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ORIGIN OF GLOBAL MELTWATER PULSES

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INTRODUCTION

The fact that frequencies measured in climate records are the same as those predicted by the astronomical theory of climate change is undisputed (Hays, Imbrie & Shackleton 1976). However, the mechanisms by which these small changes in seasonal insolation are amplified into glacial cycles remain a fundamental mystery of the Earth's climate system. The Barbados postglacial sea-level record is sufficiently detailed to resolve, for the first time, the rates as well as the magnitude of continental ice melting (Fairbanks 1989, 1990) (Fig 30.1A). The Barbados meltwater discharge curve is not smooth but pulsed, with peaks at 12,000 ^{14}C years[1] and 9500 ^{14}C years (Fig 30.1B). Sea level rose more than 24 m during each of these pulses, with annual rates of sea-level rise exceeding 3 cm/yr. These enormous pulses must mark the ice-sheet response to a change in one or more of the climate amplifiers (eg, greenhouse gases and oceanic heat transports). The suspected amplifiers have different time constants and different regional sensitivities. Therefore, the discovery of both the pulsed deglaciation itself and the geographic origin of the pulses may help pinpoint the factors responsible for the timing of the large sea-level change associated with the last deglaciation, as well as the cause of previous "terminations" which recur every 100,000 ^{14}C years during the late Pleistocene Epoch (Broecker 1984).

Probably four ice sheets contribute to the shape of the global discharge curve: Laurentide; Fennoscandian; Barents; and Antarctic. Accelerator mass spectrometry (AMS) ^{14}C dating of meltwater plumes, as measured by δ^{18}O analyses of planktonic foraminifera, clearly documents the timing of the disintegration of the different continental ice sheets. Although δ^{18}O records of large meltwater plumes were discovered more than 15 years ago in continental margin sediments

[1]All ^{14}C ages are with respect to the 5568-year half-life and are uncorrected for secular changes in atmospheric ^{14}C. They are corrected for the estimated oceanic "reservoir age" of 400 years except where noted for the Southern Ocean.

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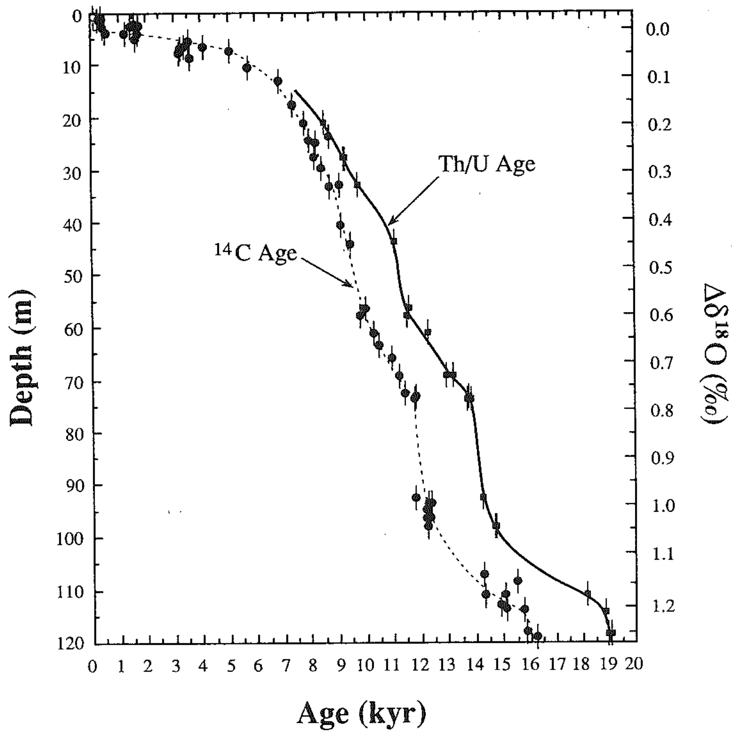


Fig 30.1A. Barbados sea-level curves based on ^{14}C-dated and $^{230}Th/^{234}U$-dated *A palmata* plotted with *A palmata* ^{14}C dates from four other Caribbean locations reported in Lighty *et al* (1982) (from Fairbanks 1990). The $^{230}Th/^{234}U$ sea-level curve is based on age determinations by mass spectrometry (squares) (Bard *et al* 1990a). Reprinted by permission of *Nature*. © 1990 MacMillan Magazines Ltd.

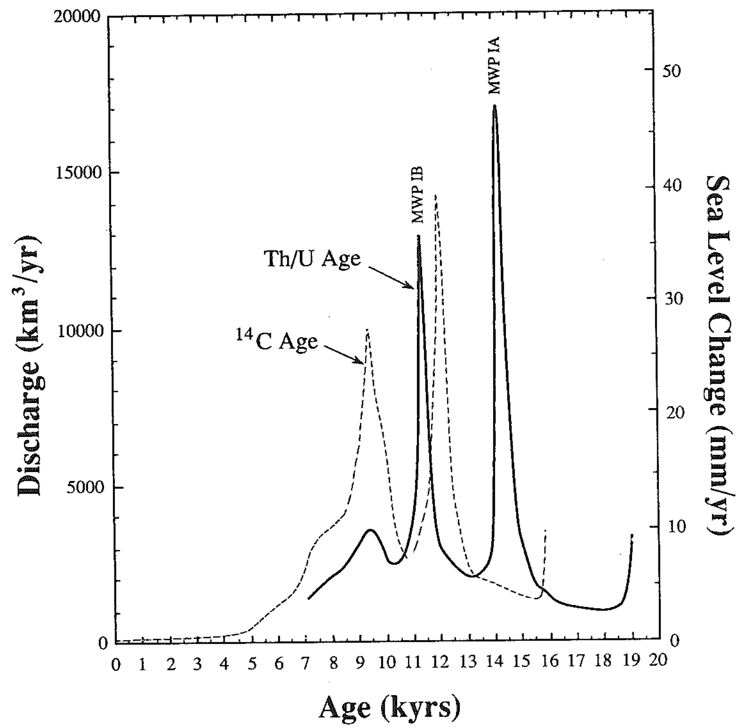


Fig 30.1B. The rate of glacial meltwater discharge calculated from the Barbados sea-level curves (from Fairbanks 1990). - - - = the discharge curve based on the differentiation of the ^{14}C-dated sea-level curve uncorrected for secular changes in the production of ^{14}C. —— = the discharge curve based on the differentiation of the $^{230}Th/^{234}U$-dated sea-level curve. The discharge units are cubic kilometers per year (left axis) and sea-level change in millimeters per year (right axis).

