Reconstructing sea surface temperature and salinity using $\delta^{18}$O and alkenone records

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The oxygen isotope ($\delta^{18}$O) composition of foraminiferal tests from deep-sea sediments is widely used as a paleoclimatic proxy, but it includes contributions from sea surface temperature, global ice volume and local salinity, which are difficult to separate. Recently a new technique for deriving paleotemperatures has been developed which is based on the abundance ratios of unsaturated alkenones in phytoplankton algalae. Here we use a combination of oxygen isotope and alkenone records in a deep-sea core from the juncture of the Arabian Sea and the Bay of Bengal to extract the salinity signal from the former record. Variations in salinity are related to the balance between evaporation and precipitation, and are thus a sensitive indicator of climate change. Our 170-kyr salinity record enables us to reconstruct changes in the Indian monsoon over this period, considerably extending earlier studies (which reached back to 18 kyr ago* ). Like these previous studies, we find that large variations in the monsoon occurred during the transition from the last glacial period to the present interglacial, but our results also provide a view of the monsoon throughout the last glacial and demonstrate the potential of this approach for reconstructing palaeosalinity.

Variations of sea surface salinity are strongly related to the evaporation–precipitation ($E - P$) balance* and thus represent a sensitive climate indicator. Due to extreme monsoonal atmospheric and oceanographic gradients (Fig. 1), the Northern Indian Ocean is characterized by two main surface water masses: high-salinity surface water (35–36.5‰) due to strong evaporation in the Arabian Sea and low-salinity water (31–34‰) in the Bay of Bengal caused by high precipitation and large river runoff, particularly during the SW monsoon*.

In order to reconstruct climate variability in the Equatorial Indian Ocean, Core MD900963 (05° 04’ N, 73° 53’ E; 2,450 m water depth; Fig. 1) was recovered on the eastern shoulder of the Maldives Islands (SEYAMA expedition, 1990; RV Marion Dufresne). Situated between the Arabian Sea and the Bay of Bengal, Core MD900963 is probably ideally located to monitor the changes of the monsoon during the last glacial cycle. A detailed $\delta^{18}$O stratigraphy was established on the surface dwelling planktonic foraminifer Globigerinoides ruber (white) for the 54-m long piston core (Fig. 2b). The timescale was obtained* by tuning the $\delta^{18}$O record to the SPECMAP record*.

We measured the alkenone unsaturation ratio $\gamma$ (refs 1, 2) on the uppermost 10 m of the core to determine sea surface temperature (SST) variations for the last 170,000 years using the temperature equation of Prahl et al.*. Alkenone measurements were carried out at intervals of 10 cm at the Department of Geoscience of the University of Bremen. Sample preparation and technical details are described elsewhere*. Preliminary observations of the coccoliths indicate that Emiliania huxleyi and other species of the family Gephyrocapsaceae are abundant in the first 10 m of Core MD900963 (L. Beaufort, personal communication).

SST variations during the last 170,000 years are rather small—between 25.5 and 28 °C (Fig. 2a). Low SST values of -25.5–26.5 °C are observed in oxygen isotope stages 4 and 3 with a minimum centred within stage 3. Interglacial stage 5 is characterized by high SST’s between 27 and 28 °C. For stage 6, the SST’s are ~1.5 °C higher than those observed in the stage 3 minimum.

Our data agree with SST reconstructions based on foraminifera assemblages from the tropical Indian Ocean* and in particular with the record published by Clemens et al.* The fact that these two SST time series (both obtained in the tropical Indian Ocean) agree with our results lends considerable weight to our interpretation of the monsoon variations.

**FIG. 1. Circulation patterns in the northern Indian Ocean**

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Indian Ocean) are similar gives more confidence in the reliability of the UK method. In particular, both records suggest that stage 6 was relatively warm and definitely not as cold as stage 3.

Changes in $\delta^{18}O$ recorded in planktonic foraminifera are controlled by global variations of seawater ($\delta^{18}O_{sw}$) due to the growth of continental ice sheets, and by local SST and $E-P$ changes. Sea surface salinity and local $\delta^{18}O_w$ of seawater are directly linked to the $E-P$ balance\(^{19,10}\) and a linear correlation was found between these parameters for most of the global ocean\(^{11}\). The glacial to interglacial change in the oxygen isotope ratio for foraminifera ($\delta^{18}O_{f}$) can be expressed as follows:\(^{21,23}\)

$$\Delta \delta^{18}O_f = a + b \Delta T + c(S - S^*)$$  \hspace{0.5cm} (1)

