. In other words, the model (or the Earth itself if its climate system is also intransitive) remembers its history, such that different equilibrium states may be realized even when the boundary conditions are fixed. Whether the actual Earth has a climate system that is transitive or intransitive is still being actively debated.

A specific current challenge to climatologists is to provide some insight into the changes of climate that might be expected in the future, in response to increasing levels of CO_2 and other greenhouse gases. In attempting to reduce the issue to one that is unrelated to the details of the actual greenhouse gas increase, it has become standard practice to ask what equilibrium climate might ensue in response to a doubling of atmospheric CO_2 concentration, all other features of the climate system remaining fixed. In particular, would this new climate state differ significantly from that which obtains today? In this circumstance, we would expect that the increased levels of CO_2 would induce an increase in the emission of infrared radiation from the stratosphere to the troposphere–surface system of about 4 W/m^2. This serves as a useful estimate of the magnitude of the associated climatic forcing, neglecting feedbacks associated with changes in stratospheric O_3, stratospheric temperature, tropopause height, tropospheric H_2O, clouds and other effects due to enhanced CO_2. The first-order issue that we must address is, then, whether a change in climatic forcing of 4 W/m^2 should be considered significant. In particular, can we provide insight into this problem based on information from the past?

On the basis of all the information at our disposal, it is clear that the climate system displays exceptional variability in both space and time on all scales. Over the past two million years, it is clear, in particular, that the changes in radiative forcing have been extremely modest. (For example, the June insolation variations at a period of 10^5 years associated with the change of orbital eccentricity have fallen from about 8 W/m^2 to about 4 W/m^2; annually averaged, of course, the variation in radiative forcing through time is vanishingly small). Nevertheless, the variations in global climate over this time have been extremely large. The most recent million years of climate history, for example, have been characterized by major episodic advances of continental ice sheets punctuated by relatively brief epochs of deglaciation, during which climate was relatively benign.

As we will discuss in much greater detail below, prior to about one million years ago the dominant period of variability in the climate system was the 41-Ka period of the obliquity cycle. Beginning about 900,000 years ago, a shift to a dominant period of 100,000 years occurred, an event to which many have referred to as the mid-Pleistocene climate transition. It is in this interval of time that the massive expansions and contractions of northern hemisphere land ice occurred, individual pulses of which were accompanied by rises and falls of global sea level in the range of 100–150 m.

In earlier periods (3–100 Ma ago), climate was apparently much warmer than at present. This is evidenced, as are the glacial cycles discussed above, by isotopic data from deep-sea sediments which indicate temperatures for the deep oceans as high as 15°C (e.g., Crowley and North 1991) during the Cretaceous (Brass et al. 1982). Even a cursory examination of the climate record, therefore, suggests that change, even profound change, has been the norm rather than the exception. As we will further elaborate below, the weight of the paleo evidence suggests that a change in the atmospheric concentration of CO_2 by a factor of 2 should be considered significant. However, the details of the climatic response to such a change of trace gas loading will depend critically upon a complex network of interacting positive and negative feedback loops. Circumstantial evidence from the past suggests, in particular, that the equable climates of the Eocene and Cretaceous periods were associated with levels of CO_2 that were between 2 and 3 times present levels (e.g., Crowley and North 1991).

THE 100,000-YR ICE-AGE CYCLE AND THE MID-PLEISTOCENE CLIMATE TRANSITION

No single observation of climate system behavior illustrates the importance of internal feedbacks in the determination of climate state better than the 100,000-year cycle of late Pleistocene ice volume, which has been so clearly revealed by oxygen isotope data from deep-sea sedimentary cores (Broecker and Van Donk 1970). Detailed analyses of the power spectra of $\delta^{18}O$ time series from such archives for the late Pleistocene (Hays et al. 1976; Imbrie et al. 1984) demonstrate the indelible imprint of the orbital insolation signal upon them. Since $\delta^{18}O$ is understood to be dominated by the signal associated with variations of continental ice volume (Shackleton 1967; approximately two-thirds of this variation is currently believed to be due to ice-volume change and one-third due to temperature change during the last half of Pleistocene time), and since the spectra reveal 95% significant lines at periods of 100,000 years (the dominant period for the past 900,000 years), 41,000 years, 23,000, years and 19,000 years, it should not be too surprising that such data are now widely accepted as verifying the essential idea in Milankovitch's theory of ice ages. The 100-Ka period is the dominant period on which changes in orbital eccentricity occur; the 41-Ka period is that which dominates the variation of orbital obliquity; and the lines at 23-Ka and 19-Ka periods are those that dominate changes of the precision parameter (e.g., Berger 1978). That the climate system's response to the seasonal insolation anomalies associated with changes in the Earth's orbit involve critical feedback effects, however, is made clear by the fact that the power spectrum of the seasonal insolation time series that is driving the variations of ice volume differs profoundly from that of $\delta^{18}O$. Although the ice-volume time series is dominated by the spectral line corresponding to a period near 100,000 yr, the summer seasonal insolation anomaly time series has essentially zero power at this period. Only by isolating the anomaly associated with a single summer month (June, say) may one find any significant signal at 100-Ka period in the incoming insolation signal that is apparently driving the ice-volume response. To illustrate this fact we show in Figure 16.1 (Hyde and Peltier 1985, 1987) the time series and power spectrum for the summer seasonal insolation anomaly at 65°N latitude (based on the analysis of Berger 1978).

A variety of different mechanistic models of the climate system have been proposed in an attempt to understand how this system converts the stimulus to which it is subject into a response that differs so radically from the stimulus itself. These models all embody some degree of nonlinearity, since it is rather difficult to imagine how any linear system could replicate the disparity between input and output documented above.

