RECEIVED
orig
FEB 16 1982
82-07
Research

THE PLEISTOCENE/HOLOCENE BOUNDARY IN THE RED SEA

By

Joshua H. Rosenfeld
Amoco Production Company
P. O. Box 3092
Houston, Texas 77001

ABSTRACT

Lowering of sea level during the Wisconsin glacial maximum probably isolated the Red Sea from the world ocean between 18,000 and 11,000 years B.P. A lowstand of Red Sea level some 345 meters below that of the contemporaneous ocean is suggested by hard crust deposition. Catastrophic refilling of the basin may have disrupted human occupation leading to population migration into the Dead Sea Rift Valley thereby contributing to the "flood" legend.

There is general agreement that increased salinity in the Red Sea between 20,000 and 11,000 years before the present (B.P.) was caused by limited water exchange with the Indian Ocean across a shallow sill during the period of glacially lowered sea level. Departing from this point of agreement, however, opinions are as diverse as the suggestion that complete isolation caused the Red Sea to dry up (Olausson, 1971) or, at the other extreme, that water exchange, although restricted, was never completely interrupted and that the Red Sea level never varied from that of the world's oceans (Yusuf, 1976).

Red Sea water level variation during the last 20,000 years merits further study. Data from sediment cores, regional geologic studies, paleoclimatic investigations, and the present day hydrography all suggest that isolation from the Indian Ocean indeed occurred, depressing Red Sea water level several hundred meters relative to the world ocean between 18,000 and 11,000 years B.P. Refilling of the basin during world sea level rise may have been catastrophic.

The Red Sea occupies a linear basin 1932 kilometers long whose average width is 280 kilometers (Figure 1). The basin formed between rifted sections of a pre-Oligocene craton. Lithospheric spreading and subsidence have been episodic since the Oligocene. More than 2,000 meters of salt were deposited in the trough during the Miocene "salinity crisis" (Stoffers and Kuhn, 1974) when the connected Red Sea and Mediterranean basins were subjected to prolonged dessication. The Mediterranean connection was severed at the end of the Miocene. Marine sedimentation resumed in the Red Sea when Pliocene rifting established the Indian Ocean connection. Renewed spreading during the Pleistocene produced an axial trough which is the locus of sea floor spreading and high heat flow (Gass, 1970).

Present day sedimentation reflects a moderately high productivity of calcareous pelagic, neritic and reef building organisms. Small, arid watersheds surrounding the basin result in the virtual absence of terrigenous sedimentation. Pelagic sediments accumulate at 4 to 12 centimeters per thousand years under the present regime (Milliman et al 1969). Coral reefs thrive in the Red Sea's warm and well oxygenated waters. The average annual surface water temperature is between 24° and 28° C. due to the NNW-SSE orientation of the basin axis between 12° and 29° North latitudes. Deep water temperature is almost constant at 21.8° C. (Oceanographic data from Morcos, 1971 unless otherwise stated).

The present day hydrography of the basin must be examined in order to understand the sensitivity of the Red Sea to changes in world sea level. No major drainage system feeds fresh water into the Red Sea. Precipitation in the watershed and onto the sea's surface averages 5 centimeters per year (Grasshoff, 1975). Evaporation has been calculated as 210 centimeters per year per unit area of water surface (Neumann, 1952) and this is the generally accepted figure, although estimates as high as 350 cm. per year have been published (Morcos, 1971). Therefore, a minimum deficit of two meters per year per unit area exists in the internal water supply necessary to replace evaporative loss. This replenishment takes place from the Indian Ocean through the Straits of Bab-el-Mandab at the southern end of the Red Sea.

Indian Ocean water of $36.7^{o}/oo$ salinity enters the Red Sea as a surface layer above an outflowing layer of $40^{o}/oo$ salinity. Inflow is about 0.29×10^{6} m^{3}/second and outflow is 0.26×10^{6} m^{3}/second (Figure 2); the excess influx replacing evaporation in the basin. Maximum bottom current velocity in the furrow north of Bab-el-Mandab is 60 to 70 cms. per second just above the substrate at 170 meters depth. Evaporation increases the salinity and density of the influx as it moves northward and it eventually sinks to form the southerly return flow. Under the present regime, all of the Red Sea is well mixed and oxygenated with the exception of some very local deep pockets containing hot brines up to 56° C and over $255^{o}/oo$ salinity (Brewer et al 1969). Mixing is enhanced by transverse prevailing winds.

