PALEOCEANOGRAPHY, VOL. 16, NO. 6, PAGES 658–668, DECEMBER 2001

Multidecadal variations in the early Holocene outflow of Red Sea Water into the Arabian Sea

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Abstract. We present Holocene stable oxygen isotope data from the deep Arabian Sea off Somalia at a decadal time resolution as a proxy for the history of intermediate/upper deep water. These data show an overall $\delta^{18}O$ reduction by 0.5‰ between 10 and ~6.5 kyr B.P. superimposed upon short-term $\delta^{18}O$ variations at a decadal-centennial timescale. The amplitude of the decadal variations is 0.3‰ prior, and up to 0.6‰ subsequent, to ~8.1 kyr B.P. We conclude from modeling experiments that the short-term $\delta^{18}O$ variations between 10 and ~6.5 kyr B.P. most likely document changes in the evaporation-precipitation balance in the central Red Sea. Changes in water temperature and salinity cause the outflowing Red Sea Water to settle roughly 800 m deeper than today.

1. Introduction

In order to improve our understanding of the processes driving climate change, a detailed knowledge of coupled oceanic-atmospheric dynamics is crucial. There are a number of strategic locations where variations in the evaporation-precipitation balance of marginal basins should affect the formation of intermediate water masses. The hypersaline Red Sea is a good example. To date, studies of the history of the water exchange between the Red and Arabian Seas have been carried out only from a Red Sea perspective. The intermediate/upper deepwater history in the Arabian Sea has received little attention. Variations in the $\delta^{18}O$ gradient in the Arabian Sea water column (e.g., between planktic and benthic foraminifera) have been attributed to changes in the thermocline thickness due to monsoonal strength variations [*Niitsuma et al.*, 1991], rather than to variations in Red Sea Water (RSW) outflow. On the basis of $\delta^{18}O$ variations in benthic foraminifera from the Oman Margin, *Zahn and Pedersen* [1991] were the first to conclude that owing to the lower glacial sea level, the sill depth at the Strait of Bab el Mandeb shoaled (Figure 1). Accordingly, the water exchange between the Red and Arabian Seas was reduced, and offshore Oman waters forming in the Arabian Sea replaced RSW. The restricted water exchange also resulted in salinities higher than 50 practical salinity units (psu) during glacial stages [*Hemleben et al.*, 1996; *Rohling*, 1994; *Thunell et al.*, 1988]. Modern water exchange patterns between the Arabian and Red Seas were not established prior to 6–7.5 ka [*Locke and Thunell*, 1988]. This implies that the early Holocene salinity values in the Red Sea were much higher than those today and must have influenced the history of early Holocene outflow. Here we present $\delta^{18}O$ data from middepth Arabian Sea, which document a short-term variability that can be explained by a variable inflow of RSW with unusual temperature-salinity (*T-S*) properties during the early Holocene.

2. Hydrographic Setting of the Arabian Sea

Below the monsoon-controlled surface mixed layer, four major intermediate to deep water masses dominate the Arabian Sea, i.e., Arabian Sea Water (ASW), Persian Gulf Water (PGW), Red Sea Water (RSW), and Circum-Polar Deep Water (CDW) (see Figure 2). Antarctic Intermediate Water does not penetrate further north than 5°N [*Wyrtki*, 1974; *You*, 1998] and hence does not occur in the modern northern Arabian Sea [see *van Bennekom et al.*, 1995]. The shallowest subsurface water mass in the Arabian Sea, the ASW, or Arabian Sea High-Salinity Water, forms in the northern Arabian Sea owing to winter cooling and predominantly spreads southward. Typical salinity values in the ASW are 35.3–36.7 psu, and temperatures range between 24° and 28°C [*Prasanna Kumar and Prasad*, 1999]. ASW prevails as a shallow subsurface water mass above 100-m depth in the eastern Arabian Sea [*Prasanna Kumar and Prasad*, 1999].

PGW enters the Arabian Sea as a warm and highly saline water mass through the Strait of Hormuz. Close to the PGW inlet, in the Gulf of Oman, salinity values reach 38.5 psu with temperatures around 24°C. Further downstream typical salinity and temperature ranges are 35.1–37.4 psu and 13–19°C, respectively [*Prasanna Kumar and Prasad*, 1999; *Rhein et al.*, 1997]. According to these *T-S* properties, PGW settles at 200- to 400-m water depth (Figure 2) [*Prasanna Kumar and Prasad*, 1999].

RSW forms in the northernmost part of the Red Sea [*Maillard and Soliman*, 1986; *Woelk and Quadfasel*, 1996] and advects into the Arabian Sea as a highly saline and warm intermediate water mass. During the SW monsoon, wind stress induces a strong subsurface inflow of Arabian Sea waters from the Gulf of Aden into the Red Sea, reducing the RSW outflow to minimum values. During winter, RSW outflow reaches its highest values [*Woelk and Quadfasel*, 1996]. At the exit of the Red Sea, the Strait of Bab el Mandeb, salinity and temperature values in RSW are 40 psu and 21°–22°C, respectively (Figure 2) [*Woelk and Quadfasel*, 1996]. Downstream, the RSW strongly mixes with surrounding waters. The main body of RSW passes through the Strait of Socotra flowing southward, and only a minor part continues to flow eastward (Figure 1) [*Shapiro and Meschanov*, 1991]. Off Somalia, the RSW core occurs at 800-m depth, and salinity and temperature values are ~35.6 and 11°C, respectively (Figure 2) [*Shapiro and Meschanov*, 1991]. On its track along the African coast it sinks to greater depth [*Quadfasel and Schott*, 1982]. RSW has been traced south of 25°S at 1000- to 1500-m water depth [*Gründlingh*, 1985].

CDW originates from the Southern Ocean and occurs in the deep Arabian Sea off Somalia below 1500 m. It consists of a mixture of 45% Wedell Sea water, 30% Pacific/Indian intermediate waters, and 25% North Atlantic Deep Water (NADW; for a summary, see *Emery and Meincke* [1986]). Characteristic temperature and sal-

Paper number 2000PA000592.
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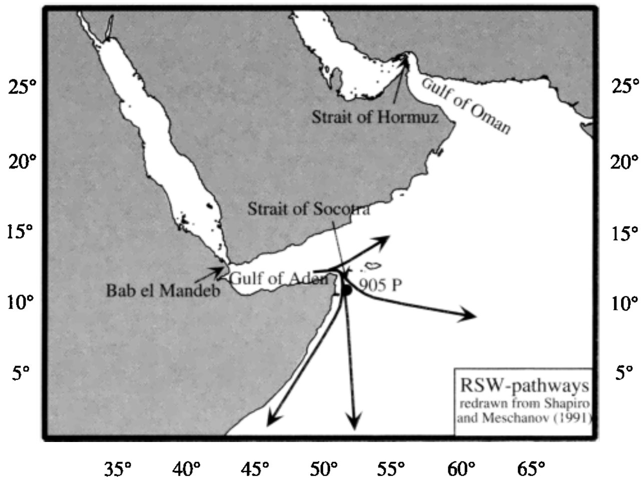

Figure 1. Location of core 905. Bold arrows are redrawn from *Shapiro and Meschanov* [1991] and indicate the main flow path of Red Sea Water (RSW) in the Arabian Sea.

inity properties in the Arabian Sea are 0.1°–3°C and 34.6–34.9 psu, respectively [*Emery and Meincke*, 1986].

3. Methodology

Piston core 905 was retrieved from a water depth of 1580 m offshore Somalia during the Netherlands Indian Ocean Program (NIOP, 1992/1993; Figure 1). The age model of core 905 is based on 12 ^{14}C dates [*Ivanova*, 2000] (Figure 3) placed at strategic depth intervals covering the last 31 kyr B.P. The average sedimentation rate in this period is around 19 cm kyr^{-1}. In the early Holocene, sedimentation rates reach up to 36 cm kyr^{-1}. To fully explore the paleoceanographic history in the western Arabian Sea, early Holocene and late glacial subsections were sampled at 0.5-cm steps, yielding an average time resolution of 10–20 years.

