ICE AND CLIMATE MODELING: AN EDITORIAL ESSAY

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Abstract. The growth and decay of ice sheets are driven by forces affecting the seasonal cycles of snowfall and snowmelt. The external forces are likely to be variations in the earth's orbit which cause differences in the solar radiation received. Radiational control of snowmelt is modulated by the seasonal cycles of snow albedo and cloud cover. The effects of orbital changes can be magnified by feedbacks involving atmospheric CO_2 content, ocean temperatures and desert areas. Climate modeling of the causes of the Pleistocene ice ages involves modeling the interactions of all components of the climate system; snow, sea ice, glacier ice, the ocean, the atmosphere, and the solid earth. Such modeling is also necessary for interpreting oxygen isotope records from ice and ocean as paleoclimatic evidence.

The study of causes of the ice ages is now seeing vigorous activity for two reasons: First are the advances in quantitative mathematical modeling of atmosphere, ice, ocean, and solid earth that are now allowing the possibility of setting up models in which all of these components interact. Second is the growth in new sources of paleoclimatic evidence concerning the ice ages, especially from ice cores and ocean sediment cores. These cores have come up with great surprises, such as the intriguing variations of atmospheric CO_2 content and dust fallout rates that suggest formerly unsuspected feedbacks in the climate system. During the next year there will be at least two conferences on causes of the ice ages: "Milankovitch and Climate, Understanding the Response to Orbital Forcing" in December 1982 at Lamont-Doherty Geological Observatory, and "Ice and Climate Modelling" in June 1983 at Northwestern University.

The causes of the Pleistocene ice ages have been a popular subject for speculation, for example in the classic work of Brooks (1927), as well as by Flohn (1974), Kellogg (1975), and Hughes (1982), among others. Now, however, many of the theories can be quantified by writing down equations that approximate them into a computer model. Models designed to look at changes in climate over thousand-year time scales must include processes that are held fixed in weather prediction models or seasonal forecast models, such as ice sheet sizes, crustal depression and rebound, and deep ocean temperatures. Because of the long times involved, they cannot take the 10-min time steps of a general circulation model to follow the atmospheric circulation in detail, and must 'parameterize' the effects of processes that act on time or space scales shorter than the smallest time step (or spatial grid) of the model. Climate models attempting to calculate the course of the ice ages vary greatly in their choices of parameterization. These models are reviewed to some extent in this issue of *Climatic Change* by Budd (1982) and Oerlemans (1982a). Most of these modeling efforts now try to include the known effects of variations in the earth's orbit on solar radiation, which were shown by Hays *et al.* (1976) to correlate with the paleoclimatic record of ice volume. These models have mostly used either a simple ice

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flow model and a more detailed atmospheric model (e.g. Pollard, 1978) or a detailed ice flow model with atmospheric effects entering only as parameterized accumulation and ablation rates. For example, the model of Budd and Smith (1981) uses two adjustable parameters which control the strength of the radiational forcing and which are chosen for "best fit for matching the historical growth and retreat of the ice sheets". It thus implicitly assumes that the ice ages *were* forced by orbital variations and is not well designed for actually testing that hypothesis. (For example, suppose the orbital variations merely *correlate* in time with the ice volume changes but are of insufficient magnitude to *cause* them. The model might not be able to make this discovery.) Others of these models, such as that of Oerlemans (1982a), exhibit 'free oscillations', meaning that the model climate can oscillate with no change in external conditions. This oscillatory behavior is usually found only for certain values of model parameters whose realistic range has not yet been pinned down, so it is not yet clear whether the real climate can exhibit free oscillations. For example, Harvey and Schneider (1982) showed that the oscillations in the model of Kallen *et al.* (1979) disappeared when a more realistic description of oceanic thermal inertia was used, allowing the ocean surface temperature to change more rapidly than the deep ocean temperature.

Climate models are usually broken into a number of components (atmosphere, ocean, land ice, sea ice) each of which can be modeled separately by *specifying* the interactions with the other components. I will discuss some of the current problems in modeling each of the components separately, as well as some of the interactions (feedbacks) which should appear when they are put together in a 'coupled' model.

