22. ON THE DEFINITION OF THE PLEISTOCENE/ HOLOCENE BOUNDARY IN DEEP-SEA SEDIMENTS

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INTRODUCTION

The terms *Holocene, Postglacial,* and *Recent* are commonly used interchangeably in the geologic literature, with an understanding that what is meant is the time span which had a climate more or less like today's. The term *Holocene* also has a formal meaning, that is, it stands for *the last 10 000 years*. This definition was adopted by the Holocene Commission of the INQUA in Paris in 1969. The Holocene Commission is now about to recommend a stratotype section. In the following I shall summarize how the Pleistocene/Holocene boundary has been defined and identified in deep-sea sediments, to provide some background for this contemplated action.

I should emphasize at the outset that little thought has been given to a logical definition of such a boundary. Instead, the boundary concept developed operationally from the observation that certain sediment properties show a rapid change at about the right depth (20 cm to 30 cm downcore in most cases) to qualify for an assumed correlation with a presumed flip-over from glacial to postglacial conditions. In some instances radiocarbon dating backed the assumption.

The success of this approach owes much to the low resolution of the deep-sea record (±2 000 years) and also to low sampling densities (usually >3 000 years). In cores with high sedimentation rates, and from areas where climatic fluctuations during the Glacial-Postglacial transition were amplified, difficulties can arise. A strong initial warming may then be seen to precede a substantial cooling, which in turn is followed by another warming (for example, see Caralp 1970, Leclaire and Vergnaud-Grazzini 1972).

Another type of difficulty arises when two or more climate-related signals show marked differences in the position of maximum change. Core V28–14 from the Irminger Sea (south-east of Greenland), for example, shows a rapid change from low to high carbonate values centered near 90 cm downcore. This level was identified as "Termination I" by Kellogg (1975). (The term "Termination I" was introduced by Broecker and van Donk, 1970, to emphasize the perceived suddenness of change at the Pleistocene/Holocene boundary.) Later it was found, however, that the maximum change in oxygen isotope composition of the planktonic foraminifer *"Globigerina" pachyderma* occurs near 140 cm. The proposed boundary then was moved to this deeper level (Kellogg *et al.* 1978). The difference in age is about 4 000 years.

The reason for the adjustment is that, by convention, the change seen in the oxygen isotope ratios of planktonic foraminifers is the master-signal to which other sediment properties are referred. The underlying presumption is that the signal closely reflects sea level change, and that it is synchronous world-wide (Shackleton and Opdyke 1973, Emiliani and Shackleton 1974). Within the limits of resolution of deep sea cores this hypothesis is indeed useful in most cases.

DEFINITION BY FAUNAL CHANGE

Historically, the first definition of the Pleistocene/Holocene boundary in deep-sea sediments was from faunal change. Schott (1935) discovered that the planktonic foraminifer *Globorotalia cultrata (= G. menardii)* was intermittently extinct in the Atlantic, based on a study of gravity cores taken during the METEOR Expedition (1925–1927). He found that *G. cultrata* last reappeared at a level some 20 cm down in his cores, and that it has been present since. He proposed that this appearance of *G. cultrata* marks the beginning of postglacial time. This level has been variously used to denote the Pleistocene/Holocene boundary. It is in fact diachronous. By comparing the *G. cultrata* appearance with oxygen isotope data one can show that this foraminifer shows up earlier in equatorial sediments than in the Gulf of Mexico or off north-western Africa. The difference is on the order of several thousand years (see Emiliani *et al.* 1975 and Thiede 1977). Also, since the *G. cultrata* disappearances and reappearances are restricted to the Atlantic, Schott's method does not work anywhere else. It should be pointed out, however, that Schott's discovery yielded the first set of reasonable deep-sea sedimentation rates at the time.

