THE ROLE OF THE OCEANS IN CLIMATIC CHANGE: A THEORY OF THE ICE AGES

PETER K. WEYL

Oregon State University, Corvallis

ABSTRACT

Changes in the surface salinity distribution in the World Ocean, by changing the extent of sea ice in the North Atlantic and Antarctic, can lead to climatic change. By reducing the water vapor flux across Central America, the salinity of the North Atlantic is reduced. If this change persists over a sufficient length of time, a glacial climate could be initiated. An examination of the "Little Ice Age" tends to confirm this hypothesis. A return to an interglacial climate may be the result of overextension of glaciers followed by stagnation of the bottom water. Stagnation is terminated by geothermal heating at the ocean floor, followed by vertical mixing of the warmed, saltier water into the subarctic gyre of the North Atlantic. This, in turn, results in a reduction of sea ice and in climatic warming.

1. Introduction

When we look at the geologic past, we are inclined to marvel that the climate has changed from what we consider as the normal. If, on the other hand, we adopt a more geologic point of view, then we might more properly marvel that conditions on the surface of our planet have remained as relatively constant as they have over the last billion years. In spite of cosmic disturbances and the volcanic and tectonic disturbances of the earth's interior, the surface of our planet has remained sufficiently hospitable to permit the gradual continuous evolution of advanced forms of life. Thus, we should admire the climatic stability and address ourselves to its study rather than try to explain the variations. While the Pleistocene variations have had a profound effect on some portions of the land, their overall effect, particularly on life in the sea, has been small.

The stability of environment on the surface of our planet over at least a billion years is largely due to the presence of a billion cubic kilometers of water. This mobile reservoir, with a large capacity for heat and chemical constituents, acts as a stabilizer against chemical and climatic variations. Elsewhere (Weal, 1966) I have examined how the hydrosphere, by interacting with the biosphere, lithosphere and atmosphere, may help stabilize the surface chemistry of the earth. Here I wish to look at the manner in which the ocean interacts with the atmosphere to stabilize the climate. In particular, I wish to look at the reasons why this stabilization was not as effective during the Pleistocene, thus giving rise to the climatic oscillations we call the ice ages.

Climate may be regarded in essence as the weather averaged over a sufficient number of years to filter out the year-to-year variations. Thus, the study of climate change is the study of the lower frequency components of weather variation. The response time of the atmosphere is much too fast to insure climatic stability. This can be seen, for example, from the work of Mintz (1967), where by using numerical models of the atmosphere and a large digital computer, he has caused the winds to come to rest. Starting the atmospheric circulation on the first day and integrating the equations of motion, he obtains a normal weather pattern on the 40th day. Thus, the memory time of the atmosphere is less than one month.

That the oceans help to stabilize the yearly fluctuations of weather becomes evident when we compare the annual temperature variations in the interior of the continents with that near and over the oceans. That the oceans play a significant role in longer time fluctuations has been pointed out in the recent studies of Bjerknes (1965) and Sawyer (1965).

In order to understand the climates of the past, it is therefore necessary to understand the ocean circulation of the past. If we restrict ourselves to the Pleistocene, the problem is simplified by the fact that the geography during that time did not differ significantly from that of today. Even so, the problem is formidable, since we do not yet understand adequately the ocean circulation of the present day. It could be argued that we must postpone a study of paleoceanography until the science of oceanography is further advanced. In the earth sciences, however, it is impossible to make global experiments. To test our understanding of the processes operating, we must, therefore, make use of the past. Thus, history re-
places experimentation, and our understanding of oceanographic processes can be tested only by checking reconstructions of the past against the geologic evidence. To study the ocean of today and to convince ourselves that we can predict the future, we must become paleo-oceanographers.

In the present paper I advance the hypothesis that the climatic fluctuations during the Pleistocene arose primarily from variations in the extent of sea ice. It has generally been supposed that sea ice is the result, rather than the cause, of climate. I will attempt to demonstrate, however, that the extent of sea ice depends more on the salinity distribution in the ocean than on heat exchange between the ocean and the atmosphere. Of particular importance is the contrast in surface salinity between the North Atlantic and the North Pacific. I shall attempt to show that this contrast could be changed by a persistent change in the atmospheric circulation near the Isthmus of Panama and that the resultant reduction in salinity of the North Atlantic, if sufficiently intense, would lead to the initiation of a glacial climate.

First, I shall make paleo-oceanographic reconstructions based on our understanding of the physics of the sea. Next, I shall draw on recent documented changes to help substantiate these reconstructions. Finally, I shall consider the glacial periods of the Pleistocene and examine the geologic evidence for support.

2. The density of sea water

The land and the sea surface form the lower boundary of the atmosphere. The exchange of heat at this boundary depends on the capacity and diffusivity for heat of the sea or the land. In the case of the ocean, heat is primarily carried to and from the sea surface by convection. The depth of convective mixing is governed by the vertical density structure. While water of uniform density will convect readily, the presence of a stable vertical density gradient (density increasing with depth) will inhibit convection.

At constant pressure, the density of fresh water is a function of the temperature and is a maximum at 4°C. The density of sea water depends on both the temperature and salinity. As the water is cooled, for waters above 24.7% salinity, the density increases monotonically to the freezing point. While the change in density with salinity is almost independent of the temperature, the thermal expansion at constant salinity decreases markedly as sea water approaches the freezing point. The relative effect of temperature and salinity on the density of sea water is shown in Fig. 1. This figure is a plot of the partial derivative of salinity with temperature at constant density, for a constant density curve passing through 0°C and 34% salinity.

At the freezing point, a salinity increase of only 0.035% is compensated by a temperature rise of 1°C. At 20°C, the salinity would have to increase by 0.39% to cancel the density change due to warming the water by 1°C. As a result, at low latitudes, temperature mainly determines the density structure near the sea surface; at high latitudes the salinity distribution becomes relatively more important.

As heat is added to the sea at the surface, water is evaporated and the sea water is raised in temperature. Evaporation increases the salinity and tends to cancel the density decrease caused by warming of the water. The fraction of the heat that must go into evaporation in order to keep the density constant is shown in Fig. 2 as a function of the temperature. At 0°C, only half the heat must be used for evaporation to keep the density constant, while at 11°C, 80% must be so used. Thus, at high temperatures the addition of heat will invariably form a less dense water at the surface, while at low temperatures, heating may produce denser, unstable water.

In summary, temperature is the dominant variable at moderate and high temperatures, but near
freezing point, salinity becomes relatively more important. Surface convection in the regions where seawater freezes will depend largely on the salinity distribution. Therefore, let us look at the distribution of salinity in the World Ocean.

3. The salinity distribution in the World Ocean

If the World Ocean is in steady state, the salinity is a three-dimensional scalar field. This is difficult to portray on two-dimensional maps. If, at a particular latitude and longitude, we proceed downwards from the sea surface, the salinity will go through one or more maxima and minima. The main features of the salinity distribution can be described by mapping the salinity along these extremum surfaces. This has recently been done by Ingham (1966) and the following description is adapted from his work.

a. A note on the maps used in this paper. The map projection used throughout this article is a gnomonic projection of the earth onto a regular icosahedron, truncated by a plane passing through 65°N. By using this projection it is possible to show the entire earth with relatively little distortion in many different ways, as the icosahedron is cut and unfolded. A disadvantage is the number of gaps that appear when the icosahedron is laid out flat. During the plotting, these gaps can be eliminated by arranging the equilateral triangles differently. A map similar to the present one was published by Buckminster Fuller (1943) as the "dymaxion globe." Fisher (1943) has discussed the projection used here in detail. While the above maps were designed to show the continents to best advantage, the present map was specifically designed to display the oceans. This was accomplished by rotating the icosahedron of Fisher about the poles, so that as many vertices as possible fall on land. To show the Arctic ocean on a single sheet, the northern triangles were truncated by a pentagon. The use of this basic map permits us to show each parameter to its best advantage and at the same time compare different maps readily. The map can be folded to approximate a globe.

b. The near-surface salinity maximum. The salinity at the near-surface salinity maximum surface is shown in Fig. 3. In Figs. 3, 4 and 5, the contours of salinity are in units of %, minus 30%. The salinity varies from a high of 37.4% in the North Atlantic to less than 34.5% in the North Pacific. The salinity maximum generally occurs between the sea surface and a depth

Fig. 1. Salinity of the near-surface salinity maximum [replotted from Ingham (1966)]. Salinity units are in %, minus 30%.

Fig. 4. Salinity of the intermediate salinity minimum [replotted from Ingham (1966)]. Salinity units are in %, minus 30%.
of 200 m. The highest salinities are located symmetrically about the equator centered at about 20° latitude. Of particular interest are the differences between the Atlantic and Pacific Oceans. While the salinity gets higher than 37%o in both the North and South Atlantic, the salinities in the North and South Pacific only become slightly larger than 35 and 36%o, respectively. The salinity in the subtropical North Pacific is anomalously low, resulting in a salinity contrast across the Isthmus of Panama of 2%o. The Indian Ocean, since it does not extend to high northern latitudes, differs from the major oceans. The equatorial undercurrents flow at the level of the salinity maximum, and at our scale of mapping produce a meridional discontinuity in the salinity maximum surface.

c. **The intermediate salinity minimum.** The salinity along the intermediate salinity minimum surfaces is shown in Fig. 4. These surfaces range in depth principally between 600 and 1000 m. Around the Antarctic continent, the salinity is a minimum at the sea surface (see Fig. 9). A subsurface minimum first appears between 50 and 60S with a salinity near 34.2%o. In the Southern Hemisphere the salinities along the surfaces are similar in all oceans, although in the Indian Ocean the salinity increases more rapidly, and the vertical variation becomes monotonic at about 10S. The salinity minimum surfaces in the Northern Hemisphere are different in each ocean.

