# 2. GENERAL COMMENTS ON THE PLEISTOCENE/HOLOCENE BOUNDARY

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## REMARKS ON THE CHOICE OF PERIOD FOR THE BOUNDARY

The Pleistocene/Holocene boundary was defined by the Holocene Commission as 10 000 $^{14}$C years B.P. Libby half-time. This is roughly the age of the Younger Dryas/Preboreal transition. A boundary of this type should be placed at the midpoint between glacial maximum and the postglacial temperature optimum, where most parameters exhibit changes. Sea-level curves (e.g. Bloom 1971, Mörner 1976) indicate that at least half of the ice stored in the continental ice sheets at the glacial maximum had melted by about 10 000 years B.P. The temperature curves (e.g. Flint 1971, Figs. 16–11) illustrate that the temperature rise was at the midpoint about the time of the Younger Dryas/Preboreal transition. However, the maximum Holocene warming in the Australia–New Zealand area occurred at 9 000 years B.P. and *c.* 6 000 years B.P. in the northern hemisphere (Salinger 1981). Considering these dates, the defined position is justified for the northern, but some 3 000 years too late for the southern hemisphere. The fact that temperature changes in the Australia–New Zealand area preceded changes in the northern hemisphere by about 3 000 years must be discussed in this context.

The extent of sea ice is a function of the salinity stratification. In earlier papers I discussed the last global deglaciation and its influence on the oceans and feed-back mechanisms. I suggested theoretical arguments for an open Arctic Ocean during the glacial maximum followed by a freezing over in deglacial times when the salinity stratification was re-established (Olausson and Jonasson 1969, Olausson 1972, Olausson 1981). The last change from open to ice-covered state in the Arctic could have been responsible for the Younger Dryas cooling, due to the great differences in the albedo between open and ice-covered seas.

A climatic deterioration during the Younger Dryas is well documented for the northern hemisphere, at least north of 30°N (Olausson 1969). The Weichselian deglaciation may therefore be divided into two phases in the north, while elsewhere it may have been either a period of more or less continuous, or a sudden, more intense, warming followed by more gradual climatic amelioration. The last trend will be illustrated further.

A major source of cooling in the southern hemisphere is the Antarctic ice sheet and the surrounding sea ice. The former was only about 10 per cent larger than today during the ice ages and, thus, fairly comparable in size. However, the surrounding pack ice varies far more in size, ranging between $24 \cdot 10^6 km^2$ (September) and $18 \cdot 10^6 km^2$ (February) through the years. This pack ice is a result of the salinity stratification around the Antarctic. The stratification is weaker than that of the Arctic Ocean, allowing formation of only a meter (or so) of sea ice (Weyl 1968). The deep salinity maximum (<34.9 o/oo) at 1–2 km depth has its origin in the effluent, very dense, Mediterranean water. This water encompasses the Antarctic continent. The present volume of the Mediterranean outflow is $1.6 \cdot 10^6 m^3 sec^{-1}$. As shown in earlier papers, the eastern Mediterranean was stagnant *c.* 11 000–8 000 years B.P. (Olausson 1961, 1965, 1969). The western basin was, so far as is known, not stagnant and the bottom-water exchange with the Atlantic (and the subsurface outflow) was, thus, restricted to this basin. The western Mediterranean is, at present, responsible for only 25% of the bottom-water formation and some 30 % of the salinity increase of the Mediterranean surface water due to evaporation. Both the volume and salinity could have been appreciably lower, the last being also due to outflowing, rather low-saline surface water from the eastern Mediterranean. Therefore, the deep salinity maximum could have been more or less absent in the south Atlantic during a few thousand years around 9 000 years B.P. This lack of salinity stratification should, in turn, have appreciably reduced the area of the sea ice around the Antarctic and, due to changes in albedo (from about 70 to about 10 per cent) during southern summers, improved the climate in the southern hemisphere. The extent of the sea ice around the Antarctic is more than twice that in the Arctic Ocean and any changes would, thus, exert a stronger influence on the climate than would the Arctic sea ice. This is equal to a semiglobal temperature change of 2°C during the northern summer (cf. my explanation of the post-Messenian climatic deterioration, Fig. 2:1).