(Kennett & Shackleton 1975; Emiliani, Rooth & Stipp 1978; Leventer *et al* 1982), they were notoriously difficult to date because of terrestrial contamination. AMS ^{14}C dating of monospecific planktonic foraminifera from these high accumulation rate cores reveals meltwater discharge records in unprecedented detail (Broecker *et al* 1990b; Jones & Keigwin 1988). When compared with terrestrial evidence for glaciation, it is possible, for the first time, to document the sequential melting of the great continental ice sheets.

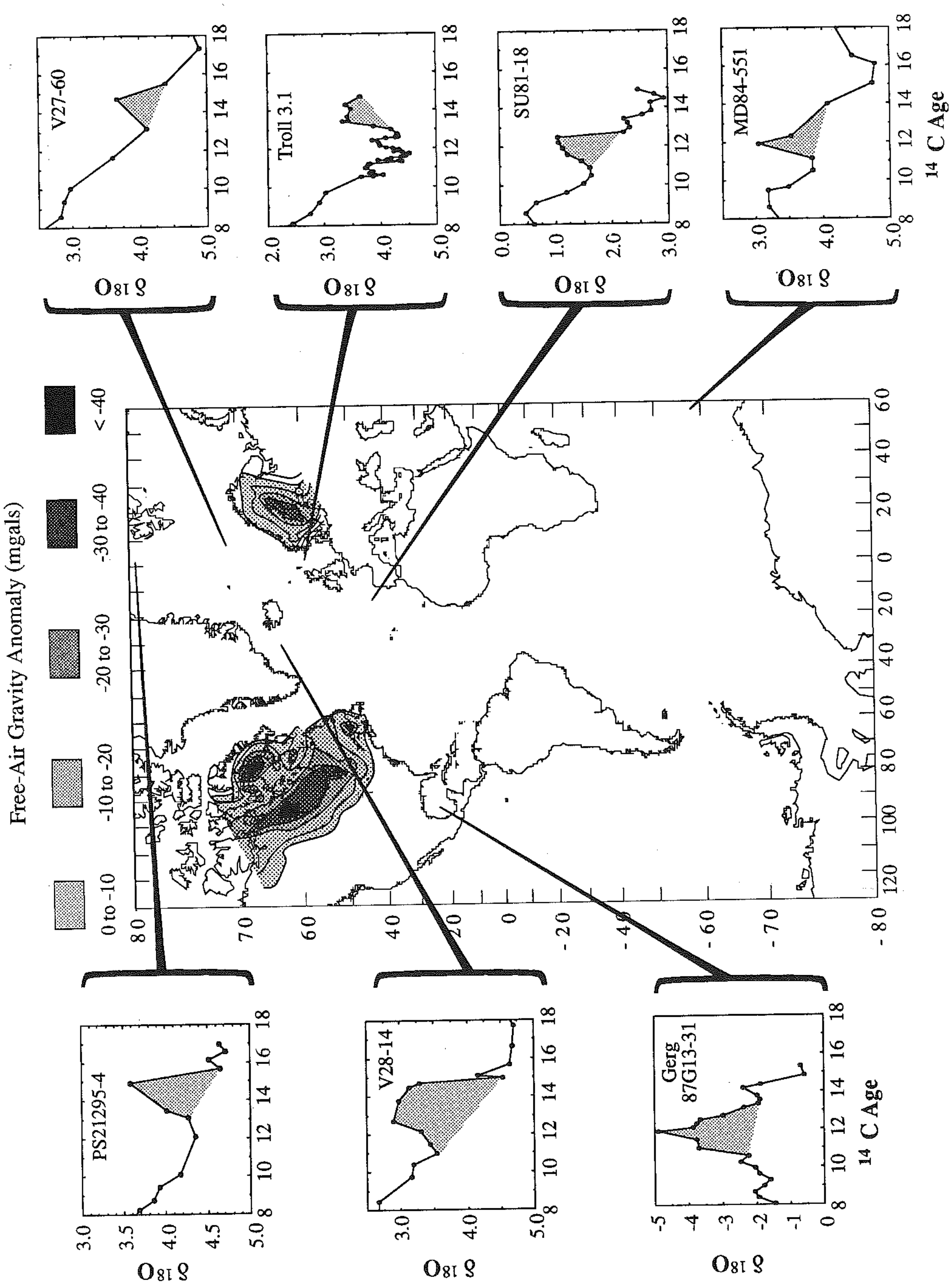
In this chapter, we examine the geographic origin of the two meltwater pulses by first documenting the timing of the meltwater plumes for the different continental ice sheets to determine which ice sheets were primarily responsible for the pulses. Analyses of the $\delta^{18}O$ of modern surface waters combined with ocean model experiments show that $\delta^{18}O$ oscillations that have the appearance of meltwater plumes can be created by ocean/climate processes. For example, changes in the production rates of North Atlantic Deep Water (NADW) may create $\delta^{18}O$ oscillations that are indistinguishable from meltwater plumes. Next, we examine the timing of several proposed amplifiers of climate change and, with the aid of General Circulation Models (GCMs), assess the regional sensitivity to these potential amplifiers. One difficulty in this approach is that comparing amplifiers of climate change to proxy records of climate change requires a standardized chronology. Most of the proxy climate records are reported in uncorrected radiocarbon years, whereas records of potential forcing, such as the CO_2 records from ice cores or astronomical calculations of local insolation, are measured or calculated in sidereal years. Here, it is important to note that the Barbados sea level record is measured in both ^{14}C years and $^{230}Th/^{234}U$ years (Bard *et al* 1990a, b; Fairbanks 1990). This new radiocarbon calibration is critical for postglacial climate research for several reasons. First, radiocarbon is not calibrated to the tree-ring chronology for the deglaciation. Second, abundant evidence from fossil trees and lake sediments indicates time intervals in the postglacial, spanning 300 to 400 years long, which had nearly constant radiocarbon ages. Third, climate researchers traditionally have assumed that sidereal and radiocarbon years are more or less interchangeable when comparing climate records (Prell 1984). The Barbados ^{14}C calibration (through $^{230}Th/^{234}U$) indicates that this assumption may be in error by as much as 2500 years (Bard *et al* 1990a, b).