Following standard sample preparation techniques (washing and sieving), the stable oxygen isotope values of the planktic foraminifer *G. ruber* and the epibenthic foraminifer *C. kullenbergi* were measured with a Finigan MAT 252 mass spectrometer coupled to an automated carbonate preparation line (Bremen type; for detailed methods, see *Ivanova* [2000]). The stable oxygen isotope records (Figures 3 and 4) are average values of replicate analysis: three individual analyses of four tests of *G. ruber* and of up to five (in rare cases one) single-specimen analyses of *C. kullenbergi* (size fraction 250–350 μm). The internal reproducibility for $\delta^{18}O$ measurements is ±0.05‰, resulting in an analytical detection limit for $\delta^{18}O$ variations of 0.1‰. The significance of the observed benthic $\delta^{18}O$ variations will be discussed in section 4. The data set is online available via the National Geophysical Data Center in Boulder, Colorado.

4. Intermediate/Upper Deep Water Dynamics During the Last 20 kyr B.P.

There are three main findings of the pre-Holocene $\delta^{18}O$ record of the epibenthic foraminifer *C. kullenbergi* (Figure 3). First, a 0.9‰ reduction in $\delta^{18}O$ is recorded during Termination Ia. This represents a 0.3‰ larger drop in $\delta^{18}O$ than was inferred from changes in the global ice volume during Termination Ia according to *Fairbanks* [1989]. Second, an extreme $\delta^{18}O$ increase by 0.8–0.9‰ to almost glacial type $\delta^{18}O$ values of ~3.5‰ marks the Younger Dryas (YD). This point is based on a single measurement due to limited sample availability. Around that time the deepwater connection between the Red Sea and the Arabian Sea began to reestablish [*Almogi-Labin et al.*, 1996; *Fenton et al.*,

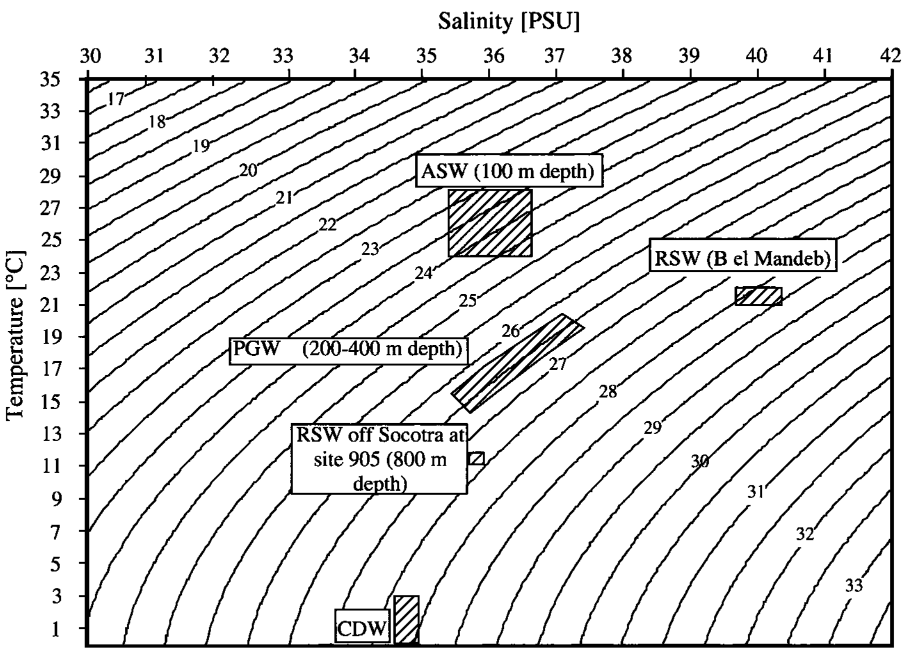

Figure 2. Temperature-salinity (*T-S*) field with contoured density lines (δ_θ values given) according to *Cox et al.* [1970]. Hatched fields indicate the *T-S* properties of the main modern subsurface, intermediate, and deepwater masses in the Arabian Sea. Data sources are as follows: ASW, Arabian Sea Water [*Prasanna Kumar and Prasad*, 1999]; PGW, Persian Gulf Water [*Prasanna Kumar and Prasad*, 1999; *Rhein et al.*, 1997]; RSW (Bab el Mandeb and off Socotra, own data); and CDW, Circum-Polar Deep Water (own data). For further explanation, see text.

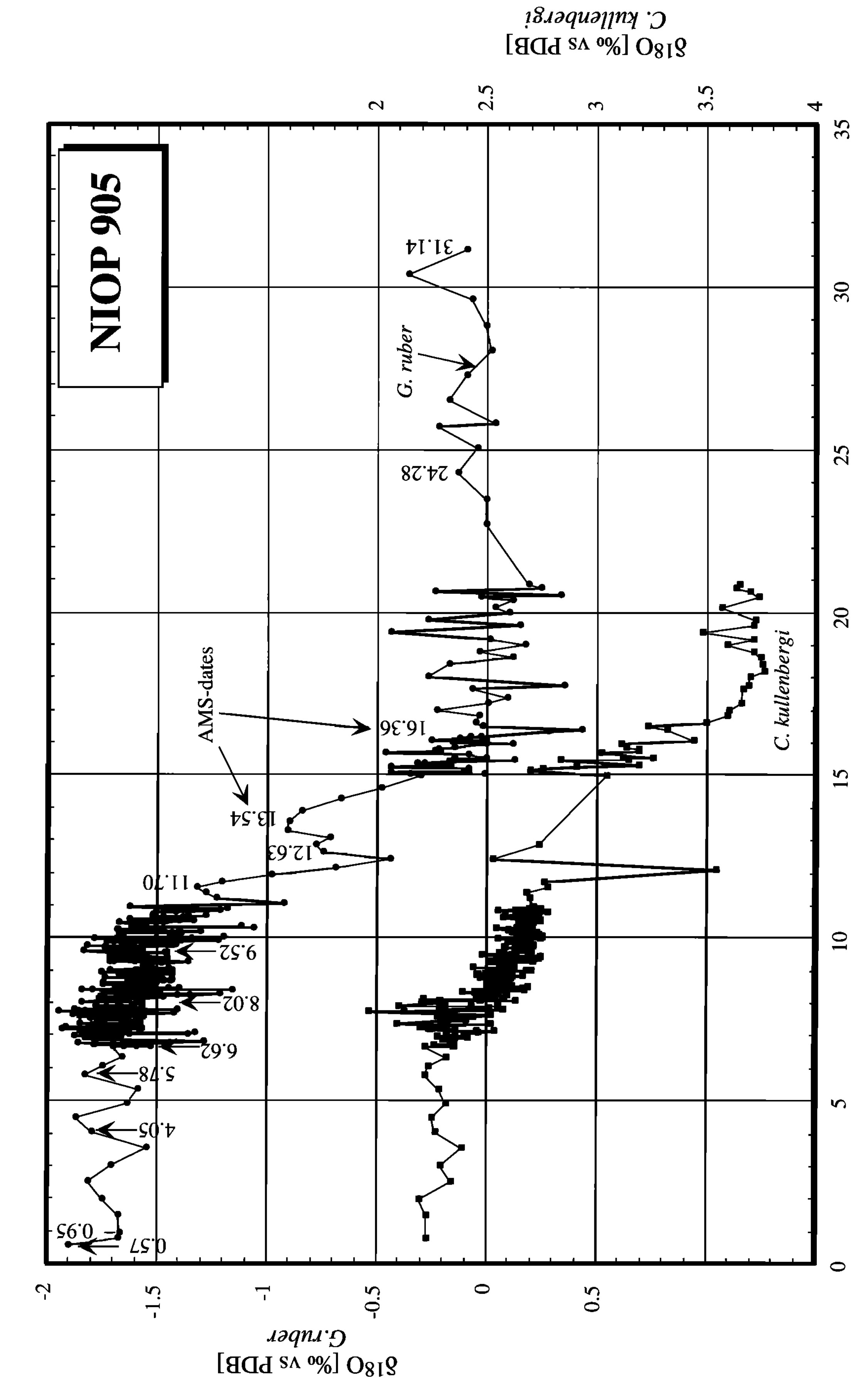

Figure 3. $\delta^{18}O$ records of the planktic foraminifer *Globigerinoides ruber* and the epibenthic foraminifer *Cibicidoides kullenbergi* over the last 35 kyr B.P. Accelerator mass spectrometry (AMS) ^{14}C dates are indicated by arrows (data source is *Ivanova* [2000]). PDB, Peedee belemnite; NIOP, Netherlands Indian Ocean Program.