In the Atlantic the southern minimum extends north to about 45N with a marked increase in salinity due to the outflow from the Mediterranean. There is a second, minor, minimum surface in the far North Atlantic. In the Pacific, on the other hand, the southern minimum only extends to about 20N. A symmetrical northern surface extends from the subarctic to the equator. The water on this surface has a lower density than that on the southern minimum, and as a result, in the region of overlap (0 to 20N), the northern minimum surface lies above the southern one. In this region the two are separated by a salinity maximum surface which is not shown. There is a small salinity minimum surface in the northwest Indian Ocean.

d. **The deep salinity maximum.** The salinity along the deep salinity maximum surface is shown in Fig. 5. Its depth generally ranges from 1–2 km. The high salinity water has its origin near the outflow of the Mediterranean and then extends south in the Atlantic Ocean. The high salinity water enters the Southern Ocean with a salinity below 34.9%o and wraps itself around the Antarctic continent. An examination of IGY data from the Southern Ocean suggests that the surface may not be continuous but rather consists of boluses of warmer, saltier water (see Fig. 9). The salinity maximum surface penetrates the Indian Ocean only to about 20S and is absent in the Pacific north of 55S. A second minor maximum surface exists in the north Indian Ocean. In the Pacific, the salinity below the intermediate salinity minimum increases monotonically to the ocean floor.

c. **Summary.**

1. The greatest differences in the salinity distribution are between the North Atlantic and the North Pacific.
2. The near-surface salinity is much higher in the Atlantic than in the other major oceans.
3. The near-surface salinity contrast is particularly marked across Central America.
4. The deep salinity maximum has its origin in the North Atlantic and wraps itself around the Antarctic continent.
5. The Pacific, except in its southernmost region, has no deep salinity maximum surface.
6. The North Atlantic has no major intermediate salinity minimum surface extending to the equator.

Figs. 3, 4, and 5 [after Ingham (1966)] give a brief summary of the salinity distribution in the World Ocean of today. It is my thesis that the salinity distribution has a profound effect on the formation of
sea ice in high latitudes, and, hence, on climate. I shall now discuss the relation of salinity to sea ice in the Arctic and Antarctic.

4. Salinity and sea ice

a. The Arctic. The maximum extent of ice cover in the Arctic Ocean in March is shown in Fig. 6. Also indicated is the summer location of the 34\% surface salinity curve. The data are taken from the U.S. Navy Hydrographic Office charts (1958) and from Chart VI of Sverdrup et al. (1942). In the Pacific Ocean, where the water north of 40°N has salinities less than 34\%, the sea ice limit in March is at about 60°N. In the North Atlantic, on the other hand, portions of the ocean remain ice free at latitudes as high as 75°N. Here, the 34\% summer surface salinity curve tends to lie just north of the March ice limit. The northward displacement of the ice limit and the 34\% isohaline are the result of the advection of warm salty water by the North Atlantic currents (see Fig. 6).

It might be thought that the ice limit is controlled mainly by the advection of warm water and that the high surface salinity of the North Atlantic is merely a result of that advection. I shall now try to show that this interpretation is not correct but, rather, that the distribution of salinity has a stronger effect on the ice limit than the distribution of temperature.

Before Nansen's historic drift in the ice of the Arctic Ocean (1893-1896), it was generally believed that this ocean was a shallow, island-strened sea (Gregory, 1897). It was argued that, if the Arctic Ocean were deep, it would not be possible for sea ice to form, because the entire water column would first have to be cooled to the freezing point. If the Arctic had a uniform vertical salinity distribution, this would be correct. Actually, however, because of its salinity stratification, the Arctic, while it has oceanic depth, appears to the atmosphere as if it were only about 100 m deep.

A typical vertical temperature-salinity distribution for the Arctic Ocean, based on data from the Russian airplane N-169 expedition in April 1941 (Zubov, 1943, p. 413), is shown in Fig. 7. In order to bring out the structure over the entire depth range, the salinity and temperatures are plotted against the cube root of the depth. It can be seen that the salinity increases from about 31 to 35\% between 30 and 200 m. The resultant large increase in density prevents convective mixing. Thus, surface sea ice is in good thermal contact with only the upper 30 m of the water column.

The contrast between the conditions in the ice-covered portions of the Arctic and the open portions of the North Atlantic are well shown in a cross section from Greenland southeast to the Norwegian Sea (for location of section see Fig. 6) shown in Fig. 8. This section for July-August shows a stable density stratification. The reason for the stability, however, differs. In the northern sector near the Greenland coast, the vertical increase in density is mainly due to an increase in salinity with depth. In the southern portion,
on the other hand, the salinity is almost uniform, and the density variation is due to a decrease in temperature with depth.

The response of the water column to winter cooling will differ in the two sectors. In the northern portion, the surface temperature will rapidly reach the freezing point because the salinity stratification prevents downward convection below 30 m. Because of the uniform salinity of the southern portion, convective mixing will continue until the entire water column has been cooled to the freezing point. Thus, while the surface temperatures of the two portions differ by at most 10°C, the convective depths differ by a factor of 100. If the southern portion had a sufficient salinity increase at 30 m to give stability, the extraction of some 30 kcal cm⁻² would start ice formation at the surface. Let us assume that below this, to a depth of 3 km, the water is, on the average, only 1°C above the freezing point. In the absence of a salinity increase, the formation of ice would require the extraction of an additional 300 kcal cm⁻².

The above example shows that the vertical salinity structure controls the amount of heat that must be extracted before sea ice can form. In this simple example we have ignored the effect of horizontal advection. It is clear that the horizontal circulation caused by cooling an isohaline water column to the sea floor would lead to advection of warmer water and hence further inhibit the formation of sea ice. Thus, the difference in the ice limit of the Atlantic and Pacific sectors of the Arctic Ocean is primarily the result of the differences in surface salinity between the two sectors. The salinity differences in high northern latitudes, in turn, are a result of the general difference in surface salinity of the two oceans (see Fig. 3) and the difference in circulation.

b. The Antarctic. In a sense, the Antarctic is the antithesis of the Arctic. The former is a continent surrounded by water, while the latter is an ocean surrounded by continents. As a result of the strong circumambient Antarctic oceanic circulation, a large asymmetry between the Atlantic and Pacific sectors of the Southern Ocean cannot be established. The charts of the salinity extremum surfaces demonstrate this. Nevertheless, I will show that the high North Atlantic surface salinity plays a decisive role in limiting the extent of sea ice around the Antarctic continent.

Fig. 8. Density-temperature-salinity section from Greenland (I) to the Norwegian Sea (II).

Fig. 9. Temperature and salinity section south of 45° at 97°E. International Geophysical Year data.
The surface salinities in the Southern Ocean are much more uniform than in the Arctic and never fall below 33‰. Low surface salinities are the result of seasonal melting of sea ice. Unfortunately, the pertinent hydrographic data were generally obtained during the southern summer season when the seas close to the Antarctic continent are accessible. To determine the effect of the thermohaline structure of the water on ice formation, we must, therefore, resort to calculations. Neglecting horizontal movement of the water, I shall calculate the changes produced in the water column by heat extraction at the surface. While it is not possible to present a detailed study, to illustrate the behavior of the water column, however, I have selected a typical south-north section, one obtained by the Soviet research vessel Ob in April 1957, at about 97°N.1

The S-N temperature and salinity sections at 97E are shown in Fig. 9. We note in high latitudes that temperature and salinity maxima exist at depth. The salinity maximum is the one mapped in Fig. 5 and has its origin in the North Atlantic. It wraps itself completely around Antarctica, although a more detailed study of the salinity distribution suggests that the maximum surface may not be continuous but rather may consist of a series of bulges. The region of high surface temperature gradient at about 53S is the Antarctic convergence.

Let us now consider what happens to the temperature-salinity section as winter cooling occurs. We shall assume the water to be stationary and will neglect dilution of the sea water by precipitation. As the surface water is cooled it will become denser and mix downward. Because of the slight thermal expansion of sea water at low temperatures and the increase in salinity with depth, downward mixing will be limited to shallow depth. Once sea ice forms, a new process comes into play. As sea water freezes, only a small fraction of the salt is incorporated into the ice. The salt content of sea ice depends on its mode of formation and age [see Zubov (1943, p. 140)]. For rapid freezing the initial salinity may be as high as 10% but as the ice ages, the salinity is reduced. The salt rejected by the ice is added to the underlying water and increases its salinity. Due to this salt addition, the water becomes denser and convection can now extend to a greater depth.

In the Arctic Ocean, this added convection is not significant due to the steep vertical salinity gradient. In the Antarctic, however, this effect is very important. As ice continues to freeze, convection extends to greater and greater depth. Once the salinity maximum is penetrated, the upward mixing of the warmer saltier water will lead to melting. Thus, extreme cooling does not lead to the formation of a thicker ice layer, but rather, tends to remove the ice previously formed. To see this, we must study the convective mixing quantitatively.

A convenient method for determining the onset of sea ice formation has been developed by Bulgakov (1942). He first plots the salinity, averaged from the surface to some depth z, against z. Next he constructs a $\sigma_T$ scale ($\sigma_T$ is a density scale used by oceanographers and is equal to the density at a pressure of one atmosphere minus one, multiplied by a thousand) related to the previously chosen salinity scale such that $\sigma_T$ corresponds to the density at the freezing point of water of that salinity. The relationship between the two variables is

$$\sigma_T = 0.815 - 0.15.$$  

Next, $\sigma_T$ is plotted as a function of depth.