G. cultrata is an extreme example of a more general phenomenon. In most regions of the world ocean the transition from glacial to postglacial conditions is accompanied by marked changes in the composition of the fossil assemblage of planktonic foraminifers. In equatorial regions the trend is from species associated with equatorial upwelling toward species more typical of low fertility regions. In the subtropics cool-water forms recede, and in middle and high latitudes *"Globigerina" pachyderma* and

Globigerina bulloides tend to become less abundant in the Postglacial sediments. In places, the faunal change is quite dramatic (Phleger *et al.* 1953, Parker 1958). Other marker signals which have been applied to define the boundary are changes in coiling ratio of certain foraminifers (Ericson 1959, Bandy 1960) and other morphological variations. Radiolarian faunas also have been used to define the boundary (Frerichs 1968, Duncan *et al.* 1970, Bjørklund *et al.* 1979).

The changes in faunal composition are not synchronous over the North Atlantic (Ruddiman and McIntyre 1973). Elsewhere, on the whole, the question of synchroneity has not been adequately addressed. Where checked, some differences in the timing of faunal change and of oxygen isotope change are usually evident (Imbrie *et al.* 1973). Such differences may or may not be real in a palaeoclimatic sense, as benthic mixing can produce spurious phase shifts between transitional signals (Hutson 1980).

DEFINITION BY OXYGEN ISOTOPES

Emiliani (1955) showed that the oxygen isotope composition of foraminifers from surface waters traces the warm-cold cycles of the ice ages. Two main effects are present: an ice-effect due to the locking up of ^{16}O-rich water in polar ice caps (and corresponding enrichment of sea water with ^{18}O) and a temperature effect (growth in cold water leads to ^{18}O enrichment). The two effects, while additive, are not exactly synchronous: while the ice effect spreads rapidly throughout the ocean, the temperature effect depends on time-transgressive shifts in climatic zone boundaries. Fortunately, in most cases, the ice effect dominates, so that there is a strong presumption in favor of synchroneity, with regard to oxygen isotope signals.

During deglaciation freshwater was introduced at a high rate on the top of the ocean. It is likely that this water, which floats on sea water, retarded the mixing processes of the ocean, so that salinity stratification developed (Worthington 1968, Olausson 1969, Berger *et al.* 1977). If so, the oxygen isotope signals of planktonic and benthic foraminifers cannot be synchronous (Olausson 1965, Pastouret *et al.* 1978).

At first glance it would seem an easy task to establish the case for or against synchroneity of planktonic and benthic species for the various ocean basins. This is not the case because of the afore-mentioned disturbance by benthic mixing. Oxygen isotope curves of different species have to be "unmixed" to be comparable with the same core. Even the simplest equation available to do this contains terms for the relative abundance of the carrier species (Berger *et al.* 1977). In virtually all published data, the abundance information is not available. In cases where it is, the abundances commonly refer to a size class different from that on which the oxygen isotopes were determined, and generally abundances do not refer to total sediment (as required) but to a portion of the sand fraction.

In summary, although better than other signals, oxygen isotope transitions are somewhat diachronous. The error is on the order of several centimeters, that is, one to two thousand years. This essentially represents current limits of the resolution of the deep-sea record.

DEFINITION BY CARBONATE CONTENT AND PRESERVATION OF CALCAREOUS FOSSILS

In the Pacific the postglacial record is generally marked by a lower carbonate content than the underlying late glacial portion (Arrhenius 1952, Broecker and Broecker 1974). In the Atlantic, one observes the inverse relationship (Olausson 1960). The reason for the change in carbonate content near the Pleistocene/Holocene boundary, in the Pacific as well as in the Indian Ocean, is that carbonate disolution greatly increased in the early Holocene (Olausson 1967, Berger 1970). This increase closely followed a marked preservational peak during deglaciation (Berger and Killingley 1977). In the Atlantic, the carbonate stratigraphy of the transition largely reflects a change in dilution from terrigenous matter (Broecker *et al.* 1958, Volat *et al.* 1980). Changes in dissolution also may play a role (Olausson 1969, Berger 1973, Gardner 1975), and a preservational peak is evident in cores from off north-western Africa, centered on the boundary (Berger 1977, Thiede 1977).

Dilution depends on the proximity of sediment sources and the efficiency of transport agents. The dilution effects are probably not synchronous across latitudinal belts and between major sedimentary basins. As far as the carbonate saturation of the deep-sea, which produces the dissolution cycles, changes are probably synchronous within the deep Pacific and within the limits set by deep-ocean mixing rates.