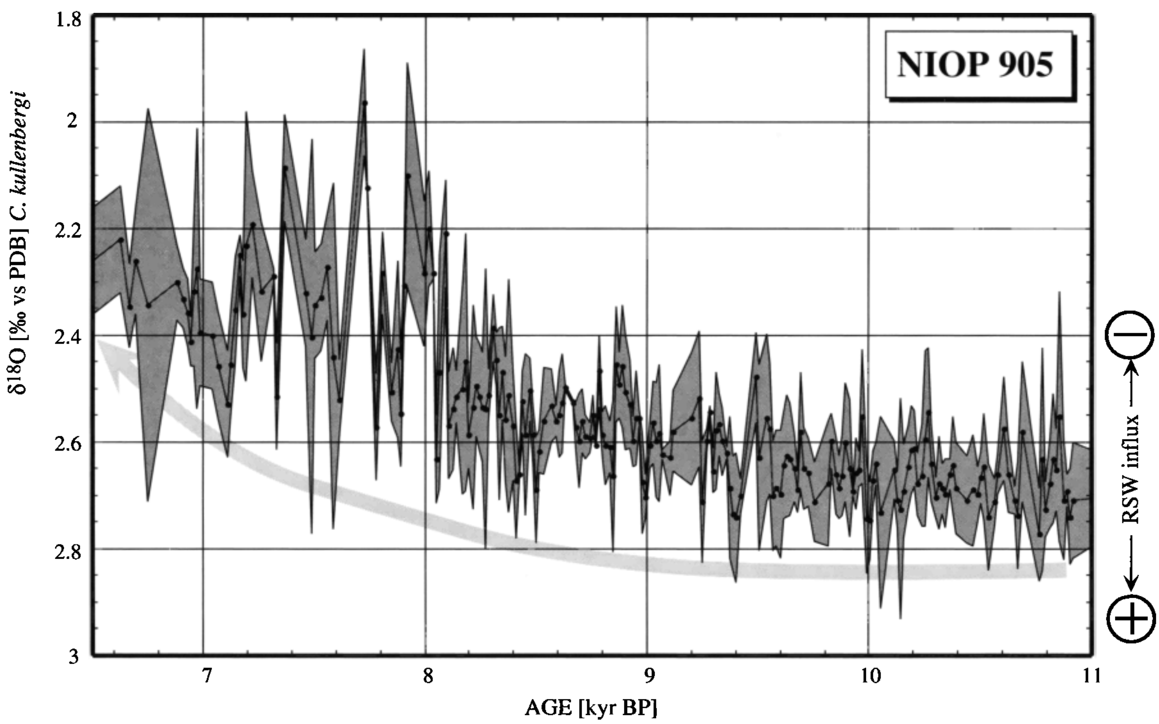

Figure 4. Detailed $\delta^{18}O$ record of the epibenthic foraminifer *C. kullenbergi* between 11 and 6.5 kyr B.P. Shaded area marks the standard deviation of the multiple analyses. For single analysis a constant standard deviation of 0.1‰ was used. For further comments, see text.

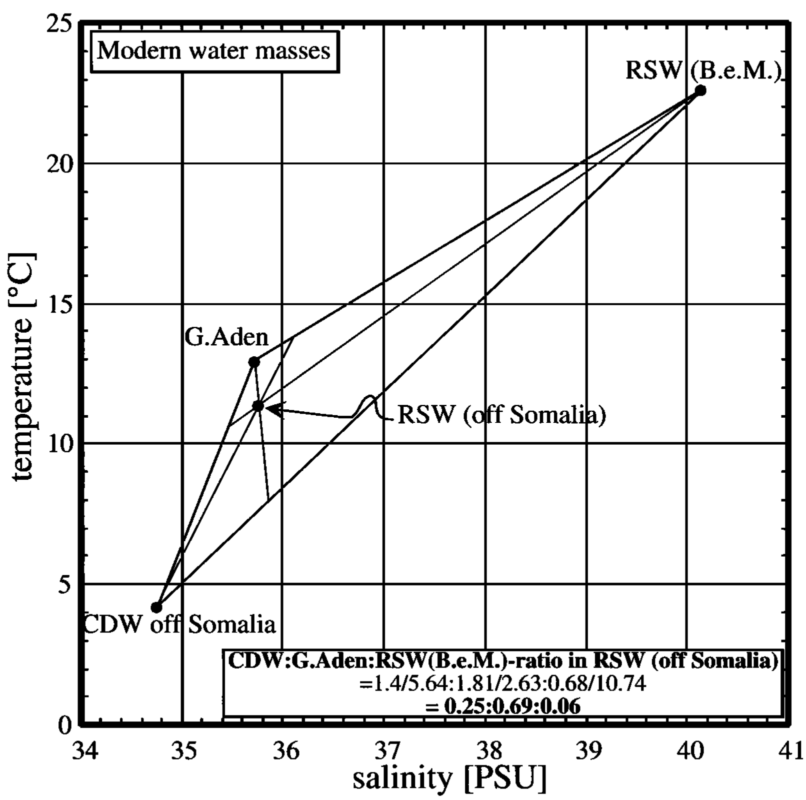

Figure 5. *T-S* distribution of the main water masses contributing to RSW occurring off Somalia. Proportion calculations are based on triangulation and given at the bottom of the graph. RSW (B.e.M.), RSW (Bab el Mandeb); G. Aden, Gulf of Aden. For further explanation, see text.

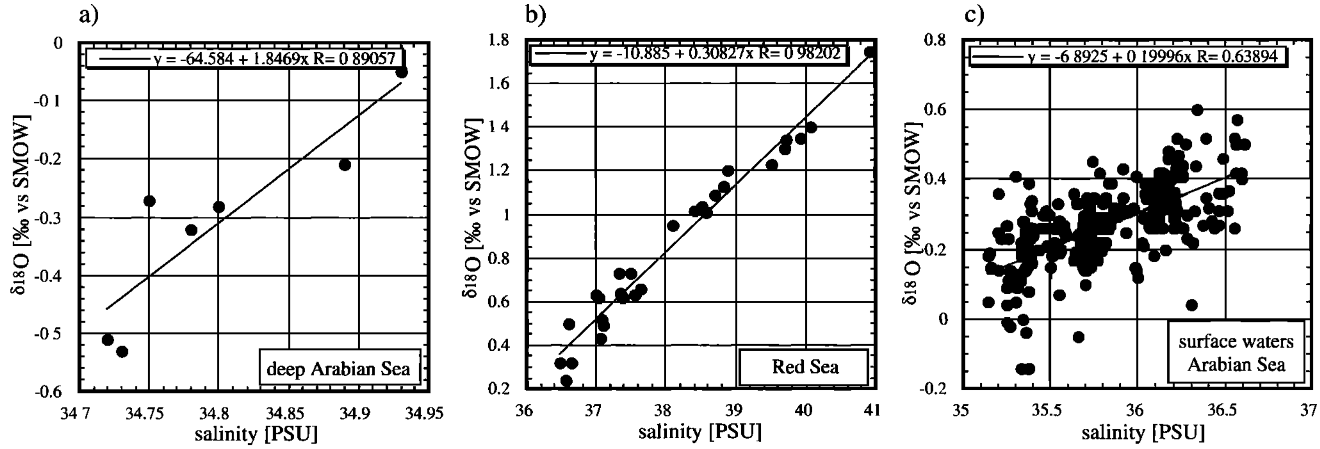

Figure 6. $\delta^{18}O$/S relation in water masses occurring in the Arabian Sea: (a) water masses from the Arabian Sea below 2000 m, (b) surface water from the Red Sea, and (c) surface water from the Arabian Sea. Linear regression lines/coefficients are given. Data source is unpublished data from the Netherlands Indian Ocean expedition.