Comparison of Global Meltwater Discharge Curve and Meltwater Pulses Measured in the Circum-Atlantic

The meltwater discharge estimates derived from the Barbados sea-level curve must reflect the global discharge from the large continental ice sheets and has been adequately described elsewhere (Fairbanks 1989, 1990). An independent sea-level record from New Guinea, spanning the interval 7000 to 11,000 ^{14}C years, is concordant with the Barbados results (Chappell & Polach 1991). The discharge curve has been computed in radiocarbon and $^{230}Th/^{234}U$ years (Fig 30.1B), which we equate to sidereal years based on cross-calibration to the dendrocalibrated radiocarbon time scale (Stuiver *et al* 1986) and annual band counting in living corals (Bard *et al* 1990a, b; Edwards, Taylor & Wasserburg 1988). The two prominent peaks in meltwater discharge, termed meltwater pulse (MWP) IA and IB, are dated at 12,000 ^{14}C years (14,100 $^{230}Th/^{234}U$ yrs) and 9500 ^{14}C years (11,300 $^{230}Th/^{234}U$ yrs), respectively. The MWPs reached values of 17,000 km^3/yr for MWP-IA and 13,000 km^3/yr for MWP-IB.

Since the Barbados sea-level and discharge curves were derived, numerous meltwater plumes have been measured in the circum-North Atlantic and Gulf of Mexico and dated by AMS ^{14}C. The plumes are identified by anomalously low $\delta^{18}O$ values measured in high-accumulation-rate cores along the continental margins of the North Atlantic. Figure 30.2 illustrates the $\delta^{18}O$ records for six cores in the circum-Atlantic which are reported to be examples of meltwater plumes from the Barents, Fennoscandian and Laurentide ice sheets. There is a clear distinction between the timing of meltwater plumes from these three ice sheets. The results from two cores in the Norwegian Sea (V27-60 and PS21295-4) document a meltwater plume centered between 15,500 and 14,500 ^{14}C years (Jones & Keigwin 1988; Lehman *et al* 1991). This meltwater plume is believed to mark the early disintegration of the Barents ice sheet (Jones & Keigwin 1988). A $\delta^{18}O$ record from the North Sea (Troll 3.1), adjacent to the Fennoscandian ice sheet, documents a meltwater plume between 14,200 and 13,200 ^{14}C years (Lehman *et al* 1991). The most dramatic and best-dated MWP is found in the Gulf of Mexico, recording an MWP centered at 12,000 ^{14}C years (Broecker *et al* 1990b). In several of the Gulf of Mexico records and all MWP records outside of the Norwegian Sea, a dramatic $\delta^{18}O$ decrease occurs between 10,000 and 9000 BP. The shift is generally associated with warming in the North Atlantic, so it does not usually result in an "overshoot" of $\delta^{18}O$ values. Thus, the MWP after 10,000 ^{14}C years is difficult to define by available $\delta^{18}O$ records alone.

Fig 30.2. Oxygen isotope records of planktonic foraminifera showing pulsed and sometimes sequential melting of different continental ice sheets plotted on a map of observed free-air gravity anomalies over Canada and Fennoscandia (from Peltier 1990). The AMS-^{14}C-dated meltwater plumes are from Broecker *et al* (1990b) Gerg 87G13-31; Jones and Keigwin (1988) PS21295-4; Lehman *et al* (1991) V27-60, V28-14 and Troll 3.1 (shallow-water benthic foraminifera were analyzed in this core); Bard *et al* (1987) SU81-18; Labracherie *et al* (1989) MD84-551.


Free-Air Gravity Anomaly (mgals)
0 to -10
-10 to -20
-20 to -30
-30 to -40
< -40
V27-60
Troll 3.1
SU81-18
MD84-551
PS21295-4
V28-14
Gerg 87G13-31
δ 18O
14 C Age

It is important to note that small shifts in the isotherms and/or changes in salinity unrelated to MWPs can produce a $\delta^{18}O$ anomaly indistinguishable from glacial meltwater. Changes in the seasonal flux of a foraminiferal species can also produce a $\delta^{18}O$ transient. Examination of the predicted $\delta^{18}O$ of calcite for the modern surface ocean shows the sensitivity of this tracer to changes in the evaporation/precipitation balance and water mass mixing. It is well known that the $\delta^{18}O$ of foraminiferal calcite tests are predominantly a function of temperature and $\delta^{18}O$ of sea water, as well as some minor effects associated with the ecology and life cycles of different species. Figure 30.3A shows an example of the modern salinity field for northern hemisphere spring (March, April, May), which was used to compute the $\delta^{18}O$ of sea water for this season based on measurements of the $\delta^{18}O$-salinity relationship for all of the major salinity gradients (Fig 30.3B). A set of seven $\delta^{18}O$-salinity equations was used to compute the