As long as the density curve is to the left of the average salinity curve, cooling will cause convective mixing. The depth at which the two lines intersect is the depth at which the in situ density is equal to the density of the water column, mixed to the depth and cooled to the freezing point. Below this depth, stability cannot be eliminated by cooling. Therefore, the inter-

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1 Data from IGY Oceanography Rept. 2, May 1961, IGY World Data Center A, Texas A & M College, 111-112.
section of the two curves represents the depth to which convection reaches before freezing at the surface can take place. Typical curves for the Arctic and for the Antarctic are shown in Fig. 10. Note that while the arctic curves remain well separated below their intersection, the Antarctic curves cross again at about 600 m. This implies that if we were able to mix the Antarctic water column to a depth of 600 m, no surface sea ice would form.

Now we must consider the effect of salt addition due to the formation of sea ice. While there is no exact relationship between the thickness of ice frozen and the amount of salt added to the underlying water, the value is approximately 3 gm salt cm$^{-2}$ m$^{-1}$ of ice. In Fig. 10, I have indicated the change in the average salinity to depth $z$ produced by freezing of 1 m of ice. Thus, to mix the water column to a depth of 100 m we would have to freeze about 1 m of sea ice at the Antarctic location and about 4 m of ice in the Arctic.

Let us define as the salinity deficit at a depth $z$, the amount of sea salt that has to be added per square centimeter of the water column, in order that the water mixed, to depth $z$ and cooled to the freezing point, will have a density equal to that found at depth $z$. Fig. 11 is a plot of the salinity deficit vs. $z$ for the Antarctic station of Fig. 10. Due to a lack of precision in the oceanographic data, it is not possible to get an exact measure of the salinity deficit. The important point, however, is that after reaching a maximum of about 2.5 gm cm$^{-2}$, the salinity deficit decreases. Thus, the maximum thickness of ice that could form at this location is only about 1 m. Further ice formation would be prevented by rapid upward convection of heat in the water column. Instead, if cooling at the surface persists, the ice thickness will decrease. To show that this situation is not unique at the station selected as an example, I have made salinity deficit calculations at a series of stations in the Southern Ocean. The data are taken from the IGY data. The maximum salinity deficits and the depths at which they occur are shown in Fig. 12. In general, the deficits do not exceed 3 gm cm$^{-2}$, suggesting that the average thickness of sea ice formed should not exceed 1 m. Precipitation and ice rafting will tend to increase that figure. Observations from ice breakers operating in the Ross Sea during the Antarctic summer show large variations in ice thickness, ranging from open water to about 3 m with an average of around 1 m.

The formation of Antarctic bottom water has been discussed by Fofonoff (1956). He points out that bottom water is formed by mixing water from the salinity maximum with surface water on the shelves of the Antarctic. However, Fofonoff does not specifically investigates the effect of the salinity increase due to the freezing of sea ice. From the present discussion it is clear that both the presence of the deep, high-salinity water and the freezing of ice is required to form bottom water. Bottom water formed under ice will differ from that formed at the atmosphere-liquid water interface in that equilibration of oxygen and carbon 14 with the atmosphere will be strongly inhibited. Thus, the relatively low oxygen values (Sverdrup et al., 1942, p. 748), and carbon-14 values (Broecker, 1963, p. 92) of the Antarctic bottom water can be readily explained.

Deacon (1963) in his review of the oceanography of the Southern Ocean points out that there is a scarcity of hydrographic data showing the formation of bottom water on the shelves. This lack of data is a result of the difficulty of navigation during the late Southern Hemisphere winter when bottom water formation.
should be at a maximum. While direct observations would, of course, be preferable, it is possible to use summer observations and predict the effect of cooling on the thermohaline structure by calculations.

In summary, the salinity maximum water in the Southern Ocean that had its origin in the North Atlantic has two effects. First, its presence limits the thickness and extent of sea ice around the Antarctic continent. Second, by vertical mixing due to surface ice formation, it leads to the formation of Antarctic bottom water. A decrease in salinity of North Atlantic deep water would lead to an increase in ice thickness and extent around the Antarctic and to a decrease in the rate of formation of bottom water and hence in a decrease in the thermohaline circulation in the Southern Ocean. While the response of the Arctic to changes in North Atlantic salinity would be relatively fast, the response of the Antarctic would be delayed by hundreds of years due to the time required for the North Atlantic deep water to reach high southern latitudes.

5. Sea ice and climate

The presence of sea ice alters the heat balance of the atmosphere in two important ways. It changes the albedo of the sea surface and impedes the transfer of heat between the ocean and the atmosphere. While the albedo of the open sea is less than 20%, the albedo of the Arctic sea ice is between 60 and 80% in February and between 40 and 60% in August (Orvig, 1963). As a result, the heat absorbed in summer is less over the ice than over open water. The radiation absorbed depends not only on the surface albedo, but also on the cloudiness. At 75N in July, Orvig estimates a rate of absorption of 310 and 240 by day^{-1} in the Atlantic and Pacific sectors of the Arctic Ocean, respectively.

Sea ice reduces the rate of transfer of heat between the ocean and the atmosphere. The thermal conductivity of air-free sea ice is 0.0054 cal sec^{-1}(°C)^{-1} cm^{-1}. The presence of air bubbles, particularly in the upper layers of the ice, reduces this by as much as a factor of 5 (Zubov, 1943, p. 160). The freezing and melting of ice has the effect of further reducing the heat exchange between the sea water and the atmosphere. In summer, the heat is used to melt ice and so forms a stable, low salinity layer of water under the ice. In winter, cooling increases the ice thickness and so reduces the rate of heat transfer. As long as ice is present, the surface temperature of the water is essentially independent of the air temperature. The only changes result from the small variation with salinity of the freezing point of ice.

The net effect of the presence or absence of sea ice on the surface air temperature can be seen in Figs. 13 and 14. Fig. 13 shows the area in the Arctic that has a mean July temperature below 40F, while Fig. 14 shows the area having a January surface air temperature below -10F. The temperature difference between the Pacific and Atlantic sectors is particularly marked in winter. The summer difference is reduced by the cooling effect of the Greenland ice cap. Note that Greenland is surrounded by low salinity surface water (Fig. 6) due to the East Greenland Current (Fig. 23).

Any extension of sea ice will reduce the warming of the polar seas, both by reducing the absorption of solar radiation and by reducing the advective heating of the
atmosphere by ocean currents. A reduction in sea ice on the other hand, will lead to a warming of the polar atmosphere.

6. Why is the Atlantic saltier than the Pacific?

In the arguments that follow, I shall examine the effect on world climate of changing the salinity distribution of the Atlantic Ocean to one more like the Pacific. I shall attempt to show that such a change would lead to a glacial climate. In order that this argument be pertinent and not just an academic exercise, it is necessary to show that a redistribution of salinity in the World Ocean could have taken place during the Pleistocene. Because of the short span of time involved (of the order of a million years), the geography of the earth during the Pleistocene was essentially what it is today. The only significant changes were in response to continental glaciation; namely, changes in sea level to about 100 m below the present level and isostatic adjustments of the land surfaces that were glaciated.

If the salinity contrast between the oceans is the result of the different distribution of land masses about these oceans, then the present differences must have existed through the Pleistocene. The lowering of sea level results in an increase of the world average salinity from 34.7 to about 33.9%. The critical factor, however, is the contrast in the surface salinity between the Atlantic and Pacific.

The salinity distribution of the World Ocean has been discussed by Dietrich (1957, p. 158). He points out that the water vapor transport across the Isthmus of Panama contributes significantly towards maintaining the salinity of the North Atlantic higher than that of the North Pacific. This transport and its effect on the salt and water budget of the oceans has been investigated quantitatively by Deffeyes (1966). Because of its pertinence to the present discussion, I shall review his treatment of the salinity problem.

It is convenient to divide the surface of the earth into two hydrographic provinces, the Atlantic and Arctic Ocean and their drainage areas, and the rest of the world (Fig. 15). Water exchange between these provinces takes place by the ocean currents flowing through the Bering Strait and through the straits between the Antarctic continent, and South America and Africa. In addition, water can be transported in the vapor phase by winds blowing across the divides separating the hydrographic provinces. The vapor pressure of water doubles approximately for every 10°C rise in temperature. To carry a significant flux of water across the divides between the hydrographic provinces requires warm air of high relative humidity combined with high wind velocities that do not reverse direction with the seasons. These requirements are met only in areas where the moist warm trade winds are able to cross the divide at low elevations.

A survey of the divide suggests that the main gap for significant water vapor transport is located in Central America. Using the upper air data from Panama City for the years 1955–1964, Deffeyes estimates the water vapor flux across the Isthmus from the Atlantic to the Pacific to average $0.1 \times 10^8$ m$^3$ sec$^{-1}$ which, incidentally, equals about 10% of the flow of all the world's rivers.

Assuming that the World Ocean at the present time is in a steady state, Deffeyes arrives at the fluxes for salt and water between the divides shown in Table 1.

Let us now assume that at some time $T$ the water vapor flux across the Isthmus of Panama is cut off.

<table>
<thead>
<tr>
<th>Water Transport (10$^8$ m$^3$ sec$^{-1}$)</th>
<th>Salt Transport (10$^8$ metric tons sec$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Bering Strait: 0.66</td>
<td>35.7</td>
</tr>
<tr>
<td>Central America: -0.16</td>
<td>6.0</td>
</tr>
<tr>
<td>Drake Strait: 1.50</td>
<td>5.2</td>
</tr>
<tr>
<td>South Africa: -1.50*</td>
<td>6.25</td>
</tr>
</tbody>
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* Assuming South Africa = (Bering Strait + Central America + Drake Strait).