AGE OF BOUNDARY

The first radiocarbon dates which allowed an assessment of the Pleistocene/Holocene boundary in deep-sea cores were made by Rubin and Suess (1955, 1956). They appear together with oxygen isotope stratigraphies in Emiliani (1955). By the conventional method of setting the boundary at the midpoint

TABLE 22:1. Radiocarbon ages (by interpolation) of Glacial-Postglacial mid-transition in box cores from the western equatorial Pacific. Ages (based on bulk carbonate) are listed in Berger and Killingley (1982). Oxygen isotope data *(Pulleniatina)* and levels of maximum preservation are given in Berger *et al.* (1977). Cores are listed in order of depth of water. All depths are shallower than 3 400 m.

Core ERDC (no.)	Oxygen isotope transition (cm)	Radiocarbon age (ka)	Maximum preservation (cm)	Radiocarbon age (ka)
92	23.5	12.7	26.5	14.0
88	20.0	14.7	21.5	15.8
112	25.0	10.7	23.5	10.2
120	27.5	12.2	26.0	11.6
83	24.5	9.6	23.5	9.3
79	22.5	11.2	19.5	8.1
123	22.0	10.5, 9.4	26.5	12.2, 11.1
125	23.5	10.3	21.5	9.3
Mean	23.6 ± 2.2	11.3 ± 1.7	23.6 ± 2.6	11.3 ± 2.4
Without Core 88		10.8 ± 1.2		10.8 ± 1.9
Median		10.7		11.1

of the transition between full Glacial and full Postglacial conditions, the dates on three Atlantic cores can be read as follows: Core A 179-4, 10.0 ka; Core A 172-6, 12.5 ka; Core A 180-73, 11.0 ka. The average age of this level, then, is near 11 000 ^{14}C years B.P. This age agrees with subsequent determinations by Ericson *et al.* (1956) and Broecker *et al.* (1958). The paper most commonly cited for the 11 000 years age is the one by Broecker *et al.* (1960), in which a number of radiocarbon ages from various environments and from several laboratories are summarized. Based on the deep-sea cores considered, these authors propose "that a large portion of the Atlantic Ocean underwent a temperature increase of 6° to 10° during a period of less than 2 000 years and that the midpoint of this change was within 300 years of 11 000 years ago" (*ibid.*, p. 435). The presumed temperature increase refers to the oxygen isotope data of Emiliani (1955).

Dating of the transition in the Pacific and Indian Ocean yielded ages ranging from 7 000 years (Blackman and Somayajulu 1966) to 13 000 years B.P. (Shackleton and Opdyke 1973), with most estimates falling between 9 000 and 12 500 years B.P. (Bandy 1960, Conolly 1967, Frerichs 1968, Duncan *et al.* 1970, Bé and Duplessy 1976, Vincent 1976, Shackleton and Vincent 1978).

The best available data on the age of the Glacial-Postglacial transition in the Pacific Ocean are those on box cores from the western equatorial Pacific. There are eight good cores above the lysocline. Oxygen isotope data are given in Berger *et al.* (1977) and radiocarbon determinations are listed in Berger and Killingley (1982). The ages for the midpoints of maximum