A detailed bathymetric survey of the Bab-el-Mandab area was conducted by the German Research Vessel Meteor in 1964 (Morcos, 1971). The major bathymetric features are represented in Figure 3 (Coleman, 1973). Approaching from the north, the floor of the Red Sea shallows rapidly at about 15° N from over 1,000 meters to about 500 meters. Further decrease to 100+ meters occurs gradually in a furrow between the shallower shelves. 140 kilometers north of the narrowest part of the strait is the sill at Hanish Island. The 100+ meter depth here is the shallowest between the Red Sea deep basin and the Indian Ocean. Continuing southward, an elongate trench of up to 200 meters depth leads to an incision with very steep sides forming a narrow throat parallel to the axis of the strait. This incision shallows in the strait itself to a depth of 170 meters from whence it deepens to 300 meters south of Perim Island and continues down the continental slope into the Indian Ocean.

The amount of Late Pleistocene global sea level change is constantly being adjusted, but it is agreed that maximum eustatic lowering during the Wisconsin Ice Age was around 100 meters. Milliman and Emery's (1968) curve is used in this study (Figure 4). Their data suggest that world sea level was depressed by more than 80 meters between 18,000 and 11,000 years B.P. The removal of a water layer of this thickness from the Bab-el-Mandab/Hanish Island channel would result in an isostatic sea bottom rebound of about 20 meters (Olausson, 1971). This adjustment would be rapid, since mantle viscosity and crustal thickness are both reduced in this zone of lithospheric spreading. Therefore, it appears that the Red Sea was isolated from the Indian Ocean when world sea level declined 80 meters, with progressive limitation of water exchange during the interval immediately approaching total isolation.

The possibility of recent changes in bathymetry through tectonic, sedimentary and volcanic processes must be considered. The bathymetry certainly reflects some tectonic control but it is unrealistic to assign much of this to the last 11,000 years. Present day bottom currents seem sufficiently strong to prevent accumulation of fine sand size particles yet are too slow to account for extensive erosional deepening of the channel. This precludes recent shoaling of the strait through slow sedimentation. Possible slumping of material into the channel cannot be evaluated at this time nor can the contribution of effusives from nearby volcanic centers. Recent peralkaline basaltic and andesitic eruptions on the islands of Hanish and Zukur have been noted on geomorphic grounds (MacFayden 1932, Gass et al 1965) although no historical eruptions are on record. Therefore, at this time, the available data suggests that the shallow sill depth is not due to post-glacial tectonic, volcanic or sedimentary processes.

During the postulated period of isolation, the hydrologic regime would be altered not only by the loss of Indian Ocean water, but also by contemporaneous rainfall pattern and air temperature. Cool and wet conditions existed in the northern part of the Red Sea basin during the glacial maximum. Fossil snowlines of the Sinai Peninsula have been interpreted to indicate a mean annual temperature of 15° C; i.e. 10° C lower than at present (Gvirtzman et al 1975). Algal tufa shoreline deposits of Lake Lisan, found 220 meters above the present level of the

Dead Sea in the Jordan Rift Valley, have been dated radiometrically at 23,000 years B.P. These sediments contain fresh water diatom remains (Buchbinder et al 1974), indicating higher than present rainfall in the Jordan River watershed. No spillover of fresh water into the Red Sea through the Gulf of Aqaba was possible, however, due to the 400 meter high drainage divide between Lake Lisan and the Gulf. Paleo-rainfall and temperature data for the central and southern Red Sea basin are still equivocal (Assad. reply by Deuser, 1976).

Evidence for climatic variation is found in Red Sea sediments, although paleoclimatic effects are difficult to isolate from those of increased salinity. Both phenomena affected faunal assemblages and oxygen isotope ratios. Evidence from taxonomic analysis and oxygen isotope ratios has been synthesized in order to arrive at an average Red Sea surface water temperature of 13° to 14° C for the period from 23,000 and 13,000 B.P. (Berggren, 1969), i.e. 10° to 14° C cooler than at present.

Evaporation rate is not only a function of air and water temperature but is also dependent on the relative humidity and wind intensity of the overlying air. Wind action increases the water surface area and removes saturated air from the interface (Knauss, 1978). Although the pleni-glacial Red Sea's surface temperature was lowered, it is still likely that a regime of high insolation and moderate to low relative humidity prevailed during its isolation. Therefore, an initial evaporation rate of 125 centimeters per year seems reasonable for the Red Sea at that time. Assuming an average rainfall of 25 centimeters per year (5 times present rate), a 100 centimeter per year net evaporative loss of water may have existed per unit area. If evaporation were a linear function, complete dessication would have all but emptied the basin in 2,000 years. However, evaporation rate decreases as solutes are concentrated, an effect which may have eventually produced a still stand of the water level. To further determine the level at which the water level may have equilibrated, the sediments must again be consulted.