If the fluxes of water remained the same, the water level in the Atlantic would rise at a rate of about 5 cm yr$^{-1}$ relative to the rest of the World Ocean. As a result, within about a year, the sea water flows would readjust themselves and come to a new equilibrium. This new equilibrium for water flow, however, would lead to a nonequilibrium in salt transport. The salt content of the Atlantic would begin to decrease at a rate of about 10$^{10}$ metric tons yr$^{-1}$.

The initial change in surface salinity produced will be concentrated on both sides of Central America. Before a new steady state can be set up, the salinity in the Bering Strait and the Southern Ocean must be changed. This involves not only transport of the altered water by the wind-driven circulation, but also transfer of water from one major gyre system to another. Assuming a time delay of 100 yr and that the salinity of the Atlantic is altered to a depth of 100 m, we would obtain an average decrease in Atlantic surface salinity of 1%.

It is of interest to compare the salinity response at the present time with that in the late Tertiary when there was a Central American seaway (Simpson, 1950). The salinity of the water flowing through the seaway would respond rapidly to changes in the atmospheric water transport across Central America. As a result, a new steady state would be set up before the gross salinity of the oceans had changed significantly. In order that changes in atmospheric water transport change the salinity distribution between oceans significantly, the connecting seaways must be poorly coupled to the regions of atmospheric transport. For the present geography, the Atlantic-Pacific salinity distribution is sensitive to changes in wind pattern. However, the difference between the Pacific and the Indian is not.

The monthly average water vapor flux across Central America for the years 1956-1964 is shown in Fig. 16. It can be seen that there is a strong annual variation with minima in June and October. The greatest flux is in the Northern Hemisphere winter. Fig. 17 is a plot of the average summer and winter surface isobars for the region of interest. Unfortunately, the calculation of the water vapor flux had to be based on the upper air data from the single station at Panama City. To refine the estimates of the flux and its variability, added stations, particularly ones located near the divide, would be useful. The variation from year to year over the period investigated by Deffeyes is shown in Fig. 18. It is clear that the flux is variable with a large annual cycle and variations from year to year. The available record is too short to deduce long term trends. The data suggest, however, that minor changes in the atmospheric pressure distribution can lead to a marked change in the water vapor transport. If these changes persist over times of the order of a hundred years, they could lead to significant changes in the salinity distribution between the oceans. Deffeyes estimates that complete blocking of the Panama flux for 600 yr would remove the salinity contrast between the Atlantic and Pacific.

The major water vapor transport across the divides between the hydrographic provinces in areas remote from seaways is in the area of Panama. There may also be a significant flux from the eastern Mediterranean to the northeastern Indian Ocean (M in Fig. 15). While the work of Starr and Peixoto (1958) shows that this flux also acts to increase the salinity of the Atlantic, no analysis that considers the topography and, hence, permits a calculation of the flux across the divide, is available.
What would be the climatic consequences of this change? A simple answer can be obtained by drawing the maximum ice limit in the Atlantic at a comparable latitude to that in the Pacific (Fig. 24). Since the atmosphere cannot tell the difference between sea ice and ice covered ground, the apparent distributions of land and sea in the northern parts of the two oceans is then quite similar. As a result we would expect the average atmospheric surface isotherms to be roughly symmetrical about the pole with values similar to, or lower than, the Pacific values. It takes only a little imagination to suggest that these conditions will lead to extensive glaciation, particularly about the North Atlantic. The change in albedo due to the addition of land ice will reinforce the cooling effect. (See Fig. 22 for the maximum extent of continental glaciation during the Pleistocene.)

In the above simple analyses we consider the oceanic salinity as a flip-flop with the North Atlantic changing from its present salinities to a Pacific-like distribution. It is possible, however, to look at the change more detail. During the early stages we can use recent climatic data as a guide. As the North Atlantic proceeds further to the Pacific-like glacial phase, we can turn to the evidence from deep sea cores for confirmation. The larger the departures become from today's environment, the less sure will be our interpretation. It is hoped, however, that an attempt at a reconstruction will enable us to look at the evidence in a new way and will suggest the gathering of new data. These new data will undoubtedly force revisions in the reconstruction. I hope, however, that our understanding and data are sufficiently good so that the recon

Fig. 18. Average yearly water vapor flux across the Isthmus of Panama (Deffeyes, 1966).

Fig. 19. Average monthly water vapor flux at Panama City as a function of elevation (Deffeyes, 1966).

The importance of the elevation of the divide on the flux may be seen in Fig. 19 which shows the relative flux at Panama City as a function of elevation.

7. The consequences of reducing the North Atlantic surface salinity

Let us assume that the water vapor flux from the Atlantic hydrologic province is significantly reduced for several hundred years so that the salinity in the North Atlantic becomes more like that in the Pacific.
construction below is not entirely a work of science fiction.

Unfortunately, adequate salinity data for the North Atlantic goes back only to the beginning of this century. The data to 1939 have been studied by Smed (1943). By separately averaging the periods 1902–1917 and 1919–1939, he was able to show a significant change in the salinity of the North Atlantic between the two averages. In the northeastern Atlantic the salinity increased by about 0.1%. The change in the summer average is shown in Fig. 20 for each 5° × 5° square. For the same time interval, the average salinity of the northward flowing water through the Faroe-Shetland Channel increased by 0.08% both at the surface and a depth of 200 m (observations for the interval April-September). This salinity increase is due to either a general increase in salinity of the subtropical Atlantic or a change in the North Atlantic circulation, or both.

If the hypothesis of sea ice control by surface salinity is correct, the salinity increase in the northeast Atlantic during the first four decades should have led to a decrease in sea ice and a warming of the climate in the North Atlantic area. This appears to be the case. The data listed in Table 2 are taken from the work of Koch (1945).

Mitchell (1963) has investigated the variation in average world surface temperature for the time interval of interest. The temperature change by 10° latitude bands from the period 1900–1919 to 1920–1939 is plotted from his data in Fig. 21. It can be seen that the dominant change is a temperature increase in high northern latitudes. The temperature change of individual stations in the Arctic region is shown in Fig. 22. We note that the largest increases occurred in the North Atlantic sector. Of particular interest is the correspondence between the areas of temperature increase during the interval and the maximum extent of Pleistocene glaciation [shaded area in Fig. 22 after Daly (1934)].

While the above data from the period 1900–1940 do not prove a causal relationship among salinity, ice and climate, they are in agreement with the hypothesis. Schell (1956) has investigated the correlation between Arctic ice and the air and sea temperature in the North Atlantic. His abstract reads:

**Table 2. Ice in the North Atlantic [after Koch (1945), pp 126 and 258].**

<table>
<thead>
<tr>
<th>Weeks with ice around Iceland</th>
<th>Mean area of ice in summer in Eastern and Southern Greenland and Spitzbergen waters</th>
</tr>
</thead>
<tbody>
<tr>
<td>1900–1920</td>
<td>9.5</td>
</tr>
<tr>
<td>1920–1938</td>
<td>1.5</td>
</tr>
</tbody>
</table>

**Fig. 22. Average winter surface temperature change 1900–1919 to 1920–1939 in units of 0.1°F (Mitchell, 1963) and maximum extent of Pleistocene glaciation (Daly, 1939).**
It is shown that the mean sea-ice limit in the Greenland and Barents Seas during the period April-August is a substantial measure of both the sea-surface temperature near the Faroes and the east coast of Iceland, and the air temperature at Iceland, during the subsequent period September-February.

Furthermore, it is shown that decadal variations of Arctic ice are a measure of decadal trends of ocean phenomena and weather in the region of the northern North Atlantic Ocean.

If we attempt to go further back in time, we no longer have any observational data on salinity. We are limited, therefore, to asking if the climatic changes could have been brought about by a change in North Atlantic salinity. Of particular interest is the “Little Ice Age” from 1600-1900.

The atmosphere-ocean interaction associated with the “Little Ice Age” has been investigated by Bjerknes (1965) and Lamb (1964). The climatic data clearly show that this cool period in the North Atlantic area was associated with an expansion of glaciers and an extension of sea ice. The question to be answered is whether the ice was merely the result of the cooler climate or if the ice resulted from lower surface salinities and thus led to a cooler climate.

The critical oceanic areas are the European arctic and subarctic comprising the Norwegian, Greenland and Barents Seas. The physical oceanography of this area has recently been reviewed by Lee (1963). The oceanic circulation is the result of three forces:

1. The action of the wind stress on the sea surface (wind-driven surface currents).
2. Horizontal density gradients of the deep water leading to overflow into adjacent oceanic basins (the thermo-haline circulation).
3. Sea level differences set up by water flow and the difference between evaporation and precipitation (maintenance of the water budget).

Excursions in temperature, or salinity, will have different effects on the thermohaline circulation. An increase in surface cooling will result in greater formation of bottom water and, hence, in more overflow. To maintain the water level, more surface water must flow into the subarctic from the North Atlantic. The advection of heat and salt by this water will lead to negative feedback. Since the distances involved are of the order of 1000 mi and the current drift is of the order of 5 mi day⁻¹, the time constant of the negative feedback will be about a year. It will thus tend to prevent a succession of extreme years. That a negative feedback mechanism must be operative is shown by the curve of mean summer ice quantity from 1898-1938 (Koch, 1945, p. 126). The years of extremely large or small ice area are followed by more average conditions. The only exceptions are the years of extreme ice years 1906-1907 and 1917-1918. Negative feedback operates both because of a direct advection of heat and because of the effect of salt advection on the ice limit.

