**References and Notes**

1. J. D. Hays, J. Imbrie, N. J. Shackleton, *Science* **194**, 1121 (1976).
2. M. A. Kominz, G. R. Heath, T.-L. Ku, in preparation.
3. R. N. Kerr, *Science* **201**, 144 (1978).
4. W. Q. Chin and V. Yejevich, *Colo. State Univ. (Fort Collins) Hydrol. Pap. 65* (1973), pp. 52–61.
5. W. S. Broecker and J. Van Donk, *Rev. Geophys. Space Phys.* **8**, 169 (1970).
6. A. L. Berger, *Quat. Res. (N.Y.)* **9**, 139 (1979); *Nature (London)* **269**, 44 (1977).
7. A. Papoulis, *Probability, Random Variables and Stochastic Processes* (McGraw-Hill, New York, 1965).
8. G. M. Jenkins and D. G. Watts, *Spectral Analysis and Its Application* (Holden-Day, San Francisco, 1969). Coherence spectra shown in this work were calculated from lagged, smoothed autocorrelation and autocovariance functions, utilizing the Oregon State University OS-3 ARAND (Analysis of Random Data) System Documentation.
9. Use of an arc tangent scale for coherence results in a constant confidence interval for all frequencies (*8*). Proof of the validity of these confidence intervals with more than 20 degrees of freedom is given in J.S. Bendat and A. G. Peirsol, *Random Data: Analysis and Measurement Procedures* (Wiley-Interscience, New York, 1971).
10. N. J. Shackleton and N. D. Opdyke, *Quat. Res. (N.Y.)* **3**, 39 (1973).
11. K. Hasselmann, *Tellus* **28**, 473 (1976).
12. See also C. Frankignoul and K. Hasselmann, *ibid.* **29**, 289 (1977).
13. T. M. L. Wigley, *Nature (London)* **264**, 629 (1976).
14. Supported by NSF grant OCE77-23472. We thank G. R. Heath and T. C. Moore for periodic reviews of the manuscript.

13 October 1978; revised 19 January 1979

## Warmth of the Subpolar North Atlantic Ocean During Northern Hemisphere Ice-Sheet Growth

Abstract. *Two 10,000-year periods of Northern Hemisphere continental ice-sheet growth stand out prominently within the last full interglacial-to-glacial cycle. During the first half of each rapid ice-growth phase, the subpolar North Atlantic from 40°N to 60°N maintained warm sea-surface temperatures comparable to those of today's ocean. The juxtaposition at latitudes 50°N to 60°N of an ''interglacial'' ocean alongside a ''glacial'' land mass, particularly along eastern North America, is regarded as an optimal configuration for delivering moisture to the growing ice sheets.*

Orbital variations are now clearly established as a significant primary cause of ice-age climatic cycles (*1*). The intermediate stages between cause and effect remain conjectural. How are orbital variations actually translated into regional climatic changes on the earth's surface? What is the mechanism of glacial growth and decay, and, more specifically, what is the mechanism by which water is extracted from the oceans and transported to growing ice sheets? This report addresses the latter question by focusing on the subpolar North Atlantic, the ocean closest to the Northern Hemisphere ice sheets.

Isotopic data from deep-sea cores with high accumulation rates (*2*) delineate two very large and rapid phases of ice growth during the last climatic cycle. One, the boundary between isotopic substages 5e and 5d at about 115,000 years before present (B.P.), marks the last glacial inception (increase of global ice above modern values). The second, the boundary between stages 5 and 4 at roughly 75,000 years B.P., can be regarded as the time of temperate-latitude transition into glacial conditions (*3*). During each of these intervals, roughly half the interglacial-to-glacial net ice mass accumulated within approximately 10,000 years (*4*).

