# Ice Ages and the Thermal Equilibrium of the Earth, II

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**Abstract**

The energy required to sustain midlatitude continental glaciations comes from solar radiation absorbed by the oceans. It is made available through changes in relative amounts of energy lost from the sea surface as net outgoing infrared radiation, sensible heat loss, and latent heat loss. Ice sheets form in response to the initial occurrence of a large perennial snowfield in the subarctic. When such a snowfield forms, it undergoes a drastic reduction in absorbed solar energy because of its high albedo. When the absorbed solar energy cannot supply local infrared radiation losses, the snowfield cools, thus increasing the energy gradient between itself and external, warmer areas that can act as energy sources. Cooling of the snowfield progresses until the energy gradients between the snowfield and external heat sources are sufficient to bring in enough (latent plus sensible) energy to balance the energy budget over the snowfield. Much of the energy is imported as latent heat. The snow that falls and nourishes the ice sheet is a by-product of the process used to satisfy the energy balance requirements of the snowfield. The oceans are the primary energy source for the ice sheet because only the ocean can supply large amounts of latent heat. At first, some of the energy extracted by the ice sheet from the ocean is stored heat, so the ocean cools. As it cools, less energy is lost as net outgoing infrared radiation, and the energy thus saved is then available to augment evaporation. The ratio between sensible and latent heat lost by the ocean is the Bowen ratio; it depends in part on the sea surface temperature. As the sea surface temperature falls during a glaciation, the Bowen ratio increases, until most of the available energy leaves the oceans as sensible, rather than latent heat. The ice sheet starves, and an interglacial period begins. The oscillations between stadial and interstadial intervals within a glaciation are caused by the effects of varying amounts of glacial meltwater entering the oceans as a surface layer that acts to reduce the amount of energy available for glacial nourishment. This causes the ice sheet to melt back, which continues the supply of meltwater until the ice sheet diminishes to a size consistent with the reduced rate of nourishment. The meltwater supply then decreases, the rate of nourishment increases, and a new stadial begins.

# Introduction

It has long been recognized that ice ages correspond to major perturbations of the energy budget of the earth. A common assumption has been that some external disturbance(s) produces continental glaciations, but it has also been realized that large ice sheets must produce major climatic effects of their own.

While most ice age theories are concerned primarily with "trigger factors" responsible for the initiation of glacial and interglacial intervals, this model is concerned mainly with the internal dynamics of glaciations. Glacial intervals are initiated by the formation of a large perennial snowfield in the subarctic; it is of little consequence to this model just why such a perennial snowfield forms. We know that the Laurentide Ice Sheet could not have formed without such a snowfield forming first. Whether it formed in response to solar variations, an Antarctic ice surge, orbital variations of the earth, an open Arctic Ocean, volcanic activity, or any arbitrary *deus ex machina*, the course of the subsequent glaciation was largely determined once the Laurentide Ice Sheet began to form. Once the continental ice sheets appeared, their own effects on the global energy budget probably far surpassed the direct effects of the triggering mechanism that caused or allowed the initial snowfield to form.

The point of view pursued here is that the climatic perturbations of ice ages are almost entirely produced by terrestrial causes that are themselves the result of continental glaciation, that solar output need not vary, and that the sequence of climatic events characteristic of continental glaciations results from changes in the relative importance of the various planetary energy sinks and energy storage and transfer mechanisms. The implication for man in an age when many of his activities are increasing at an exponential rate is significant: we may well be able to start an ice age, but we probably cannot stop one once it begins.

# Energy Required for Continental Glaciations

The energy fluxes associated with a glaciation are in response to the large heat sink added to the global energy budget when an ice sheet forms. The ice sheet, because of its high albedo, absorbs less solar energy than an unglaciated surface but still loses heat by outgoing infrared radiation. Outgoing energy expenditures must be met either by cooling or by solar energy, received either directly from the sun or advected by the atmosphere from elsewhere.