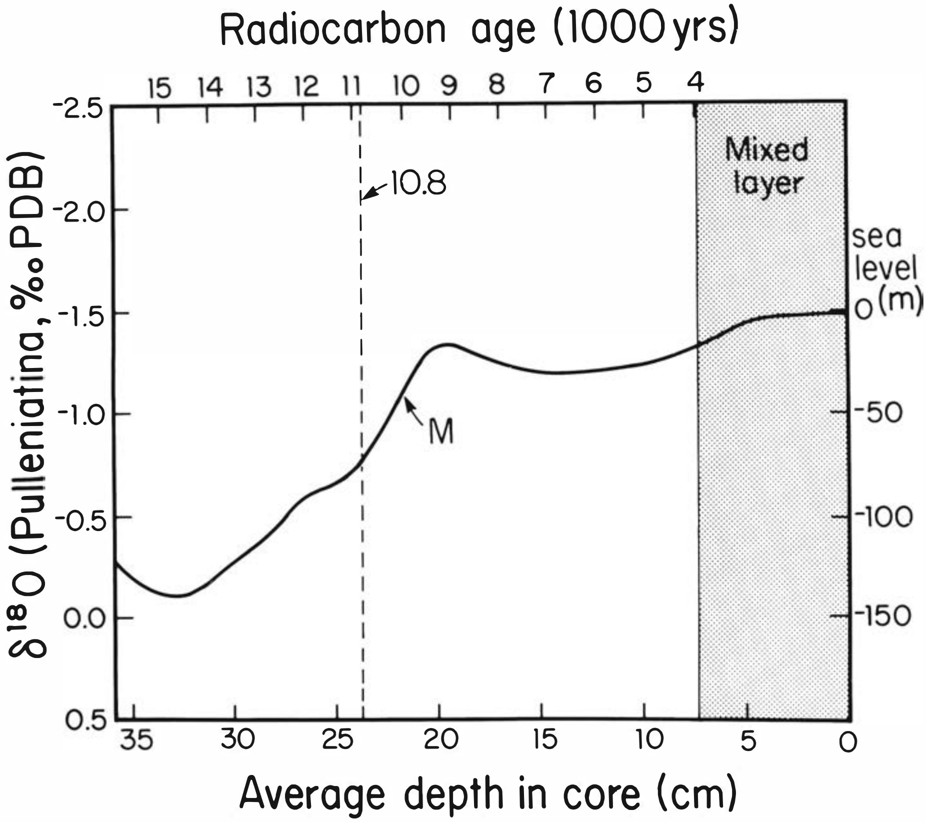

Fig. 22:1. Composite oxygen isotope signal of the Glacial-Postglacial transition as given by Berger *et al.* (1977). The curve is based on data from 8 box cores taken in the western equatorial Pacific, above the lysocline. Oxygen isotopes are those of *Pulleniatina obliquiloculata,* a planktonic foraminifer which lives in subsurface waters. More than 80 determinations are contained within the curve. The radiocarbon scale is based on about 25 analyses of bulk carbonate (listed in Berger and Killingley 1982). The broken line labelled "10.8" represents my best estimate of the position of the transition midpoint. If one assumes (albeit incorrectly) that the plotted curve represents the output, after mixing, of a step function, then the broken line marks the position of the step. M marks the spot where deconvolution of the curve produces maximum overshoot ("meltwater-spike" of Berger *et al.* 1977). The recording of palaeoceanographical information starts near 5 cm, within the lowermost part of the mixed layer. The sea-level scale is based on the convention that 0.1 o/oo in the $\delta^{18}O$ signal corresponds to about 10 m of sea-level change (Shackleton and Opdyke 1973).

change of oxygen isotopes are shown in Table 22:1, as are the ages for levels of maximum preservation. The overall average age is 11 300 ^{14}C years B.P., both for the oxygen isotope mid-transition, and for the preservation maximum. Allowing for the fact that Core ERDC 88 has an unusually low sedimentation rate within this set of cores and seems winnowed and otherwise somewhat disturbed (Berger and Killingley 1982), the average

age may be recalculated without the values from this core. The result is an average age of 10 800 ^{14}C years B.P. for both the age of the oxygen isotope transition and the age of the preservation spike (see Fig. 22:1).

This age, obviously, is not significantly different from that of Broecker *et al.* (1960). All these age determinations must be viewed with caution because of mixing processes on the sea-floor. There are at least three effects, all tending to increase the ^{14}C-age of events recorded in deep-sea sediments. Firstly, the true age of deglaciation is probably younger than recorded in the sediment. A change in concentration of a tracer such as radiocarbon is first felt at the bottom of the mixed layer, within sediment that is already several thousand years old. Hence a correction toward a younger age may be in order (Peng *et al.* 1977, Berger and Johnson 1978, Johnson 1980). Secondly, the bulk sediment age can differ substantially from the coarse fraction age (Suess 1956, Olsson and Eriksson 1965). Commonly, the fine sediment is older, due to redeposition processes (see, however, Ruddiman *et al.* 1980a). The third effect is due to the ^{14}C age of the water within which plankton shells are formed, which adds several hundred years to the apparent age of newly formed sediment (Erlenkeuser 1979). The ^{14}C age of 10 800 years B.P. here proposed is therefore, a maximum age. The real ^{14}C age of the transition very likely lies between 9 000 and 10 000 years B.P.