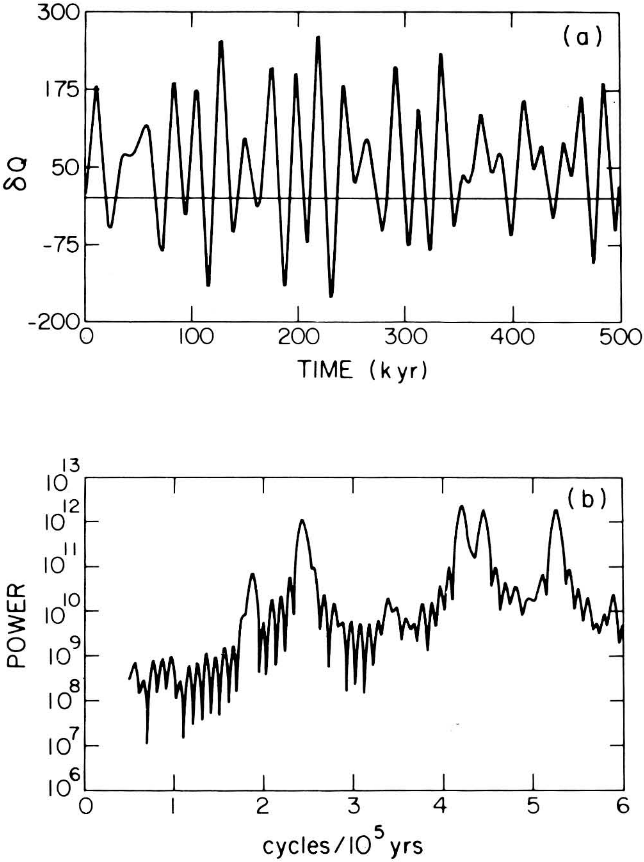

Figure 16.1 (a) The variation of caloric summer seasonal insolation at 65°N latitude over the past 500 Ka computed from Berger (1978a, 1978b). (b) The power spectrum of 1000 Ka of this insolation anomaly time series. Negligible power is observed at 100 Ka, while strong peaks are observed at 41 Ka, 23 Ka, and 19 Ka. From Hyde and Peltier (1985).

However, aside from explaining the origin of the 100-Ka component in the ice-volume variability, such models are also obliged to explain why this signal was essentially absent prior to about 900 Ka ago. This further fact is illustrated in Figure 16.2a (from Deblonde and Peltier 1991a), where the $\delta^{18}O$ record from ODP 677 (presented in Shackleton et al. 1990) is shown. The lower panels of this figure display a number of time-dependent statistics that have been employed to characterize the rather sharp transition in the record that occurs at mid-Pleistocene time. Even visually it is apparent that the 100-Ka cycle is essentially absent prior to 900 Ka ago. This fact is quantified in Plate I, in which we show the digital sonogram (time-dependent power spectrum) for ODP 667 on the time scale of Shackleton et al. (1990). Inspection of this figure

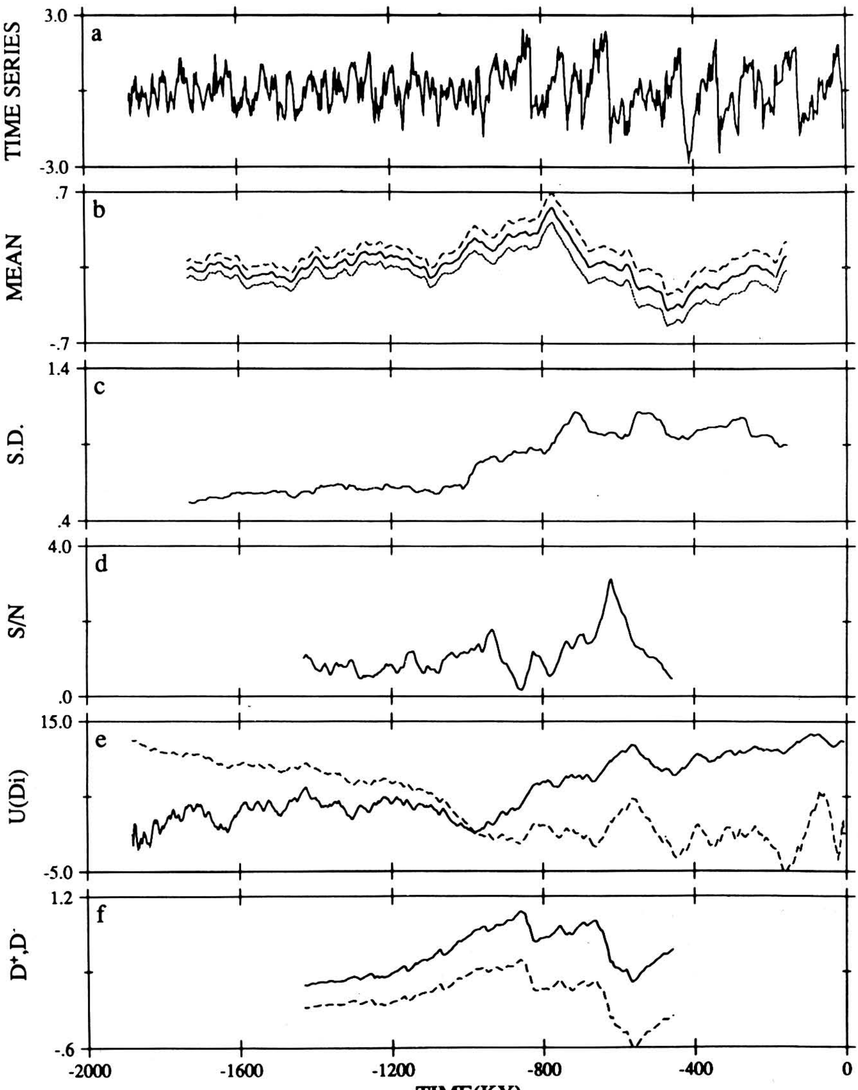

Figure 16.2 Statistical analysis of ODP site 677 $\delta^{18}O$ record with the depth-age model taken from Shackleton et al. (1990). (a) Time series with linear trend removed. (b) Running mean of the time series shown in (a) with segment length of 300 Ka. (c) Changing standard deviation of the time series shown in (a) computed for a segment length of 300 Ka. (d) Signal over noise ratio averaged over a range of segment lengths of 300 to 450 Ka (the confidence level is 99%). (e) Mann-Kendall rank statistic, $U(D_i)$, for forward (solid) and retrograde (dashed) time series. (f) Test for jump in variance, D^+ (solid line) and D^- (dashed line), averaged over the same segment length as in (d). From Deblonde and Peltier (1991a).

