Reports

Glacial and Pluvial Periods: Their Relationship Revealed by Pleistocene Sediments of the Red Sea and Gulf of Aden

Abstract. *Oxygen isotope analyses of planktonic foraminifera from the Red Sea and Gulf of Aden indicate that during periods of maximum continental and polar glaciation in the late Pleistocene, the Red Sea was subject to strong evaporation. Between glacial maximums the salinity of the Red Sea was equal to or below that of the open ocean. This suggests that high-latitude glacial periods corresponded in time to interpluvial stages in the present-day desert belt of northern Africa, whereas high-latitude interglacial periods coincided with pluvial stages.*

The relationship in time between high-latitude glacial and interglacial periods on the one hand and low-latitude pluvial and interpluvial periods on the other has been the subject of extensive discussions. Arguments both for and against a glacial-pluvial correspondence have been advanced (*1*). There is no general solution to the problem because arid and humid climate belts shift during glacial-interglacial transition periods, so that a wet period at one place may be synchronous with a dry period at another. For any specific location, however, the problem usually reduces to correlating isolated pluvial-interpluvial records to distant glacial-interglacial records.

Pleistocene sediments of the Red Sea and Gulf of Aden offer an opportunity to study glacial-pluvial relationships in the latitudes of the Red Sea (13° to 28°N) by analyzing the oxygen isotope ratios of foraminiferal tests. The Gulf of Aden freely exchanges water with the Indian Ocean and is thereby closely coupled to the variations of salinity and $^{18}O/^{16}O$ of the world's oceans in response to waxing and waning of polar and continental ice sheets. The Red Sea, on the other hand, by virtue of its narrow and shallow connection to the Gulf of Aden at the Strait of Bab al Mandab, has very limited exchange of water with the Indian Ocean. At times of lowered world sea level this exchange was further reduced, but even today the salinity and ^{18}O content of Red Sea water are higher than those of the ocean because of a low ratio of precipitation and continental runoff to evaporation (*2, 3*). When a more humid climate prevailed in the area, the precipitation/evaporation ratio must have been higher and salinity and ^{18}O content correspondingly lower. Inasmuch as planktonic foraminifera build their calcium carbonate tests in or very close to isotopic equilibrium with the water in which they live (*4*), the relative fluctuations in $^{18}O/^{16}O$ of the two adjacent seas in the past can be detected by analyzing tests of the same foraminifera species from the sediment of both areas and taking into account the effects of temperature on the equilibrium.

We have analyzed single-species samples of 20 to 50 individuals each of the planktonic species *Globigerinoides ruber* and *Globigerinella siphonifera* from Deep-Sea Drilling Project sites 228 in the Red Sea (19°05.16′N, 39°00.20′E) and 232 in the Gulf of Aden (14°28.93′N, 51°54.87′E). The results obtained for sediments deposited during the past 250,000 years are shown in Fig. 1. To facilitate comparison, the results from the two sites are plotted on the same time scale, based on data given in the *Initial Reports* of the Deep-Sea Drilling Project (*5*). The data for these two and other species give very similar but systematically offset curves, the relative positions of which on the $\delta^{18}O$ scale (*6*) are the same for both sites. The Gulf of Aden data match earlier data for a Caribbean core (*7*) in most major and many minor details, so that the major deglaciation stages of the last 250,000 years, designated terminations I, II, and III by Broecker and van Donk (*7*), were identified with considerable confidence. The major difference for which we have no satisfactory explanation at this time is the absence from our data of the pronounced sawtooth pattern of the oxygen isotope curves from other parts of the world.

Figure 2 shows the differences between $\delta^{18}O$ values of Red Sea and Gulf of Aden samples as a function of time. Because of small-scale uncertainties in the correlation and the presence of drilling disturbances, especially in the Red Sea cores, the smaller peaks created by one or a few points on either side of the zero lines in Fig. 2 may be spurious. The shape of the wider peaks is influenced by the sampling density and fossil assemblage in any interval. There is little doubt, however, that the wider peaks represent real differences which persisted over longer periods and that the peak heights indicate the extent of those differences. Profiles of the two species shown are very similar. The top portion of the Red Sea core contained only *G. ruber* in sufficient numbers, so that this core portion is missing from the *G. siphonifera* profile.

Comparison of the data from the Gulf of Aden with those from the Red Sea reveals the following features: (i) The maximum amplitude of the $\delta^{18}O$ fluctuations is considerably larger in the Red Sea than in the Gulf of Aden. (ii) Maximum $\delta^{18}O$ values occurred in the Red Sea just prior to and during major deglaciation periods I, II, and III. (iii) During these times foraminiferal tests in the Red Sea acquired a $\delta^{18}O$ around 2 per mil greater than the value for tests of the same species in the Gulf of Aden. (iv) In the intervals between the $\delta^{18}O$ maximums, foraminiferal tests in the Red Sea acquired a $\delta^{18}O$ around 1 per mil lower than that of tests of the same species living in the Gulf of Aden (Fig. 2).