2000; *Hemleben et al.*, 1996]. Hence the benthic $\delta^{18}O$ maximum spike may represent the first sign of RSW in the Arabian Sea. Finally, a second $\delta^{18}O$ reduction of 0.8–0.9‰ occurs during Termination Ib. The overall glacial-early Holocene $\delta^{18}O$ reduction is 1‰, which is 0.2–0.3‰ smaller than the widely accepted 1.2–1.3‰ related to global ice volume change [*Fairbanks*, 1989; *Labeyrie et al.*, 1987]. Between 10 and ~6.5 kyr B.P, $\delta^{18}O$ values generally decrease from ~2.7 to 2.2–2.3‰ (Figures 3 and 4). The latter is the level that persists through the remainder of the core. Superimposed on these changes are ultra short-term $\delta^{18}O$ variations with a decadal/centennial time resolution (Figure 4). Between 10 and 8.1 kyr B.P the amplitude of these changes reaches up to 0.3‰. The short-term $\delta^{18}O$ fluctuations occurring between 8.1 and ~6.5 kyr B.P have significantly higher amplitudes. There are three main oscillations with amplitudes of up to 0.6‰ in this time period (Figure 4) with minimum spikes that center around 7900, 7730, and 7370 kyr B.P. The amplitude of the fluctuations decreases approaching ~6.5 kyr B.P.

Before proposing possible explanations for these findings, we must assess the effect of bioturbation on the paleorecords. Bioturbation of sediments resembles a low-pass filter [e.g., *Trauth et al.*, 1997]. In general, the higher sedimentation rate will improve the potential of high-frequency climate change to be preserved, particularly if bioturbation depth is constant. The absolute effect of bioturbation on the preserved record of core 905 is difficult to assess. Given the high sedimentation rate of up to 36 cm kyr^{-1} we believe it is conservative to assume that its influence is minor. In order to assess the variability in $\delta^{18}O$ recorded by each sample, the standard deviation was calculated (an error of ±0.1‰ was assumed for each single analysis). Figure 4 shows that between 10 and 8.1 kyr B.P, although some of the $\delta^{18}O$ variation is within error, there is significant variation, in particular on centennial timescales. This finding suggests that a substantial part of the variability shown reflects the natural variability within the time period covered by each sample. Accordingly, we assume that the small $\delta^{18}O$ variations, although clearly close to the detection limit of the method, reflect the natural variability off Somalia. The high-amplitude variability between 8.1 and ~6.5 kyr B.P records major changes in climate and/or oceanography.

5. Discussion

A series of explanations are required to explain why benthic foraminifera at 1580-m water depth off Somalia record: (1) a 0.2–0.3‰ smaller drop in $\delta^{18}O$ during Termination I than was inferred from the global ice volume change and (2) a general decrease by 0.5‰ between 10 and ~6.5 kyr B.P, which has (3) decadal/centennial scale $\delta^{18}O$ variations superimposed upon it that have amplitudes of up to 0.3‰ between 10 and 8.1 kyr B.P and up to 0.6‰ between 8.1 and ~6.5 kyr B.P. The processes governing $\delta^{18}O$ variations of calcareous organisms depend on the temperature and $\delta^{18}O$ of the ambient seawater. $\delta^{18}O_{water}$ varies with salinity. Hence, in order to understand the causes of the $\delta^{18}O$ variations, an assessment of current and possible past *T-S* and $\delta^{18}O$ properties of water masses in the Arabian Sea is required.

5.1. RSW Mixing in the Arabian Sea: Implications for $\delta^{18}O$ and Salinity

The *T-S* properties of RSW change as it advects through the Strait of Bab el Mandeb and mixes with local subsurface waters in the Gulf of Aden [*Shapiro and Meschanov*, 1991]. In the open Arabian Sea, RSW also mingles with CDW. A simple three-end-member *T-S* mixing model illustrates the different contributions of water masses to the modern RSW as it is found at roughly 800-m depth off Somalia (Figure 5). This model includes source RSW (Bab el Mandeb properties), subsurface waters from the Gulf of Aden, and the CDW. A simple triangulation implies a pure RSW content of ~6%, a Gulf of Aden (GOA) water content of ~69%, and a ~25% admixture of CDW to explain the mixed RSW off Somalia (see Figure 5). In order to assess possible variations in admixture rates, we use a new set of $\delta^{18}O$ data from the Red Sea and the northern Arabian Sea (Figures 6a–6c). The $\delta^{18}O$/salinity correlation slope for the Arabian Sea below 2000 m is >1.8 (Figure 6a), a number that is slightly higher than the value of 1.5 given by *Zahn and Mix* [1991]. However, "our" slope is based on only a few data points compared to that of *Zahn and Mix* [1991]. We therefore adopt a deepwater $\delta^{18}O$/*S* slope of 1.5 for the discussion below. In contrast, the $\delta^{18}O$/*S* slope in surface waters is generally much lower. For surface waters in the northwestern Arabian Sea it is only 0.1–0.2 with a large standard deviation (Figure 6c), and in the Red Sea the $\delta^{18}O$/*S* slope is 0.3 (Figure 6b), a value almost identical to the value published by *Craig and Gordon* [1965]. The weak $\delta^{18}O$/*S* correlation in surface waters from the Gulf of Aden implies that variations in the supply of this water mass would have little effect on the $\delta^{18}O$ of the resulting mixed RSW but could potentially cause major changes in the *T-S* properties. Consequently, any large $\delta^{18}O$ variations found in mixed RSW cannot result simply from differing proportions of Gulf of Aden waters.

Table 1. Parameters for Numerical Contribution Experiments[a]

Water Mass	Modern				Experiment 2			Experiment 3		
	Temperature	Salinity	Density	$\delta^{18}O$	Temperature	Salinity	Density	Temperature	Salinity	Density
CDW	4.3	34.93	27.7	−0.05	-	-	-	-	-	-
RSW (B.e. Mandeb)	22	40.13	28.1	1.43	22	47.5	33.8	22	45.5	32.2
RSW off Somalia	11	35.61	27.25	0.29	11.1	41.8	32.1	8.3	39.7	30.9

[a] Data for modern T (potential T), S, and $\delta^{18}O$ water values are from the Netherlands Indian Ocean Program (G. Ganssen, unpublished data, 1992–1993). Modern temperature of Red Sea Water (RSW) (Bab el Mandeb) is used in experiments 2 and 3 (see also text). Salinity for RSW (Bab el Mandeb) is taken from the most probable RSW contribution/salinity vector (mean value; see text). T and S values for RSW off Somalia are deduced from the RSW (Bab el Mandeb) contribution value and differential diffusion factor (for temperature only). Density (sigma theta) values are based on the individual T and S parameters given for the modern situation and for experiments 2 and 3. CDW, Circum-Polar deep water.

Based on the outlined $\delta^{18}O$ properties, two models may explain the observed variations off Somalia. In the first scenario the oxygen isotope record reflects in situ T/S variations in the CDW (hypothesis 1), a scenario that we reject (see section 5.2). In a second scenario, early Holocene RSW was generally more saline and denser and accordingly settled deeper in the water column. In this view the variations in $\delta^{18}O$ reflect evaporation-precipitation changes in the Red Sea (hypothesis 2).

5.2. Hypothesis 1: CDW Variations

In the first hypothesis we assume that the observed $\delta^{18}O$ pattern in the benthic record reflects in situ T-S changes in CDW, which is the prevailing modern water mass at site 905. If we translate the most striking $\delta^{18}O$ variation of 0.6‰ (between 8 and ~6.5 kyr B.P) solely to a temperature change in CDW, this would imply variations of ~2.4°C (using a $\delta^{18}O/T$ ratio of 0.25 [*O'Neil et al.*, 1969]; for discussion, see *Zahn and Mix* [1991]). Comparable temperature changes for NADW occur on glacial-Holocene timescales [*Jung*, 1996; *Labeyrie et al.*, 1987]. In relation to these extreme changes we argue that it is unlikely that this magnitude of temperature change occurred during the relatively stable Holocene. Furthermore, if we would ascribe a temperature change of 2.4°C off Somalia solely to T variations in NADW (25% contribution to CDW), this would imply >9°C warmer NADW, a highly unlikely scenario. Similarly, advocating that the induced T variation of CDW was solely caused by Antarctic Bottom Water (AABW) would imply a ΔT of >5°C in the latter. This is an equally unlikely explanation.

It is also difficult to explain the observed high-amplitude and short-term $\delta^{18}O$ oscillations by salinity changes in deepwater masses. Given the $\delta^{18}O_{water}/S$ correlation of 1.5 in deep water [*Zahn and Mix*, 1991], the $\delta^{18}O$ amplitude of 0.6‰ implies a ΔS of 0.4–0.5 psu (see Figure 2). Such a change is equivalent to 40% of the inferred salinity drop in deep water on glacial-interglacial timescales. The Holocene is a time without large-scale melting of continental ice sheets; hence, such a salinity variation is highly unlikely to have occurred in deep waters.