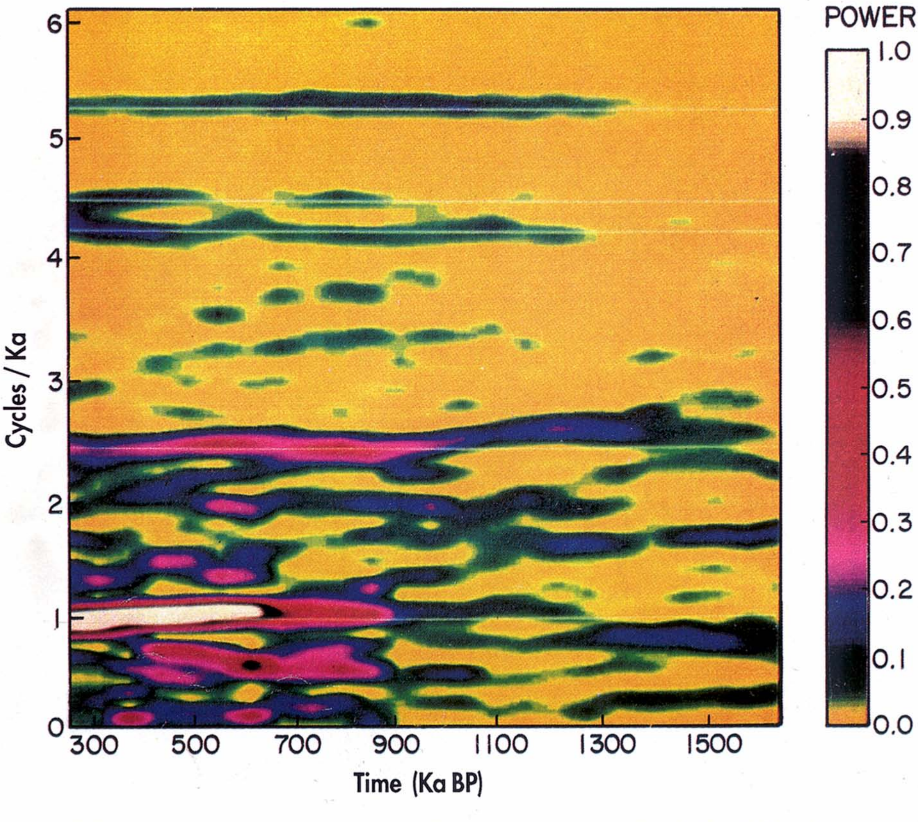

Plate I Time-dependent power spectrum (digital sonogram) of the ODP 677 $\delta^{18}O$ time series (part a of Figure 16.2) on the time scale of Shackleton et al. (1990). A segment length of 500 Ka has been employed for the purpose of computing the running spectrum.

reveals a number of further interesting aspects of the way in which the 100-Ka line rather suddenly appears as the dominant feature of the climate record, even though no substantial variation in the orbital insolation signal accompanies it.

There currently exist two relatively simple models that have been advanced to "explain" both the origin and the internal dynamics of the 100-Ka cycle. These two models, by Maasch and Saltzman (1990) and Deblonde and Peltier (1991a), "explain" the 100-Ka cycle as being due, respectively, to a free oscillation of the system that could exist even in the absence of the insolation stimuli and a forced oscillation that would die out if the stimulus were removed. The former model is of the so-called "dynamical systems" type, which is based upon a coupled set of nonlinear ordinary differential equations with only three degrees of freedom. The latter is based upon a set of coupled nonlinear partial differential equations that are derived from a two-dimension (zonally averaged) description of the interaction between an applied insolation anomaly time series, a large-scale ice sheet (whose nonlinear dynamics are described explicitly and whose mass balance is parameterized in terms of the applied insolation anomaly), and the underlying Earth, which is assumed capable of adjusting isostatically under the weight of the ice load (e.g., Peltier 1982). For the former model, the 100-Ka cycle may be induced to appear by hypothesizing the action on the climate system of a long time scale additional forcing, which causes the sudden appearance of the 100-Ka cycle at 900 Ka B.P.. In the latter model, the appearance of the 100-Ka cycle is associated with the activation of a feedback loop that impacts the mass balance only when the time rate of change of ice volume exceeds some threshold. Such a feedback loop was first hypothesized by Pollard (1982) as being associated with amplification of meltback due to calving into proglacial lakes. In the more detailed model of Deblonde and Peltier (1991b), the feedback is left generic (unspecified in detail as to its precise cause). It should be understood that in both the model of Maasch and Saltzman (1990) and that of Deblonde and Peltier (1991b), some additional forcing must be invoked beyond that associated with orbital insolation variations to understand why the 100-Ka cycle appears. We find the slow decrease of atmospheric CO_2 concentration that apparently has been occurring over the past several tens of millions of years (Crowley and North 1991) to be a good candidate for the nonorbital contribution to the forcing that is required in the context of either of these models to stimulate the onset of the 100-Ka cycle. The ability of such simple models to mimic the $\delta^{18}O$ time series (and therefore the variations of Pleistocene ice volume) is demonstrated in Figure 16.3. This figure presents a synthetic ice history from the model of Deblonde and Peltier (1991a) that may be compared to the $\delta^{18}O$ time series from ODP 677 previously shown on Figure 16.2. On the lower plates of Figure 16.3, the same statistical tests are applied to the synthetic record as were previously applied to ODP 677. Except for the apparent 400-Ka component of the model response in the earliest Pleistocene, which is not evident in ODP 677, the synthetic time series fits the observed time series rather well.