Both ice-growth phases are closely related in time to Northern Hemisphere summer insolation minima (*1, 5*), substantiating Milankovitch's theory (*6*) that cool summers are critical in preserving large portions of the annual snowfall through the ablation season. As noted by numerous researchers, the growing glaciers provide positive feedback to enhance the high-latitude cooling by increasing the albedo over land. But the growth of these extensive bodies of ice also implies an expansion of the polar anticyclone normally positioned over ice cover in high latitudes of the Northern Hemisphere. This expansion of dry cold air would reinforce the normal high-Arctic aridity and slow or stop the rapid growth of ice sheets unless opposed by other parts of the climatic system (*7*). Specifically, the intervals of rapid ice growth that occurred twice during the last climatic cycle demand a significant moisture source for intervals of 10,000 years.

Ewing and Donn (*8*) hypothesized an ice-free Arctic Ocean to maintain the moisture supply to expanding ice sheets, but subsequent examination of Arctic sediments indicated no break in Arctic ice cover during the late Quaternary (*9*). Other paleoclimatologists have looked toward warm oceans at much lower latitudes (the Gulf of Mexico and western subtropical North Atlantic) as primary moisture sources (*10*). Recently, several papers have stressed the importance of the subpolar North Atlantic as the most proximal moisture source to the Laurentide Ice Sheet (*7, 11, 12*).

Our evidence (Fig. 1) shows that the subpolar North Atlantic (40°N to 60°N) maintained relatively warm ''interglacial'' sea-surface temperatures through much of the two major intervals of ice growth (*4*). This conclusion is based on oxygen isotopic evidence of ice growth from benthic Foraminifera and on transfer function estimates of sea-surface conditions derived from planktonic foraminiferal assemblages. The co-occurrence of these two separate signals within the same levels of deep-sea cores permits a precise comparison of their relative timing during intervals well beyond radiometric dating. To date, we have paired isotopic and faunal data across the stage 5-4 boundary from ten subpolar North Atlantic cores and across the substage 5e-5d boundary from five cores (*4*); two examples from each level are shown in Fig. 1.

During the substage 5e-5d transition, the sea surface in the subpolar Atlantic maintained temperatures as high as today's or 1° to 2°C warmer until about halfway into the ice-growth interval. The ocean surface finally cooled late in isotopic substage 5d, several thousand years after the midpoint of the ice-growth phase. This lag in ocean-surface cooling of about 3000 to 4000 years behind ice growth has been found so far in cores from 41°N to 54°N and from 15°W to 40°W.

During the stage 5-4 transition, the sea-surface temperatures again remained high during most of the period of rapid glacier ice accumulation (Fig. 1). Several cores even show increasing temperatures until the midpoint of ice growth. At this isotopic transition, the ocean cooling lags an estimated 4000 to 5000 years behind ice growth. This pattern has been observed in cores from 41°N to 59°N and from 15°W to 50°W (*4*).

As shown in Fig. 2, the subpolar North Atlantic at the midpoint of ice growth during the stage 5-4 transition was nearly as warm as today. Estimated sea-surface temperatures were generally within 1° to 2°C of modern values. We have detected marked oceanic warmth (19°C in summer) within 600 km of the present-day Newfoundland coast during the stage 5-4 transition (*4*).

There are two lines of evidence that much of the large ice volume indicated by the stage 5-4 isotopic data accumulated at the same longitudes as (and perhaps even directly alongside) this warm subpolar ocean: (i) ice-rafted detritus entered the subpolar Atlantic Ocean in great abundance late in this ice-growth transition (*3*), and (ii) end moraines from ice-growth phases on Baffin Island