The energy requirements of an ice age may be appreciated by considering the amount of heat released during condensation and freezing of the snow that formed the ice sheets. Flint (1971, Table 4E) estimates that the maximum ice volume during Quaternary time removed $47.3 \times 10^6$ km$^3$ more water from the oceans than is now found in glaciers. The energy released by the precipitation of that much snow is

$$47.3 \times 10^{21}\mathrm{cm}^3 \times 677\ \mathrm{cal\ g}^{-1} \times 1.0\ \mathrm{g\ cm}^{-3} = 3.20 \times 10^{22}\mathrm{kcal}\ . \quad (1)$$

The "heat of glaciation" thus determined is over 37 times the amount of energy absorbed annually by the earth-atmosphere system (Sellers, 1965, p. 68). Other figures provide even more impressive comparisons. Using data of Sellers (1965) and Sverdrup and others (1942), the heat of glaciation is 121 times the amount of energy used annually for oceanic evaporation and 1165 times the amount of heat currently transferred from oceans to the continents as latent heat.

Even these figures must be taken only as minimum values, for the figure given above for maximum glacial volume takes no account of ice that melted before the maximum, nor of ice that accumulated afterward ( Adam, 1973b; Newell, 1974; Newell and Herman, 1973).

Newell (1974, p. 118) has described the heat of glaciation as “excess heat which must be radiated away.” While this is correct in a technical sense, it is also misleading with respect to the causation of ice ages, rather like regarding food expenditures as excess money that must be spent in order to balance a budget. The heat of glaciation is in fact not excess at all, but is *required* to balance the energy budget over the ice sheet.

There are four possible sources for the heat of glaciation: local absorbed insolation, local cooling, and advected sensible and latent heat. Local absorbed insolation is low because of the high albedo. Local cooling contributes some energy, and also decreases the energy deficit by cutting outgoing radiative energy losses. This cooling increases the energy gradient between the ice sheet and surrounding warmer continental and oceanic areas just enough to balance the energy budget with advected sensible and latent heat, and the advected latent heat produces a significant by-product: snow.

# The Oceans as a Glacial Energy Source

About 88.5% of the heat of glaciation is the heat of condensation of water and has its source in the oceans that are the source of water for the snow. (The balance comes from the heat of freezing.) The energy balance of the oceans is thus of critical importance for the growth of ice sheets and the dynamics of ice ages.

When the oceans are considered as a whole, advection is eliminated as an energy source; the two remaining significant oceanic energy sources are incoming solar radiation and stored heat.

Cooling of the oceans during times of glaciation is complicated by the fact that they are being continuously warmed and cooled at the surface. The surface temperature of the ocean represents a local balance between warming and cooling processes. Local warming processes include insolation, freezing, advection, and inflow of warm water from land; the cooling processes are net outgoing infrared radiation, sensible and latent heat losses, advection, vertical mixing of warm water with cooler water beneath, inflow of cold water from land, and melting of ice.

During a glaciation, ocean surface temperature is controlled by negative feedback mechanisms. The warmer the ocean relative to the ice sheets, the stronger the energy gradient between them. This intensifies the atmospheric circulation and the rate at which heat and water are removed from the ocean. If the ocean should cool too much, the gradient will diminish and so will the rate of heat loss, so that the surface temperature will increase again to the point where a steady state prevails.

The local heat balance varies seasonally. When warming processes are dominant, sea surface temperature rises; general oceanic cooling is thus limited to the season when cooling processes predominate. However, during much of the cooling season, the heat loss involves heat stored during the preceding warming season and is a seasonal phenomenon rather than a contribution to secular cooling. Only after heat stored during the warming season is lost can decreases in long-term oceanic heat storage occur.

These long-term heat losses are restricted to a short period near the end of the cooling season and are thus limited in the speed with which they can affect the temperature of the oceans. Abrupt temperature changes are also prevented by the sinking of cooled water beneath the surface as soon as its density increases sufficiently. Water that sinks cannot cool further until it is once again at the surface, and there it will once again be affected by both heating and cooling processes.