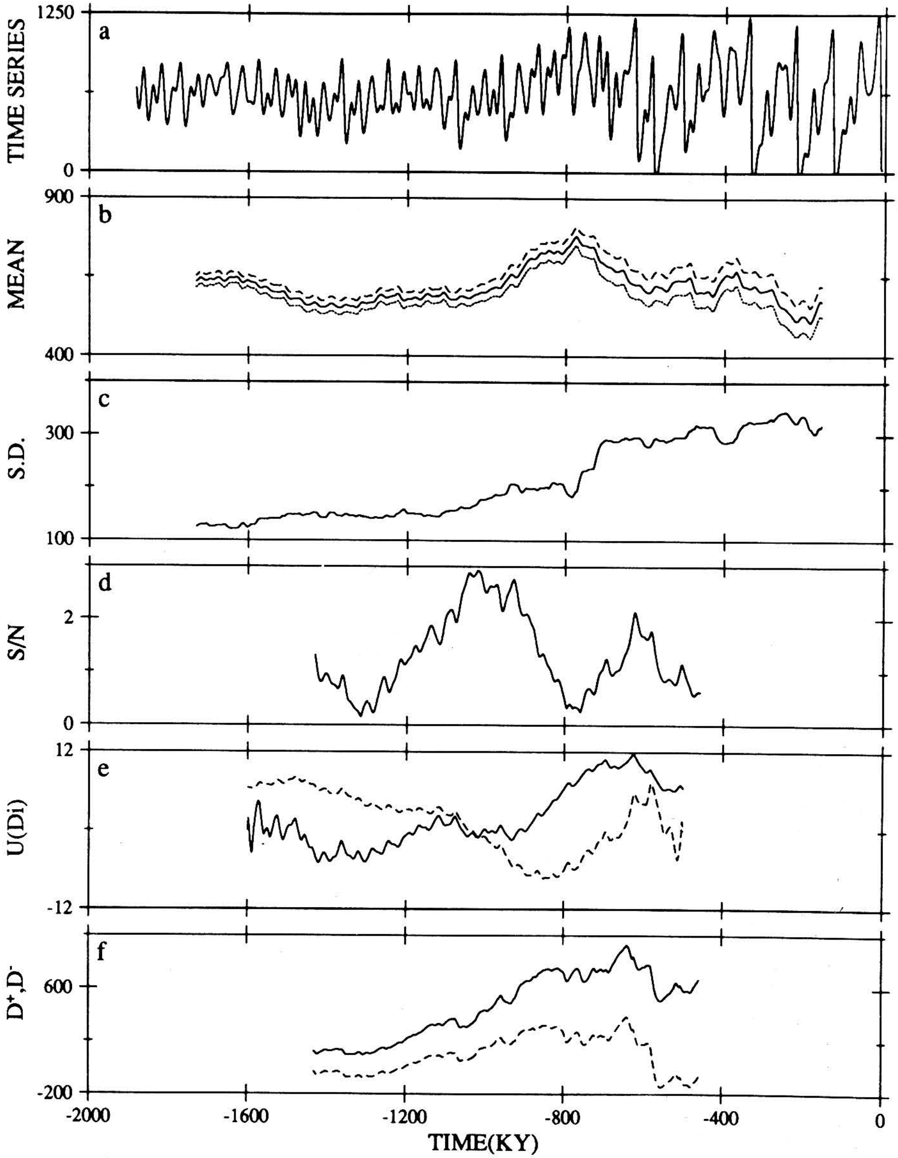

Figure 16.3 Same as Figure 16.2 but for a simulated ice volume time series (plate a) based upon the model of Deblonde and Peltier (1991a).

A number of more complex models of the 100-Ka cycle have also been proposed to describe more accurately the feedbacks that are operative within the system and which support the dominance of this cycle in late Pleistocene time. The model described by Gallée et al. (1991), for example, is based on a zonally averaged formulation in which, when it is subject to the Milankovitch variations of solar insolation, the predicted ice volume rises slowly and collapses rapidly in a way that is similar to that implied by the $\delta^{18}O$ data from deep-sea sedimentary cores (Labeyrie et al. 1987; Broecker and Van Donk 1970) and by the ice-core record from Vostok (e.g., Jouzel et al. 1987). In this model, the rapid retreats of continental ice that mark the terminations of each glacial cycle are driven by a sharp decrease of the albedo of the ice induced by the "aging" of the snow on its surface. The model of Deblonde and Peltier (1991b), on the other hand, is based on a two-dimensional global seasonal energy balance model due to North et al. (1983), in which separate subdomains are imbedded wherein the dynamics of the accumulation and flow of ice in response to applied orbital insolation anomalies are described separately. When it is initialized with ice-free (with the exception of Greenland) Northern Hemisphere conditions beginning 122,000 years ago (the last interglacial), this model successfully predicts the observed locations and the maximum thickness achieved by both the North American and the Northwest European ice complexes as implied by the geophysically inferred ICE–3G model (Tushingham and Peltier 1991, 1992). With the same minimal set of interacting physical ingredients as were employed in the simpler model of Deblonde and Peltier (1991a), however, the termination that commences near 21 Ka B.P. is not predicted. Just as in the simpler model, an additional feedback loop must be added. Recently completed additional tests have demonstrated that neither the observed variation of CO_2 (Barnola et al. 1987) nor the variation of high-latitude heating associated with the varying strength of the thermohaline circulation (see below) can by themselves induce the observed full deglaciations. One feedback, in the context of this model, has proven effective in the simulation of complete terminations is the modulation of ice-sheet albedo due to the intense rise in atmospheric dust concentration, which appears to be a generic feature of the paleo record prior to each major termination of ice volume (De Angelis et al. 1987). Although this influence is similar to that of the seasonally "snow aging" invoked by Gallée et al. (1991) to cause terminations, it is derivative of an entirely different cause. GCM experiments run with the GISS AGCM also suggest that high tropospheric dust loading at glacial maxima over the desert regions may have led to tropospheric warming at high latitudes and thus contributed to rapid terminations. How are we to decide which, if either, of these feedback loops is actually involved in the termination process?

We hope that this brief review of theories of the 100-Ka cycle will help to put in evidence the fundamental issues that are raised by investigations of climate system variability extracted from paleo records. In what follows we will "zoom-in" on the separate subcomponents of the global climate system, which paleo observations suggest were intimately involved in the most recent termination event of the present ice age. As we will attempt to make clear, fundamental issues remain concerning the relative timing of the imprint of termination upon these different subsystems, including the global thermohaline circulation of the oceans, the terrestrial biosphere, and the trace gas component of the global atmosphere. In each of these cases, the paleodata will be shown to shed considerable light on the functioning of the system, as well as upon the timing of variations in its level of "activity."

MODES OF CIRCULATION IN THE GLOBAL OCEAN

That there may be more than one single mode of deep circulation of the oceans is an idea first postulated by Stommel (1961) on the basis of a rather simple box model. Bryan (1986) later showed this same "intransitivity" to be characteristic of a very detailed oceanic general circulation model (OGCM) (see Duplessy and Maier-Reimer, this volume). The possible importance of this intransitivity to the understanding of late Pleistocene glacial history has been stressed by Broecker and Denton (1990), among others, especially in the context of the terminations mentioned above. This has spawned a small industry devoted to the construction of simplified models of oceanic behavior that might be suitable for application to the understanding of long time-scale climate variability (e.g., Marotzke et al. 1988, Marotzke 1989; Stocker and Wright 1991).

That much simpler hydrodynamic systems than oceans may exhibit such intransitive behavior has, of course, been well understood on the basis of the observed behavior of laboratory systems. The most famous of these is the differentially heated rotating annulus which, beginning with the very important early work of Fultz (1961), has continued to deliver extraordinarily useful insights into the behavior of nonlinear systems in general. With the outer cylindrical wall heated and the inner cylindrical wall cooled, the rotating system exhibits two principle regimes of behavior, which may be distinguished in the plane with the thermal Rossby number and the Taylor number as coordinate axes. At low rotation rates for all levels of heating, the flow driven by the differential heating takes the form of a zonally symmetric (Hadley) circulation. With the differential heating (intended in this laboratory system to mimic the differential heating between equator and poles to which the planetary atmosphere is subject) set at intermediate levels, and for sufficiently rapid rotation, the steady solution loses stability to a zonally wavy solution (via the hydrodynamic instability process), with the zonal wave number of the physically realized solution being higher, the lower the heating rate. When the heating rate is set to be close to the boundary separating two distinct wavy regimes, the flow is then observed to execute unpredictable excursions from the regime dominated by one wave number to the regime dominated by the adjacent wave number. This is an excellent example of a dynamical system evolving under the influence of a "strange attractor." Similar behavior might be anticipated for the dynamical behavior of the ocean as a whole in response to the buoyancy flux that drives its large-scale overturning.