In attempting to interpret these differences in terms of their climatic significance, one must consider the effects of changes in both temperature and isotopic composition of the water on the isotopic composition of foraminiferal tests. Assuming equilibrium, or a small but systematic deviation from equilibrium, between the isotopic compositions of the calcium carbonate of the foraminiferal tests and the water in which they were formed (*4*), one can place limits on the variation in either temperature or isotopic composition of the water. If the isotopic composition of the water in the Red Sea and the Gulf of Aden was identical during the three major glaciations of the past 250,000 years, then the isotopic data for the foraminifera require that during those periods the near-surface temperature was 9°C colder in the Red Sea than in the Gulf (*8*). Clearly, this is a highly unlikely postulate, considering the physiography and close proximity of these two seas. One would expect any temperature difference to be in the opposite sense because of the near or complete decoupling of the Red Sea from the glacially influenced open ocean during times of lowered sea level. Evaporative increase of the ^{18}O content of the Red Sea during those times is a more plausible explanation, which is also supported by sedimentological evidence (*9*). Even today, depending on the latitude, Red Sea surface water is enriched in ^{18}O due to evaporation by up to 1.1 per mil relative to Gulf of Aden surface water (*3*). Evidently, during the periods of maximum glaciation the climate in the area of the Red Sea was, on the average, considerably drier than today.

We can estimate how much the salinity rose in order to obtain an indication of the degree of dryness. To this end, we assume the same mean temperature in both the Red Sea and the Gulf during glacial maximums and, relative to today's conditions, an increase of 1 per mil in $\delta^{18}O$ and 3.5 per mil in salinity in the Gulf of Aden due to high-latitude ice formation (*3*). Then, based on the 2 per mil difference observed in the foraminifera (Fig. 2), the $\delta^{18}O$ of surface water in the Red Sea must have been 2 per mil higher than that in the Gulf of Aden. From Craig's relationship (*3*) between $\delta^{18}O$ and salinity in the Red Sea, the salinity in the Red Sea under those conditions can be calculated as 46.4 per mil. For each degree of difference between the mean temperatures of the two seas, the calculated Red Sea salinity would shift by 0.76 per mil in the same sense; that is, if the average temperature during a glacial maximum was 1°C higher in the Red Sea than in the Gulf of Aden, the salinity rose to 47.2 per mil. For reasons given above, it appears likely that the average temperature in the Red Sea was higher than that in the Gulf of Aden (as it is today), rather than lower. Thus a surface salinity of 46.4 per mil in the Red Sea during glacial maximums can almost be considered a minimum value.

Less obvious perhaps, but no less significant, are implications for the interglacial climate in the area of the Red Sea. The incomplete record for the interglacial periods in Fig. 2 contains two major sections during which $\delta^{18}O$ of the foraminifera in the Red Sea was close to 1 per mil lower than that of their contemporaries in the Gulf of Aden. As above, the difference might be interpreted as due to temperature only, and would then indicate temperatures 4° to 5°C higher in the Red Sea than in the Gulf of Aden. This possibility cannot be rejected offhand, but it implies identical surface salinities in both seas, which could happen only under climatic conditions appreciably wetter than prevail in the area today. If, however, the Red Sea was less than 4°C warmer than the Gulf, the salinity in the Red Sea must have been lower than that in the Gulf, a condition possible only if the Red Sea, in contrast to today's situation, had an excess of freshwater input over evaporation. Considering the small catchment area of the Red Sea, this would imply a very humid climate in that area. At times when the foraminifera in the Red Sea had a $\delta^{18}O$ 1 per mil lower than those in the Gulf, the salinity in the Red Sea could have been higher than that in the Gulf only if the temperature was higher by more than 4° or 5°C. At present, temperature differences of that order occur only for a brief period in late summer (*10*). If such occurrences were responsible for the differences in $\delta^{18}O$ observed in the fossil foraminifera, most or all of the tests must have been formed during "blooms" or