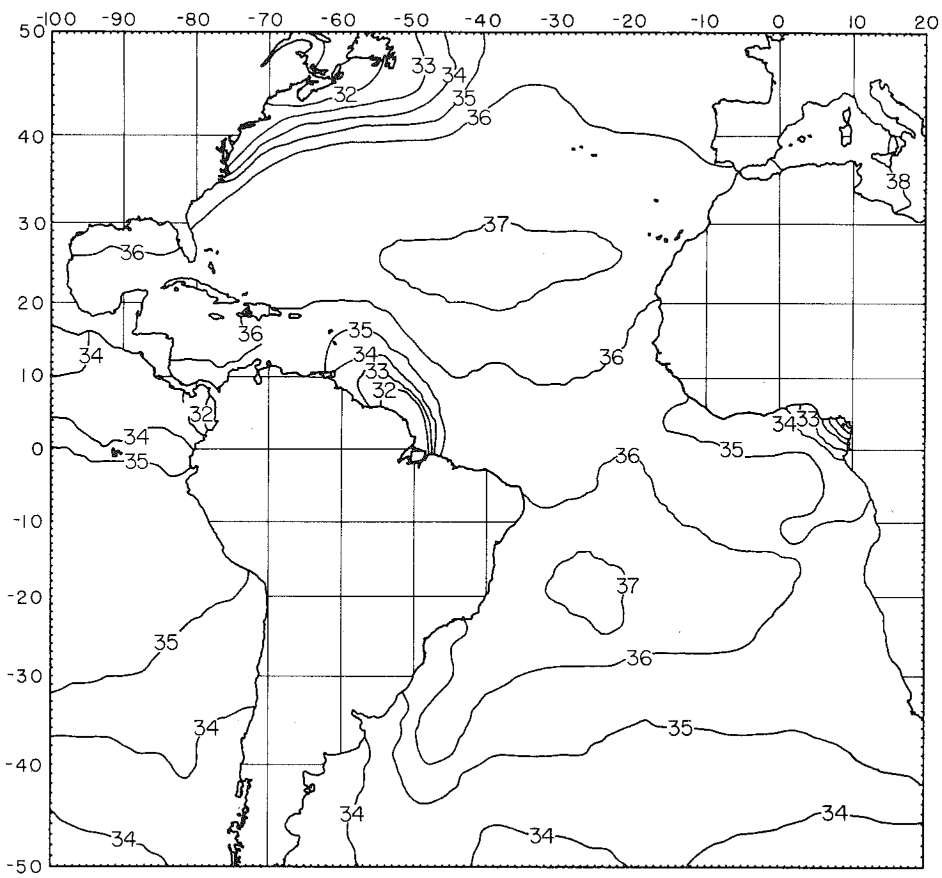


Fig 30.3A. Map of modern Atlantic surface water salinity averaged for the months of March, April and May (from Levitus 1982)

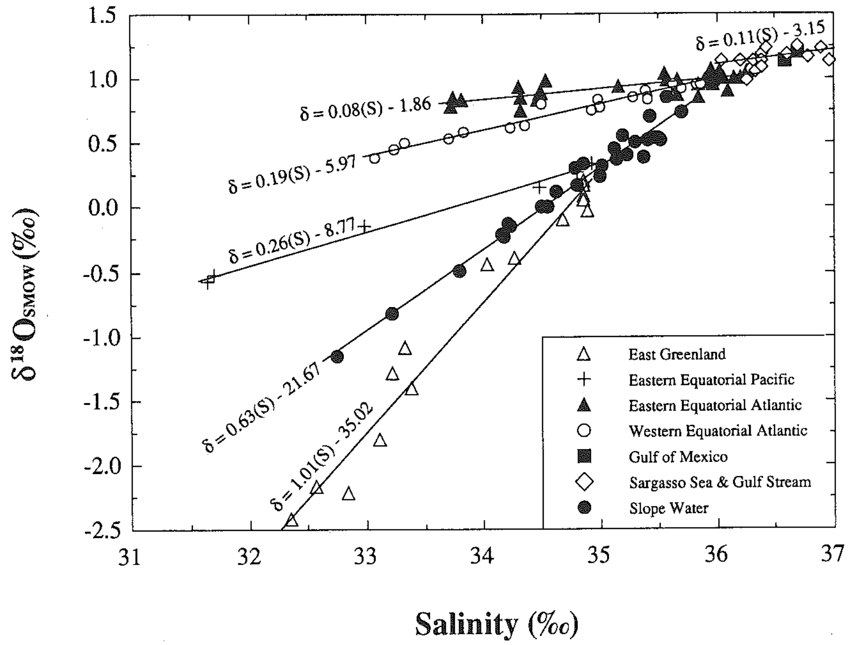


Fig 30.3B. Plots of $\delta^{18}O$ and salinity measurements across all major salinity gradients in the Atlantic and eastern tropical Pacific (Fairbanks, ms in preparation)

surface-water $\delta^{18}O$ on a 1° × 1° grid (Fairbanks, ms in preparation) for each season. Combining the estimated $\delta^{18}O$ of sea water with the measured monthly temperature from Levitus (1982) (Fig 30.3C), we computed the estimated $\delta^{18}O$ of calcite for each month. Figure 30.3D is the computed annual average $\delta^{18}O$ of calcite, and illustrates that it is easy to create $\delta^{18}O$ transients that are not MWPs. In the subpolar region, the effect of decreasing temperature on the $\delta^{18}O$ of calcite dominates the opposite effect of decreasing $\delta^{18}O$ of sea water poleward. In the polar waters, where surface water is freezing, the salinity has a greater influence on the isopleths of $\delta^{18}O_{calcite}$. This relationship adds some confidence to the meltwater origin of the Fram Strait core PS21295-4 (Fig 30.2). However, the best way to verify that a $\delta^{18}O$ anomaly is an MWP is to trace its increasing amplitude toward the source. The 4‰ $\delta^{18}O$ pulse from the Gulf of Mexico (Gerg 87G13-31) is the only MWP that is of indisputable origin. Although all of the other $\delta^{18}O$ anomalies in Figure 30.2 have been reported in the literature as MWPs, their relationship to meltwater awaits verification.