The climatic optimum occurring 10 000–8 000 years B.P. in Australia and New Zealand (see Summary by Salinger 1981) and the coeval isotope maximum in the Antarctic ice dome C (Lorius *et al.* 1979) could result from the eastern Mediterranean stagnation and, indirectly from the deglaciation in the north.

I therefore conclude that the hydrographical changes in connection with the deglaciation affected the amelioration and caused both cooling and warmings of nearly semiglobal extent. Within this deglacial period a global boundary can be placed independently of the consequences of the feed-back

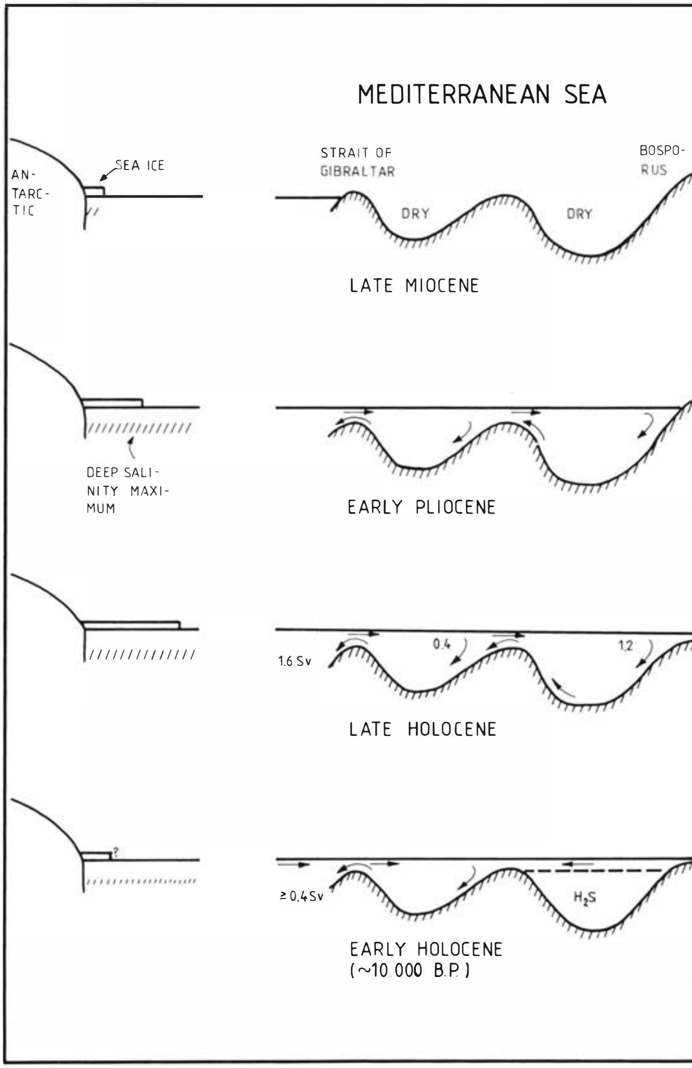


Fig. 2:1. The pack ice surrounding the Antarctic results from the salinity stratification in the underlying waters. The deep salinity maximum has its origin in the effluent, very dense, Mediterranean water. During the late Miocene (the Messianian ”Salinity crisis”) this efflux was stopped, and during the early Holocene stagnation in the eastern Mediterranean (Olausson 1961) the outflow diminished considerably.

When the water connection with the Atlantic was re-established after the “Salinity crisis” a salinity stratification around the Antarctic was developed. This led in turn to an increase in the extent of the sea ice there, and, by albedo changes, to a climatic deterioration. The early Pliocene cooling is proved by e.g. Kennett (1977).