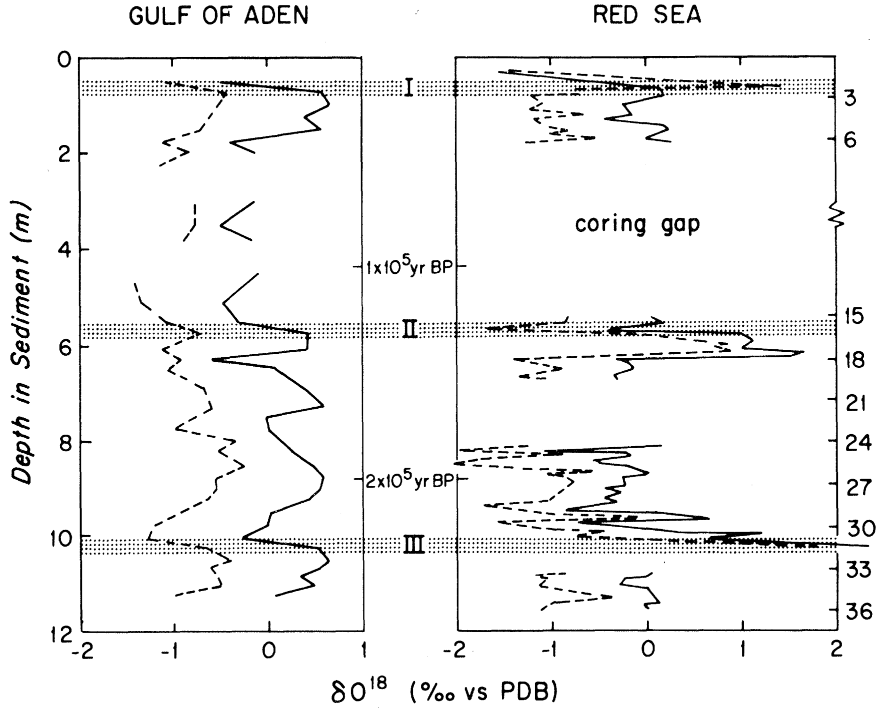

Fig. 1. Correlation of $\delta^{18}O$ variations in planktonic foraminifera *Globigerinoides ruber* (dashed lines) and *Globigerinella siphonifera* (solid lines) deposited at Deep-Sea Drilling Project sites 232 in the Gulf of Aden and 228 in the Red Sea over the past 250,000 years. Horizons I, II, and III are terminations (major deglaciation phases) noted by Broecker and van Donk (*7*) in the Caribbean. The Gulf of Aden curves represent 34 and 37 samples, respectively; the Red Sea curves 71 and 67 samples, respectively. Gaps in the profiles are due to incomplete core recovery during drilling or badly disturbed core material unsuitable for analysis.

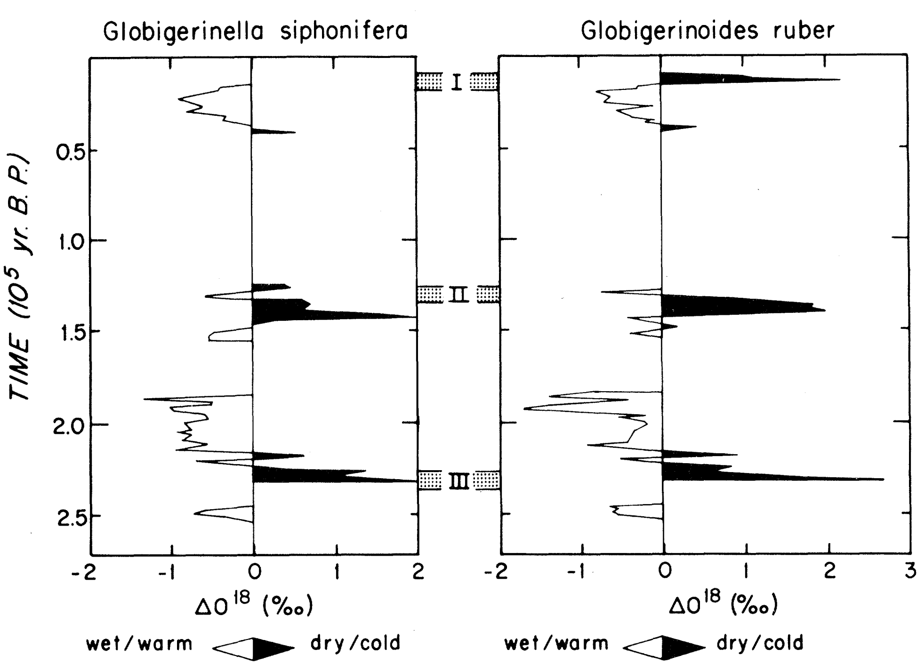

Fig. 2. Enrichment (black) and depletion (white) in ^{18}O of planktonic foraminifera in the Red Sea relative to their contemporaries in the Gulf of Aden. Horizons I, II, and III are as in Fig. 1. Gaps indicate intervals for which comparable data are lacking. Wet/warm and dry/cold refer to climate in the Red Sea area.

population explosions that coincided exactly with the periods of maximum temperature differences between the two seas. We do not have sufficient information to reject this possibility but do not consider it very likely.