Complete dessication of 2,000 meters of seawater precipitates 12 centimeters of inorganic calcium carbonate, 2 meters of calcium sulfate, and 40 meters of halite and bittern salts per unit area in that order (Borchert, 1965). Piston cores taken by the Swedish Deep Sea Expedition (Olausson, 1971), the R/V Vema (Friedman, 1972) and the R/V Chain (Berggren and Boersma, 1969, Milliman et al 1969) sampled a lithified aragonitic and high magnesian calcite crust or "hard layer" dated by C^{14} at 13,000 to 11,000 years B.P. (Ku et al 1969). The following general description is from Gevirtz and Friedman, 1966. The hard crust consists of thin layers from several tenths of a millimeter to 5 millimeters thick composed of a feltlike matrix of aragonite needles growing syntaxially around pelletlike accretionary centers and pteropod tests with which they are in optical continuity. Extremely rare gypsum crystals occur in the layers. Total thickness of the horizon is not known since retrieval is generally in the form of broken chunks in a lime mud matrix. This fragmentation, ascribed to wave action by Olausson (1971), is more likely a result of dredging and coring processes (Milliman et al 1969). The hard layers have not been described in cores taken at water depths shallower than 516 meters (Gevirtz and Friedman 1966) although they are ubiquitous in deeper water cores.

Various interpretations have been given to sedimentary aragonite needles. They are found in lagoonal muds on the Andros Bank of the Bahamas and in the shallow waters of the Persian Gulf where local "whitenings" of the water are caused by rapid photosynthetic uptake of CO_2 in warm water (Cloud, 1962, 1965), with the possible involvement of carbonate secreting algae (Purdy, 1963). Inorganic aragonite whitenings occur at intervals of about 5 years in the hypersaline Dead Sea where they are triggered by peak summer temperatures after HCO_3^- build up from the Jordan River (Friedman 1965, 1972; Gevirtz and Friedman, 1966). The inorganic origin of the Red Sea hard crust aragonite is suggested by its strontium content of 1.07% (Milliman et al 1969) compared to 0.25% strontium in the aragonite of co-existing pteropod tests.

Aside from rare gypsum crystals, no evaporitic calcium sulfate has been sampled in any of the cores. Anhydrite layers which occur in sediments of the hot brine deeps are due to oxidation of sulfide in the presence of calcium ions within the brine pools (Bischoff, 1969). Halite has not been seen in any cores.

Evaporation apparently proceeded to the point where inorganic calcium carbonate and some gypsum were precipitated. This implies about 60% volume loss and a final salinity of 80 to $90^{0}/oo$. This would result in a lowstand about 500 meters below present day sea level. Removing this volume of water would elevate the Red Sea floor about 75 meters, assuming an isostatic rebound effect of 15%. Therefore, the final water surface of the Red Sea at steady state during isolation would be 345 meters below sill level. At this depth, almost all of the water would be confined to the axial trough with extensive exposure of the shelves.

Yusuf (1977) objects to this interpretation since sparse foraminiferal tests have been found in the hard layers. These may be wind transported or otherwise reworked.

Further evidence for the depression of Red Sea water level is equivocal. Bottom levels of minor offshore canyons (Sharms) in the Red Sea and Gulf of Aqaba have been correlated to a pleniglacial erosional base level of -130 meters (Gvirtzman et al, 1975) yet a sounding of -197 meters is shown in the Sharm of Port Sudan in one of their accompanying diagrams. In the same study, subaerially deposited Pleistocene conglomeratic alluvium was found at least 120 meters deep in the Gulf of Aqaba.

Turning now to the question of refilling the basin once world sea level rose sufficiently to breach the sill, the following calculations apply.

The present volume of the Red Sea is 0.215×10^6 km^3, from which an 80 meter deep layer must be subtracted in order to arrive at the isolation volume (V_1).

Therefore:

	0.215×10^6 km^3	= Present volume of Red Sea
Minus	0.028 " "	= Water above 80 meter isobath
	0.187 " "	= Basin volume at isolation
Minus	0.075 " "	= Water in basin after 60% evaporation.
	0.112 " "	= Total water deficit below sill level

Assuming various channel configurations and applying the combined Manning-Chezy equation for flow velocity, the channel discharge and time required to replace this water deficit can be estimated (See Table 1).