A change in the North Atlantic surface salinity, on the other hand, will result in positive feedback. The lowered surface salinity will reduce the rate of formation of bottom water in the Greenland-Norwegian Seas and thus reduce the advection of saline water from the North Atlantic. As a result, the ice limit will progress southward and reinforce the cooling due to the decreased advection of warm Atlantic water.

As might be expected, the review of Lee (1963) shows that the intensity of the currents has a strong seasonal and year-to-year variation. Since the feedback relationships among climate, ice and ocean circulation are different for the wind-driven surface currents and the thermohaline overflow, we must estimate the relative importance of these two processes. Unfortunately, adequate data are not available to obtain yearly average flows due to the two mechanisms. Representative data are given in Table 3. About all that can be said is that the two terms are of the same order of magnitude.

The schematic surface circulation in the North Atlantic and North Pacific are shown in Fig. 23. The January location of the dominant atmospheric surface pressure lows, the Alaskan in the Pacific and the Icelandic in the Atlantic, are also indicated. The feedback relations between the climate and the wind-driven surface currents will not be as direct as the relations with the thermohaline circulation. An intensification of the Icelandic low will tend to increase the northward advection of heat and salt, while a southward displacement of the low will tend to reduce advection.

My interpretation of the processes leading to the “Little Ice Age” is as follows. The surface salinity of

<table>
<thead>
<tr>
<th>Water balance for the arctic-subarctic</th>
</tr>
</thead>
<tbody>
<tr>
<td>Surface currents [from Lee (1965)]</td>
</tr>
<tr>
<td>(10⁶ m³ sec⁻¹)</td>
</tr>
<tr>
<td>Inflow of Atlantic water</td>
</tr>
<tr>
<td>Shetland-Iceland</td>
</tr>
<tr>
<td>Inflow Bering Strait</td>
</tr>
<tr>
<td>Runoff</td>
</tr>
<tr>
<td>Precipitation-evaporation</td>
</tr>
<tr>
<td>Outflow Denmark and Davis Strait</td>
</tr>
<tr>
<td>Overflow [from Worthington and Volkman (1965)]</td>
</tr>
<tr>
<td>Faroe-Iceland</td>
</tr>
<tr>
<td>Denmark Strait</td>
</tr>
</tbody>
</table>

(June 1958 regular)
the North Atlantic is reduced, due to a decrease in the water vapor flux across Panama. This leads to an expansion of sea ice in the Atlantic sector with resulting cooling. Due to the lower surface salinity in the Norwegian-Greenland-Barents Seas, the rate of formation of bottom water and, hence, the thermohaline part of the northward advection of heat and salt, is reduced. The "Little Ice Age" was terminated by an increase in North Atlantic surface salinity resulting in a retreat of sea ice and general warming.

Bjerknes (1965) has suggested an alternate mechanism by which an increase in salinity may lead to warming. He points out that an increase of the salinity of the water moving into the Norwegian Sea would lead to the formation of bottom water at higher temperatures. This would lead to an increase in the intensity of the thermohaline circulation and, hence, to an increased advection of heat.

Having shown that the recent minor climatic variations are at least not inconsistent with my thesis, I must now investigate the Pleistocene glacialations. Let us assume that the surface salinity of the North Atlantic is reduced more strongly and for a longer period of time than we assumed for the "Little Ice Age". As a result, at first the thermohaline circulation in the Greenland-Norwegian sea is reduced and, because of an expansion of the low surface salinity area in the Atlantic portion of the Arctic, the sea ice limit moves southward. Positive feedback comes into play and the cooling and advance of the sea ice is intensified.

The decrease in surface salinity will lead to a reduction in salinity of the water at the deep salinity maximum (Fig. 5) which results from a mixture of the deep water overflowing across the Faroe-Greenland ridge and water from the Mediterranean. This decrease would be particularly marked if there were also a blocking of the water vapor export from the eastern Mediterranean to the Indian Ocean. After several hundred years, this reduction in salinity will be felt in the Southern Ocean. As a consequence, the salinity deficit near the Antarctic will increase and the rate of bottom water formation will decrease, leading to an extension and thickening of sea ice around Antarctica.

This southern response to the salinity decrease in the North Atlantic did not occur during the "Little Ice Age" because of its short duration. As a result the cooling was largely restricted to high northern latitudes [see Mitchell (1963)]. Once the southern response sets in, the Southern Hemisphere will also experience cooling and the deep circulation of the World Ocean is reduced.

That there was indeed an extension of sea ice around Antarctica is shown by the work of Conally and Ewing (1965). They find that during the glacial periods, ice rafted detritus extended further north than in interglacial periods. Sediment core V 17-121 of the Lamont Geological Observatory, collected at 44S, 52W, contains three layers with more abundant large grains. At the present time, the ice rafted glacial marine sediment at this longitude extends only about 65S.

Global cooling due to the extension of sea ice and the reduction in meridional heat transport by the deep circulation of the ocean will lead to glaciation. The increase in albedo due to snow cover will further increase the general cooling. So far, as long as the North Atlantic salinity remains depressed, the mechanisms have led to positive feedback reinforcing the cooling trend. Now, however, we get stronger negative feedback as water is stored on land in the form of ice. The distribution of glacial ice (Fig. 22) suggests that the water was derived mostly from the North Atlantic. Therefore, the salinity of this ocean must increase and tend to reverse the cooling trend. Negative feedback must have set in well before sea level dropped to its glacial minimum, about minus 100 m, for this amount of water removal leads to an increase of the average oceanic salinity from 34.7 to 35.9 %. To see how negative feedback may have been prevented, we must consider the oceanic surface circulation.

As previously indicated, the present surface circulations are shown schematically in Fig. 23, where the North Pacific circulation is from Döös and Hed (1962, Fig. 109). In this ocean the saline warm water from the Kuroshio current is kept south of the 40th
North Atlantic where the cold, left coiling form dominates at the present time is shown in Fig. 23 [after Ericson and Wollin (1964, Fig. 7)].

If our reconstruction of the surface currents during the glacial period is correct we would expect the right coiling *Globigerina pachyderma* to be restricted to the area south of 40°N. D. Ericson has kindly supplied the information on coiling direction given in Table 4, the data for the glacial interval being indicated in Fig. 24. Ericson writes:

> All we can tell from the data is that the 7°C isotherm must have been between 32°N and 46°N at 18° 30'W longitude and that at 43° 30'W longitude it must have been south of 40° 30'N latitude, but perhaps not much south because the coiling below the top at 40° 34' is only 69% left, whereas it is 96% left at 41° 31'N. Unfortunately, this zone, the vicinity of 40°N, has been cored hardly at all since 1959. If more cores were taken there, I feel fairly sure that the former position of the isotherm could be fixed with reasonably good precision.

Ericson’s data thus confirm our reconstruction of the glacial circulation in the North Atlantic. We would expect that the surface circulation in the North Pacific was not very different during glacial times from what it is today. This seems to be confirmed by core data. D. Ericson (private communication) further writes:

> “We are now working on the planktonic foraminifera in long cores from the Pacific. We are finding that there is a striking difference between the Pleistocene record of the Atlantic and the Pacific. The abrupt, well-defined changes in the Atlantic are absent in the Pacific cores.”

The reconstruction of the glacial surface circulation in the North Atlantic suggests that once the Icelandic low has been displaced sufficiently far south, the changed circulation will prevent the warm salty water of the Gulf Stream extension from reaching the sea ice, thus inhibiting negative feedback due to land

<table>
<thead>
<tr>
<th>Core no.</th>
<th>Latitude (°N)</th>
<th>Longitude (°W)</th>
<th>Depth in core (cm)</th>
<th>Coiling direction</th>
</tr>
</thead>
<tbody>
<tr>
<td>A 157-11</td>
<td>41° 31'</td>
<td>43° 24'</td>
<td>0</td>
<td>mostly R</td>
</tr>
<tr>
<td>A 157-13</td>
<td>40° 34'</td>
<td>43° 51'</td>
<td>0</td>
<td>mostly R</td>
</tr>
<tr>
<td>R 5-36</td>
<td>46° 55'</td>
<td>18° 35'</td>
<td>0</td>
<td>mostly R*</td>
</tr>
<tr>
<td>S18-4</td>
<td>32° 50'</td>
<td>18° 32'</td>
<td>entire</td>
<td>100% R</td>
</tr>
</tbody>
</table>

* Top of core is deep sea foraminiferal lutite, going to glacial marine at 30 cm.
glaciation. The latitudinal variation of the Icelandic low since 1800 has been discussed by Lamb (1964). In particular, he shows a northward shift of about 2.3° between the 40-yr averages 1880-1919 to 1915-1954. To prevent negative feedback due to continental glaciation, a southward displacement of about 10° from the present position would be required.

Fletcher (1967) writes: "The Icelandic and Aleutian lows are at about the same latitude." His charts (Figs. 29 and 30) show the approximately 10° difference in latitude, and the present discussion indicates that a southward shift of the Icelandic low to the latitude of the Aleutian low would have a profound effect on the circulation of the North Atlantic.

8. The deep circulation of the ocean

So far I have attempted to infer the response of the climate to changes in the salinity of the North Atlantic. It now becomes necessary to investigate the interaction between the deep circulation of the ocean and the climate. Paleo-oceanography can be studied in two ways. By looking at the ocean of today, we can deduce the forces that drive the present circulation. We then investigate how these forces would be affected by climatic change and so infer the circulation of the past. Another procedure is to look at the evidence from the past that is preserved in the deep sea sediments. We then use these data to reconstruct the past circulation. As we shall see, the interpretation of the data is difficult, unless we already have some ideas about the circulation of the past. I shall begin, therefore, with a reconstruction based on the dynamics of the ocean circulation and then compare reconstruction with the geologic data.