In earlier papers (Adam, 1969, 1973a) I advanced the thesis that heat stored in the oceans during interglacial periods was the energy source that fueled the subsequent continental glaciations. However, that source by itself is quantitatively insufficient to account for the estimated volume of the ice sheets. The present volume of the oceans is about $1.37 \times 10^9$ $km^3$ (Sverdrup, Johnson, and Fleming, 1942); if the heat released during a glaciation were entirely derived from stored oceanic heat, the oceans would have cooled by more than 23°C throughout their depth. Since the average temperature of the oceans is far below 23°C, it is clearly impossible for all of the heat of glaciation to have been stored in the oceans prior to the onset of glaciation.

If stored heat is not sufficient to account for the evaporation of the water transferred from the oceans to the ice sheets, the only other reasonable possibility is incoming solar radiation. At first glance this would appear to require that incoming solar radiation must increase in order to supply the extra energy to fuel a continental glaciation, in accord with Simpson’s (1934, 1940) hypothesis, but this is not necessary. The thesis advanced here is that ice ages can occur without any causative variations in the solar constant, and that changes in the amount of energy available for evaporation from the oceans represent changes in the relative importance of the various energy sinks for solar radiation absorbed by the oceans.

# Glaciation

I have argued (Adam, 1969, 1973a) that continental glaciation commences when a perennial snowfield forms in the subarctic regions for any reason. Such a snowfield replaces a relatively 1ow-albedo surface with a high-albedo surface and is thus a new planetary heat sink. Because the snowfield cools rapidly, a strong energy gradient forms between the snowfield and the warm ocean. Energy is transferred across this gradient as latent heat, which both alleviates the local energy deficit and nourishes the developing ice sheet.

It is during the initial period of glacial growth that heat stored in the oceans above the thermocline is important. The warm surface layers of the oceans act as a sort of thermal capacitor, storing heat that can be readily tapped as an energy source whenever a sufficiently strong energy gradient develops between the ocean and a suitable continental heat sink.

However, it was shown above that the heat required to account for the ice accumulated during glaciations vastly exceeds the heat that could be stored in the oceans under realistic conditions. Once the stored heat is consumed, the ice sheets may be well established, but some other energy source must be invoked to sustain them.

This energy source may be found by examining the various local energy sinks for the radiation absorbed from the sun by the ocean. Advection merely redistributes heat within the ocean and can be eliminated by considering the ocean as a whole; the remaining channels for loss of energy from the ocean surface are storage, latent heat of evaporation, sensible heat loss to the atmosphere, and outgoing infrared radiation.

The oceans cooled during times of waxing glaciation, as has been shown by many lines of evidence (for example, Emiliani, 1966; Broecker and van Donk, 1970), although there is a possibility that Atlantic bottom temperatures may have increased slightly (Schnitker, 1973; Newell, 1974). This paper assumes that bottom warming has been minor in its effect on the oceanic energy budget, so that long-term storage of incoming solar energy may be eliminated as a major heat sink during glaciations.

Outgoing infrared radiation is governed by the Stefan-Boltzmann law (Sellers, 1965, p. 40):

$$I_{\uparrow} = \epsilon \sigma T^4 \,, \quad (2)$$

where $I_{\uparrow}$ is the outgoing infrared radiation, $\epsilon$ is the infrared emissivity, $\sigma$ is the Stefan-Boltzmann constant, and T is the absolute temperature in °K. The composition of ocean water cannot have changed radically between glacial and interglacial periods, so significant changes in infrared emissivity of the oceans seem unlikely. The energy loss attributable to outgoing infrared radiation is thus proportional to the fourth power of the absolute temperature.