Influx would have had to be rapid enough to gain ground on renewed evaporation. Moreover, it would have had to do so quickly enough to produce a well mixed water column. If not, density stratification would have trapped saline residual water in the basin under incoming normal marine water. Such stagnation is marked by sapropelitic sediments deposited under anoxic conditions. However, the stratigraphic interval immediately overlying the hard layers shows a regime such as exists today; i.e. lime muds with abundant remains of pelagic foraminifera, diverse pteropod taxa, and coccoliths. There are no sapropels outside of the deep brine holes where stagnation and density stratification exist today (Berggren, 1969). These brines are not residual and are probably younger than 10,000 B.P. (Deuser and Degens, 1969). A further indication of rapid filling of the basin is a layer of crushed microfossil shells at the transition between hard layers and overlying normal marine sediments (Deuser and Degens, 1969). The presence of this fragmented zone, the sharp change in lithology, the absence of sapropel, and the sudden changes in fauna and oxygen isotopic composition of the calcareous tests all lead to the conclusion that the influx was sufficiently rapid to sweep shell fragments along as bedload or in suspension, and to cause mixing of the denser residual water and incoming Indian Ocean water. This change in sedimentation has been dated at 11,000 years B.P (Ku et al, 1969). Climatic change was rapid at the end of the last glaciation (Broecker, et al 1960) and sea level rise may have reached decimeters per year during the surging Valders ice advance (Emiliani, 1975). Data at hand do not allow further speculation on the time elapsed for equalization of Red Sea and world ocean levels, but a reasonable estimate may be between 5 and 250 years (Table 1)

The conclusions I have reached are presented in Figure 5 with a proposed sea level variation curve for the Red Sea superimposed on the world eustatic curve of Milliman and Emery.

A further analysis of the hypothesis may be possible when detailed submarine examination produces data on rock types, sediment distribution, bedforms, reef distribution and possible mass movements of the channel walls around the Strait of Bab-el-Mandab. Closer definition of the limits of hard crust deposition will constrain the lowstand even further.

The deeper segments of some piston cores (Ku et al, 1969) and Deep Sea Drilling (Site Reports DSDP V. 23) indicate that the events described herein are only the latest in a series of alternate high salinity and normal marine conditions through the Pleistocene. This latest event is of particular interest, however, since it occurred at a time of rapid human cultural evolution in the surrounding region and, perhaps significantly, coincides with arrival of the Eurafrican Natufian culture in the Jordan River Valley and the founding of Jericho in the tenth millenium B.C. (Melaart, 1965, p. 22, 23). Thus, a tenuous link may be established between a catastrophic flood and the earliest urbanized inhabitants of the region who were, perhaps, capable of incorporating this event into their oral tradition and eventually, into the Biblical flood account.

TABLE 1

$$U = 1.49/n\ (R^{2/3})(S^{1/2})$$

where: U = Velocity (meters per second)

n = Manning Coefficient (.030 for straight stream on a plane bed)

R = Hydraulic Radius (Channel area/wetted perimeter)

S = Slope (100 meters per 20 kilometers or .005)

Channel Width	Depth	R	S	n	U	Discharge m^3/sec	Filling Time in years
1,000m	20m	19.23	.005	.030	26m/sec	520,000	6.8
100	20	14.29	.005	.030	20m/sec	40,000	88.8
100	10	8.33	.005	.030	14m.sec	14,000	253.7

REFERENCES

Assad, G. 1977; Sea Straits and Glacial Periods in the Red Sea. Reply by Deuser, W., Science V. 195 p. 90.

Berggren, W. 1969; Micropaleontologic Investigations of Red Sea Cores, Summation and Sunthesis. in Hot Brines and recent heavy metal deposits in the Red Sea, p. 329-335.

Berggren, W. and Boersma, A. 1969; Late Pleistocene and Holocene Planktonic Foraminifera from the Red Sea, in Hot Brines and recent heavy metal deposits in the Red Sea, p. 282-298.

Bischoff, J. 1969; Red Sea Geothermal Brine Deposits, Their Mineralogy, Chemistry and Genesis, in Hot Brines and recent heavy metal deposits in the Red Sea, p. 368-401.

Borchert, H. 1965; Principles of Ocean Salt Deposition and Metamorphism, in Chemical Oceanography, 1st Ed. Academic Press p. 205-176.

Brewer, P; Densmore, C; Munns, R. and Stanley, R. 1969; Hydrography of Red Sea Brines, in Hot Brines and recent heavy metal deposits in the Red Sea, p. 138-147.

*HBRHMDRS = Hot Brines and Recent Heavy Metal Deposits of the Red Sea. Degens and Ross (Eds.) Springer-Verlag, 600 pp.

Broecker, W; Ewing, M. and Heezen, B. 1960; Evidence for an Abrupt Change in Climate Close to 11,000 Years Ago. Am. J. Sci. V. 258, p. 429.

Cloud, P. 1965; Carbonate Precipitation and Dissolution in the Marine Environment. in Chemical Oceanography, 1st Ed. p. 127-158.

Coleman, R. 1974; Geologic Background of the Red Sea. in DSDP V. 23 p. 813-819 and Map Pocket.