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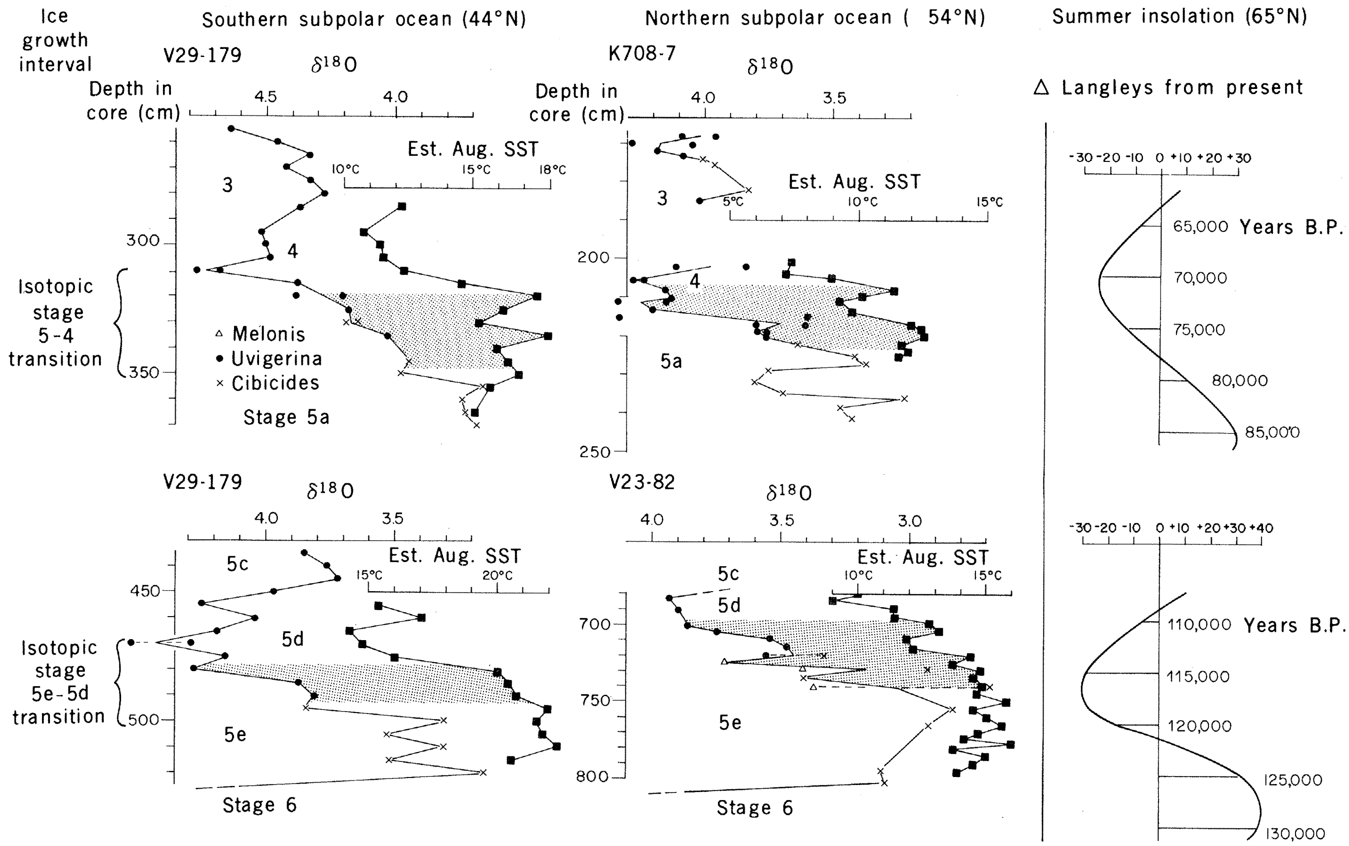


Fig. 1. Downcore record of estimated August sea-surface temperature (*SST*) and ice growth derived from studies of planktonic foraminiferal assemblages and benthic foraminiferal oxygen isotopic ratios, respectively. All parameters are plotted by convention with warm interglacial to the right and cold glacial to the left. Analytical techniques are discussed in (*4*). Cores are located in Fig. 2. Two records each of the isotopic stage 5e-5d and 5-4 transitions are shown. The subpolar ocean surface stayed warm during most of each ice-growth interval; the major warm-ocean lag is marked by stipple. Insolation is from (*20*).