CLIMATE FORCING FACTORS

According to the astronomical theory of climate change, the Pleistocene ice ages were caused primarily by changes in the seasonal distribution of incoming solar

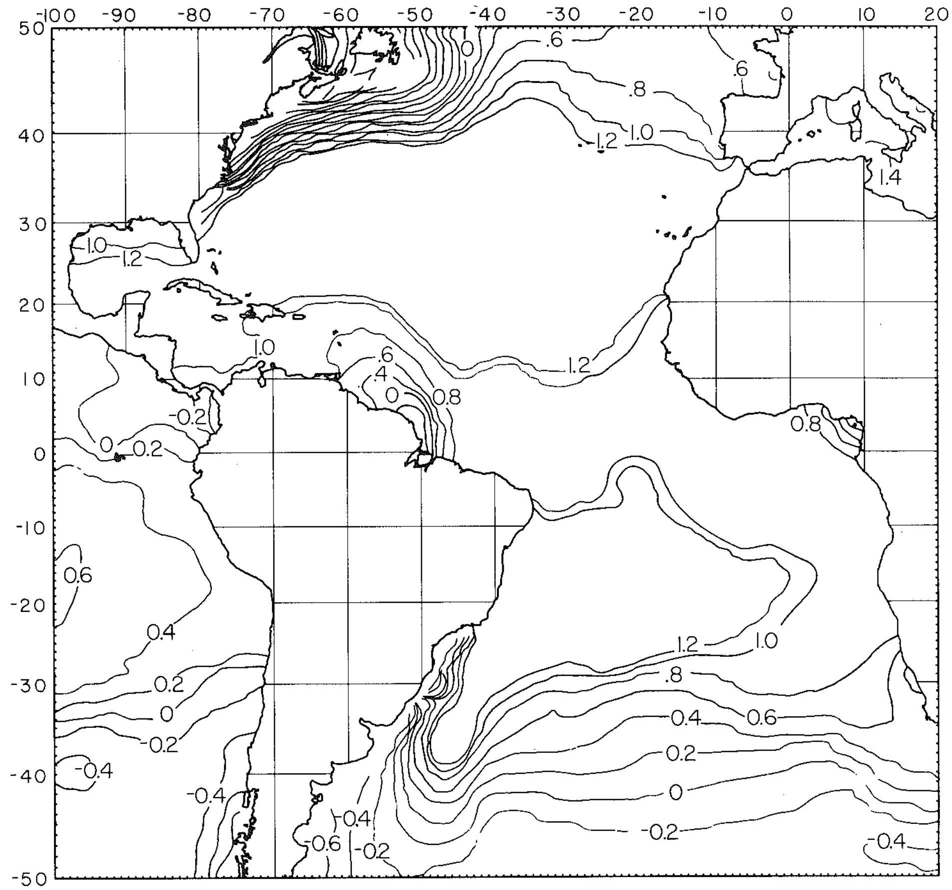


Fig 30.3C. The estimated $\delta^{18}O$ of modern Atlantic surface water averaged for the months of March, April and May. The map is based on the salinity *vs* $\delta^{18}O$ equations in Figure 30.3B applied to the salinity map in Figure 30.3A.

radiation associated with orbital variations (Milankovitch 1941). The northern hemisphere ice sheets accumulate in the 40°–80° N latitude range, whereas the Antarctic ice sheet is located poleward of 70° S. Insolation changes associated with the 41,000-year tilt cycle increase poleward in both hemispheres, while the 23,000-year precessional cycle dominates insolation changes at middle and low latitudes and is out of phase between hemispheres. Figure 30.4 shows the summer insolation at 50°, 60°, 70° and 80° N latitude, as calculated from the equations of Berger (1978). Summer insolation between 24,000 and 11,000 years increases by 11% at 50°N compared to 15% at 80°N. The relationship between insolation and a "filtered" sea-level curve is reasonably good for the last deglaciation, now that the sea-level curve is corrected to sidereal years. However, we know from comparison of insolation to the present-day sea level,

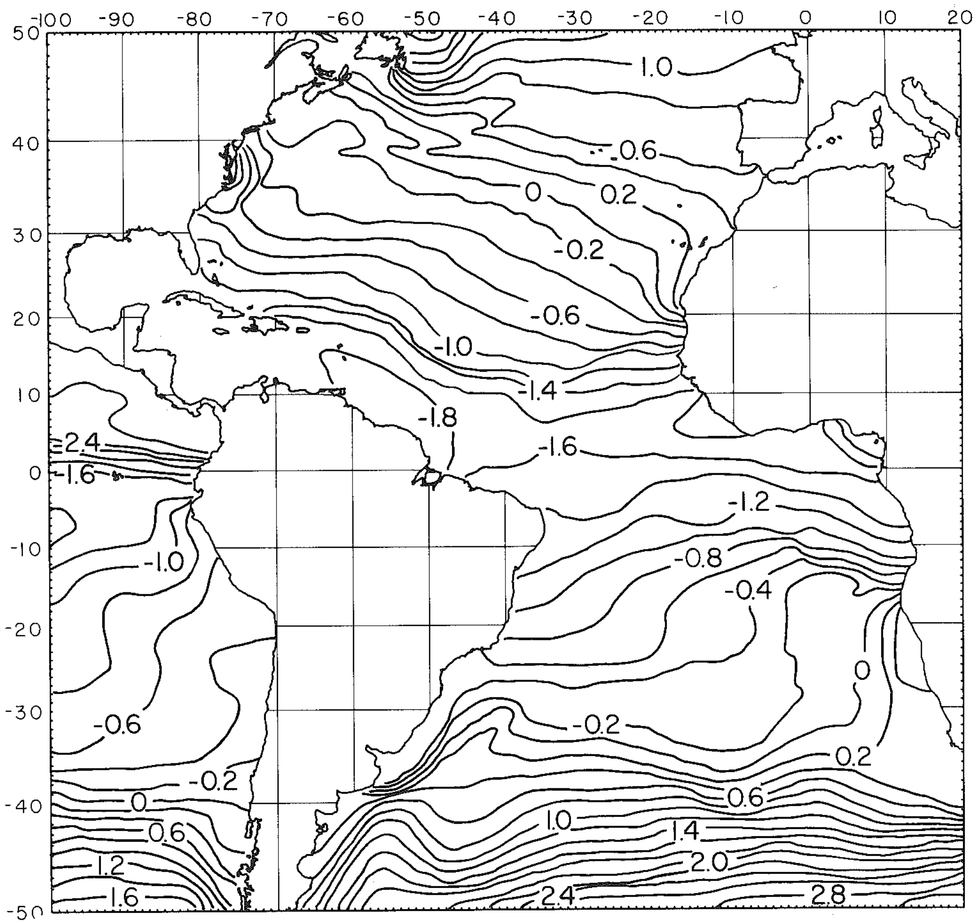


Fig 30.3D. The predicted annual average $\delta^{18}O$ of calcite is based on the "paleo-temperature" equation of Epstein *et al* (1953) using the 1° × 1° gridded temperature and $\delta^{18}O_{sea\ water}$ values.