Unfortunately, our mechanistic understanding of the deep circulation of the ocean is in a primitive state. Simple theoretical models have been explored by Stommel (1958) and Robinson and Stommel (1959). The usual approach is to consider the ocean to be in a steady state. This is a good assumption for the present ocean, since its turnover time is only of the order of one thousand years, while sea level has been at its present high stand for the last 5000 yrs (Broecker, 1965), and the climatic changes during this period have been minor compared to the glacial-interglacial variations. When we consider the response of the deep ocean to glacial changes, however, a steady state model is not adequate. Lack of time and space does not permit me to develop the subject in detail. I hope, however, that the following simple considerations will provide insight into the behavior of the deep ocean circulation during the Pleistocene.

Let us consider the energy of the ocean. It contains potential and kinetic energy. The potential energy manifests itself in the departure of the surfaces of constant density from the horizontal. This is illustrated for the Pacific in Fig. 25, a meridional density section from Antarctica to Alaska, simplified after Reid (1965). If the ocean were to come to rest without diffusion of heat and salt, the isopycnals would become horizontal with density increasing downwards. Since energy is dissipated by eddy viscosity, work must be done to maintain the present potential energy.

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![Figure 25](https://example.com/figure25.png)

**Fig. 25.** North-south density profile of the Pacific Ocean from Antarctica to Alaska. [From Reid (1965, Fig. 6).]
The Coriolis force, since it acts normal to the direction of motion, does no work.

In order to maintain the distribution of density, it is necessary to form bottom water which sinks in the Antarctic region. In addition, surface density gradients must be maintained by the exchange of heat and water across the sea surface. If these were the only active forces, the density section (Fig. 25) would look very different. Since most of the ocean has a stable density stratification and since the solar heat is absorbed in the top few tens of meters, the ocean would be filled with water of maximum density, except for a thin skin of lower density water not exceeding a depth of 100 m. The actual depth of the low density layer is at least an order of magnitude greater. Thus, a force that depresses the low density layer, and hence increases the potential energy of the ocean, must exist.

a. Vertical eddy diffusion. Since the surface density of the subtropical and tropical water is low because of its elevated temperature, a mechanism that is able to transfer heat down from the sea surface exists. Let us, therefore, look at the depth of the 10C isotherm in the Pacific (Fig. 26). This chart is redrawn from the study of Muromtsev (1963). In winter the 10C isotherms intersect the sea surface at approximately 42N and 45S. The isotherm has two deep troughs with depths in excess of 600 m near the western margins of the Pacific, at a latitude of about 30°. The isotherm becomes shallower towards the east and toward the equator.

In the North Pacific the maximum depression of the 10C isotherm lies inside the curve where the western boundary current, the Kuroshio, turns east. In the South Pacific the region of depression and the western boundary currents are more complex [for a description of the currents see Weylki (1962)]. Because of the stable density stratification heat cannot be transported downwards by sinking. This leaves vertical eddy diffusion as the only possible transport mechanism. Since mixing leads to an increase in entropy it might be thought that vertical mixing would proceed spontaneously. Due to the small molecular diffusion constants of heat and salt (10^{-3} and 10^{-2} cm^2 sec^{-1}, respectively) however, this is not the case on the time-space scale of the ocean. Only eddy diffusion is rapid enough to be significant.

Consider the step change in density at a depth d shown in Fig. 27a. For small changes in temperature and salinity, density is a linear function of these variables. Let eddy mixing convert the step change in density (in T and S) into a linear gradient in density over a distance 2d (Fig. 27b). The potential energy is given by

\[ PE = -g \int_a^b \rho(z) z \, dz. \]
Integrating the density distributions in Figs. 27a and b, we find that diffusion has increased the potential energy of the water column by an amount $\frac{1}{2} \Delta \rho \phi d^2$. Eddy diffusion diminishes the scale of the density gradients sufficiently so that molecular diffusion can become effective. The resulting increase in entropy makes the increase in potential energy irreversible.

That vertical diffusion, even at the sea surface, on a scale of tens of centimeters and several months, does not take place spontaneously is shown by the practices of the makers of solar salt (Ver Planck, 1958). Precipitation during the rainy period forms a fresh water layer on top of the brines in the salt evaporating ponds of San Francisco Bay. When the rains stop, this fresh water layer is pumped off and concentration of the brines continues. The formation of a thin stable layer of sea water on top of denser evaporite brines was also observed in the Pekelmeer on the island of Bonaire, Netherlands Antilles (Defays et al., 1965, Fig. 10).

The energy necessary to drive the vertical eddy diffusion is derived from the wind-driven surface currents, particularly the Western Boundary currents. The depression of the 10°C isotherm by the meanders of the Gulf Stream can be seen clearly in the 1960 Gulf Stream survey (Fuglister, 1963, Fig. 4). In three isolated patches, as well as near the northwestern margin of the stream, the isotherm is depressed below 900 m. The depression of the low density layer in the Atlantic is shown well in the charts of the depths of the isopycnals published by Montgomery and Pollack (1942). The general distribution is similar to that in the Pacific (Fig. 26).

In addition to the areas of vertical eddy diffusion associated with the Western Boundary currents, there exists a diffusion zone associated with the strong circumpolar Antarctic currents in the Southern Ocean. Because of the lower temperatures, this cannot be seen on the charts of the 10°C isotherm. It manifests itself in the depression of the isopycnals near 55° in the section of the Pacific (Fig. 25). The Indian Ocean section (Fig. 9) shows a steep drop of the isotherms in the same region. Because of geography, this feature is absent in the Northern Hemisphere. The resultant difference between the intermediate waters of the North and South Pacific has been discussed in detail by Reid (1965).

The energy to produce vertical eddy diffusion, and so drive the thermohaline circulation, is derived from the turbulence in the regions of current shear of the wind-driven surface currents. Thus, there is a nonlinear interaction between the wind-driven and the thermohaline circulations. Theories of the wind-driven ocean circulation [for a compilation of papers see Robinson (1963)] have assumed either a uniform density, a fixed density distribution, or fixed values for the coefficient of vertical eddy diffusion. These theories, therefore, are not capable of treating the time-dependent interaction between the two types of circulation. Until an adequate theory is developed, we are limited to qualitative insight derived from a knowledge of the mechanisms.

In summary, the relationship between the wind-driven and the thermohaline circulations is as follows. If the density of the ocean were uniform, the wind-driven currents would extend with uniform velocity to the ocean bottom. This is because in a uniform density field the horizontal pressure gradients generated at the surface cannot be compensated for at depth. If thermohaline processes were acting in the absence of wind stress, horizontal density gradients could be generated only in the surface skin of the sea, and there would be no deep circulation. A combination of the two processes, however, leads to an active circulation at all depth. The turbulence produced by the wind-driven circulation forces the low density surface layer downwards. This produces horizontal density gradients at depth which limit the downward extent of the surface currents and drive the thermohaline circulation. If, due to climatic change, the thermohaline density gradients in the surface layer should be decreased, the thermocline (the region of low density water where the temperature has a large vertical gradient), will be deepened. This increase in depth will feed energy back into the thermohaline circulation. An increase in surface thermohaline processes, on the other hand, will drive the thermocline toward the surface and so decrease the energy pumped into the thermohaline circulation. Negative feedback, thus provided, reduces the amplitude of the variations in the deep circulation, in response to climatic change.

b. Geothermal heating. In addition to thermohaline
density gradients produced at the surface and vertical eddy diffusion generated by the wind-driven circulation, the ocean is heated at the bottom by the geothermal flux. If there is no advection of bottom water, it is heated, and as a result, the density decreases with time. The time rate of change of density for a 1-km column is shown in Fig. 28 as a function of the temperature. At the present time, the main effect of geothermal heating is to keep the water in isolated basins in a state of vertical convection, provided the vertical temperature gradient is close to adiabatic. This prevents stagnation in the Arctic Ocean Basins (Coachman, 1963) and in the deep ocean trenches.

Geothermal heating becomes extremely important when, in response to climatic change, the maximum density of the water formed at the surface decreases. In this situation, the geothermal heat flux limits the time during which the bottom water stagnates, for sooner or later, its density is reduced to the new maximum density of the surface water.

9. The formation of bottom water during the glacial period

At the present time, bottom water forms near the Antarctic and in the Norwegian Sea (Coachman, 1963). Both types of bottom water are formed near or under sea ice and have temperatures below 0°C. According to the hypothesis proposed in this paper, a glacial climate develops because of an expansion of sea ice cover. The sea ice expansion, in turn, is the result of a stable salinity stratification near the ice limit. If this hypothesis is correct, bottom water during the glacial period cannot form near sea ice and, as a result, must be significantly warmer than 0°C.

a. The paleotemperature data. We can obtain paleotemperature information on the bottom water by looking at the oxygen isotopic composition of benthic foraminifera obtained from deep sea cores. The ratio of O\(^{18}/O\(^{16}\) in the carbonate tests of these animals depends on both the isotopic composition of the water in which they live and on the temperature. During the glacial period, a fraction of the water now in the ocean was held in the continental ice sheets. As a result, the average isotopic composition of the oxygen in seawater during the glacial period was different from its present average. Let us first estimate this change.