The work of Bryan (1986) was the first to point out that large-scale numerical models of the oceans, which have traditionally been driven by the prescription of wind stress, temperature, and salinity at the surface, were capable of this behavior. He pointed out that such models were strictly intransitive in the sense that, when driven by freshwater fluxes, they may either remain in their existing state of circulation or flip into a completely different state when driven by the same freshwater flux, but initializing with a slight perturbation of the surface salinity. At the Max Planck Institute in Hamburg, the first experiments were conducted in which small amplitude fluctuating white noise was superimposed upon the main freshwater flux. These integrations demonstrated the existence of a relatively long period oscillation in the resulting response, especially in the Atlantic, which might possibly be connected to the fast $\delta^{18}O$ variations observed in Greenland ice cores. These experiments have also been reported by other groups and the same result derived.

The extent to which paleodata may be brought to bear on the issue of whether or not such variations in the mode of thermohaline circulation are actually observed is extremely interesting and constitutes a main thrust of modern research in the area of paleooceanography. In fact, two distinct modes of deep ocean circulation have been tentatively identified: that associated with the modern ocean (based upon both hydrological data and GEOSECS geochemical data) and that associated with the last glacial maximum (LGM). The latter mode of circulation is characterized by:

- reduced North Atlantic Deep Water (NADW) flow compared to Antarctic Bottom Water (AABW) flow (Curry et al. 1982; Streeter and Shackleton 1979; Boyle and Keigwin 1982);
- enhanced intermediate water flux in the three oceans (Duplessy et al. 1988; Duplessy and Labeyrie 1989);
- a sharp separation between intermediate waters more ventilated than today, and deep waters that are less ventilated than today (Kallel et al. 1988); and
- significantly lower deep water temperatures (Labeyrie et al. 1987). Among the major unknowns concerning the oceanic circulation at the LGM we must note especially the following: (1) the circulation of the Southern Ocean, especially concerning the areas of deepwater formation, deep convection, and surface-water sinking; (2) the location of the main zones of intermediate water formation in both hemispheres; and (3) the mode of intermediate water ventilation in the Pacific.

It is perhaps likely that modes of oceanic circulation other than the "full glacial" and "modern" modes, which have been recognized to date in the paleo record, also exist (Duplessy and Shackleton 1985; Boyle and Keigwin 1982; Boyle 1984), and it is important to realize that the mere recognition of differences between LGM and modern circulation patterns is insufficient evidence to prove bistability of the climate system. Paleooceanographic reconstructions have, however, shown that the modern mode of oceanic circulation also prevailed during the previous interglacial about

125 Ka ago (Duplessy et al. 1984). $\delta^{18}O$ and $\delta^{13}C$ analyses of benthic foraminifera demonstrate that both the temperature and dynamics of the circulation also changed at the end of the Eemian period about 115 Ka ago (CLIMAP Project Members 1984; Duplessy and Shackleton 1985; Boyle 1988a,b). However, the glacial mode of deep-water circulation prevailed only during the period from 65,000 to 15,000 years ago in the Atlantic ocean, i.e., during isotope stages 2, 3, and 4 (Boyle and Keigwin 1982; Duplessy et al. 1988). For the period of 115,000 to 65,000 years ago, the nature of the circulation is still poorly documented; however, available evidence shows that it exhibited characteristics intermediate between the modern and LGM modes (Duplessy and Shackleton 1985, Duplessy et al. 1988). We do not yet know, however, whether this corresponds to a well-defined and distinct circulation pattern or whether the ocean is in fact able to exhibit a continuum of dynamical states between the modern and LGM extremes. To model properly past circulation changes of the ocean using an OGCM, we require knowledge of the freshwater fluxes at the ocean–atmosphere interface (Broecker 1989b, 1990 Duplessy et al. 1991).

Concerning the circulation changes associated with the last glacial–interglacial transition, we note that over time scales of several centuries the atmospheric CO_2 concentration is slave to ocean circulation and chemistry (Broecker 1982, 1989a; Boyle 1986). Paleooceanographic data show that oceanic changes continued for several thousand years and that we are not yet able to describe accurately the complete set of changes that accompanied the last deglaciation. Nevertheless, based upon present evidence, we may safely assert that the first evidence of circulation change is provided by the observed difference in ^{14}C age between planktonic and benthic foraminifera from the same sediment level. This provides an estimate of the ventilation age of the deep water (Broecker et al. 1988, 1990; Duplessy et al. 1991). By this means, the ventilation age is observed to decrease in the deep and intermediate Pacific ocean shortly following the LGM, although we do not yet know the mechanism whereby the circulation rate is enhanced (Shackleton et al. 1988). Deep-water formation in the Norwegian Sea did not begin until much later, following the first phase of rapid deglaciation about 12,000 years ago (Sejrup et al. 1984). The modern thermal structure of the Pacific ocean was established only when the deglaciation event was essentially complete, namely about 9000 years ago (Labeyrie et al. 1990). A major effort should be undertaken in the future to understand more fully the processes by which the ocean shifted from its glacial to its interglacial mode of operation, by analyzing a set of high sedimentation rate deep-sea cores from the major ocean basins.

Another type of climate instability involves the small ice-cap instability (SICI), which is related to albedo discontinuities at the snow–ice transition. This phenomenon has been studied for over twenty years and a stability proof has been developed for some of the cases (e.g., Crowley and North 1989). It may be possible to test for this instability at certain intervals in the geologic record. For example, at the end of the last interglacial, ice volume started to increase while the North Atlantic was still in its present circulation mode. High resolution studies of unmixed $\delta^{18}O$ records might provide some evidence for abrupt transitions that implicate SICI or some other instability mechanism.

From a somewhat different perspective, we note that the role of the tropical oceans is frequently neglected in considering the influence of ocean circulation on climate, in spite of the very large surface area which they cover. This is clearly due to the fact, discussed above, that the polar oceans are the locations of the formation of deep water and thus the ultimate motors that drive the thermohaline circulation. Small changes in the tropical oceans may nevertheless be rather effective agents of climate change; for example, the climatic effect of the most important greenhouse gas (H_2O) increases exponentially with temperature, and the effect of phase transformation (a release of the heat of condensation) adds to the greenhouse effect.