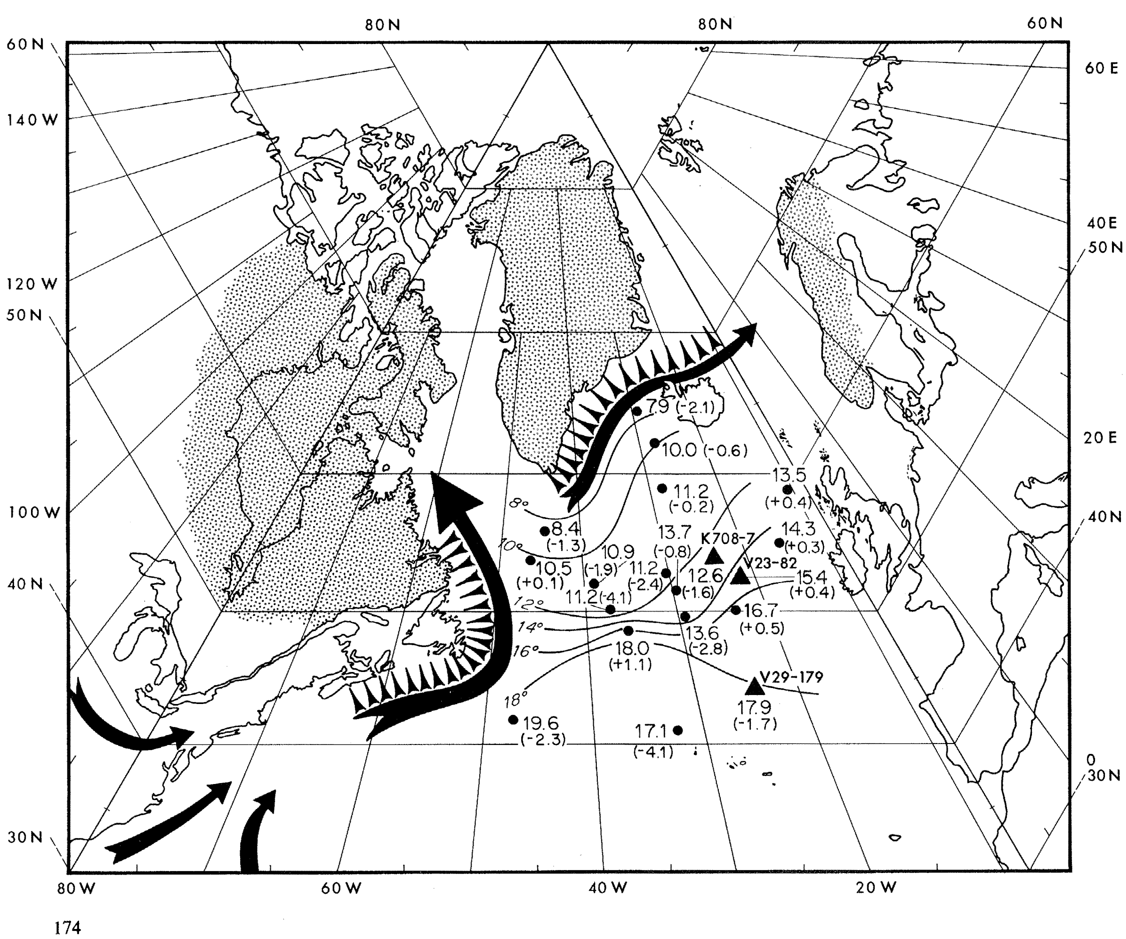


Fig. 2. Map of circulation near the midpoint of the ice-growth phase at the isotopic stage 5-4 boundary. Estimated sea-surface temperatures and departures from modern temperatures are plotted next to cores studied in (*4*). Inferred ice cover (stipple) over eastern North America was positioned adjacent to warm subpolar ocean, creating strong albedo contrast and thermal gradients (wedges). Cyclonic storm tracks (arrows) tended to follow thermal fronts toward ice-sheet centers.

roughly dated by uranium series and amino acid stratigraphy interfinger with very warm shallow marine facies (*13*). Furthermore, there is a strong circumstantial argument: where else could one put a volume of ice that by stage 4 reached 75 to 90 percent of the stage 2 full-glacial maximum? It seems necessary to assume that large portions of the late-Wisconsin areal extent of Laurentide ice had already been achieved during the stage 5-4 transition. We conclude that snow and ice fields reached to at least 50°N over eastern Canada at the stage 5-4 transition.