In the model presented here, heat stored in the oceans above the thermocline is used to nourish developing continental ice sheets until it is depleted. This depletion, however, corresponds to a decrease in the surface temperature of the oceans and produces a parallel decrease in outgoing infrared radiation, as shown in

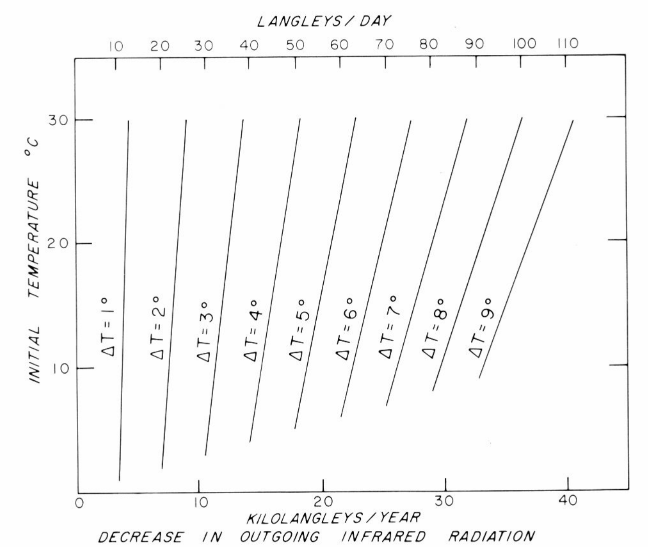


**Figure 1.** Decrease in outgoing blackbody radiation for a given temperature decrease and initial temperature. l langley = 1 calorie/cm$^2$.

Figure 1. Over the range of temperatures found in the oceans (-2–30°C), each 1°C drop in surface temperature will produce a drop in outgoing infrared radiation of 1.31–1.47%.

The process is somewhat complicated by the presence of long-wave counterradiation from the atmosphere back to the ocean. Counterradiation is strongly influenced by the amount of water vapor in the atmosphere; indeed, the effect is so strong that with a fully saturated atmosphere, the net outgoing infrared radiation actually decreases with increasing temperature (Sellers, 1965).

However, during times of glaciation the increased ocean–ice sheet energy gradient produced a more vigorous atmospheric circulation than at present, and it thus seems likely that air over the ocean (particularly the North Atlantic) was vigorously mixed, so that the decrease in outgoing infrared radiation from the sea surface (caused by declining ocean surface temperatures) was greater than the decrease in counterradiation from the atmosphere to the ocean (caused by decreasing atmospheric water vapor content). The result was a decrease in energy lost by the ocean as outgoing long-wave radiation during glacial periods.

If insolation did not change but sea surface temperature declined, then the energy retained by the sea through decreased outgoing infrared radiation was available to augment other heat loss processes. If storage of heat in the oceanic deep water was not of major importance, the bulk of the newly available heat must have augmented sensible and latent heat losses from the sea surface. More energy than at present was available for evaporation and warming air.

# The Bowen Ratio and Sea Surface Temperatures

The relative amounts of heat leaving the ocean as sensible and latent heat are controlled by various climatological parameters; most important are the temperatures of the water surface and the overlying atmosphere, and the water vapor content of the atmosphere. The ratio of sensible to latent heat loss is known as the Bowen ratio (Jacobs, 1951). Once a continental ice sheet forms, its rate of nourishment is limited by the amount of energy available to evaporate water from the oceans. The Bowen ratio measures the partition of oceanic nonradiative energy losses into a latent heat fraction, which can generate potential glacial nourishment, and a sensible heat fraction, which cannot. Changes in the climatology of the Bowen ratio during the Pleistocene are thus a sensitive measure of the health of continental glaciations.

The Bowen ratio may be expressed as

$$R = \frac{H}{\mathrm{LE}} = 0.64\frac{P}{1000}\frac{t_w - t_a}{e_w - e_a} \;, \quad (3)$$

where *H* is sensible heat, LE is latent heat, *t* is temperature in degrees C or K, *e* is water vapor pressure in millibars, *P* is total air pressure in millibars, and the subscripts w and a refer to the sea surface and to an arbitrary distance above the water surface, respectively (Jacobs, 1951). Since the heat and water transfers we are concerned with take place at the sea surface, we may safely set the pressure term *P*/1000 equal to unity and drop it from the equation.