During the recent past (the 41-year period, 1949–1989), the surface temperature in the Tropics has risen by 0.4°–0.5°C (see Flohn, this volume), which should have been accompanied by increases in evaporation, winds, the strength of the extratropical circulation, and the average temperature of the tropical troposphere (from 3–12 km altitude). Disregarding what triggered this process, this example demonstrates the important role that water vapor plays as an amplifier of climatic processes, one that could also have been effective during abrupt climatic change, as, for example, during the transition from full glacial to Holocene conditions. The Quaternary ice core described in Figure 19.11 (Flohn, this volume) provides information beginning from about 50 years prior to the most marked glacial advance of the Little Ice Age (following A.D. 1550) and documents a large increase of precipitation with an Atlantic moisture source. This moist period ends by about 1700, simultaneously with the arrival of the most intense period of the Little Ice Age in Europe. Later cold periods occurring until about 1850 were accompanied by a dry period in the Andes of Peru. This sequence of events might be taken to suggest that tropical ocean evaporation played a leading role in initiating the glacier response in the first half of the 16th century.

TERRESTRIAL BIOSPHERIC VARIATIONS THROUGH THE GLACIAL–INTERGLACIAL TRANSITION

Overpeck (this volume) summarizes recent research directed toward refining our understanding of the changes in continental vegetation that might be expected in a world characterized by double the present concentrations of atmospheric trace gases. This research builds upon a variety of prior activities directed towards an understanding of the changes in vegetation that accompanied the last glacial–interglacial transition. The point is made that possible future changes in surface vegetation might be of the same order as those that accompanied the glacial–interglacial transition itself. Significant variations in the terrestrial biosphere potentially involve significant changes in the quantity of carbon stored in this component of the Earth system and it is therefore important that we investigate whether and to what extent such changes may be associated with the change in CO_2 concentration that has been observed in the ice-core record (e.g., Raynaud et al. 1988) to have accompanied the glacial–interglacial transition. Because trace gas concentration is a determinant of climate state, terrestrial vegetation could be implicated in a significant climate feedback loop, making this an extremely important issue that we shall attempt to briefly review. It should also be recognized that the biosphere may influence climate through means unconnected with trace gases (see Overpeck, this volume), for example through changes in surface albedo and its role as a land surface stabilizer.

First, we note that there are large differences among the estimates of the change in terrestrial carbon storage (biomass, soil organic carbon, and peat) between the LGM and present (or more accurately, pre-forest clearance). K.C. Prentice and Fung (1990) estimated little change; Adams et al. (1990) estimated that C storage decreased by over 1000 Gt (i.e., a decrease of more than 30%), whereas the estimate based on an interpretation of the change in global ocean $\delta^{13}C$ is around 500 Gt (Duplessy et al. 1988; I.C. Prentice and Sarnthein, this volume). Any data-based estimate of LGM carbon storage (e.g., Adams et al. 1990) is difficult to constrain. For many areas, paleobotanical data are sparse and poorly dated. There is particular concern about Amazonia and Indonesia (including the exposed Sunda shelf), which are key areas in terms of possible changes in C storage. There are, fortunately, ongoing efforts to obtain ^{14}C-dated pollen records from these areas extending back to 18 Ka.

The estimate of Prentice and Fung (1990) uses a model approach (a GISS GCM experiment, with CLIMAP boundary conditions and a modified Holdridge climate–vegetation classification scheme) to predict the distribution of vegetation in regions at 18 Ka. Use of a model approach is probably unavoidable because we may never have enough data to obtain reliable estimates entirely from them alone. On the other hand, the Prentice and Fung results were not compared extensively with paleodata and may be inconsistent with some data (Adams et al. 1990).

There are also a number of general problems with all land-based estimates of 18 Ka C storage to date, namely:

1. They are sensitive to variations in the values assumed for the biomass and (especially) soil C storage in each vegetation type; these values themselves are not well constrained.
2. There is a major assumption involved in applying these values in the past, e.g., we know from pollen evidence that many vegetation types had a different species composition at 18 Ka, and, we might guess, different C fluxes and storages also.
3. Existing studies have not taken into account possible direct effects of the low CO_2 at 18 Ka. This would have meant generally lower net primary production and possibly also lower net ecosystem production and lower equilibrium C storage. However, it should be noted that there is no necessary relationship among these variables. Organic matter decomposition rates are also crucial and are limiting in cold climates.

4. It has been assumed that C storage is now and was also at 18 Ka and in pre-forest clearance time, in equilibrium with climate. This is adequate for biomass and *may* be a reasonable approximation for soil C storage (assuming residence time on the order of 1–2 Ka). However, it is not reasonable for peat, which has accumulated to the extent of perhaps 450 Pg during the Holocene.

The marine-based estimates of around 500 Pg C transfer cannot be definitively rejected, but these estimates may also be confounded, e.g., by a change in oceanic DOC (Siegenthaler's suggestion), an increase in continental plant $\delta^{13}C$ linked to the lower available pCO_2 (through possible changes in the ratio of C_3 to C_4 plants), and possibly other mechanisms (Prentice and Sarnthein, this volume). Despite all these uncertainties, it seems unlikely that the terrestrial biosphere could have been a net contributor to the CO_2 increase during the glacial–interglacial transition.

Group discussion emphasized the importance of real (i.e., pollen or plant macrofossil-based) and well-dated information of glacial-age vegetation in the Tropics. It is suspected that tropical forests may have been significantly changed at 18 Ka (see Overpeck, this volume). The evidence for this seems to be strong in Africa but less strong in Amazonia and Indonesia, where there is some evidence for aridity but not enough data to allow a quantitative assessment. The issue has been clouded by the acceptance of the "refugium hypothesis" to account for local centers of diversity in Amazonia; this hypothesis is questioned by many ecologists and paleoecologists. Discussion also emphasized the need for process-based modeling, not just of past climates but also of the ecological processes determining above- and below-ground C storage as a function of CO_2 and climate. Life is obviously not in equilibrium with climate. Major evolutionary events, such as the invention of carbon fixation and, later, of oxygen evolution, had enormous effects on the internal dynamics of the atmosphere–Earth system. Less tumultuous events continued as, for example, continents collided and terrestrial species from one invaded the other. It would therefore seem dangerous to assume that the biosphere will interact with the rest of the system in a manner which depends only on climate (CO_2), etc., and not on time, or evolutionary stage.

When marine fauna and flora are analyzed to infer past climate, a great deal of effort is expended in buffering the calibrations against "noise" in the abundance of individual species. This "noise" probably reflects complex microevolutionary changes in the competition between species which probably operate on all time scales and not only between individual species, but between major groups. This could well have a significant effect on climate and may represent a significant component of unforced low-frequency climatic variability.

FEEDBACK AND CONTROL OF THE CONCENTRATIONS OF CO_2 AND CH_4

Because the physical and chemical controls on the concentrations of CO_2 and CH_4 are distinct, though both trace gases are currently increasing in concentration and are strong contributors to potential greenhouse warming, it will be useful to address the issues concerning them separately. Once again we will look to the paleorecord for insights concerning the feedbacks in which these gases are involved.