With land ice reaching southward to 50°N and a warm ocean flowing northward to 60°N, a narrow and unusually strong albedo gradient must have existed along the Canadian maritime provinces during the stage 5-4 transition (Fig. 2). Summer stands out as the season of really extreme thermal contrast, with, in effect, the ocean in an "interglacial summer" and the land surface in a "glacial winter." Our data (Fig. 2) do not rule out the possibility of 19°C sea-surface temperatures directly alongside an ice-covered coast. A large fraction of the total pole-to-equator thermal difference was thus encompassed within this narrow band.

This intense thermal gradient during the stage 5-4 isotopic transition has several important paleoclimatic implications. It forms part of an optimal configuration for delivering moisture to the Laurentide Ice Sheet. Thermal gradients at the earth's surface are regions of natural cyclogenesis, either causing the intensification of already existing low-pressure centers passing through from the south and west (Fig. 2) or creating new storms in situ by drawing on local oceanic latent heat (*14*). Thus, this configuration would establish a storm track northward and northwestward into an ice-free Labrador Sea (*15*) and toward the accumulation centers of the Laurentide ice mass. Other storms fed by still considerable oceanic warmth reaching to at least 59°N in the central and eastern North Atlantic may have enhanced glaciation over Scandinavia in a similar way. The even warmer subpolar ocean at the 5e-5d transition may have aided moisture delivery to ice centers at still higher Arctic latitudes.

The thermal gradient off Newfoundland also has elements of positive internal feedback. Modern climatological studies show that strong thermal gradients over eastern North America and the western North Atlantic lead to increased poleward flow of surface westerlies and, consequently, enhanced northward advection of North Atlantic drift flow into the subpolar Atlantic (*14, 16*). Thus, once falling summer insolation levels allow snow and ice to survive over lower latitudes of eastern Canada, the thermal-albedo contrast along Labrador and Newfoundland will help to maintain a warm subpolar ocean until other factors eventually intervene.

Our evidence constitutes a test of the mechanistic aspects of several previous theories of glaciation. We strongly support the general picture developed by Barry and co-workers (*7*), who stress the need for a warm open North Atlantic Ocean as a moisture source. We also support Lamb (*10*), who argues the need for a strongly blocked (meridional) atmospheric circulation to transport moisture from low latitudes. We broadly support Adam (*11*), who looked to the thermal (energy) gradient along eastern North America to fuel the process of glaciation (*12*).

We disagree with the shared assumption of Weyl (*17*) and Johnson and McClure (*12*) that an initial pack-ice advance across the Norwegian Sea and into the North Atlantic (due to lowered salinity) was the means of initiating glaciation. Instead, the major cold water (and potential pack-ice) advance into the Northeast Atlantic—marked by strong decreases of summer sea-surface temperature in Fig. 1—clearly follows the first half of ice growth shown at both the 5e-5d and 5-4 isotopic transitions (Fig. 1). Also, subpolar Atlantic salinities south of Iceland remained far too high during early ice accumulation at the stage 5-4 boundary to permit pack-ice formation (*4*).

Our evidence also directly contradicts Newell (*18*), who theorizes that cooler waters upwelling at the equator and advecting northward would cause an initial sea-ice growth, resulting in increased high-latitude albedos and eventually glaciation. Our findings also seem to disagree with Flohn (*19*), whose detailed theory included as an early feature a major cooling of the Florida Current as it emerged into the Atlantic.

In summary, we speculate that, along with low summer insolation, a warm subpolar North Atlantic Ocean may be a necessary condition for rapid and extensive Northern Hemisphere ice growth. We support the contention that a vigorous meridional atmospheric circulation directed northward along an anomalously strong surface thermal gradient off the East Coast of North America is the circulation regime most compatible with the process of rapid glaciation over North America. Today, the subpolar Atlantic is in a warm interglacial mode ready for the next phase of rapid Northern Hemisphere ice growth when insolation passes below some critical threshold.