Snowmelt

It is worth mentioning some of the details of initiation and growth of ice sheets which so far have been hidden in the parameterizations of coupled models. The problem of ice sheet initiation is well-illustrated by Figure 1 of Adams and Rogerson (1968), showing the snow depth at Knob Lake. This location is close to the site of final disappearance of the Labrador-Ungava ice dome about 6000 yr BP, shown in the maps of Denton and Hughes (1981). The snow depth builds up steadily from September to April, then melts rapidly in a 6-week period, leaving the ground snow-free for nearly four months. Now what would be required to melt the snow so much more slowly that some would still remain at the end of the summer? Williams (1979) has studied this question and found glacial inception in Northern Canada rather difficult, requiring a summer temperature decrease of 10–12 K (or its radiation equivalent). This same difficulty was encountered in the models of Budd and Smith (1981) and Birchfield *et al.* (1982) which thus emphasized the need for a region of high elevation to act as a 'seed' for the ice sheet. Koerner (1980) has however pointed out that "the extension (in time) of the annual period of snow cover generally is more important to the feedback process (by increasing albedo) than the specific lowering of the equilibrium line", suggesting that "a decreased variability of summer climate, and hence the disappearance of 'anomalously' warm summers, may be an integral part of the glacierization process".

Snowfall

However necessary solar radiation changes affecting snow melt may be for the ice ages, they cannot do everything. Precipitation changes must also be considered, as discussed by Loewe (1971) and Brinkmann and Barry (1972). At the center of the Keewatin Dome west of Hudson Bay, which once had an ice thickness of 4260–4960 m (in the minimal and maximal reconstructions of Denton and Hughes), the mean annual precipitation is now only 25 cm. So even if this all fell as snow, never melted, and did not flow, it would still take 17 000–20 000 years to build up such a dome. Where the extra precipitation came from is a difficult puzzle. In the nearby Alaskan tundra, the evidence is for *drier* conditions than today (Barry, 1982), as would be expected if the Arctic Ocean had a shorter melt season 18 000 years before present (k yr BP) than now. It may have had almost no melt season at all, if it was covered with ice shelves instead of sea ice as Denton and Hughes suggest. The Atlantic may have been the main source of moisture. Ruddiman *et al.* (1980) showed that the North Atlantic between 40° and 60° N maintained warm sea surface temperatures (within 1–2 K of today's temperatures) during the first 10 000 years of the Northern Hemisphere ice sheet growth beginning about 75 000 years ago.

However, there actually may be little need for increased precipitation. There is not general agreement that the Keewatin ice thickness reached even the 'minimal' size of Denton and Hughes; Andrews (1982) reviewed other reconstructions of the Laurentide ice sheet and showed that isostatic sea level data suggest a thickness of only 1800–2200 m at 18 k yr BP, and this may be a partial solution to the 'puzzle'. Andrews also summarized evidence from glacial geology to make a severe criticism of Denton and Hughes' ice-sheet reconstruction.

Snow and Ice Albedo

The positive climatic feedback caused by high snow albedo has been emphasized in most models. But snow albedo is quite variable, varying between about 50 and 90%. Koerner (1980) has emphasized that snow albedo varies gradually in time (season) and space in the tundra regions where the Laurentide ice sheet might start. The reasons for this variability of snow albedo are explained by Wiscombe and Warren (1980) and Warren and Wiscombe (1980), who did radiative transfer calculations to find the effects on albedo due to snow age, depth, sun angle, cloud cover, and impurities. Sea ice albedo undergoes an even larger seasonal cycle than does snow albedo, varying from 80% for snowcovered ice or 60% for bubbly ice to 30% for puddled ice (Langleben, 1971). The fact that these albedos drop systematically during the melting season may be just as important as the fact that they are high: Birchfield and Weertman (1982) found their climate model much more sensitive to orbital changes with the more realistic variable snow albedo than with a fixed albedo. (But even they did not allow for variable sea-ice albedo, so when this is included the model may be even more sensitive.) The Antarctic ice sheet is not necessarily a good example of conditions at the surface of the North American ice sheet. Whereas the Antarctic snow does not melt, and its albedo varies little through the year, the southern portions of the Laurentide ice sheet undoubtedly underwent a seasonal cycle, since the albedo of melting snow is lower than that of new snow. Benson (1967) estimated that the dry snow zone on the North American ice sheet extended no farther south then 60° latitude.