Changing the proportion of AABW input to mixed RSW also does not help to explain the observed changes. This is because T and S characteristics within the deep Atlantic Ocean have counteracting effects on the $\delta^{18}O$ value of benthic foraminifera [*Zahn and Mix*, 1991]. Any possible minor increase in $\delta^{18}O$ by a higher AABW influx is counteracted by its generally lower $\delta^{18}O$ compared to NADW, resulting in a zero net change in the mixed water [see *Zahn and Mix*, 1991]. Furthermore, the small Last Glacial Maximum (LGM)-Holocene drop in $\delta^{18}O$ of 1‰ off Somalia compared to the global ice volume effect of 1.2–1.3‰ [*Fairbanks*, 1989; *Labeyrie et al.*, 1987] also suggests a water mass change at site 905 rather than in situ changes in CDW. This observed relative rise in $\delta^{18}O$ implies a cooling in CDW by >1°C or an increase in salinity by roughly 0.15 psu. Both implied changes would be, however, in conflict with the general global LGM-Holocene temperature rise and deep ocean freshening due to the decay of the large glacial ice sheets [*Jung*, 1996; *Labeyrie et al.*, 1987; *Schrag et al.*, 1996]. Furthermore, NADW contributes 25% to CDW [see *Emery and Meincke*, 1986]. A temperature rise of 2°–2.5°C in NADW at Termination I [*Jung*, 1996; *Labeyrie et al.*, 1987] implies that CDW waters should have induced a $\delta^{18}O$ drop off Somalia larger than the global ice volume change rather than a smaller shift. Accordingly, one or both of the other contributing water masses must have cooled much more than inferred above to counterbalance the NADW change. A scenario, however, of cooling in CDW would be in conflict with recently published evidence for a temperature rise in the deep Pacific across Termination I [*Matsumoto and Lynch-Stieglitz*, 1999]. On the basis of these arguments we therefore reject in situ CDW changes as the explanation for the $\delta^{18}O$ record from the deep Arabian Sea during the early Holocene.

5.3. Hypothesis 2: RSW Advection/Property Changes During the Early Holocene

The major reason to consider the role of RSW in producing variations in $\delta^{18}O$ off Somalia is the potential that RSW properties may change quickly and significantly. In contrast to the relatively slow changes in the deep ocean, water masses formed in a semi-enclosed basin, such as the Red Sea, will be sensitive to changes in the evaporation-precipitation balance (E-P balance). Such changes will be carried quickly and relatively unaltered into the Arabian Sea.

Compared to today, the glacial water exchange between the Red and Arabian Seas was strongly diminished [*Rohling and Zachariasse*, 1996] owing to a shallower sill depth at the strait of Bab el Mandeb. The combination of strong evaporation and reduced water exchange resulted in generally higher salinities in the Red Sea during the last glacial period. Estimates of the glacial salinity reach up to 55 psu [*Hemleben et al.*, 1996; *Thunell et al.*, 1988]. On the basis of the $\delta^{18}O/S$ relation in the Red Sea of 0.3 and the glacial-Holocene difference in salinity of roughly 15 psu, glacial RSW may have been up to 4.5‰ heavier in $\delta^{18}O$ than today. Sedimentary sections with very low abundances of planktic foraminifera are viewed as records of such extreme glacial conditions in the Red Sea when salinity exceeded the maximum tolerance level of planktic foraminifers [see *Fenton et al.*, 2000]. In two cores, *Locke and Thunell* [1988] report that the youngest aplanktic zone extends into the Holocene up to ~7 kyr B.P. *Hemleben et al.* [1996] concluded that the establishment of full modern-type water exchange occurred during the early Holocene around 8 kyr B.P. Although both estimates deviate, they imply that formation of RSW with salinities significantly higher than those today may have continued well into the Holocene. A higher salinity probably resulted in a higher RSW contribution to the RSW mix off Somalia, the latter being denser than today. Any outflow of

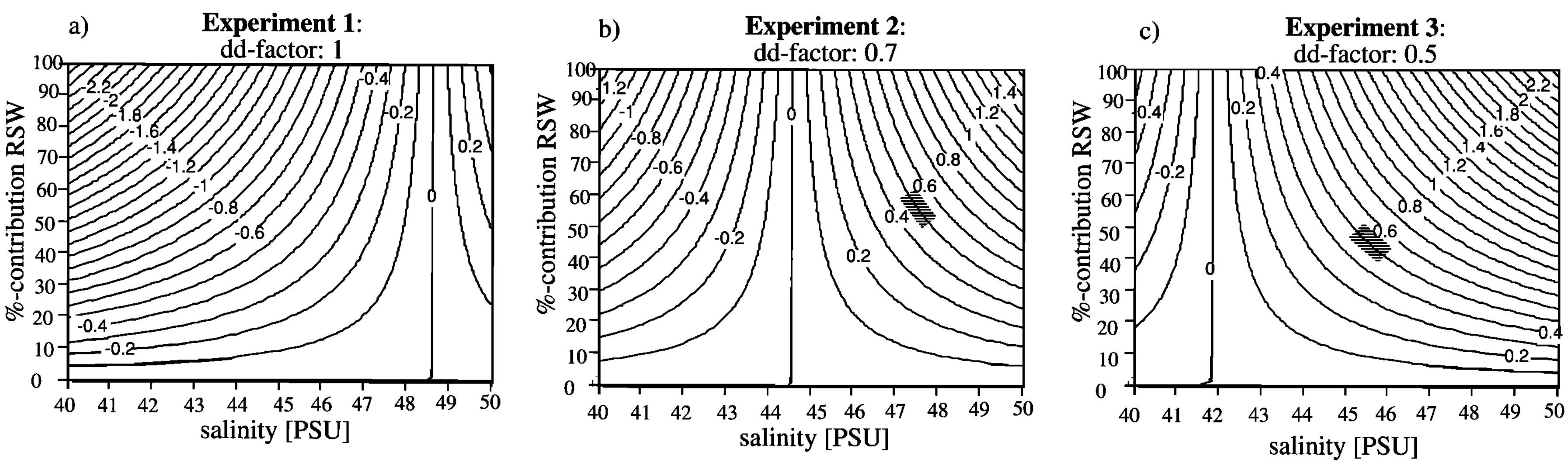

Figure 7. Graphic illustrations of three numerical experiments to assess the net effect on $\delta^{18}O$ values in *C. kullenbergi* using different double-diffusion values (contoured for $\delta^{18}O$). All experiments run assuming the modern RSW temperature of 22°C. (a) Temperature and salinity mix at equal rates (dd, double-diffusion ratio = 1). (b, c) Temperature mixes faster than salinity (dd = 0.7 and dd = 0.5, respectively). Compare Table 1.