where $\Delta \delta^{18}O_f$ is the measured deviation between modern and past $\delta^{18}O_f$ values of planktonic foraminifera, $a$ represents global variations of sea water $\delta^{18}O_{sw}$ (variable through time due to changes of the global ice volume), $b$ is the slope of the $\delta^{18}O_f$ versus temperature relationship, $\Delta T$ is the local temperature variation, $c$ is the slope of the $\delta^{18}O_w$ versus salinity relationship, $S$ is the past local salinity and $S^*$ is the modern local salinity augmented with the global salinity change due to changes of the global ice volume. From equation (1), we obtained $S$ by the following equation:

$$S = S^* + (\Delta \delta^{18}O_f - a + b \Delta T)/c$$  \hspace{0.5cm} (2)

The salinity $S^*$ varies through time in response to the sea level variations (for example 120 m of sea level lowering during the Last Glacial Maximum (LGM) compared with a mean ocean depth of about 3,800 m). Consequently $S^*$ can be expressed as:

$$S^* = S_0^* + (SL \cdot 35/3000)$$  \hspace{0.5cm} (3)

where $S_0^*$ is the modern local salinity and 35% the modern mean ocean salinity, SL = sea level changes expressed in m below modern level (lowering being counted as positive).

Therefore, the past local salinity can be estimated by the following equation:

$$S = S^* + (SL \cdot 35/3000) + (\Delta \delta^{18}O_f - a + b \Delta T)/c$$  \hspace{0.5cm} (4)

In order to provide an index of local salinity variations due to the local changes of the $E-P$ balance, we also calculated $\Delta S$ values expressed as:

$$\Delta S = S - S^* = (\Delta \delta^{18}O_f - a + b \Delta T)/c$$  \hspace{0.5cm} (5)

The seasonal range of SST is very small (about 1–2 °C) at the location of core MD900963\(^{1,21}\) and sediment trap studies\(^{2,24}\) demonstrate that the $G. ruber$ bloom does not lag significantly the abrupt increases in wind speed which corresponds to the beginning of the monsoons. Consequently, it is reasonable to use the U37 SST record to obtain $\Delta T$, and to apply temperature correction to the $\delta^{18}O_f$ record ($b = 0.2$ for living $G. ruber$ in the Indian Ocean\(^{21}\)).

Labeary et al.\(^{20}\) and Vogelsang\(^{27}\) reconstructed a mean global seawater $\delta^{18}O_{sw}$ record for the past 200 kyr which is only affected by global ice volume changes. We normalized this seawater $\delta^{18}O_{sw}$ record to the global $\delta^{18}O$ value of 1.2‰, which has been established for the last deglaciation\(^{28}\). To estimate local salinity variations, changes in the global seawater $\delta^{18}O_{sw}$ (term $a$ in equation (2)) have to be removed from the SST corrected $\delta^{18}O_f$ record of MD900963. Since a point-by-point subtraction of two independent isotopic records is very sensitive to phase differences and thus to timescale details, we first correlated the global seawater $\delta^{18}O_{sw}$ reconstruction to the timescale of the temperature corrected isotope record of Core MD900963 before subtracting it (Fig. 2c, we used the Analyserie software developed by D. Paillard & L. D. Labeary, CFR Gif-sur-Yvette).

In order to reconstruct the $\delta^{18}O_w$ versus salinity relationship (term $c$ in equation (2)), we used 52 seawater $\delta^{18}O_w$ and salinity measurements between 20° S and 21° N from the Arabian Sea (J. C. Duplessy, personal communication). Because no seawater $\delta^{18}O_w$ measurements are available for the Bay of Bengal, a $\delta^{18}O_w$/salinity relationship was established by using 22 $\delta^{18}O_w$ $G. ruber$ core top measurements\(^{1}\). To represent both data sets in the same graph (Fig. 3) we converted the $\delta^{18}O_f$ core top measurements into seawater $\delta^{18}O_w$ values by using the temperature equation of Duplessy et al.\(^{29}\) established for living $G. ruber$ in the Indian Ocean.

The $\delta^{18}O$/salinity slope of 0.37 of the composite data set is probably the best value of $c$ to be used in equations (4) and (5) to quantify the salinity variations at the location of Core MD900963. In order to illustrate the uncertainty associated with the conversion of $\delta^{18}O$ into surface salinity, we also used the individual slopes obtained for the two basins Arabian Sea and Bay of Bengal (dashed lines on Fig. 4a and 4b). Taking into account the standard reproducibility ($\pm 1$ sigma) on the SST ($\pm 0.3$ °C; ref. 17), on $\delta^{18}O_{sw}$ ($\pm 0.07‰$; ref. 29) and on $\delta^{18}O_w$ ($\pm 0.05‰$; ref. 30), it is possible to estimate an average analytical error of about $\pm 0.3‰$ for the reconstructed $\Delta S$ salinity values. This indicates that only the major trends of the $S$ and $\Delta S$ records should be interpreted in term of palaeoceanographic variations.