In considering cold airmasses passing over large bodies of warmer water such as oceans, we may assume that air becomes saturated with moisture and warmed to the temperature of the underlying water. The amount of heat that can be transferred by evaporation is limited by the saturation vapor pressure at the sea surface, while the amount of heat used to warm the air is limited by the initial temperature difference between the air and the sea. If we introduce the value

$$A = \frac{0.64}{(e_w - e_a)} \quad (4)$$

the Bowen ratio reduces to

$$R = A\,(t_w - t_a) \;. \quad (5)$$

The minimum value of *A* at any temperature will be for the case in which the air is dry $(e_a = 0)$ ,

$$A_{\min} = 0.64/e_w \;, \quad (6)$$

and this leads to

$$R_{\min} = A_{\min}\,(t_w - t_a) \;. \quad (7)$$

This minimum value of the Bowen ratio corresponds to the maximum possible rate of evaporation and latent heat flux.

Figure 2 shows minimum possible Bowen ratios plotted against temperature for several values of initial temperature difference between water and air. Figure 2 shows that as sea surface temperatures decline during a glaciation, more and more of the available energy (absorbed insolation minus net outgoing infrared radiation) is used to warm the air masses passing over the sea, and less and less is available for evaporation.

The increase of the Bowen ratio with decreasing sea surface temperature is a primary reason for the termination of midlatitude continental glaciations. These ice sheets require large amounts of snow in order to exist under the high insolation they receive in middle latitudes, but their long-term effect upon the oceans whence they derive their nourishment is to cool them. As cooling progesses, the minimum Bowen ratio increases, and an ever greater proportion of the oceanic heat is lost as sensible rather than latent heat. Although a strong energy gradient may be maintained between the ice sheets and the oceans, less and less of the energy transferred across the gradient is latent heat, which is required to maintain the glacier. As the rate of nourishment of the ice sheet declines, so does its steady-state size, until it finally disappears or retreats to a polar position where the rate of nourishment from a cool ocean can balance ablation.

To end a glaciation, it is not necessary for sea surface temperatures to fall everywhere; all that is required is that the ocean that acts as a precipitation source for the ice sheets must cool, and particularly those areas that are closest upwind from the ice sheets. It is likely that the storms that nourished the continental ice sheets were rather warm storms in order to carry the necessary water vapor to the ice sheets (Lamb and Woodroffe, 1970; Barry, 1966); much of the cooling must have taken place after the airmasses reached the ice sheets, for otherwise the moisture and energy would have been lost along the way. Warm, wet airmasses could not have survived a passage across a wide expanse of cool ocean. In the case of the ice sheets of the northern hemisphere, once the northern North Atlantic cooled sufficiently, the ice sheets could not maintain themselves and disappeared.

# Stadials and Interstadials

In previous papers (Adam, 1969, 1973a), I suggested that a negative feedback relationship between excess glacier size and nourishment acts through the effect of glacial meltwater on ocean surface temperatures, and that this mechanism is a cause of the oscillations between stadial and interstadial conditions observed in the stratigraphic record. Those papers assumed that stored heat from the oceans was an essential energy source throughout a glaciation, and that a surface layer of meltwater impeded the loss of stored heat from the ocean to the atmosphere. Although stored heat cannot have such an important role in fueling ice ages (because there is not enough of it), glacial meltwater nevertheless affects the energy budget of the oceans in such a way as to produce stadial/interstadial oscillations, in the manner proposed below.

The meltwater layers postulated here result in large part from the inability of a growing ice sheet simply to grow to a steady-state size and then stop; the growing ice sheet must overshoot its steady-state size to some degree (Šegota, 1963; Adam, 1969, 1973a). As long as the ice sheet is smaller than its steady-state size, it has a positive mass budget, and as long as it has a positive mass budget, it must increase the convexity, or steepness, of its profile. It follows that when the ice sheet first attains its steady-state size, it has the profile of an advancing ice sheet and will continue to advance beyond the steady-state size. Rapid melting will ensue; the effects of the meltwater thus produced on the oceans and on the various elements of the oceanic energy budget are described below.