The oxygen isotopic composition is expressed as a \( \delta \) value where \( \delta O\(^{18}\) \) is the difference in the O\(^{18}/O\(^{16}\) ratio from that of a standard expressed in parts per thousand. The mean \( \delta \) value of the additional ice during the glacial period has been estimated as:

<table>
<thead>
<tr>
<th>Source</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Emiliani (1955)</td>
<td>-15</td>
</tr>
<tr>
<td>Olausson (1965)</td>
<td>-33</td>
</tr>
<tr>
<td>Weyl (this paper)</td>
<td>-40</td>
</tr>
</tbody>
</table>

Emiliani assumed that the composition of the ice would be similar to that of the present day precipitation falling on the same areas, while Olausson derived his estimate from a consideration of the elevation and surface temperatures of the ice sheets. He obtained an upper limit on the value by assuming that the isotopic composition of the ice was independent of depth below the ice surface.

At the present time, the isotopic composition of surface ice in the Antarctic varies from -20 to -30 at the periphery to -50 at the south pole. Values from Greenland range from -20 to -35 (Emiliani, 1966). The isotopic composition of precipitation depends on the fraction of the original water vapor that makes up the precipitation on the ice sheet. As the water vapor content of the air is reduced, the \( \delta O\(^{18}\) \) of the residual vapor becomes more and more negative. The isotopic composition of the ice caps thus depends on the storm tracks that produce the precipitation. Due to the change in North Atlantic circulation during the glacial period (Fig. 24) and their larger areal extent, the northern ice sheets must have had a more negative value than the present Greenland ice. The Antarctic values are probably a better estimate of glacial conditions. Due to the motion of the ice, most of the ice volume will consist of ice deposited at the higher elevations and more central positions. The marginal surface ice can make up only a small portion of the total volume. Averaging the two Antarctic ice values gives -35. The weighted mean should be more negative than this, leading to my guess of -40.

During the glacial period, sea level was about 100 m below its present stand. Allowing for isostatic adjustment of the sea floor, this leads to a removal of 130 m of water or 47 \( \times 10^6 \) km\(^2\) of water out of a present oceanic volume of 1.37 \( \times 10^6 \) km\(^2\). This leads to a change in the \( \delta \) value of the glacial ocean of +1.4. This can be compared with estimates of +0.4 by Emiliani (1955) and +1.05 by Olausson (1965).

Emiliani (1966) has introduced another argument to minimize the isotopic change of the average ocean water during the glacial period. He writes: “An isotopic variation of the bottom water larger than that given by the foraminifera would require bottom temperatures during glacial ages to be higher than during interglacial ages. This is a clear impossibility because of the rapid vertical mixing of the ocean today and, if at all, even faster mixing during glacial ages.”

I cannot follow Emiliani’s argument. If there were rapid vertical mixing in the ocean today, the bottom temperature should reflect the average temperature of the sea surface and be above 10°C. The bottom temperature depends on the temperature of the surface water that has the maximum density and, hence, sinks.
or is connected to the bottom. At the present time, this is slightly above the freezing point. If the hypothesis here proposed is correct, not only can the bottom water during the glacial period have been warmer than today, but it must have been so, in order to permit a maximum expansion of sea ice.

The oxygen isotopic variation in the present ocean has been reviewed in detail by Craig and Gordon (1965). A summary of the data from their work is shown in Table 5. The values are relative to Standard Mean Ocean Water (SMOW). The world average oceanic value, keeping in mind that the surface water accounts for only about 4% of the total water, is between -0.1 and -0.2. The deep water in the Pacific is a good measure of the present day 80°S average and probably also was a good measure of the glacial average. It is, therefore, important to look at the 80°S values of benthic foraminifera from the Central Pacific during the glacial phase. Emiliani (1955) found the following glacial-interglacial 80°S changes in cores from the Central Pacific:

Core 58: -0.08 ± 0.16
Core 60: +0.51 ± 0.22
Average: +0.2 ± 0.2

At the present time, the bottom temperature at the location of the cores is 1.7°C. Correcting for the glacial change in isotopic ratio of the water, we find that the glacial value for the benthic foraminifera was -1.2, giving a glacial temperature of 6°C. Using Olausson’s estimate we get a bottom temperature of 4.7°C. There can be little doubt that the temperature of the bottom water in the Pacific during the glacial period was significantly warmer than it is today. A more accurate estimate will have to await a meteorological reconstruction of precipitation during the glacial period.

In view of the large isotopic variations of the surface water, even at the same salinity (Table 5), the interpretation of the isotopic data from pelagic foraminifera, particularly in the Atlantic, is much more difficult. Both the surface currents and the dynamics of water vapor exchange among the ocean, the atmosphere and the ice sheet must be considered.

Typical glacial-interglacial changes in 80°S for pelagic foraminifera are of the order of 1.5 (Emiliani, 1966). As pointed out by Olausson (1965), these variations do not differ significantly from the isotopic change of the average ocean water. The intense glacial cooling (at the order of 3°C) of tropical surface waters claimed by Emiliani are, therefore, not supported by the evidence.

In this connection, a study of the instrumental record from the “Little Ice Age” by Lamb (1964) is pertinent. Comparing the temperatures in the Atlantic from 1780-1820 with those from 1887-1899 and 1921-1938, he finds significant warming in the waters north of 40°N, associated with climatic warming. Between 15 and 40°N, on the other hand, the subtropical waters became cooler (Lamb, 1964, Fig. 3). He attributes this to a more southward track of the Gulf Stream during the “Little Ice Age.” During the glacial period, this effect must have been even stronger (see Fig. 24), and so the surface cooling claimed by Emiliani is suspect on purely climatological grounds. This leaves the faunal changes documented by Ericson (Ericson and Wollin, 1964) to be explained by ecological factors other than temperature. That the distribution of one of their fossils (Globorotalia truncatulinoides) depends on neither temperature nor salinity is shown by its present North Atlantic distribution (their Fig. 6).

b. Glacial bottom water. The removal of 47 × 10^8 km³ of water as glacial ice will increase the average oceanic salinity from 34.7 to 35.9%. If, as at the present time, a major portion of water in the ocean is bottom water, its salinity must have been close to the average value. Thus, glacial bottom water must have had a salinity

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**Table 5. Oxygen isotope data for the Present Ocean [from Craig and Gordon (1965)]. 80°S values in ‰ relative to SMOW.**

<table>
<thead>
<tr>
<th>Category</th>
<th>Sample</th>
<th>Salinity (‰)</th>
</tr>
</thead>
<tbody>
<tr>
<td>A. Surface waters</td>
<td>33</td>
<td>34 35 36 37</td>
</tr>
<tr>
<td>N. Equatorial Atlantic</td>
<td>0.5</td>
<td>0.65 0.8 1.1</td>
</tr>
<tr>
<td>Norwegian Sea</td>
<td>0.0</td>
<td></td>
</tr>
<tr>
<td>N. Pacific</td>
<td>-0.7</td>
<td>-0.15 0.4</td>
</tr>
<tr>
<td>S. Pacific</td>
<td>-0.4</td>
<td>0.25 0.5</td>
</tr>
<tr>
<td>Indian</td>
<td>-0.4*</td>
<td>0.35 0.6*</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Subsurface waters</th>
<th>80°S (‰)</th>
<th>Salinity (‰)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Antarctic bottom water</td>
<td>-0.4</td>
<td>34.7</td>
</tr>
<tr>
<td>N. Atlantic deep water</td>
<td>0.1</td>
<td>34.9</td>
</tr>
<tr>
<td>Pacific bottom water</td>
<td>-0.1</td>
<td>34.7</td>
</tr>
<tr>
<td>Indian bottom water</td>
<td>-0.2</td>
<td>34.7</td>
</tr>
<tr>
<td>Southern salinity minimum</td>
<td>-0.15</td>
<td>34.5</td>
</tr>
</tbody>
</table>

* Extrapolated values.
circulation in the Atlantic is shown in Fig. 30. The deep circulation in the Pacific would be similar to that of today, except that the bottom water would be formed in the North Atlantic rather than in the Antarctic.

The reconstruction of the glacial circulation will undoubtedly require modification as our understanding of the present ocean is improved. Also, new and existing geologic data will have to be integrated into the scheme to describe the circulation in more detail. Fig. 30 should be considered as a first crude attempt in paleo-oceanography rather than as a final reconstruction.

10. Changes in the deep circulation in response to climatic change

So far I have considered the deep circulation only in its present and glacial state. We must now consider how the deep circulation changes from one phase to the other.

a. Interglacial-glacial. The reduction of North Atlantic salinity stops the formation of Arctic bottom water and reduces the salinity of North Atlantic deep water. The resultant lowering of salinity in the Antarctic stops the formation of Antarctic bottom water. Meanwhile, a subarctic gyre forms in the North Atlantic and the formation of North Atlantic intermediate water is shifted southward. As water is extracted to form the glacial ice caps, the salinity of the North Atlantic increases, resulting in an increase in density of the North Atlantic intermediate water, which now becomes Glacial bottom water. Meanwhile, the formation of intermediate water in the Southern Ocean continues, although the area of formation is shifted northward by expanding Antarctic sea ice.

The changes in the Indian and Pacific Ocean circulation are smaller. Antarctic bottom water is replaced by the warmer, saltier Glacial bottom water that forms in the North Atlantic. As a result, the average salinity in the Indian and Pacific Ocean increases, although the relative salinity probably does not change significantly. However, shifts in the atmospheric circulation may change the distribution of evaporation and precipitation somewhat and hence lead to minor changes in the surface salinity distribution.

b. Glacial-interglacial. Three possible electrical analogs for the Pleistocene climatic fluctuations exist: the oscillator, the univibrator and the flip-flop. In the first case, the climatic changes lead to auto-oscillations without requiring an external trigger. In the case of the univibrator, external triggering causes the climate to change to a glacial phase and then return to an interglacial one. In the flip-flop, the return to interglacial conditions requires external triggering.
Broecker (1966) has presented a flip-flop model, triggered by changes in insolation due to geometrical variations in the earth-sun system. In my hypothesis, the transition to glacial conditions is triggered by a change in the atmospheric moisture transport. As the salinity in the North Atlantic is depressed, the formation of a subarctic gyre makes the sea ice limit insensitive to further salinity variations. We must search, therefore, for another mechanism to prevent the glacial climate from becoming permanent.