During the afore-mentioned stagnant phase of the Mediterranean a conciderable reduction of the sea-ice area took place. This is here judged to be the cause of the very early warming in the Australia–New Zealand area (Salinger 1981), preceding that in the northern hemisphere by about 3 000 years, and the abnormally early isotope maximum in the Antarctic ice dome C (Lorius *et al.* 1979).

1 Sv (Sverdrup) = $10^6$ $m^3$ $sec^{-1}$.

from the oceans. The age 10 000 years B.P. fits well with the beginning of the Preboreal and the onset of the warming in Australia and New Zealand. Various sea-level curves show that at least half of the ice volume had melted by 10 000 years B.P., and that the hydrographical conditions were, thus, closer to the present conditions than to those during the glacial maximum.

In a global perspective the choice of period for the Pleistocene/Holocene boundary is justified.

## ON THE CHOICE OF TYPE LOCALITY AREA AND CRITERIA FOR THE BOUNDARY

It may be discussed whether a near-shore deposit is preferable as a stratotype section rather than deep-sea sequences. Referring to hydrographical changes and the rapid rise/decrease of many parameters in deep-sea cores, I formerly suggested that the boundary should be defined by means of deep-sea cores (Olausson 1969). However, as the resolution possible with near-shore deposits is much higher than in deep-sea oozes, they may be a better choice. My intention was then to find a marker, useful for the boundary definition in both areas. Meltwater injections could be such an instrument.

The meltwater spikes can be recognized in both shelf and deep-sea deposits. There are two large meltwater spikes in our cores, one in the Alleröd, and the other in the Preboreal substage. Only in near-shore deposits is this high resolution possible. The low resolution of deep-sea records ($\pm$2 000 years, see Berger, Chapter 22) preclude discernment of the intervening short interval (Younger Dryas $\sim$ 800 years) with a varied meltwater supply to the oceans.

The increased melting of the ice sheet at the onset of the Preboreal is a direct response to a temperature rise, and its result, a meltwater spike, is here considered as the prime signal of the Pleistocene/Holocene boundary. This signal can also be traced in deep-sea sediments (see Chapters 10 and 22). The second reaction to an amelioration is found in the fossil content (see, further, Chapter 20). However, various feed-back mechanisms can locally or semiglobally alter the general warming trend (see the discussion above).

## DATING PROBLEMS

The dating of the Pleistocene/Holocene boundary is a problem. Many molluscs seem to be reworked and it may be difficult to select *in-situ* shells.

In the future, benthic foraminifers can perhaps be used for dating purposes with the help of an accelerator technique. Radiocarbon datings of the Pleistocene/Holocene interval may also deviate markedly from calendar years due to past changes in the $^{14}C$ activity.

It is probably possible to affix the boundary to the Swedish varve chronology, which is being revised. The ice recession from Stockholm to Medelpad and its connection with present time is under review. The duration of the former part of the varve chronology has increased by about 100 years (Strömberg 1981). The connection with present time is still problematic, with an estimated error of $+200 \pm^{300}_{200}$ years (Fromm 1970). The chronology of the ice recession from the Fennoscandian terminal moraines to the Stockholm area is complicated. A control of this portion of the Swedish varve chronology is in progress. An important event is the final drainage of the Baltic Ice Lake, which occurred when the ice receded from the Fennoscandian terminal moraines, opening the connection between the Baltic and the Skagerrak. The lowering of 26–28 m of the Baltic Ice Lake is presumed to have occurred at 10 163 years B.P. (Donner and Eronen 1981), or 10 200 years B.P. (see Fredén, Chapter 3).

During the drainage of the Baltic Ice Lake some $10^4$ $km^3$ of water was discharged into the Skagerrak. The subsequent deglaciation, during the subsequent millennium, uncovered an area of about $10^6$ $km^2$ of Scandinavia (Bloom 1971). If we assume that this ice sheet was 0.5 km thick on average, we arrive at an idea of the amount of meltwater ($\sim 5 \cdot 10^5$ $km^3$) which was predominantly discharged through the Väner basin into the Skagerrak giving rise to the Preboreal meltwater spike in our cores.

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