Clouds

But snow albedo is not all-important. For the planetary albedo, cloud variations are probably at least as important as snow variations. The change from spring to summer in the Arctic Ocean sees snow-covered sea ice under clear sky replaced by melt-pond-covered sea ice under stratus clouds, with little change in the planetary albedo. The cloud conditions may well have been different during the glacial maximum. One likely negative feedback on temperature is that a colder atmosphere containing less water may result in optically-thinner clouds with reduced albedo, as suggested by Paltridge (1980) and Charlock (1982). Changes in cloud amount and cloud top temperature are probably harder to predict but are no less important (e.g. Schneider *et al.*, 1978). Climatic feedback involving clouds thus still remains one of the biggest uncertainties in climate modeling. As such, the subject has been deemed worthy of a four-day symposium at the International Union of Geodesy and Geophysics meeting in Hamburg in August 1983. Cloud-climate feedbacks will eventually also have to be included in models of the ice ages.

Carbon Dioxide

Other feedbacks are now appearing which are likely to affect the *magnitude* of glacial-interglacial climatic changes, however these changes are initially forced. The lowered CO_2 content of the atmosphere toward the end of the Wisconsin (Neftel *et al.*, 1982) very likely helped cool the climate, and it may well have been the result of the glaciation itself. Broecker (1982) showed how this might happen. Deposition of phosphate-rich biological sediments on shelves when sea level rises, and erosion of these shelves when sea level drops, causes the phosphate content of ocean surface water to be higher during glacial times when sea level is low. The biological productivity of surface water, limited by phosphate, would thus be higher in glacial times and would fix more CO_2, reducing the CO_2 concentration in surface waters and thus also in the atmosphere. Another mechanism for increasing atmospheric CO_2 as sea level rose is that of Berger (1982): coral reef growth precipitates $CaCO_3$ from bicarbonate, thus acidifying ocean surface water and releasing CO_2.

Deserts

The cooling of the oceans during the ice age probably led to a weakened hydrological cycle at glacial maximum, according to the general-circulation-model experiments of Manabe and Hahn (1977). This result is generally supported by evidence of fossil dunes

(Sarnthein, 1978) which indicates greatly expanded deserts in Asia, Africa, and Australia 18 k yr BP. In contrast to this global pattern, the North American desert was certainly wetter at that time, a puzzle that Manabe and Hahn were unable to explain. The expanded deserts had higher surface albedo than does the vegetation that replaced them, a possible positive climatic feedback. The dust from these deserts was blown about the world and found its way to the great ice sheets. The dust content of Antarctic ice (in the Byrd Station core) was 5–10 times Holocene levels, and that of northern Greenland (Camp Century) was up to 100 times present levels (Thompson, 1977). That enhanced amount of dust at Camp Century ~~is still a factor of 10 too small to have even a 1% effect on~~ could reduce snow albedo by 1-5% at the most sensitive wavelength. The dust deposition rate was probably higher in the lower-latitude portions of the Northern Hemisphere ice sheets, but the snow accumulation rate was also higher, so the dust content may still not have significantly reduced the snow albedo. If it did, this dust could lead to a negative feedback on the ice advance similar to one suggested by Kellogg (1975).

Antarctic Ice Sheet

Budd and Smith (1982) pointed out the need to include the Antarctic ice sheet in any model that tries to reproduce ice-volume records, because its size responds to Northern Hemisphere climatic changes. Its response time (e-folding time ~ 30 000 years) is longer than that of the North American ice sheet (~10 000 years), so that its growth and decay can be out of phase with Northern Hemisphere ice. Melting is negligible on the Antarctic ice sheet, so its size is limited by sea level; it can expand onto the continental shelf when sea level drops due to Northern Hemisphere ice growth (Figure 4 of Oerlemans, 1982b). Proper modeling of the ice flow will be aided when accurate bedrock topography is available; only about one-half of the continent has so far been surveyed by radio-echo sounding (Drewry *et al.*, 1982). It is difficult to measure whether Antarctica is presently gaining or losing ice mass, partly because of the uncertain amounts of melting and accreting ice on the bottom of ice shelves which are being measured by Zotikov *et al.* (1980). Etkins and Epstein (1982) use the current rising sea level and increasing length of the day to estimate that Antarctica is losing mass quite rapidly: an average over the continent of 8 cm yr^{-1} for the last 40 years. This has not been confirmed by more direct observations. Although the Antarctic ice sheet's rough symmetry about the pole suggests simple treatment in models, the flow is by no means radial (Table 3.2 of Budd *et al.* (1971)) and probably requires rather detailed modeling. It is now apparent that most of the Antarctic ice drains through narrow fast-flowing *ice streams*. For example, the Byrd Glacier, surveyed by Hughes and Fastook (1981), is only 25 km across but drains 1–3% of East Antarctica, flowing at nearly 1 km yr^{-1}.