as well as from longer records of sea level, that insolation alone does not correspond directly to ice volume or its sea-level equivalent. In the next section, we examine the timing and regions of influence of two suspected amplifiers of the insolation forcing function, in order to assess their possible roles in generating the meltwater discharge curve (Fig 30.1B).

One popular explanation for the non-linear response of climate change is variability in the production rate of NADW (Stommel 1961; Rooth 1982; Broecker, Peteet & Rind 1985; Bryan 1986; also see Broecker & Denton 1989, for a review). Several characteristics of the NADW thermohaline circulation provide the basis for this hypothesis. The process of NADW formation leads to excessive heat loss from the ocean to the atmosphere in the northern hemisphere

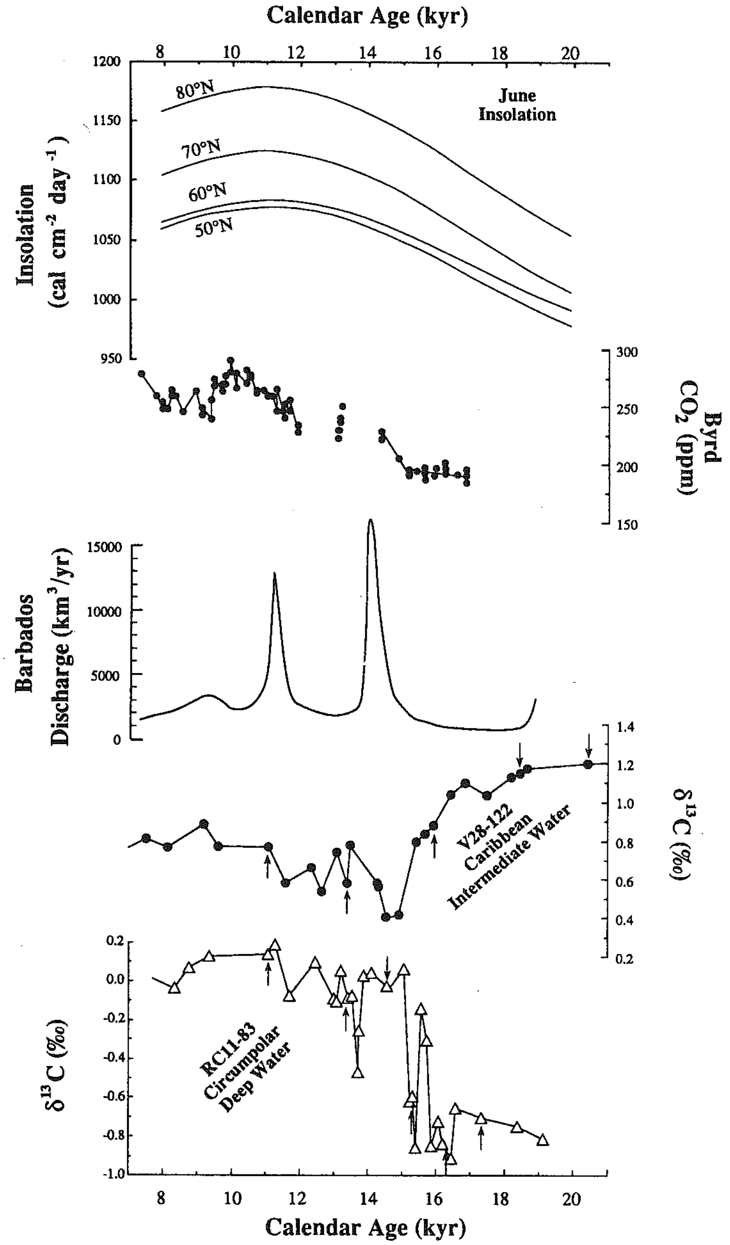


Fig 30.4. Record of atmospheric CO_2 from Byrd (Antarctica) polar ice core (Neftel *et al* 1988; Staffelbach *et al* 1991) compared to global meltwater discharge curves (Fairbanks 1990) and northern hemisphere summer insolation (Berger 1978), all reported in calendar or $^{230}Th/^{234}U$ years. Comparison of the Southern Ocean record of NADW flux (core RC11-83 benthic foraminifera $\delta^{13}C$ record) with a mid-depth $\delta^{13}C$ time series of Atlantic intermediate water chemistry (core V28-122 benthic foraminifera $\delta^{13}C$ record) (Charles & Fairbanks, ms) adjusted to calendar years using the ^{14}C calibration of Bard *et al* (1990a, b), showing the coincidence of NADW production and deglaciation. AMS-^{14}C-dated levels are indicated by arrows.

winter. The heat released upon NADW formation is equivalent to 35% of the solar energy incident on the North Atlantic polar latitudes today, and therefore, changes in the production of NADW might explain large, rapid changes in European continental climate (Broecker & Denton 1989). NADW also provides a crucial source of heat and salt to other oceans, especially the Southern Ocean, where a convergence of oceanic heat is required to melt sea ice on an annual basis (Gordon 1986). Further, NADW variability would affect oceanic nutrient and CO_2 distributions on a global scale, and therefore, may indirectly change atmospheric CO_2 content. Finally, modeling studies suggest that it may be possible for the NADW cell to switch catastrophically between stable on and off modes (Bryan 1986; Manabe & Stouffer 1988; Maier-Reimer & Mikolajewicz 1989).