Glacial–Interglacial Variations of CO_2

Polar ice studies have shown that atmospheric CO_2 was 80 to 100 ppm lower during the LGM than during the Holocene. There is a general consensus among carbon cycle modellers that oceanic processes must have been responsible for these glacial–interglacial CO_2 variations. Changes of the terrestrial biomass (perhaps excepting humus) went the wrong way, as discussed above. According to the majority of studies (reviewed briefly above), biomass most likely increased from glacial time to the Holocene and, therefore, CO_2 would have been withdrawn from the atmosphere–ocean system. This alone would have caused a drop of the atmospheric CO_2 level. The atmospheric CO_2 concentration is determined by the physical-chemical conditions in the ocean's surface waters (pCO_2, alkalinity, temperature, and salinity).

Temperature and salinity variations were too small to explain more than a 10 to 20 ppmv decrease in atmospheric CO_2. A main mechanism for changing the concentration of pCO_2 in surface water is the so-called biological pump, i.e., the downward transport of carbon by dead organic particles which then are oxidized at depth and returned to inorganic carbon (pCO_2). A stronger biological pump could therefore withdraw CO_2 from the surface waters and the atmosphere. The biological productivity is limited in the large oceanic regions by lack of nutrients (P, N) but not so in the high-latitude southern ocean; therefore, changes in that region have been suggested to be responsible for the CO_2 variations. For an extreme scenario in which all available surface nutrients would be used up by marine biology, carbon cycle model studies yield an atmospheric level of 150 to 170 ppmv (not considering land biomass changes). For the opposite extreme of a dead ocean (no biological carbon pump), the simulated CO_2 level is in the range of 450 to 550 ppmv. The observed LGM concentration is 180–200 ppmv, that is, near to the lower limit of the range that is possible if the strength of the biological carbon pump (but nothing else) is varied in carbon cycle models. If, based on a whole-ocean $\delta^{13}C$ change of ca. 0.3‰, the glacial-to-Holocene change in terrestrial biomass is assumed to be 500 Gt C, this alone would have decreased atmospheric CO_2 at LGM by roughly 30 ppm. To explain the LGM CO_2 level by the biological carbon pump would then mean that surface water nutrients would have to have been zero in the whole ocean. This would mean that $\delta^{13}C$ in surface water should have been higher ($\delta^{13}C$ being anticorrelated with the pCO_2 concentration) and the Cd/Ca ratio (a proxy for the nutrient phosphate), smaller. These predictions are, however, not confirmed by deep-sea sediment studies on planktonic foraminifera. Furthermore such an active biological pump would have created extensive zones with zero dissolved oxygen in disagreement with data from the sedimentary record. Therefore, the biological pump discussed above seems not to have been solely responsible for the observed change of CO_2. Some other mechanism must also be invoked to explain the observed low glacial CO_2 concentration. One possible mechanism might involve a change in the oceanic alkalinity budget, since lower CO_2 concentration can be caused by higher surface-water alkalinity.

The required alkalinity changes might themselves be caused by a stronger dissolution of carbonate sediments at depth, for instance, due to a different depth distribution of pCO_2 in the glacial ocean (more aggressive deep waters, as in the "bottom heavy" hypothesis of E. Boyle). Another conceivable way to change oceanic alkalinity would involve changes of the river input of organic and inorganic carbon to the oceans; this has not been studied in detail to date. A problem with this mechanism is that sediment-water interaction could change total-ocean alkalinity within a period of only several Ka, while ice-core data show that the CO_2 increase occurred early during the most recent two deglaciations, approximately simultaneously with the southern hemisphere warming and several Ka prior to the most recent change of ice volume (see Raynaud and Siegenthaler, this volume). To constrain oceanic alkalinity changes, more information on variations of the lysocline depth in different ocean basins is required. On the other hand, detailed model studies on the river input sedimentation cycle must also be performed.

According to current understanding, a combination of different mechanisms seems to have been responsible for the observed glacial–interglacial CO_2 variations. To identify these mechanisms, a number of questions will have to be addressed:

1. What were the global mean changes of ^{13}C in surface water and in the ocean as a whole in time? How did ^{13}C in atmospheric CO_2 change?
2. Is it possible that the Redfield ratios C/P and C/N in marine biomass were different during the ice age than today? What controls these Redfield ratios during the formation of biomass, and is there a fractionation of nutrients versus carbon during the remineralization?
3. Shifts between plankton communities producing $CaCO_3$ shells (e.g., coccoliths) and those not producing $CaCO_3$ shells (e.g., diatoms) may have an impact on oceanic alkalinity. Were there such shifts on a large scale, and by what factors are they controlled?
4. Can the changes in oceanic alkalinity be better constrained from sediment data? Could changes in the riverine input of dissolved and particulate matter have affected atmospheric CO_2? Note that during low sea-level periods, accumulation of river sediments occurred at great depths, not on the (then not water-covered) continental shelf, as today.
5. A significant fraction of the global new production occurs in relatively small, fertile areas of the ocean (coastal reprocessing zones). There is good evidence for a sizably larger ice-age productivity in these low-latitude regions. Could this have significantly influenced atmospheric CO_2?
6. A new method to reconstruct pCO_2 levels is based on ^{13}C analyses on sedimentary organic carbon. The basis for this method should be studied further, and potential applications should be investigated.

On one hand, global reconstructions of ocean paleoproductivity by means of accumulation rates of organic matter (likewise by means of planktonic foraminifera assemblages) show that low- and mid-latitude productivity was increased by a factor of 2–5 in the high-productivity upwelling belts during the LGM, except for the high-productivity belt offshore of southern Arabia. On the other, productivity generally was slightly decreased, by an average of about 20%, in the low-productivity subtropical gyres and in ice-covered portions of the ocean (Sarnthein et al. 1988, see also Prentice and Sarnthein, this volume). If one summarizes the areas with increased and decreased export productivity during the LGM, one arrives at a LGM-to-Present decrease by about 2–4 Gt C yr^{-1}, depending on different values of modern productivity published by various authors. The value of this difference contradicts most recent model calculations and is mainly based on narrow belts of the ocean surface.

Increased glacial productivity values were clearly linked to an increased nutrient reservoir in the surface ocean, as deduced from $\delta^{13}C$ values of near-surface dwelling planktonic foraminifera (*Globigerinoides ruber*); lower productivity values paralleled lower to equal nutrient concentrations.