If a layer of cool meltwater floats on a warmer ocean, it warms rapidly until it reaches the temperature of the underlying water. Most of this warming is accomplished by diffusion of heat from the underlying water rather than by incoming radiation because of the high heat capacity of the ocean. The underlying water thus cools. Because of the nature of double-diffusive convection (heat can diffuse into the surface layer about 100 times as fast as salt; see Turner, 1965), the meltwater layer maintains its low density and physical integrity while the heat flux from the underlying ocean warms it.

The heat added from below promotes convective overturning within the surface layer until the entire meltwater layer is as warm as the underlying ocean. Thereafter, however, further insolation warms it from the top, adding a thermal density stratification to the already strong salinity density stratification. Both factors inhibit the downward mixing of absorbed solar heat, raising the surface temperature of the ocean in the presence of a surface meltwater layer above the value that would be found without one. This effect should be most pronounced during the summer, both because meltwater production is at a maximum at that time and because summer insolation over the ocean warms it at a maximum rate.

The immediate effect of a surface meltwater layer is thus to warm the sea surface to a summer temperature higher than would occur in the absence of the meltwater layer. Because the atmosphere is warmed by the sea surface, this will result in warm summers both at the sea surface and on the surrounding continents because summer is the time of both surface layer development and insolation maxima.

What is of great importance here is that increased summer temperatures at the sea surface produce an increase in both outgoing infrared radiation and sensible heat losses from the sea surface. In the absence of a surface layer, this energy would be stored in the upper layers of the ocean and released during the winter, providing energy and moisture to nourish the ice sheets. The surface layer thus acts to diminish the amount of energy available from the annual energy budget for winter ice sheet nourishment.

It is not necessary for the surface meltwater layer to persist from year to year, as was suggested earlier (Adam, 1969, 1973a). Winter storms may weaken or destroy the layer each year, but it will re-form each spring when the ice sheets begin to melt again.

The degree to which the meltwater layer affects the oceanic energy budget should be roughly proportional to the amount of melting during the year. This in turn is controlled by how large the ice sheet is compared to its steady-state size, so that when the ice sheet overshoots its steady-state size, the resulting excess meltwater acts to decrease the steady-state size by decreasing the amount of energy available for nourishment. This produces interstadial conditions with warm summer temperatures. Once the ice sheet melts back to a size near the interstadial steady-state size, the amount of meltwater produced each year diminishes. This reduces the effect of the meltwater layer on the oceanic energy budget and increases the rate of nourishment and the steady-state size again, leading to a new stadial interval.

The influence of surface layer formation on ice sheet regimen must have been particularly important in the Gulf of Mexico, where the effects of a surface meltwater layer have been reported by Kennett and Shackleton (1974). During times of maximum advance of the Laurentide ice sheet, the meltwater from the periphery of the ice between the Rocky Mountains and the Appalachians flowed into the gulf, which occupied a critical position as a low-latitude heat and moisture source for the Laurentide ice. The feedback link between the Laurentide ice and the Gulf of Mexico may even have been a major control of the climate of the whole northern hemisphere during glacial times because of the huge size of the Laurentide ice sheet and its dominant influence on the circumpolar energy budget.

This link may also explain the absence of a late-glacial ice readvance in North America to match the well-known younger Dryas glacial advance of the Scandinavian ice sheet. At the time of the Two Creeks interval about 11,800 y.a., the drainage of the Laurentide ice sheet from present Lake Superior eastward was diverted from the Gulf of Mexico into the St. Lawrence lowland and the North Atlantic (Hough, 1963). This event must have significantly augmented surface layer formation in the North Atlantic and may have been responsible for the onset of the Allerød warm period in northem Europe about 11,700 y.a. (Iversen, 1973). The Two Creeks interstade was followed by the Valderan substage, which diverted the drainage of the Laurentide ice sheet west of Michigan back into the Mississippi drainage about 11,000 y. a. (Hough, 1963); this reduced surface layer formation in the North Atlantic and may have caused the Younger Dryas cooling and glacial readvance (Iversen, 1973). The Laurentide and Scandinavian ice sheets thus need not have been in phase with each other during their deglacial histories.