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References and Notes

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Historical Dates for Neolithic Sites of Southeast Europe

Abstract. *Direct comparison of the radiocarbon contents of charcoal samples with those of bristlecone pine wood samples dated by tree rings shows that a full-fledged Neolithic with pottery and all the domesticated animals, except the horse, was present in southeast Europe as early as the 65th century B.C. The chronologies for the stratigraphic sequences of the settlements of Achilleion and Anza, based on a total of 37 La Jolla radiocarbon measurements, cover almost 1000 years.*

Neolithic sites have been excavated in the Aegean area from the beginning of this century (*1*), but their chronological context was not clearly understood for decades. The steady increase of systematic investigations of stratified mounds and caves (*2*), coupled with the advent of radiocarbon dating, brought on drastic revisions for interpreting the true age of the southeast European Neolithic (*3*–*5*). It is now clear that a full-fledged Neolithic stage of animal and crop domestication, including hard-baked pottery, existed in southeast Europe as early as the middle part of the seventh millennium B.C. At this stage, emmer wheat, barley, peas, and lentils were cultivated and all animals were domesticated except the horse. Obsidian trade was known; fine stone carving and ceramic techniques attest a high level of development, as does the architecture of solidly built rectangular houses of pisé or mud-brick walls and plastered floors. This fully developed agricultural civilization of Europe was essentially contemporaneous with those of Anatolia and Mesopotamia.

Of the 20 systematically excavated mounds in Thessaly and Macedonia, the settlements of Achilleion and Anza are of exceptional value for chronological studies. The mound of Achilleion is located on the edge of the Karditsa Plain of Thessaly, Greece. It was excavated in 1973–74 (*4*). Anza is located in Ovče Polje, in the River Vardar basin, in eastern Macedonia, and was excavated in 1969–71 (*5*). Both locations are multiple-strata sites yielding stratigraphic evidence of millennial duration, which covers complete sequences of the Sesklo civilization in Greece (Achilleion) and Starčevo–Early Vinča in Yugoslavia (Anza).

The long-lived bristlecone pine (*Pinus aristata*) of California has made it possible to establish a continuous absolute tree-ring chronology going back to the sixth millennium B.C. Prior to 4000 B.C. the chronology is based on 28 individual wood specimens (*6*). Cross-dating of these specimens has now made it possible to continue the sequence to beyond 6200 B.C. From this early sequence a limited number of samples, each comprising ten annual tree rings, consisted of more than 18 g of wood and were suitable for carbon-14 determinations.

Recently, European workers have been able to extend their tree-ring chronology, essentially from European oaks, to about 300 B.C. (*7*). Comparisons of samples from these two tree-ring series showed that irrespective of their place of origin, samples of wood grown at the same time have essentially the same carbon-14 content. Effects from in situ production of carbon-14 and from the physiology of wood growth must be negligible (*8*).

The experimental errors encountered when the carbon-14 contents of two wood samples are compared arise not only from statistical counting errors, but also from uncertainties in the counter calibrations. These calibration errors are practically avoided if measurements of the two samples are carried out in the same laboratory, preferably at the same time. In this way the most accurate single radiocarbon dates are obtained. In principle, it is possible to obtain results of even greater accuracy if series of samples from so-called floating tree-ring sequences are available. Age determinations of this type have been carried out for Neolithic settlements of Swiss lake dwellers (*9*). Unfortunately, floating tree-

Table 1. Conventional radiocarbon ages (*12*) of three samples from Achilleion (Table 2) and of six tree ring–dated bristlecone pine (BCP) wood samples listed in sequence of measurement. Values in the last column are given with their 1 σ statistical counting error (*12*). Abbreviations: LJ, La Jolla; B.P., before present.

LJ number	Sample	Archeological field number	Tree-ring age (years B.C.)	Conventional radiocarbon age (years B.P.)
3306	BCP wood		6030–6020	7125 ± 55
3308	BCP wood		5960–5950	6958 ± 100
3325	Charcoal	B-5 L-21		7287 ± 52
3310	BCP wood		5990–5980	7124 ± 52
3328	Charcoal	B-1 L-19		7307 ± 52
3311	BCP wood		6000–5990	7194 ± 53
3312	BCP wood		6020–6010	7156 ± 52
3329	Charcoal	B-1 L-26		7368 ± 52
3314	BCP wood		6050–6040	7103 ± 51
3313	BCP wood		5980–5970	7053 ± 51
Average of seven BCP wood samples			5999	7100 ± 21
Average of LJ 3325 and LJ 3328 (IIa2)			6211	7297 ± 36