Oceans

Another major uncertainty in modeling the ice ages is the role of the oceans, recently reviewed by Rooth (1982). He showed that the growth and decay of ice sheets should cause salinity anomalies that can greatly affect oceanic circulation patterns, both directly and also through the changing pattern of aridity mentioned above. For example, a decrease of freshwater runoff from North America as the ice built up would increase the salinity of the North Atlantic, allowing enhanced thermohaline circulation. This could explain the warm sea-surface temperatures found by Ruddiman *et al.* (1980) which were mentioned above. Another potentially important oceanographic process is one mechanism of bottom water formation in the Weddell Sea which is apparently dependent on the geography of one submarine canyon (Foldvik, 1982). Conceivably, this mode of production of bottom water might be altered with a drop in sea level and expansion of Antarctic continental ice. The ocean thermal structure also plays a crucial role in modeling the seasonal cycle of sea ice.

Sea Ice

Sea ice undergoes a large seasonal cycle and exhibits large interannual variations in its extent. Because it also contributes greatly to the albedo-temperature feedback, it is a very sensitive part of the climatic system. Manabe and Stouffer (1980), for example, found the sea ice region to exhibit the largest temperature changes of any geographical region when the atmospheric CO_2 was quadrupled in a general circulation model. The variations in sea ice concentration and extent are now being monitored by observations from satellite of microwave emission from the ice (Zwally *et al.*, 1982). Modeling of the growth and decay of sea ice has been reviewed by Hibler (1980). Both modeling and observational studies have focussed more on the Arctic than on the Antarctic sea ice, so it is worth pointing out that they do differ significantly.

At its maximum extent, Antarctic sea ice covers an area as large as South America. Unlike the Arctic ice, nearly all of it melts away during the summer. The divergent flow of ice away from the Antarctic continent means that there is more open water than in the Arctic. The prolific algal population in the Antarctic ocean visibly colors the snow on top of sea ice (Meguro, 1962). There is also a striking lack of melt puddles on bare Antarctic sea ice, observed consistently in three different sectors by Meguro (1962), Spichkin (1966), and Ackley (1979). Spichkin attributed this to a low relative humidity of the overlying air. Andreas and Ackley (1982) support Spichkin's contention with some simple energy-budget modeling, but it seems that a weaker radiational forcing (because clouds are thicker in the Antarctic Ocean than in the Arctic Ocean) could also contribute. In any case, the absence of puddles in the Antarctic probably keeps the sea-ice albedo up to about 60%, as opposed to 30–40% for puddled Arctic ice.

There have been a few attempts to model the seasonal cycle of Antarctic sea ice, especially to explain the rapid melting of the ice from November to January. Recent controversy has focussed on the heat flux from the ocean to the ice, but these comprehensive models have other defects as well. Parkinson and Washington (1979) assumed a maximum snow depth of only 3 cm, which is a factor of 10 too small (Meguro, 1962; Petrov, 1967; Ackley, 1979). Pease (1975) failed to get the sea ice to melt rapidly enough in her model, no doubt partly because she kept the albedo fixed too high (80%). Gordon (1981) was able to obtain rapid retreat of the ice in a thermodynamic model only by assuming a large average heat flux of 30 W m^{-2} from the ocean to the ice. Hibler and Ackley's (1982) dynamic-thermodynamic model, by contrast, required only 2 W m^{-2} for this heat flux, but they were simulating only the Weddell Sea, which is anomalous because of its preponderance of multi-year ice. Allison (1981) and Allison *et al.* (1982) have now measured these heat fluxes for each month of the year, but only very close to the coast, finding an annual average of 10–15 W m^{-2}, much larger than the 2 W m^{-2} found in the Arctic. The measurements to date have been in 'anomalous' regions, either at the coast by Allison or in the Weddell Sea by Ackley. What is now needed are measurements of the sea-ice energy budget in the main Antarctic pack, from a ship or an ice island.