# Deglaciation

The final disappearance of low-latitude continental ice sheets occurs when the oceans cool to the point that they cannot supply enough snow to maintain the glaciers. When this occurs, there is a long time lag while the ice sheets melt away. During this time, the oceans warm, and an obvious question is why the warming oceans during the early part of an interglacial do not initiate a resurgence of ice growth before the ice sheets disappear entirely. How does the northern hemisphere emerge from an ice age?

Two answers are suggested here. First, the rapid increase in surface temperatures noted in deep-sea cores at the end of the last glaciation must have been initially confined to a thin layer near the surface, and much of this heat may have been present only during the summers. The heat now stored above the thermocline in the North Atlantic cannot have suddenly appeared at the beginning of the Holocene; rather, it must have accumulated gradually, and it may still be accumulating as a very slow deepening of the thermocline. During the early Holocene, when a sizable ice sheet still existed, the small amount of stored heat must have been insufficient to allow reglaciation.

The second reason is that the postglacial warming of the North Atlantic was not an instantaneous event. Ruddiman and McIntyre (1973) have demonstrated that the retreat of polar waters from the North Atlantic was time-transgressive and lasted from about 13,500 BP to at least 6500 BP. The waning Laurentide ice sheet, and perhaps the Scandinavian ice sheet as well, was separated from the warming waters of the central North Atlantic by a broad band of cooler water that prevented a high rate of nourishment. Ruddiman and McIntyre believe that glacial meltwater played a relatively minor role in delaying the warming of the northern reaches of the North Atlantic, but the direct and indirect effects of meltwater may have been of considerable importance, either through effects on local oceanic energy budgets outlined above or by affecting the formation of sea ice (Olausson, 1972; Olausson and Jonasson, 1969).

Evidence useful in testing the model proposed here should be forthcoming from wider use of the deep-sea biostratigraphic record to study changes in seasonal temperature contrasts (Prell, 1975) and thermocline structure (Hecht, 1973). Seasonal temperature contrasts should reach maximum values during interstadials and the early part of interglacials, while thermocline depths should be greatest during the latter part of interglacials.

# Summary

The energy to fuel an ice age can be obtained by a redistribution of the solar energy received by the oceans in response to the large heat sinks created by the formation of continental ice sheets. An ice age is triggered when a perennial snowfield forms in the subarctic. This sets up a large, high-albedo heat sink and creates a large energy gradient between itself and warm oceans that have been storing heat during an interglacial. Both energy and moisture are transferred across this gradient as latent heat and water vapor; the vapor precipitates as snow, and the energy is released to satisfy the local energy deficit caused by the high albedo. During the initial formation of the ice sheet, stored heat from the oceans augments current solar radiation to provide the required energy, but the stored energy is soon depleted and ocean surface temperature drops. This permits a decrease in energy lost by the ocean surface as outgoing long-wave radiation, and the energy thus conserved is then used for evaporation and sensible heat transfers to the atmosphere and thence to the ice sheets. The glaciation finally terminates when the sea surface temperature drops too far, increasing the Bowen ratio to the point where most of the available energy leaves the oceans as sensible heat, and the glaciers starve for lack of nourishment.

Interstadials are produced when large pulses of glacial meltwater produced by ice sheet overextension enter the ocean as surface layers during the summers. The surface layers inhibit vertical mixing and confine the absorbed solar radiation to a relatively thin layer near the surface, raising its temperature above that which wou1d prevail in the absence of a surface layer. This produces relatively warm summers and causes higher summer energy losses from the ocean both as sensible heat and as outgoing long-wave radiation. This energy is then not available during the winter to fuel the glacier, and the interstadial is thus maintained until the glacier melts back to a smaller steady-state size. When this occurs, production of meltwater is diminished, summer energy losses are cut, and more energy is then available for winter evaporation, glacier nourishment, and a new stadial.

# Acknowledgements

I thank R. G. Barry, T. J. Blasing, D. M. Hopkins, J. M. Mitchell, Jr., G. I. Smith, and J. Williams for manuscript reviews, W. Sellers for helpful discussions, and L. M. Gould for continuing encouragement and moral support.

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