Changes in the flux of North Atlantic Deep Water as measured in the Southern Ocean

Although investigators of deep-sea sediment cores in the North Atlantic have tried to delineate the history of NADW production during the last deglaciation, the chronology of the records analyzed and the generalities of the results remain uncertain (Mix & Fairbanks 1985; Boyle & Keigwin 1987; Jansen & Veum 1990; Keigwin & Jones 1989). Oppo and Fairbanks (1987, 1990) and Charles and Fairbanks (ms) show that the Southern Ocean is one of the best regions in which to study the details of NADW variability and the consequent climatic effects. Taking advantage of the global-scale mixing of deep water that occurs in the Antarctic Circumpolar Current, Charles and Fairbanks (ms) used benthic foraminiferal $\delta^{13}C$ from a Southern Ocean core (RC11-83, 41°36′S; 09°48′E, 4718 m) with a high sedimentation rate to document the timing of NADW flux from the Atlantic Ocean during the last deglaciation (Fig 30.5). The deglacial time scale for RC11-83 is derived from 11 AMS ^{14}C dates on seven levels in the core (Charles & Fairbanks, ms). The AMS ^{14}C measurements were made at the University of Arizona NSF Accelerator facility for Radioisotope Analysis by A J T Jull and D J Donahue.

The RC11-83 benthic $\delta^{13}C$ record (Fig 30.4) indicates that deglacial changes in the nutrient chemistry of the deep Southern Ocean were rapid and large. An abrupt 1‰ shift from low values characteristic of glacial conditions to high values occurred from 12,600 to 12,200 ^{14}C years. This rapid rise in Southern Ocean $\delta^{13}C$ values was most likely associated with a change in status of NADW, switching from the "off" mode to the "on" mode. The RC11-83 record also shows significant oscillations superimposed on this basic NADW off-on switch. A prior oscillation occurred between 13,300 ^{14}C years and 12,800 ^{14}C years, and a low $\delta^{13}C$ event occurred at 11,800 ^{14}C years.

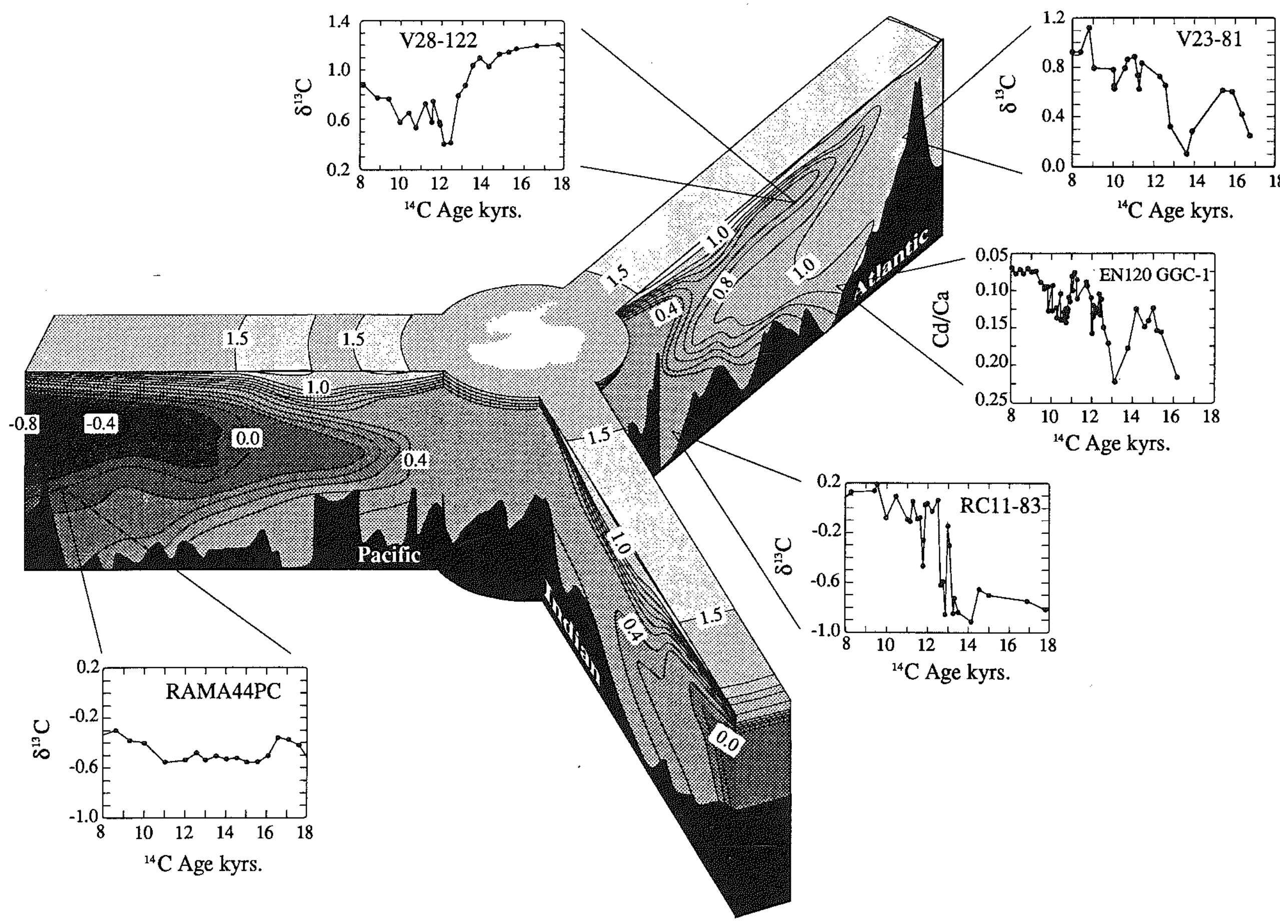


Fig 30.5. GEOSECS $\delta^{13}C$ data from the different ocean basins (Kroopnick 1985) with postglacial times series of $\delta^{13}C$ in sediments dated by AMS ^{14}C for the North Atlantic (Jansen & Veum 1990), intermediate-depth Atlantic (Oppo & Fairbanks), Southern Ocean (Charles & Fairbanks, ms) and Pacific Ocean (Keigwin 1987). Also shown for the North Atlantic are Cd/Ca in sediments dated by AMS (Boyle & Keigwin 1987).