Models will have to be accurate in predicting ice thickness distribution and ice areal coverage. This is because air-sea heat exchange and absorption of solar radiation occurs mainly over leads (Andreas, 1980) and thin ice (Maykut, 1978), which may cover only a small fraction of the sea-ice domain.

Antarctic Ice Surges

The amount of oceanic heat flux available to melt *icebergs* will also be crucial in any attempted modeling of the climatic effects of Antarctic surges. Flohn (1974) thought that a complete surge into the ocean of one of the West Antarctic drainage basins (about 5% of the total Antarctic ice mass) over a 25 year period could trigger a Northern Hemisphere ice age. The ocean covered with a bright snow surface was thought to cool the climate sufficiently to initiate an instantaneous glacierization. Flohn probably exaggerated these effects by assuming an albedo of 75% for the icebergs, characteristic of new snow. As the icebergs move north and start to melt and break up, their albedo would actually drop to that of bubbly ice, at most 60% albedo. More importantly, Flohn ignored the fact that most of this ice would be covered by the thick clouds of the Southern Ocean and so have little effect on planetary albedo. The surge also may proceed more slowly than Flohn assumed. Most of Antarctica is drained by fast flowing ice streams. Weertman and Birchfield (1982) pointed out that these streams are already moving at surge velocities; they may not be able to speed up much more. (For the same reason, they doubt that a catastrophic West Antarctic surge can occur at the *present* time. Such a surge, causing a rapid rise of sea level, had been suggested (Mercer, 1978) as a consequence of the warming induced by an enhanced CO_2-'greenhouse'.) However, there is ample evidence in the paleoclimatic record for rapid climatic changes in the past (indeed, this was the topic of this year's meeting of the American Quaternary Association), so a quantitative modeling of the climatic effects of surges would be worthwhile.

Climate 'Transitivity'

The Milankovitch mechanism has been especially popular because it involves a known climatic forcing – variations in solar isolation due to small changes in the Earth's orbit – and (unlike Flohn's mechanism) does not require the climate to be 'intransitive' (possess more than one stable state for a given set of external conditions). Other forcings which can cause climatic change in a 'transitive' climatic system are sustained changes in solar output and changes in atmospheric content (e.g. CO_2, volcanic ash). [However, the atmospheric CO_2 content, as mentioned above, may not be an 'external' cause but may actually be affected by ice sheet sizes. The same may be true for the level of volcanic activity, as suggested by Rampino *et al.* (1979). Gow and Williamson (1971) found much more volcanic ash in the Byrd Station core in ice corresponding to the last glacial maximum. A nearby source was indicated, suggesting that the Antarctic volcanoes were more active at that time, possibly responding to the increased ice load (Kyle *et al.*, 1981).] In contrast to these sustained causes of climatic change, *impulsive* forcings (ice sheet surges, single volcanic explosions, meteorite impacts) which are not maintained indefinitely will have no long-lasting effect unless the climate is intransitive. Even for a 10-km diameter meteorite impact that may have marked the Cretaceous-Tertiary boundary 65 million years ago, Toon *et al.* (1982) find in a simple climate model that, even though so little sunlight reaches the surface that the continents freeze for up to two years, the climate soon recovers after the dust falls. The reason for this resilience is the seasonal cycle. All it takes to destroy the incipient 'instantaneous glacierization' is one summer. Most proposed impulsive forcings are of smaller magnitude than the seasonal change in solar radiation. One thus suspects that some of the mathematical climate models which exhibit intransitivity such as 'free oscillations' may cease to do so when they include a realistic seasonal cycle.