Emiliani and Geiss (1957) have suggested that land glaciation may lead to overshooting. Cooling and addition of glacial ice is furthered by the elevation produced by the accumulated ice sheets. As the land surface is lowered by the excess ice slowly sinks to reestablish isostasy, the ice volume is reduced. Most of the melt water will be added to the North Atlantic and tend to reduce its salinity. The reduction in salinity would reduce the maximum density of surface water in the North Atlantic and so stop the formation of Glacial bottom water. Bottom water formation can recommence only after geothermal heating has raised the density of the bottom water to the new surface values. Once this state is reached, the vertical mixing of the surface water with the warmer, saltier Glacial bottom water may provide the necessary conditions to return the climate to an interglacial phase. It is important, therefore, to examine the geologic evidence to see if stagnation near the termination of glacial periods could, in fact, have occurred.

During stagnation, biological respiration in the deep water continues until all the oxygen is used up. Without oxygen, the deep sea fauna become extinct and organic material is preserved in the sediment. First, let us look at the present rate of oxygen consumption in the deep ocean.

Using crude assumptions, Weyl (1965) has estimated the present residence time of oxygen in the deep Pacific as 4000 yr. A more reliable estimate can be obtained by combining the deep oxygen data compiled by Wooster and Volkman (1960) with natural radiocarbon data (Bien et al., 1966). We find the following:

<table>
<thead>
<tr>
<th>Location</th>
<th>$\Delta^{14}N$</th>
<th>Age (yr)</th>
</tr>
</thead>
<tbody>
<tr>
<td>near New Zealand</td>
<td>4.6</td>
<td>260</td>
</tr>
<tr>
<td>N. E. Pacific</td>
<td>3.4</td>
<td>120</td>
</tr>
<tr>
<td>Total</td>
<td>-1.2</td>
<td>120</td>
</tr>
</tbody>
</table>

At this rate, the 4.6 ml per liter of oxygen would be used up in about 5000 yr while more oxygenated water with 7 ml per liter would require 7500 yr before it becomes exhausted.

Contrary to early expectations, the bottom of the sea is not inhabited by survivors from the distant geologic past. In a review paper, Brunn (1956) states that "the true abyssal fauna is derived from the bathyal fauna, from which such species can live under the high pressure and at lower temperatures have spread downwards." Discussing the deep sea mollusk fauna, Clarke (1962) writes:

In fact, on the basis of our present knowledge concerning which feeding types are likely to be successful in the abyssal environment, one may speculate that if the whole abyssal molluscan fauna were wiped out today, in a few million years re-population from shallower regions would produce a fauna which would be very much like the present one.

The faunal evidence thus suggests that the abyssal fauna could have been reintroduced into the deep fairly recently. Whether the few million years of Clarke could be reduced to the 10,000 yr since the last glacial remains to be investigated. At least, the depletion of oxygen near the termination of the last glacial period does not appear to be contradicted by the biologic data.

Pleistocene deep sea sediments in the open ocean have not revealed any layers of organic rich sediments such as would be deposited under anoxic conditions. This does not preclude short periods of oxygen depletion. Benthic animals, by burrowing, could reoxidize thin sediment layers (up to 25 cm). At a sedimentation rate of 3 cm (1000 yr)$^{-1}$ (Emiliani, 1955), the maximum time for anoxic conditions is less than 10,000 yr.

A possible clue to stagnation could be obtained from $^{14}C$ data. $^{14}C$ is produced in the atmosphere and is mixed with non-radioactive carbon in the atmospheric and the oceanic reservoir. The oceans contain about 50 times as much carbon as the atmosphere. As pointed out by Broecker et al. (1960), the $^{14}C$ concentration in the atmosphere depends on the rate of oceanic mixing. At the present time, the average $^{14}C$ content relative to $^{14}C$ in the ocean, is about 16% less than that of the atmosphere. Stagnation of the deep ocean, or a reduced deep circulation, would lead to less mixing into the ocean and hence to a higher $^{14}C$ concentration in the atmosphere. This would be reflected in lower apparent $^{14}C$ ages compared to the true time scale. So far such comparisons have been carried back only 6000 yr (Damon et al., 1966).

A drastic change from stagnant conditions may lead to apparent time reversals in long continuous cores. The older layers during stagnant conditions, because of the high $^{14}C$ content of the atmosphere, would look younger than the superposed later layers during times when mixing was more effective. To check for this effect, continuous cores deposited under anoxic conditions should be examined in detail.
From the above, it appears that while long periods of stagnation (longer than 15,000 yr) are clearly excluded, short time stagnation could have occurred. It should be possible to obtain a more definite answer in the near future. At a bottom temperature of 5°C, the increase of density per kilometer water column is 0.5σρ units in 10,000 yr (Fig. 28). This is sufficient to raise the density of 2.5 km of water from the density of Glacial bottom water to that of Arctic bottom water.

Let us assume that maximum development of glaciers is followed by partial melting, thus reducing the North Atlantic salinity. This stops the formation of Glacial bottom water. Stagnation follows, and the oxygen in the deep is consumed. The reduction in the deep circulation will lead to a deepening of the thermocline by vertical eddy diffusion. At the same time, convection in the Antarctic region will tend to deepen the Glacial intermediate water. Thus, the volume of the ocean that derives its oxygen from the Glacial bottom water is reduced, and the thickness of the column of Glacial bottom water that has to be heated by the geothermal flux is decreased.

In the absence of a quantitative theory, it is not possible to evaluate the importance of the above effects. It appears likely, however, that the oxygen of the bottom water is depleted before geothermal heating has raised the bottom water temperature sufficiently to permit convective mixing to the surface. Due to year-to-year fluctuations, the termination of stagnant conditions is probably not sudden. Rather, as the bottom density decreases, the rate of bottom water formation will increase. This process will probably start in the subarctic gyre of the North Atlantic. Vertical convection will add warm salty bottom water to this gyre. As a result, vertical mixing will be intensified, and the increase in temperature and salinity will push the sea ice limit northward.

The new bottom water will have a density closer to that of the Glacial intermediate water and will be of higher temperature and salinity. Vertical diffusion in the Antarctic region can lead, therefore, to the formation of Antarctic bottom water with a resultant retreat of the Antarctic sea ice. The reduction in sea ice in the North Atlantic and the Antarctic leads to warming and a melting of the glaciers. The addition of 30 × 10^6 km^3 of melt water to the North Atlantic over a period of 10,000 yr gives a flux of 0.1 × 10^9 m^3 sec^-1. The present water vapor flux across the Isthmus of Panama is thus adequate to maintain the North Atlantic salinity and prevent an immediate return to glacial conditions.

11. Conclusions

My study of the role of the ocean in climatic change started only in July 1965. Having rashly promised to say something on the subject at the INQUA-NCAR symposium, I began to look at the problem. The present paper should be considered, therefore, a progress report rather than a final statement. That the ocean, particularly its salinity distribution, plays a role in climatic change has been demonstrated. If that role is of primary importance as suggested here, or only a contributing factor, can only be established after further investigations.

We cannot hope to understand the causes of climatic stability or change by restricting ourselves to any one field of earth science. Nature is ignorant of the ways our universities are organized, and to solve the problems she poses we must integrate the findings of many specialists. I have attempted such a synthesis. However, the lack of time and previous experience makes the present study incomplete. Nevertheless, I hope that it will lead to a better understanding of the factors that control the natural environment in which we have evolved and in which, hopefully, we will continue to live.

Only by studying the past can we hope to predict the future evolution of the climate. The understanding gained may ultimately permit man to control the climate. At the present time, we are interfering significantly with natural processes without knowing whether such interference exceeds the capabilities of nature to stabilize the environment.

I have offered the hypothesis that a glacial climate may be initiated by reducing the salinity of the North Atlantic. The salinity reduction could be the result of reducing the water vapor flux across the Isthmus of Panama for several hundred years. The change in the atmospheric circulation required to accomplish this, while minor, must be produced by some primary cause. Mitchell, in the discussion at the INQUA-NCAR symposium, has suggested that a shift in the winds may be brought about by slight warming in the subtropics, caused by astronomical variations in the earth-sun geometry (the Milankovitch cycles). Once triggered into the glacial mode, the climate could return to interglacial conditions by temporary stagnation of the bottom water, followed by geothermal heating and mixing of the warmed, saltier water into the North Atlantic subarctic region. In this hypothesis, the climate is primarily controlled by the distribution of sea ice which depends, in turn, on the salinity structure in high latitudes.

If sea ice plays a major role in controlling the climate, it may be possible to explain extreme winters. If storm activity between the period of ice melting and freezing in the Atlantic Arctic is abnormally low, the low salinity surface layer formed will not be strongly mixed with the deeper, saltier water. As a result, sea ice will freeze more rapidly and extend
further south. A preliminary comparison of the extreme winter of 1962–1963 with other years suggests that this may be the case. If further studies should confirm this, it may be possible to prevent extreme winters by artificially producing vertical mixing in the surface ocean near the ice limit.

Acknowledgments. I am indebted to my colleagues at Oregon State University and elsewhere for many helpful discussions. I am particularly grateful to K. S. Deffeyes and D. Ericson for permitting me to use their as yet unpublished results. The present work was supported in part by National Science Foundation Grant GP 3582 and by the Office of Naval Research Contract No. 1280 (10) Project NR 082-102.

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