In a core from right below the equatorial upwelling belt in the East Atlantic (2°S, 10°W), i.e., in clearly pelagic sediments, glacial productivity maxima (stages 2, upper 3, 4, 6) parallel distinct decreases in the CO_2 concentration of the surface ocean (top 10 m) by 10–20%. These variations have, for the first time, been reconstructed using ^{13}C data from the organic carbon in the sediment, which were corrected for a uniform UK-sea-surface temperature and a uniform nutrient (C = ^{13}C) content of the surface water, based on ^{13}C values of surface dwelling planktonic foraminifera (*G. ruber*).

The Glacial–Interglacial Variations of CH_4

The increase in the concentration of CH_4 over the last several centuries is even more spectacular than that of CO_2. Glacial levels of CH_4 were about 300 ppb: the interglacial value was about 600 ppb (Raynard et al. 1988, Stauffer et al. 1988, Chappellaz 1990), while the contemporary value is close to 1,800 ppb (Blake and Rowland 1986, 1988). Temporal changes in CH_4 can arise as a consequence of either a change in production or a change in loss. Previous discussions of the ice-core CH_4 record have stressed changes in sources (Chapellaz et al. 1990); however, our group felt that changes in sinks could be equally important. Production, according to present understanding, is dominated by microbially mediated processes on land (with contributions in the modern era from biomass burning and leakage of natural gas). Loss proceeds mainly in the atmosphere, by reaction with the hydroxyl radical, OH.

The concentration of OH itself depends on the column abundance of stratospheric O_3 (determining the flux of ultraviolet radiation penetrating to the troposphere), on the abundance of tropospheric O_3 (the source of O^1D), on the abundance of tropospheric H_2O (the source of OH through reaction with O^1D), on the abundance of CH_4 and CO (the primary loss reactions), and indirectly on species such as NO_2. Part of the rise of CH_4 in the contemporary atmosphere may be due to a temporal decline in OH, reflecting perhaps increases in CO due to enhanced incidence of biomass burning. Estimates of concentrations of OH, and consequently of production and loss processes for CH_4 during earlier times depend critically on assumptions made concerning the properties of O_3.

Measurements of CH_2O and formic acid (and if possible CO), found in air bubbles of ice cores in conjunction with CH_4, can provide an indirect estimate of the abundance of OH. It would be useful if such measurements could be carried out not only in ice from polar regions but also from mid-latitudes. The concentration of OH is expected to vary appreciably, both with latitude and season. The lifetime of CH_4 is set mainly by reaction with OH in the Tropics.

Studies of the isotopic composition of CH_4 can provide important clues that can be exploited to refine our understanding of the significance of the various possible sources. In particular, measurements of the isotopic composition of CH_4 at glacial time, and of its evolution from the last termination to present, would restrict the range of unknowns in present models.

The global budget of recent atmospheric methane, and particularly the relative importance of anthropogenic vs. natural sources, is still known only with great uncertainties. Most budget estimates have been constrained by the atmospheric methane sinks (oxidation by OH and uptake by soils) which seemed to be better known (Cicerone and Oremland 1988). However, just recently the OH sink has been revised by as much as 25% (Vaghijani and Ravishankara 1991; Crutzen 1991). Another constraint on the global methane budget has been derived from isotope (^{13}C, ^{14}C) observations on atmospheric methane and its sources. This, however, implies an exact knowledge of the kinetic isotope fractionation factor, which has been measured under laboratory conditions (Cantrell et al. 1990) but must be confirmed by isotopic measurements in the upper troposphere and lower stratosphere as well as in the Southern Hemisphere (Cassey et al. 1992) where the concentration variations are assumed to be mainly OH driven.

Concerning the interpretation of glacial–interglacial atmospheric methane concentration changes as derived from ice-core measurements, a much better knowledge of the recent methane budget, and particularly of its natural sources and sinks, is necessary. In this context, a closer look at sources and sinks that are of minor importance today but which might have been of relevance in the past (such as oceans and soils) should perhaps be emphasized. Bacterial methane production is strongly linked to anaerobic environments. Changes in these anaerobic conditions from the present-day situation to glacial times, with a different climate, may cause considerably different methane production rates as well as changes in the global distribution of the sources. Methane oxidation within source areas, already today, significantly influences the net methane flux into the atmosphere. It may have become even more important during glacial time. The processes involved in methane production, its final emission into the atmosphere, as well as the removal processes have to be studied in order to model the methane budget under modified climatic conditions.

The issue of microbial oxidation of methane in aerated soils as a sink for atmospheric methane has been investigated in several studies in recent years (Keller et al. 1983; Stendler et al. 1989; Born et al. 1990; Whalen and Reeburg 1990; Dörr et al. 1992). Dorr et al. (1992) found a strong correlation between soil permeability (soil texture class) and the methane uptake rate by soils, derived from parallel ^{222}Radon flux and ^{222}Radon and methane concentration measurements in the soil air at various sites in Central Europe. Methane decomposition in soils is basically controlled by the gas transport resistance within the soil, since soil temperature has been found to be of minor influence.The soils investigated in this study were classified with respect to their soil texture class; the mean observed methane uptake rates within each class were then used together with the NASA–GISS global digital data set of soil types (Staub et al. 1987) to estimate the global methane uptake rate by soils. With this parameterization, a global methane soil sink in the range of 9–56 Tg CH_4 yr^{-1} with a best estimate of 29 Tg CH_4 yr^{-1} was calculated. This corresponds to ca. 7% of the total recent global destruction rate by OH (Crutzen 1991). As the soil permeability strongly depends on soil humidity in a postulated drier climate during glacial periods, the total global uptake of atmospheric methane by soils may have been larger than in the present-day climate.

CONCLUSIONS

We have attempted in the preceding discussion to review the main issues upon which our group focused in attempting to come to grips with the ways in which proxy data records of past climates could be used to deepen our understanding of the main feedbacks operating within the climate system as a whole. The dominant signal in the climate record throughout the last 2 Ma of Earth history, especially in the time since 0.9 Ma, is that associated with the variation of continental ice volume, which may be accurately tracked using $\delta^{18}O$ data from deep-sea sedimentary cores. This signal is very strongly influenced by the variations of solar insolation associated with variations of the geometry of the Earth's orbit around the Sun. The dominant 100-Ka signal of the late Pleistocene, however, can apparently not be understood in terms of a linear model; nonlinear dynamics with strong coupling to parts of the climate system outside of the ice sheets themselves appear to be required to reconcile the observations. Many additional components of the Earth system appear to be deeply implicated in this oscillation of planetary climate. We have discussed three of the most important: the oceans, the biosphere, and the atmospheric load of radiatively active trace gases CO_2 and CH_4. Other components of the system, especially the atmospheric dust level, may have a very specific role, particularly in the physics of terminations. In providing a means to determine the relative phasing of the activity of these various subsystems during the glacial cycle, paleo records clearly provide a valuable key to the understanding of whole system function.

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