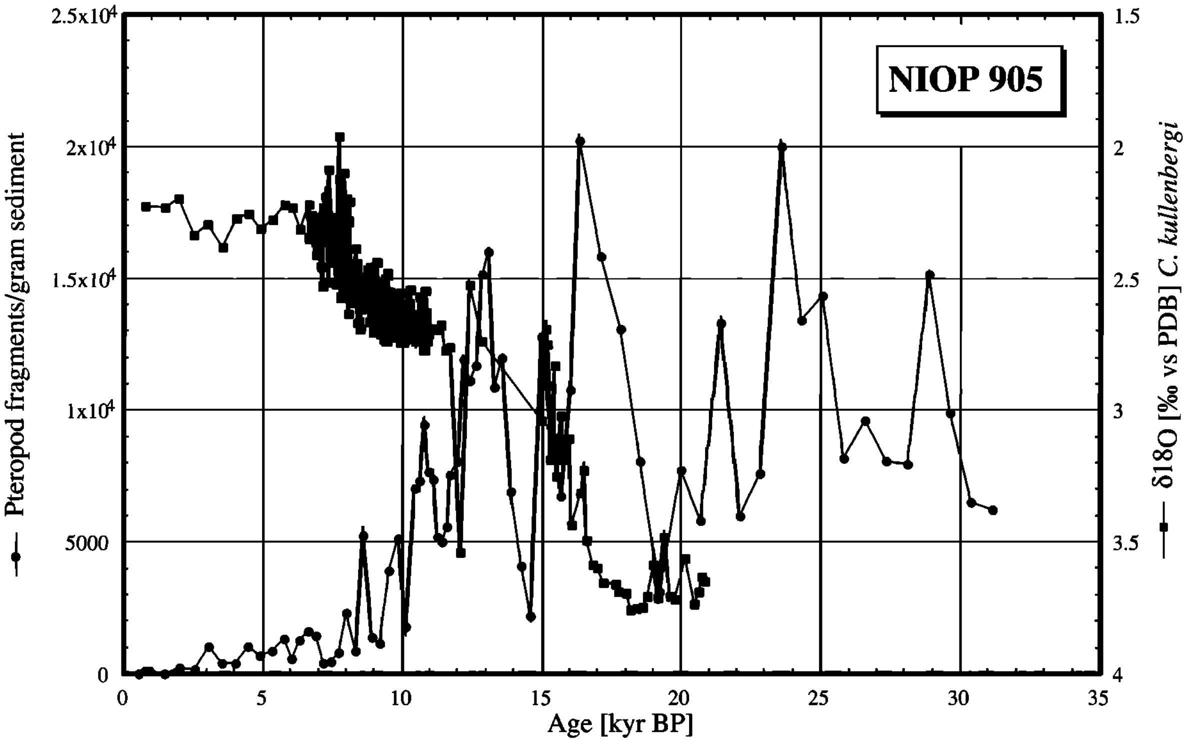

Figure 8. Comparison of the $\delta^{18}O$ record of *C. kullenbergi* (squares) with the abundance of pteropod fragments (circles) as indicators of variations in the preservation of $CaCO_3$. Data source for pteropod fragments [*Ivanova*, 2000].

RSW would therefore sink deeper and hence potentially also affect the bottom waters at site 905.

In this context the question arises of how the water exchange between the Red and Arabian Seas was re-established. Exchange was restricted prior to 12–13 kyr B.P [*Hemleben et al.*, 1996] owing to the shoaling of the sill depth at the Strait of Bab el Mandeb, indicating that RSW may have "preserved" glacial-type hypersaline properties up to that time. Hence it seems likely that the first RSW outflow into the Arabian Sea had glacial-type *T-S* properties. Accordingly, the $\delta^{18}O$ maximum spike during the Younger Dryas may document the first signs of such a hypersaline RSW outflow. This conclusion is in line with our hypothesis that the $\delta^{18}O$ variations in the deep Arabian Sea are due to an early Holocene RSW with different *T-S* properties than today.

5.4. Constraints on RSW Mixing

In the following section we assess the *T-S* properties of pure RSW during the early Holocene and the possible change in contribution to the mixed RSW found off Somalia. We use simple modeling experiments that combine the necessary *T-S*-related density enhancement and its effect on $\delta^{18}O$.

The modern density difference between mixed RSW and bottom water off Somalia is ~0.4 density units (Table 1). In order to enhance the RSW density sufficiently to allow for a deepening of 800 m (see Figure 2), *T* and *S* changes of roughly −2°C or 1 psu would be required. Unfortunately, temperature reconstructions for the Red Sea are very limited. A rough estimate of a 4°C colder northern Red Sea during the LGM is available [*Reiss et al.*, 1980]. For simplicity and in the absence of good constraints, we therefore assume that the Holocene RSW temperature remained constant at 22°C. In fact, the insolation peak during the early Holocene may have induced a temperature slightly higher than today. This would have resulted in higher evaporation and accordingly higher $\delta^{18}O$ values of surface waters, counteracting the decrease in $\delta^{18}O$ induced by the higher temperature. Even if we assume that the RSW temperature was slightly higher than today, our general conclusions below would not be affected.

A series of numerical mixing experiments were conducted to assess the net $\delta^{18}O$ change in benthic foraminifera in salinity versus percentage of RSW contribution fields (Figure 7). On the basis of the weak correlation of $\delta^{18}O$ and *S*, we assume that varying GOA contribution cannot induce high-amplitude variations in $\delta^{18}O$ in RSW (see above). For our numerical experiments we therefore assessed a simple mixing between RSW and CDW. In the first mixing experiment we assume that temperature and salinity differences between RSW and CDW are mixed at equal rates. In two further experiments, however, we take into account that with high *T-S* differences between deepwater masses, these parameters mix at different rates. This effect is due to double diffusion, where temperature is mixed 100 times faster than salinity (for review, see *Schmitt* [1994]). This process acts on molecular to centimeter scales. Its effect on the differential *T-S* mixing between large water masses is smaller, and its role in ocean circulation has been explored based on, for example, global ocean simulations. Reasonable estimates of the effect of double diffusion on differential *T-S* mixture rates in those simulations range from 0.5 (experiment 3) to 0.7 (experiment 2) [*Gargett and Ferron*, 1996; *Gargett and Holloway*, 1992; *Merryfield et al.*, 1999; *Zhang et al.*, 1998, and references therein]. Figures 7b and 7c summarize these calculations.

For our experiments we model the long-term decrease in $\delta^{18}O$ (see marker line in Figure 4) between the early Holocene ($\delta^{18}O$ values around 2.7‰) and the Holocene subsequent to ~6.5 kyr B.P. ($\delta^{18}O$ values of 2.2–2.3‰). We calculated this change of 0.5‰ using the modern $\delta^{18}O$ values of CDW (−0.5‰), RSW off Somalia (0.19‰), and RSW Bab el Mandeb (1.4‰) (G. Ganssen, unpublished data, 1992–1993). Hence the $\delta^{18}O$ patterns shown in Figures 7b and 7c indicate the relative change in $\delta^{18}O$ with respect to the modern values due to either/or variations in RSW salinity and in percentage of RSW contribution, respectively.

In our experiment 1, where *T* and *S* differences between RSW and CDW mix at equal rates, it is not possible to induce a shift of +0.5‰ in $\delta^{18}O$ within a salinity range of 40–50 psu. Higher salinity values would allow such a change, but this would imply salinity values in RSW even higher than during glacials, a highly unlikely scenario for the early Holocene. Mixing experiments 2 and 3, however, result in more reasonable changes in the contribution of pure RSW. A +0.5‰ shift is formed by varying the contribution of RSW at moderate salinity by up to 100%, to lower contributions with extremely high salinity values of up to 50 psu. We anticipate that neither of the extreme scenarios would occur in nature. It is most likely that a reasonable RSW contribution/salinity vector in Figures 7b and 7c combines the needed 0.5‰ change in $\delta^{18}O$ with minimum salinity and percent of contribution values (the hatched fields in Figures 7b and 7c indicate the most likely combinations). In experiment 2 the required RSW contribution would be 50–60% along with an initial RSW salinity around 47–48 psu. A lower RSW contribution of 40–50% and a salinity between 45 and 46 psu are inferred from experiment 3 (see also Table 1). Given the hypersaline conditions at the formation site of RSW, both scenarios appear plausible.

As pointed out above, the density difference between the modern RSW off Somalia and the bottom waters is 0.4 density units. The densities in RSW off Somalia implied by experiments 2 and 3 are 31.8 and 31.0 density units, respectively (Table 1). Both estimates significantly exceed the density of modern bottom water at site 905 of ~27.8 density units. Given this predicted density/salinity excess, we infer that between 10 and ~6.5 kyr B.P, RSW settled roughly 800 m deeper than today and is recorded in the benthic O isotope record. We interpret the general decrease in $\delta^{18}O$ to reflect a long-term salinity reduction in RSW toward that associated with the modern-day water exchange patterns between the Red and Arabian Seas around ~6.5 kyr B.P. We attribute the differences between estimates of the timing of the full establishment of modern-type water exchange between the Red and Arabian Seas (discussed in section 5.3 [*Fenton et al.*, 2000; *Hemleben et al.*, 1996; *Locke and Thunell*, 1988] and this paper) to the quality of the age control and the time resolution achieved. On the basis of the high sedimentation rate and our continuous sampling of core 905, we favor our estimate of ~6.5 kyr B.P. Hence we further conclude that around that time the density/salinity probably passed a minimum threshold value, resulting in a shoaling of RSW off Somalia and no further effect at site 905.