Figure 4a shows the downcore $\Delta S$ changes (equation (5)) whereas Fig. 4b presents the corresponding $S$ values which take into account the global salinity variation due to sea level changes (equation 4). In the following paragraph we will mainly discuss the $\Delta S$ changes which are only due to local variations of the $E-P$ balance.

Negative $\Delta S$ values of about $-1$ to $-0.5$‰ are observed between 125 and 115 kyr BP (stage 5a) and between 9 and 6 kyr
Between 160 and 140 kyr (stages 6.4 and 6.3) and between 75 and 25 kyr (stages 5.0 to 3.1), $\Delta S$ is consistently positive with values ranging between 0.5 to 1%. Three short periods of elevated $\Delta S$ are clearly present in the record with values higher than 1%: 140 to 125 kyr (stages 6.2 to 6.0), 90 to 80 kyr (stage 5.2) and 20 to 15 kyr (stage 2).

Negative $\Delta S$ during stage 5.5 and the early Holocene indicate that $E - P$ was lower than today. Because the precipitation over India and the Indian Ocean is related to the SW monsoon strength, the $\Delta S$ index suggests that the SW monsoon was stronger during these time intervals.

The higher $\Delta S$ values observed during glacial stages 6 and 2 and substage 5.2 on the Core site MD900963 were probably caused by increased evaporation and/or decreased precipitation. Consequently, it can be suggested that the dry NE monsoon was the dominating seasonal feature during these periods and/or that the humid SW monsoon was weakened. The generally high values obtained for the $\Delta S$ index (with the exception of stage 5.5 and the Holocene) suggest that in this part of the Indian Ocean the $E - P$ budget was to a large extent determined by the conditions of today during most of the last glacial cycle.

Based on pollen and foraminifera data, oxygen and carbon isotope composition of planktonic foraminifera and organic matter content, similar climatic conditions were suggested for the LGM. During the LGM the decrease in precipitation led to considerably higher salinities in the Bay of Bengal due to the lower riverine freshwater input. Our work extends these studies, and also shows that the Indian monsoon underwent large variations concomitant with the last glacial-interglacial transition. Further SST determinations and salinity reconstructions are still necessary for the two basins (Arabian Sea and Bay of Bengal) both in order to calculate and model the changes of $E$ and $P$ independently and to understand in detail the climatology of the northern Indian Ocean during glacial periods. It will also be important to study the time variations of the marine productivity in the monsoonal upwelling zones off Somalia and Arabia, and their relationships with the salinity index reconstructed in this paper.

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**FIG. 3** $\delta^{18}O$-salinity relationships for the Arabian Sea and the Bay of Bengal. Regression lines were calculated for these two basins and for the composite data set as follows: Arabian Sea, $y = -9.24 + 0.45x$ (R = 0.87); composite data set, $y = -12.68 + 0.37x$ (R = 0.95). For this graph we used 52 seawater $\delta^{18}O$ and salinity measurements between 20° S and 21° N from the Arabian Sea (J. C. Duplessy, unpublished data). For the Bay of Bengal, a $\delta^{18}O$-salinity relationship was established by using 22 $\delta^{18}O$ G. ruber core top measurements. To represent both data sets in the same graph the $\delta^{18}O$ core top measurements were converted into seawater $\delta^{18}O$ values by using the temperature equation of Duplessy et al. established for living G. ruber in the Indian Ocean, and by taking into account a $\delta^{18}O$ enrichment of about 0.85% due to late calcification.

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**FIG. 4** a, Salinity variations $\Delta S$ for Core site MD900963. To eliminate high frequency variations due to the subtraction of two independent isotope records, we applied a least square filter to the salinity curves. $\Delta S$ changes are calculated for the different slopes (C) of the $\delta^{18}O$ versus salinity relationships of the Arabian Sea (upper stippled line, $-0.28$), the Bay of Bengal (lower stippled line, $-0.45$) and the composite data (solid line, $-0.37$). $\Delta S$ values are directly related to local variations of the $E - P$ balance. b, Local salinity variations $S$ for Core site MD900963, taking into account salinity changes due to changes of global ice volume. The modern salinity of Core site MD900963 is taken as 35% (ref. 21).

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