Deuser, W. and Degens, E. 1969; Oxygen and Carbon Isotopic Ratios of Fossils from the Hot Brine Deep Area of the Central Red Sea. in Hot Brines and recent heavy metal deposits in the Red Sea, p. 336-347.

Emiliani, C. 1975; Paleoclimatological Analysis of Late Quaternary Cores from the Northeastern Gulf of Mexico. Science, V. 189 p. 1083-1088.

Friedman, G. 1965; Occurrence and Stability Relationships of Aragonite, High Magnesium Calcite and Low Magnesium Calcite Under Deep Sea Conditions. GSA Bull. V. 76 p. 1191-1195.

Friedman, G. 1972; Significance of the Red Sea in the Problem of Evaporites and Basinal Limestones. AAPG Bull. V. 56 p. 1072-1086.

Gass, I. 1970; The Evolution of Volcanism in the Junction Area of the Red Sea, Gulf of Aden and Ethiopian Rifts. Phil. Trans. Roy. Soc. Lond. "A" V. 267 p. 369-381.

Gass, I; Mallick, I; and Cox, K. 1065; Royal Society Expedition to the Red Sea. Nature V. 205 p. 952-955.

Gevirtz, J. and Friedman, G. 1966; Deep Sea Carbonate Sediments of the Red Sea and Their Implications on Marine Lithification. Jour. Sed. Pet. V. 36 p. 143-151.

Grasshoff, K. 1975; The Red Sea, in Chemical Oceanography, 2nd Ed. Academic Press p. 578-597.

Gvirtzman, G; Buchbinder, B; Sneh, A; Nir, Y; and Friedman, G. 1975 Morphology of the Red Sea Fringing Reefs. Second Symposium International sur les Coreaux et Recifes Corallienes Fossiles. p. 480-491.

Knauss, J. 1978; Introduction to Physical Oceanography. Prentice Hall

Ku. T; Thurber, D. and Mathieu, G; 1969; Radiocarbon Chronology of Red Sea Sediments, in Hot Brines and recent heavy metal deposits in the Red Sea, p. 348-359.

MacFayden, W. 1932; On the Volcanic Zebayir Islands, Red Sea. Geological Magazine V. 69, p. 63-67.

Milliman, J. and Emery, K. 1968; Sea Levels During the Past 35,000 Years. Science V. 162 p. 1121-1123.

Milliman, J.; Ross, D. and Ku, T. 1969; Precipitation and Lithification of Deep Sea Carbonates in the Red Sea. Jour. Sed. Pet. V. 39 p. 724-736.

Morcos, S. 1971; Physical and Chemical Oceonography of the Red Sea Oceanogr. Mar. Biol. Ann. Rev. V. 8 p. 73-202.

Neumann, J. 1952; Evaporation for the Red Sea. Israel Exploration Journal V. 2 p. 153-162.

Olausson, E. 1971; Quaternary Correlations and the Geochemistry of Oozes. S.C.O.R. Proc. Symposium 1967. Micropaleontology of Marine Bottom Sediments p. 375-398.

Purdy, E. 1963; Recent Calcium Carbonate Facies of the Great Bahama Bank. J. Geol. V. 71 p. 472-497.

Stoffers, P. and Kuhn, R. 1974; Red Sea Evaporites: A Petrographic and Geochemical Study. DSDP V. 23 p. 821-847.

Yusuf, N. 1976; Trocknete das Rote Meer im Jungpleistozan aus? Naturwissenschaften V. 63 p. 576-577.