Comparison with other detailed deglacial records from the deep North Atlantic shows significant differences for the interval from 16,000 to 13,500 ^{14}C years (Fig 30.5). The North Atlantic records show this region was first inundated with NADW between 17,000 and 15,000 ^{14}C years, followed by minimum NADW between 14,000 and 13,000 ^{14}C years, reaching or exceeding glacial $\delta^{13}C$ values (Boyle & Keigwin 1987; Jansen & Veum 1990). The striking transition from nutrient enrichment (low $\delta^{13}C$) to nutrient depletion (high $\delta^{13}C$), which is dated between 12,600 and 12,200 ^{14}C years in RC11-83, is apparent in all three records (Fig 30.5). Another step in the same direction, but with much lower amplitude, is apparent from about 10,000 to 9,000 ^{14}C years.

The $\delta^{13}C$ records that monitor mid-depth North Atlantic water provide further documentation of a rapid change in upper NADW nutrient chemistry from 14,000 to 12,400 ^{14}C years (Fig 30.5). In these records, the sense of change prior to 12,200 ^{14}C years is opposite to that of the deep Atlantic. Whereas relative nutrient depletion prevailed during glacial periods, a shift to more nutrient-rich conditions occurred at 12,700 ^{14}C years, synchronous with the timing of deep-ocean change. The inverse mid-depth and deep-Atlantic signals could both reflect the same phenomenon if, as in today's ocean, relatively nutrient-rich water from the Southern Ocean were drawn northward across the equator at mid-depth by the removal of surface and thermocline water to form southward-flowing NADW.

Where the various North Atlantic nutrient proxy records disagree, we suggest that RC11-83 should be taken as the more comprehensive measure of NADW flux variability for two reasons. First, deglacial sedimentation rates in RC11-83 are highest. Second, RC11-83 $\delta^{13}C$ values presumably monitor the flux of NADW relative to Pacific and Indian outflow, while mixing between NADW and Southern Ocean Water may vary independently of NADW flux in some North Atlantic locations. This issue is important for the interval from 17,000 to 15,000 ^{14}C years, because the North Atlantic records indicate a switch from high to low nutrients, suggesting that the initiation of NADW began 3000 years earlier than indicated by the Southern Ocean record. The RC11-83 record, on the other hand, shows more uniformly low $\delta^{13}C$ values until 13,000 ^{14}C years, indicating minimal NADW flux throughout the glacial period. This issue is also important for characterizing the Younger Dryas interval (11,000 to 10,000 ^{14}C years), the period during which, according to previous interpretations of a Bermuda Rise core (Boyle & Keigwin 1987), NADW production was temporarily reduced. The Southern Ocean record does not show an anomaly during this period, implying that the total flux of NADW may not have varied significantly. Thus, North Atlantic cores situated in the mixing zones between water masses may be too sensitive to small circulation changes to document accurately net changes in the NADW flux.

It is possible, however, that deviations from the simple mixing models for Circum-Polar Deep Water (CPDW) $\delta^{13}C$ can occur. For example, some evidence suggests that CPDW $\delta^{13}C$ was lower than in the deep Pacific during glacial periods, and this distribution of $\delta^{13}C$ cannot be easily explained by mixing alone unless Indian Ocean deep water compensated for the relatively nutrient-depleted Pacific water mixing in the circumpolar region. One could also invoke non-conservative effects that are unique to the Antarctic, such as proposed by Boyle (1990). A more serious problem is the observation that Cd/Ca ratios in benthic foraminifera from Southern Ocean cores do not increase nearly as much as the $\delta^{13}C$ might predict during glacial periods (Boyle 1991). We have no definitive answer for why the Southern Ocean $\delta^{13}C$ at times might be lower than the Pacific, and the discrepancy between $\delta^{13}C$ and Cd/Ca demands more research into the decoupling of these tracers from the nutrient elements in the Southern Ocean (eg, see Charles & Fairbanks 1990). Nevertheless, we make the simplest interpretation that seems justified from the modern distribution of $\delta^{13}C$ (Kroopnick 1985) and the core-top calibration of benthic foraminifera $\delta^{13}C$ in deep water (Belanger, Curry & Matthews 1981): a strong relative contribution of NADW is the only means for increasing CPDW $\delta^{13}C$ above Pacific (mean ocean) values.

Ice-core record of atmospheric CO_2 change

Antarctic ice-core specialists have emphasized the primary role of atmospheric CO_2 content in amplifying the glacial-interglacial cycles (Lorius *et al* 1990). The Antarctic CO_2 results are cited as an important independent calibration for the so-called CO_2 amplifier factor, which is incorporated in GCMs. Evaluating the magnitude of the CO_2 amplifier is at the heart of climate research today. Recently, the deglacial ice from Vostok (Antarctica) and Dye 3 (Greenland) ice cores have been reanalyzed for atmospheric CO_2, and additional data have been added to the Byrd (Antarctica) record in the critical deglacial interval (Barnola *et al* 1991; Staffelbach *et al* 1991). Re-evaluation of the air dating at Vostok indicates that the age difference between air and ice is about 6000 years during the coldest periods, rather than 4000 years as previously estimated (Barnola *et al* 1991). The long and variable closure age and poor sampling resolution of Vostok make the record unsuitable for comparison to the high-resolution ocean/climate records from the Southern Ocean, North Atlantic and Barbados. However, new data from Dye 3 and Byrd (Antarctica) are critical for this comparison (Fig 30.4).