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### References and Notes

1. W. S. Broecker, *Science* **151**, 299 (1966); J. D. Hays, J. Imbrie, N. J. Shackleton, *ibid.* **194**, 1121 (1976).
2. N. J. Shackleton, *Philos. Trans. R. Soc. London Ser. B* **280**, 169 (1977); J. Thiede, *"Meteor" Forschungsergeb. Reihe C* **28**, 1 (1977).
3. W. F. Ruddiman, *Geol. Soc. Am. Bull.* **88**, 1813 (1977); *Science* **196**, 1208 (1977).
4. ______, A. McIntyre, V. Niebler-Hunt, J. T. Durazzi, *Quat. Res. (N.Y.)*, in press; W. F. Ruddiman, *Geol. Soc. Am. Annu. Meet. Abstr.* **9**, 1150 (1977).
5. N. J. Shackleton and N. D. Opdyke, *Quat. Res. (N.Y.)* **3**, 39 (1973).
6. M. M. Milankovitch, *K. Serb. Akad. Beogr. Spec. Publ. 132* (1941) (translated by the Israel Program for Scientific Translations, Jerusalem, 1969).
7. R. G. Barry, J. T. Andrews, M. A. Mahaffy, *Science* **190**, 979 (1975); J. T. Andrews and R. G. Barry, *Annu. Rev. Earth Planet. Sci.* **6**, 205 (1978).
8. M. Ewing and W. L. Donn, *Science* **123**, 1061 (1956).
9. T.-H. Ku and W. S. Broecker, *Prog. Oceanogr.* **4**, 95 (1967); K. Hunkins, A. W. H. Bé, N. D. Opdyke, G. Mathieu, in *Late Cenozoic Glacial Ages*, K. K. Turekian, Ed. (Yale Univ. Press, New Haven, Conn., 1971), pp. 215-238.
10. H. H. Lamb and A. Woodroffe, *Quat. Res. (N.Y.)* **1**, 29 (1970); H. H. Lamb, *Climate: Present, Past and Future* (Methuen, London, 1972), vol. 1.
11. D. P. Adam, *J. Res. U.S. Geol. Surv.* **1**, 587 (1973); *Quat. Res. (N.Y.)* **5**, 161 (1975).
12. R. G. Johnson and B. T. McClure, *Quat. Res. (N.Y.)* **6**, 325 (1976).
13. J. T. Andrews and G. H. Miller, in *Quaternary Stratigraphy of North America* (papers from a symposium, Toronto, Canada, 1975), W. Mahoney, Ed. (Hutchinson & Ross, Stroudsburg, Pa., 1976), pp. 1-31.
14. M. L. Blackmon, J. M. Wallace, N.-C. Lau, S. L. Mullen, *J. Atmos. Sci.* **34**, 1040 (1977).
15. R. H. Fillon (personal communication) has evidence of open waters along the west coast of Greenland during the stage 5-4 isotopic boundary.
16. J. Bjerknes, *Arid Zone Res.* **20**, 297 (1961); H. H. Lamb, *Q. J. R. Meteorol. Soc.* **85**, 172 (1955).
17. P. K. Weyl, *Meteorol. Monogr.* **8**, 37 (1968).
18. R. E. Newell, *Quat. Res. (N.Y.)* **4**, 117 (1974).
19. H. H. Flohn, *ibid.*, p. 385.
20. A. L. Berger, *ibid.* **9**, 139 (1978).
21. We thank V. Niebler-Hunt and J. Sciarillo for the foraminiferal counts, M. Casslar and O. Anderson for the isotopic analyses, J. Durazzi and R. Fillon for helpful discussion, B. Walters for the illustrations, and S. Bowen for typing the manuscript. Supported by NSF grants OCE 76-21075 (Submarine Geology and Geophysics) for work at the stage 5-4 isotopic transition and OCE 77-22893 (CLIMAP Project—jointly funded by the Office of Climate Dynamics and the International Decade of Ocean Exploration) for work at the substage 5e-5d boundary, and by NSF grants GA 10635 and GA 19690 to the Lamont-Doherty Geological Observatory Core Laboratory of Columbia University. Lamont-Doherty Geological Observatory Contribution No. 2810.

25 October 1978; revised 14 February 1979