JHR/mlm
LR865-A

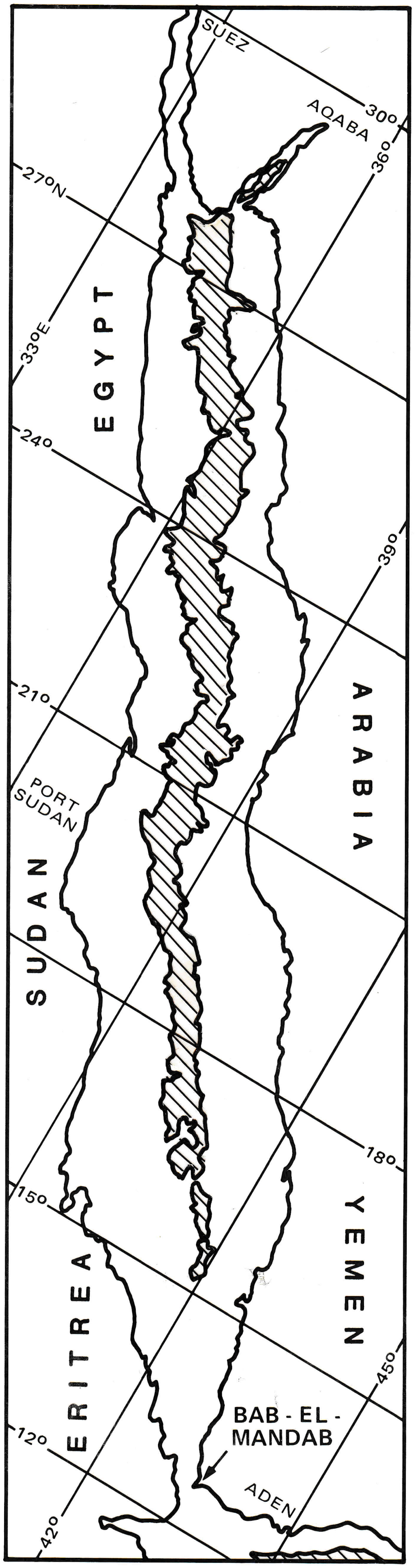

Figure 1 - The Red Sea

LEGEND

The Axial Trough Water Depths Greater Than 500 Meters Below Sea Level

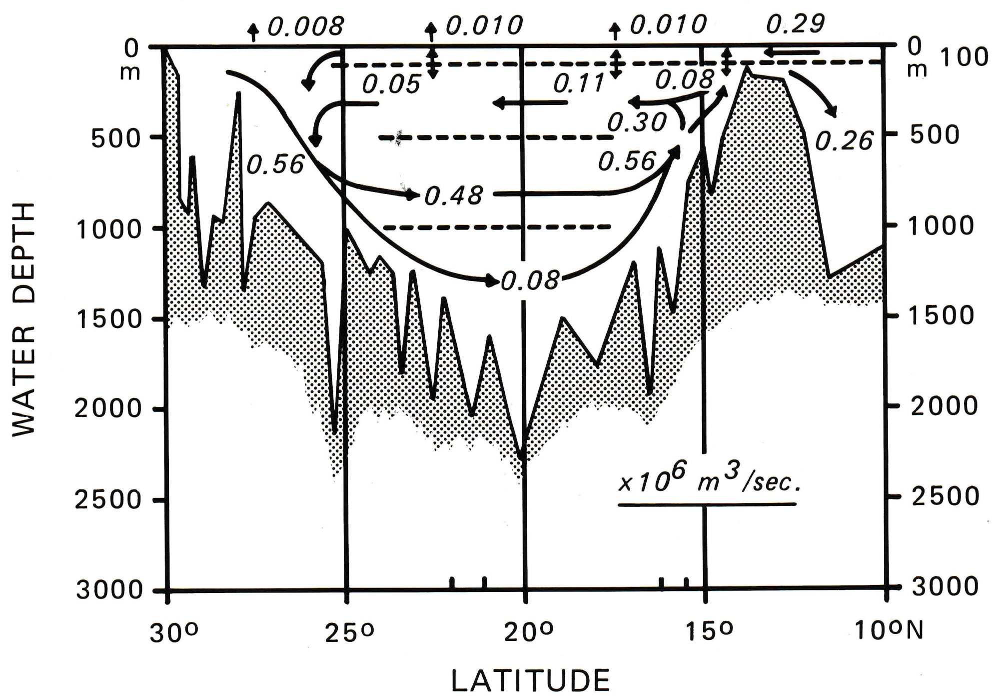

Figure 2. A schematic model for Red Sea circulation (From Morcos, 1971)