To find a reliable age for the oxygen isotope transition, short of taking recourse to more or less arbitrary corrections, one might consider using cores from near-continent regions, with high sedimentation rates. Unfortunately, in such areas downslope redeposition is common, and special precautions have to be taken. A high rate core from the Gulf of Mexico analyzed by Emiliani *et al.* (1975) shows the maximum change in oxygen isotope values near 140 cm, for which level the ^{14}C age is 13 000 years B.P., based on "bulk core material". Another high-rate core, taken off the Niger delta, and analyzed by Pastouret *et al.* (1978) shows the maximum change in the oxygen isotopes of *Globigerinoides ruber* between 11 200 and 11 500 ^{14}C years B.P. A core off north-western Africa (Lutze *et al.* 1979) shows the oxygen isotope transition near 10 500 years B.P. Apparently, cores with high sedimentation rates do not necessarily yield ages significantly different from those of deep-sea cores.

Radiocarbon ages are not true ages, because the production rate of radiocarbon from nitrogen-14 changes through time (Stuiver 1978). For the time of deglaciation the error may be considerable (Oeschger *et al.* 1980) or it may be only a few per cent (Vogel 1980). If the CO_2-content of the atmosphere fluctuated (as suggested by CO_2-content in ice cores; Berner *et al.* 1980, Delmas *et al.* 1980) considerable change in $^{14}C/^{12}C$ ratios probably occurred within the transition, depending on the carbon reservoirs partaking in the fluctuations and their $^{14}C/^{12}C$ ratios (Siegenthaler *et al.* 1980). It appears preferable, therefore, to specify radiocarbon years whenever appropriate. The age of 13 000 years B.P. proposed by Shackleton and Opdyke (1973) for the isotopic stage boundary 1–2 is not a radiocarbon age. For this reason (and because of lack of detail) it is not comparable to the various age determinations cited.

WHERE SHOULD THE BOUNDARY BE PUT?

As I have tried to show, where geologists put the Pleistocene/Holocene boundary in deep-sea sediments, and why, is based on operational arguments rather than philosophical ones. Should this practice be changed?

It is always safe, of course, to demand that reason prevail over convenience. Yet, I do not think it makes much difference to science where one decides to put the "actual" boundary. For communication, it is preferable that researchers mean the same thing when referring to the "Holocene", rather than various periods ranging from 9 000 years B.P. to 13 500 years B.P. In practice however, a reference to the Pleistocene/Holocene boundary has to be checked against the data available to the researcher using the term in any case. The task in studying the Pleistocene/Holocene transition in a deep-sea core is to ascertain the sequence of events within it, and to compare this sequence with similar ones elsewhere. A stratotype section might be useful in this task provided it shows a sequence of events of world-wide significance. Any of a number of such events could function as a marker for the definition of a boundary. Changes in oxygen isotopes of shallow water organisms, and faunal changes, are recommended for purposes of correlation.

My own preference is to put the beginning of the Holocene at a time when at least two thirds of the sea-level rise was completed, and when major readvances of continental ice sheets no longer occurred. Assuming that salinity stratification developed during deglaciation in the North Atlantic (Olausson 1965, Berger 1978, Ruddiman *et al.* 1980b), the Holocene, by this definition, would begin with the final dissipation of such stratification. As far as can be ascertained by the available information from deep-sea cores, this turning point is close to 10 000 ^{14}C years B.P., most likely somewhat younger.

As far as placing the boundary in deep-sea cores, my proposal would result in putting the boundary after the midpoint of the transition in the oxygen isotope signal of planktonic foraminifers, on the late portion of the transition, but where the rate of change is still strong.

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