Oxygen Isotopes

Climate modeling of the ice ages is ultimately tested against the paleoclimatic evidence, especially the oxygen-isotope ratios in ocean sediments and ice cores. The oxygen-isotope index $\delta^{18}O$ in fossil benthic foraminifera in ocean sediments is often interpreted as a direct measure of ice volume on land, but it is also affected by changes in deep-ocean temperatures and by variable $\delta^{18}O$ of the various ice sheets. Broecker (1978) estimated (from the discrepancy between isotope records and sea level records) that bottom-water temperatures are 2 K warmer than normal in peak interglacial time (~120 k yr BP and present). Budd (1981) estimates that a change in Antarctic ice volume gives three times the 'ice volume' signal in the ocean as does the same change in North American ice volume. This is because the average $\delta^{18}O$ of Antarctic ice is $-45^{o}/oo$, and Budd assumed $-15^{o}/oo$ for Laurentide ice. However, Dansgaard and Tauber (1969) showed, by analogy with the $\delta^{18}O$ distribution in Greenland, that Laurentide ice should have $\delta^{18}O = -30^{o}/oo$ if the Laurentide ice sheet was as high in elevation as the Greenland ice sheet is today, so Budd may have overstated the contrast. Budd's small δ-value for Laurentide ice may be correct if the Laurentide ice sheet was much thinner than the Greenland ice sheet, as suggested by the crustal rebound rates (Andrews, 1982) referred to above.

Although the $\delta^{18}O$ of snow correlates rather well with mean temperature at the snowfall site, it actually depends on the fractionation of oxygen isotopes along the path from ocean to ice sheet. The correlation with temperature may be quite good at sites on the high ice plateau, but Kato (1978) has shown that the $\delta^{18}O$ of falling snow at a coastal site varies depending on the moisture source region. In particular, Weaver and Bromwich (1982) showed that the $\delta^{18}O$ of snow at the Antarctic coast depends on sea ice extent. Because ice-age modeling is tested against $\delta^{18}O$ records, it will be important to use atmospheric transport models, glacier flow models and ocean circulation models not only to model the climatic response to orbital forcing but also to keep track of the oxygen-isotope budgets of the ice and the ocean and to model the isotopic fractionation in the atmospheric part of the water cycle. The complete analysis will be further complicated by taking into account loss by sublimation, which amounts to half of the annual snowfall on the Antarctic slope (Fujii and Kusunoki, 1982), but less elsewhere.

Dating of Ice Cores

The dating of ice cores is mostly done by stratigraphic methods. Thompson and Mosley-Thompson (1981) find an annual cycle in dust concentration that they use for dating. Wilson and Hendy (1981) used the annual cycle in sodium concentration to date the Vostok core and concluded that the annual accumulation at the last glacial maximum was half the present value. Lorius *et al.* (1979) correlated the $\delta^{18}O$ record in the Dome-C core with the oceanic $\delta^{18}O$ record of sea-surface temperature to date the ice core, and similarly concluded that the annual accumulation before 14.4 k yr BP was 75% of its present value of 3.7 cm yr^{-1}. These stratigraphic methods give excellent temporal resolution (~1 yr) in regions of high accumulation. For low-accumulation regions there can be missing years in the stratigraphic record. In these regions the stratigraphy is now being supplemented by the accelerator-mass-spectrometer measurements capable of detecting small quantities of ^{14}C in the CO_2 in air bubbles within the ice. ^{14}C can thus be useful for dating ice a few tens of thousands of years old. For even older ice in the cores of East Antarctica, methods are being developed which use longer-lived radioisotopes (B. Stauffer, personal communication). We can soon expect much longer climatic records from the ice. The new results of oxygen-isotope analysis to 1400 m in the Vostok ice core (Gordienko *et al.*, 1982) show a signal that may correspond to the last interglacial about 100 k yr BP. The ice has now been penetrated to 2038 m, but not yet analyzed, and there appears to be no impediment now to continue to the bedrock of East Antarctica.

In summary, there is now rapid progress being made in paleoclimatic evidence, climate modeling, and ice sheet modeling. Success in climate modeling of the ice ages will require collaboration among these disciplines, because it involves careful attention to details of physical processes in a variety of geophysical systems: snow, sea ice, glacier ice, the ocean, and the atmosphere.

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