Additional evidence for a deeper RSW is inferred from the abundance record of pteropod fragments in core 905. The preservation of aragonite covaries with the intensity of the monsoon winds. The stronger the SW monsoon, the stronger the induced upwelling. The latter, in turn, results in a higher productivity that leads to higher oxygen consumption at the seafloor, which causes a lower preservation rate of $CaCO_3$. A number of studies [e.g., *Ivanova*, 2000; *Sirocko et al.*, 1993] concluded that the SW monsoon peaked during the early Holocene around 9 kyr B.P. Figure 8 indicates, however, that the abundance of pteropod fragments reached a minimum around 6.5–7 kyr B.P. This is significantly later than the peak in monsoon strength. RSW, in contrast, has a high $CaCO_3$ preservation potential [see *Almogi-Labin et al.*, 1986]. Accordingly, RSW settling 800 m deeper than today may have offset any enhanced carbonate dissolution induced by the early Holocene monsoon maximum. The modern-day situation of low carbonate preservation was fully established subsequent to ~6.5 kyr B.P., a period when the RSW stopped influencing the bottom waters at site 905. Interestingly, this time coincides with the onset of maximum strength of the oxygen minimum zone (OMZ) in the eastern Arabian Sea [*Reichart et al.*, 1998]. These coincidences suggest that RSW admixture may have induced a delay of the monsoon-related buildup of a maximal OMZ.

Our favored hypothesis for a generally enhanced effect of RSW raises the question of the origin of the short-term variations, in particular, between 8.1 and ~6.5 kyr B.P. Two scenarios appear possible. In a first scenario, RSW-induced higher $\delta^{18}O$ values are modulated by varying contributions of CDW. In a second scenario the $\delta^{18}O$ variations reflect in situ changes of RSW. In the first scenario the massive minimum spikes between 8.1 and ~6.5 kyr B.P. would result from a strong CDW influx, and the maxima would reflect a strong RSW influx. As discussed above, it is very difficult to construct scenarios for changes in *T-S* properties of CDW that result in lower $\delta^{18}O$ values than the present (see above). Furthermore, scenarios (see above) indicating short-term temperature or salinity changes in CDW appear physically unlikely. Hence we conclude that the observed $\delta^{18}O$ variations most likely document changes within RSW.

5.5. Implications for the Regional Evaporation-Precipitation Balance

The observed $\delta^{18}O$ minimum spikes in the benthic foraminifera, in particular, the high-amplitude variations between 8 and ~6.5 kyr B.P., cannot result from simple *T-S* variations within RSW, since they would indicate either a significant temperature rise of 2.2°–2.3°C or a salinity drop of up to 2 psu. Both types of changes would, however, reduce the density and probably result in a shoaling of RSW. Accordingly, such variations would not be recorded in the benthic $\delta^{18}O$ record at site 905. To explain this apparent contradiction, we propose that the Red Sea region was affected by variations in the *E-P* balance with the minimum spikes caused by enhanced precipitation of rainwater enriched in ^{16}O. This water can be subsequently carried into the Arabian Sea. Variations of this type have been shown to affect the $\delta^{18}O$ in surface waters in the modern Red Sea [cf. *Ganssen and Kroon*, 1991].

Assessing the potential origin of enhanced precipitation is difficult because the Red Sea appears to be influenced by two rainfall regimes. The southern Red Sea is monsoonal dominated whereas the northern part is affected by the Mediterranean regime [*deMenocal and Rind*, 1996; *Hsü and Walace*, 1976]. We anticipate that in the early Holocene, a period when the SW monsoon strength peaked, its impact on the Red Sea would be larger than today. The generally higher precipitation during the early Holocene recorded in NE African lakes [*Gasse et al.*, 1989, 1990; *Lézine*, 1989; *Ritchie et al.*, 1985; *Van Campo et al.*, 1982] (for a recent summary, see *Gasse* [2000]) provides precipitation peaks that may have been entrained into RSW and advected to the Arabian Sea. There is additional evidence of variations in regional moisture pathways from the link between *E-P* balance changes in the Red Sea region and the formation of the Mediterranean sapropel 1 (S1). We propose that the deep convection in the eastern Mediterranean was suppressed by enhanced input of low-density runoff via the Nile River *Rossignol-Strick et al.*, 1982]. Today the Nile runoff is controlled by the African monsoon/equatorial rainfall and not by the Asian monsoon. Under the most extreme conditions the Asian monsoon would have had a regional impact on the *E-P* balance in NE Africa, so that it may have governed the entire Red Sea *E-P* balance rather than just the southern part as today. A major problem in coupling Mediterranean and Arabian Seas records is insufficient age control. Well-dated high-resolution studies from the Arabian Sea, from the eastern Mediterranean, and from continental Africa/Arabia are required to better compare the chronology of climate events and thereby improve our understanding of the decadal-scale variations. Only with such data can we understand the principle forcing functions that control the regional climate.

6. Concluding Remarks

On the basis of early Holocene variations in $\delta^{18}O$ of benthic foraminifera off Somalia, we deduce that property variations in RSW were the most likely explanation for the observed pattern. This conclusion implies that because of enhanced salinity/density, RSW settled 800 m deeper in the water column than today. The decadal-scale variations in $\delta^{18}O$ reflect variations in the evaporation-precipitation balance.

Using this case study as an example, we speculate that similar processes may also have occurred at other inlets of marginal seas (e.g., the inflow of Mediterranean overflow water to the Atlantic Ocean). Additional studies with the highest possible time resolution are required to further our understanding of the processes that are linked to the advection of water from marginal seas like the Red Sea. This approach would help to assess the impact of a deeper settling high-salinity water mass, and the according excess salt at depth, on the large-scale deep ocean circulation: a point that will lead to more precise reconstructions of the short-term dynamics in the deep ocean.

Acknowledgments. We thank E. Bard for a thorough review of an earlier version of this paper. This study is part of the EU-TMR network ERBFMRXCT960046 "The marine record of continental tectonics and erosion" and is NSG publication 2001-6901. We also thank E. Rohling and two anonymous reviewers for their constructive criticisms that greatly improved the quality of the manuscript.

References

Almogi-Labin, A., L. Luz, and J.-C. Duplessy, Quaternary paleo-oceanography pteropod preservation and stable-isotope record of the Red Sea, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *57*, 195–211, 1986.

Almogi-Labin, A., C. Hemleben, D. Meischner, and H. Erlenkeuser, Response of Red Sea deep-water agglutinated foraminifera to water-mass changes during the late Quaternary, *Mar. Micropaleontol.*, *28*(3/4), 283–297, 1996.

Cox, R. A., M. J. McCartney, and F. Culkin, The specific gravity/salinity/temperature relationship in natural seawater, *Deep Sea Res.*, *17*, 679–689, 1970.

Craig, H., and L. I. Gordon, Deuterium and oxygen-18 variations in the ocean and the marine atmosphere, in *Stable Isotopes in Oceanographic Studies and Paleotemperatures: Third SPOLETO Conference on Nuclear Geology*, edited by E. Tongiorgi, pp. 9–130, Lab. di Geol. Nucl., Cons. Naz. delle Ric., Pisa, Italy, 1965.

deMenocal, P. B., and D. Rind, Sensitivity of subtropical African and Asian climate to prescribed boundary conditions changes: Model implications for the Plio-Pleistocene evolution of low-latitude climate, in *The Limnology, Climatology and Paleoclimatology of the East African Lakes*, edited by T. C. Johnson and E. O. Odada, pp. 57–77, Gordon and Breach, Newark, N. J., 1996.

Emery, W. J., and J. Meincke, Global water masses: Summary and review, *Oceanol. Acta*, *9*(4), 383–391, 1986.

Fairbanks, R. G., A 17,000-year glacio-eustatic sea level record: Influence of glacial melting rates on the Younger Dryas event and deep-ocean circulation, *Nature*, *342*, 637–642, 1989.

Fenton, M., S. Geiselhart, E. J. Rohling, and C. Hemleben, Aplanktonic zones in the Red Sea, *Mar Micropaleontol.*, *40*, 277–294, 2000.

Ganssen, G., and D. Kroon, Evidence for Red Sea surface circulation from oxygen isotopes of modern surface waters and planktonic foraminiferal tests, *Paleoceanography*, *6*(1), 73–82, 1991.

Gargett, A. E., and B. Ferron, The effects of differential vertical diffusion of T and S in a box model of thermohaline circulation, *J. Mar. Res.*, *54*, 827–866, 1996.