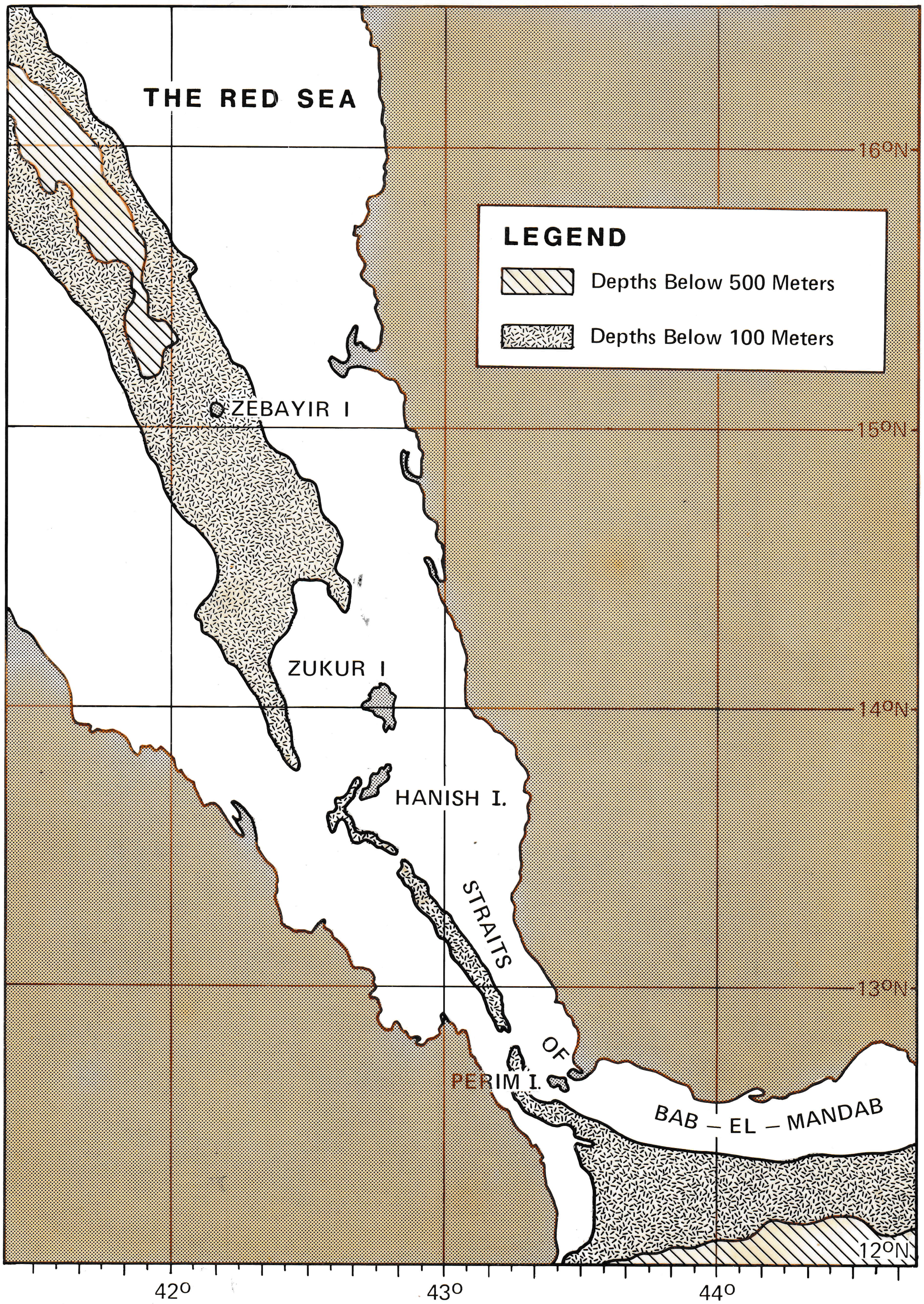

Figure 3. Bathymetry at The Straits of Bab-El-Mandab (From Coleman, 1974)

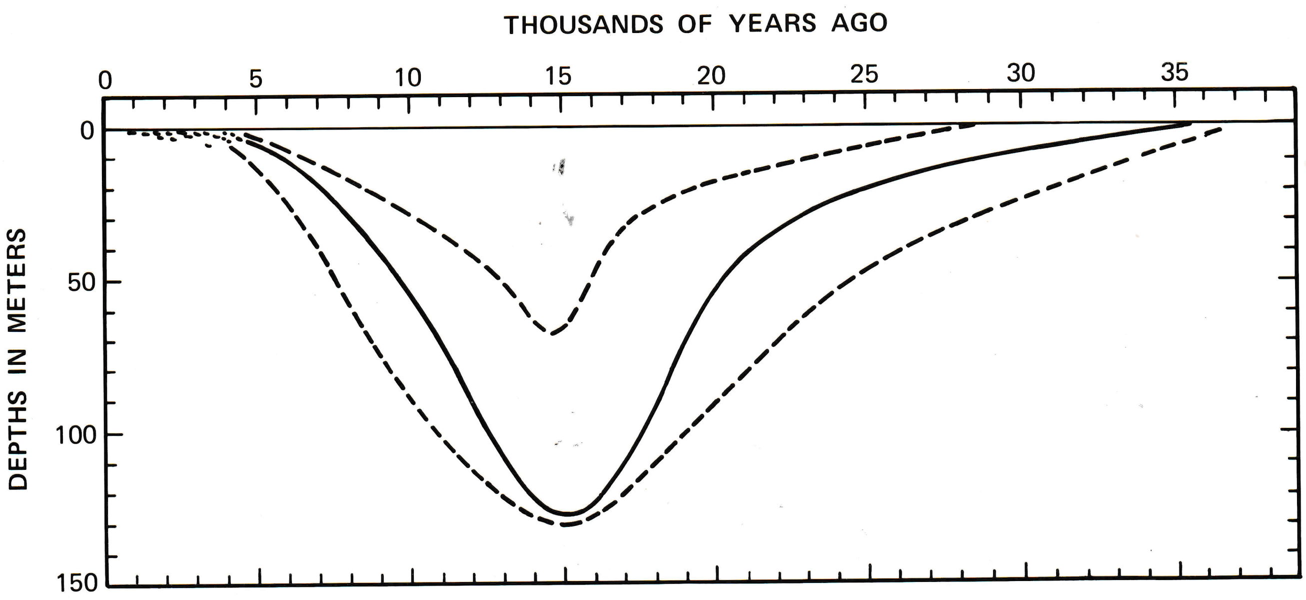

Figure 4. Depths and ages of sea-level indicators from the Atlantic continental shelf of the United States. The solid line is the inferred sea-level curve for the past 35,000 years; the dashed lines, envelope of values. (From Milliman and Emery, 1968)

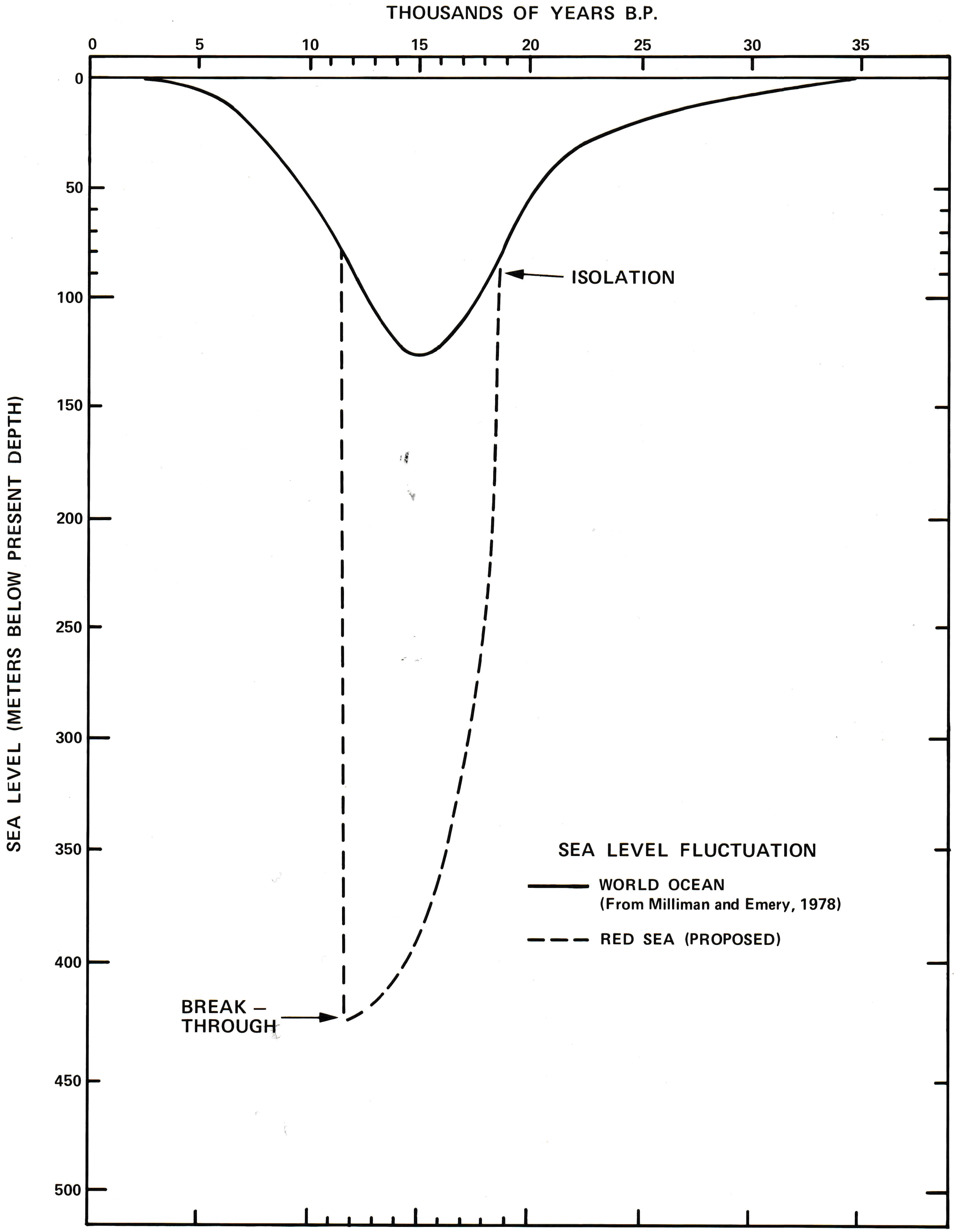

Figure 5. Proposed late Pleistocene Red Sea level fluctuation superimposed on world sea level fluctuation.