Gargett, A. E., and G. Holloway, Sensitivity of the GFDL ocean model to different diffusivities for heat and salt, *J Phys. Oceanogr.*, *22*, 1158–1177, 1992.

Gasse, F., Hydrological changes in the African tropics since the Last Glacial Maximum, *Quat. Sci. Rev.*, *19*, 189–211, 2000.

Gasse, F., V. Lédée, M. Massault, and J.-C. Fontes, Water-level fluctuations of Lake Tanganyika in phase with oceanic changes during the last glaciation and deglaciation, *Nature*, *342*, 57–59, 1989.

Gasse, F., R. Téhet, A. Durand, E. Gibert, and J.-C. Fontes, The arid-humid transition in the Sahara and the Sahel during the last deglaciation, *Nature*, *346*, 141–146, 1990.

Gründlingh, M. L., Occurrence of Red Sea Water in the southwestern Indian Ocean, 1981, *Bull. Am. Meteorol. Soc.*, *15*, 207–212, 1985.

Hemleben, C., D. Meischner, R. Zahn, A. Almogi-Labin, H. Erlenkeuser, and B. Hiller, Three hundred eighty thousand year long stable isotope and faunal records from the Red Sea: Influence of global sea level change on hydrography, *Paleoceanography*, *11*(2), 147–156, 1996.

Hsü, C.-P. F., and J. M. Walace, The global distribution of the annual and semiannual cycles in precipitation, *Mon. Weather Rev.*, *104*(9), 1093–1101, 1976.

Ivanova, E., Late Quaternary monsoon history and paleoproductivity of the western Arabian Sea, Ph.D. thesis, 172 pp., Free University, Amsterdam, Netherlands, 2000.

Jung, S. J. A., Wassermassenaustausch zwischen NE-Atlantik und Nordmeer während der letzten 300 000/80 000 Jahre im Abbild stabiler O- und C-Isotope, Christian-Albrechts-Univ., Kiel, Germany, 1996.

Labeyrie, L. D., J.-C. Duplessy, and P. L. Blanc, Variations in mode of formation and temperature of oceanic deep waters over the past 125,000 years, *Nature*, *327*, 477–483, 1987.

Lézine, A.-M., Late Quaternary vegetation and climate of the Sahel, *Quat. Res.*, *32*(3), 317–334, 1989.

Locke, S. M., and R. C. Thunell, Paleoceanographic record of the last glacial/ interglacial cycle in the Red Sea and Gulf of Aden, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *64*, 163–187, 1988.

Maillard, C., and G. Soliman, Hydrography in the Red Sea and exchanges with the Indian Ocean in summer, *Oceanol. Acta*, *9*, 249–269, 1986.

Matsumoto, K., and J. Lynch-Stieglitz, Similar glacial and Holocene deep water circulation inferred from southeast Pacific benthic foraminiferal carbon isotope composition, *Paleoceanography*, *14*(2), 149–163, 1999.

Merryfield, W. J., G. Holloway, and A. E. Gargett, A global ocean model with double diffusive mixing, *J. Phys. Oceanogr.*, *29*, 1124–1142, 1999.

Niitsuma, N., T. Oba, and M. Okada, Oxygen and carbon isotope stratigraphy at Site 723, Oman margin, *Proc. Ocean Drill. Program, Sci. Results*, *117*, 321–341, 1991.

O'Neil, J. R., N. C., and K. M. T, Oxygen isotope fractionation in divalent metal carbonates, *J. Chem. Phys.*, *51*, 5547–5558, 1969.

Prasanna Kumar, S., and T. G. Prasad, Formation and spreading of Arabian Sea high-salinity water mass, *J. Geophys. Res.*, *104*(C1), 1455–1464, 1999.

Quadfasel, D., and F. Schott, Water-mass distributions at intermediate layers off the Somali coast during the onset of the Southwest monsoon, 1979, *J. Phys Oceanogr.*, *12*, 1358–1372, 1982.

Reichart, G. J., L. J. Lourens, and W. J. Zachariasse, Temporal variability in the northern Arabian Sea oxygen minimum zone (OMZ) during the last 225,000 years, *Paleoceanography*, *13*(6), 607–621, 1998.

Reiss, Z., B. Luz, L. Almogi, A. Halicz, E. Winter, and M. Wolf, Late Quaternary paleoceanography of the Gulf of Aqaba (Elat), Red Sea, *Quat. Res.*, *14*(3), 294–308, 1980.

Rhein, M., L. Stramma, and O. Plähn, Tracer signals of the intermediate layer of the Arabian Sea, *Geophys. Res Lett.*, *24*(21), 2561–2564, 1997.

Ritchie, J. C., C. H. Eyles, and C. V. Haynes, Sediment and pollen evidence for an early to mid-Holocene humid period in the eastern Sahara, *Nature*, *314*, 352–355, 1985.

Rohling, E. J., Glacial conditions in the Red Sea, *Paleoceanography*, *9*, 653–660, 1994.

Rohling, E. J., and W. J. Zachariasse, Red Sea outflow during the Last Glacial Maximum, *Quat. Int.*, *31*, 77–83, 1996.

Rossignol-Strick, M., W. Nesteroff, P. Olive, and C. Vergnaud-Grazzini, After the deluge: Mediterranean stagnation and sapropel formation, *Nature*, *295*, 105–110, 1982.

Schmitt, R. W., Double diffusion in oceanography, *Annu. Rev. Fluid Mech.*, *26*, 255–285, 1994.

Schrag, D. P., G. Hampt, and D. W. Murray, Pore fluid constraints on the temperature and oxygen isotopic composition of the glacial ocean, *Science*, *272*, 1930–1932, 1996.

Shapiro, G. I., and S. L. Meschanov, Distribution and spreading of Red Sea Water and salt lens formation in the northwest Indian Ocean, *Deep Sea Res.*, *38*, 21–34, 1991.

Sirocko, F., M. Sarnthein, and H. Erlenkeuser, Century-scale events in monsoonal climate over the past 24,000 years, *Nature*, *364*, 322–324, 1993.

Thunell, R. C., S. M. Locke, and D. F. Williams, Glacio-eustatic sea-level control on Red Sea salinity, *Nature*, *334*, 601–604, 1988.

Trauth, M. H., M. Sarnthein, and M. Arnold, Bioturbational mixing depth and carbon flux at the sea floor, *Paleoceanography*, *12*(3), 517–526, 1997.

van Bennekom, J., M. Hiehle, J. van Ooyen, E. van Weerlee, and M. van Koutrik, CTD and hydrography, in *Tracing a Seasonal Upwelling*, edited by J. van Hinte, T. C. E. van Weering, and S. R. Troelstra, Leiden, Netherlands, 1995.

Van Campo, E., J. C. Duplessy, and M. Rossignol-Strick, Climatic conditions deduced from a 150 kyr oxygen isotope-pollen record from the Arabian Sea, *Nature*, *296*, 56–59, 1982.

Woelk, S., and D. Quadfasel, Renewal of deep water in the Red Sea during 1982–1987, *J. Geophys. Res.*, *101*(C1), 18,155–18,165, 1996.

Wyrtki, K., Physical oceanography of the Indian Ocean, *Contrib. Hawaii Inst. Geophys.*, *483*(3), 202–220, 1974.

You, Y., Intermediate water circulation and ventilation of the Indian Ocean derived from water-mass contributions, *J. Mar. Res.*, *56*, 1029–1067, 1998.

Zahn, R., and A. C. Mix, Benthic foraminiferal $\delta^{18}O$ in the ocean's temperature-salinity-chemistry field: Constraints on ice age thermohaline circulation, *Paleoceanography*, *6*(1), 1–20, 1991.

Zahn, R., and T. Pedersen, Late Pleistocene evolution of surface and mid-depth hydrography at the Oman margin: Planktonic and benthic isotope records at Site 724, *Proc. Ocean Drill. Program, Sci. Results*, *117*, 291–308, 1991.

Zhang, J., R. W. Schmitt, and R. X. Huang, Sensitivity of the GFDL modular ocean model to parameterization of double-diffusive processes, *J. Phys. Oceanogr.*, *28*, 589–605, 1998.

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(Received September 22, 2000;
revised May 16, 2001;
accepted July 2, 2001.)