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# Spatiotemporal variations of the $\delta^{18}$O–salinity relation in the northern Indian Ocean

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ABSTRACT

A new data set of oxygen isotopic composition ($\delta^{18}$O) and salinity ($S$) of surface and sub-surface waters of the northern Indian Ocean, collected during the period 1987–2009, is presented. While the results are consistent with positive $P-E$ (excess of precipitation over evaporation) over the Bay of Bengal and negative $P-E$ over the eastern Arabian Sea, a significant spatiotemporal variability in the slope (also intercept) of the $\delta^{18}$O–$S$ relation is observed in the Bay; the temporal variability is difficult to discern in the Arabian Sea. The slope and intercept are positively and negatively correlated, respectively, with the annual rainfall over India, a rough measure of river runoff. Both the slope and intercept appear to be sensitive to rainfall; the slope (intercept) is higher (lower) during years of stronger monsoon. The observed variability in the $\delta^{18}$O–$S$ relation implies that caution needs to be exercised in paleosalinity estimations, especially from the Bay of Bengal, based on $\delta^{18}$O of marine organisms.



## 1. Introduction

Oxygen isotopic composition ($\delta^{18}$O) and salinity ($S$) of the surface ocean are useful in the study of mixing of oceanic water with runoff, sea ice melting, brine release caused by sea ice formation, and even large-scale ocean mixing (Ferronsky and Brezgunov, 1989; Bigg and Rohling, 2000). They are known to covary linearly in the surface ocean in the mid- and high-latitudes, and to a lesser degree, in the tropics (Craig and Gordon, 1965). This linear relation is exploited in paleoceanographic reconstructions: $\delta^{18}$O of foraminifera from sediment cores is often used to infer past changes in $\delta^{18}$O of ocean surface (e.g., Duplessy et al., 1981), after correcting for temperature effects, either using alkenones (e.g., Rostek et al., 1993) or Mg/Ca ratios (Rashid et al., 2007). The tacit assumption here is that the spatial $\delta^{18}$O–$S$ relation observed today is equally valid for the past, the validity of which has been questioned by Rohling and Bigg (1998). Evaporation ($E$), precipitation ($P$), continental runoff ($R$), upwelling/advection and diffusion are the major physical processes that determine the $\delta^{18}$O–$S$ relation in tropical oceanic surface waters; $\delta^{18}$O and salinity are known to (i) increase with evaporation, (ii) decrease with higher precipitation and continental runoff and (iii) vary by mixing due to advection and diffusion. Thus the slope and intercept of the $\delta^{18}$O–$S$ relation might vary seasonally and geographically, as the controlling processes are season-, climate- and location-dependent (Benway and Mix, 2004). The northern Indian Ocean is an ideal region for investigating such variations as it receives monsoon runoff through rivers of Himalayan and peninsular origin.

Published $\delta^{18}$O data pertaining to the northern Indian Ocean are too limited to enable detection of seasonal and interannual variations in the $\delta^{18}$O–$S$ relation, if there are any. Paul et al. (1999) inferred from 15 GEOSECS data pairs that the $\delta^{18}$O–$S$ relation for the tropical Indian Ocean as a whole was $\delta^{18}O=0.18S–5.74$. LeGrande and Schmidt (2006) estimated a slope of 0.16 for the Indian Ocean. A compilation of the available data in 2001 (e.g., Table 1 of Delaygue et al., 2001) showed 112 pairs of $\delta^{18}$O–$S$ data from the Arabian Sea (including 7 from GEOSECS), some of which suffered from systematic shifts and were not used by them for analysis. Delaygue et al. (2001) added a set of 78 additional data pairs, mostly from the open Bay of Bengal, with only ~6 samples representing the coastal Bay. Most of their Arabian Sea samples derived from the Gulf of Aden, which prevented them from deciding whether the observed $\delta^{18}$O–$S$ relation ($\delta^{18}O=0.27S–9.2$) applied to the Arabian Sea as a whole, on the basis of latitudinal similarity, or if the Gulf of Aden acted only as an end member, with the Arabian Sea having quite a different $\delta^{18}$O–$S$ relation. They also claimed that a simple box model that ignored mixing between boxes could successfully explain their observations. Somayajulu et al. (2002) reported 5 data points from the northern Bay of Bengal, which showed a relation: $\delta^{18}O=0.23S–7.6$. Here we report 152 new data pairs for surface water (0–2 m depth), 43 for sub-surface (50 m) and 16 for deep (200–700 m) samples collected during eight different cruises conducted in the northern Indian Ocean during 1987–2009 and discuss the seasonal and spatial variations of the $\delta^{18}$O–$S$ relation and its implication for

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paleoceanographic studies. We also examine the validity of some of the conclusions of Delaygue et al. (2001) in the light of new data.

The climatology, hydrography and circulation of the northern Indian Ocean have been summarized by many authors (Schott and McCreary, 2001; Delaygue et al., 2001; Rao and Sivakumar, 2003; Jyothibabu et al., 2010). In the following we give a very brief account of the monsoon circulation and hydrography of the northern Indian Ocean and discuss earlier models pertaining to the $\delta^{18}O$–*S* relation.

Unlike other oceans, the northern Indian Ocean, being bounded by the Asian land mass to its north, suffers a strong seasonal reversal in the wind direction, well known as the monsoon circulation. During June–September (the southwest or summer monsoon), strong winds blow from ocean towards land resulting in intense upwelling in the northwestern Arabian Sea. Evaporation over the Arabian Sea (275 cm $yr^{-1}$) is estimated to be higher than that over the Bay of Bengal (250 cm $yr^{-1}$; Prasad 1997). Rainfall over the Indian subcontinent is derived from the northward movement of the Inter-Tropical Convergence Zone (ITCZ) during June–September and its withdrawal during October–November (Gadgil, 2003). The rainfall varies seasonally, regionally and annually over the region. Most Indian plains receive a major fraction (~80%) of the annual rainfall during summer. The highest rainfall occurs along the west coast and northeastern regions of the Indian peninsula. During December–February (the northeast or winter monsoon), cool and dry air from the Himalayas enhances evaporation in the northeastern Arabian Sea causing convective mixing (Madhupratap et al., 1996), while there is little rainfall over most of India; southeastern India gets its major share of rain during this period.

A limited set of $\delta^{18}O$ values of precipitation over the Indian subcontinent is available from the web site for Global Network on Isotopes in Precipitation (IAEA/WMO, 2006). Araguas-Araguas et al. (1998) have shown that there is an 'amount effect' in the monsoon precipitation, i.e., higher amount of rainfall is more depleted in $^{18}O$ ($\sim -1.5$‰ per 100 mm of monthly rainfall). $\delta^{18}O$ values of direct precipitation over the Bay of Bengal and the Arabian Sea are $\sim -2$‰ and $\sim -1$‰, respectively (Bowen and Revenaugh, 2003). Reported $\delta^{18}O$ values of the runoff to the Bay of Bengal from the Ganga river system are similar (~ –6‰) (Ramesh and Sarin, 1992; Somayajulu et al., 2002). Additionally, the Bay receives runoff from the Brahmaputra, Irrawady, Mahanadi, Godavari and Krishna rivers (Lambs et al., 2005). All this adds up to $1.6 \times 10^{12}$ $m^3$ $yr^{-1}$, much higher relative to the runoff of $0.3 \times 10^{12}$ $m^3$ $yr^{-1}$ to the Arabian Sea from rivers Narmada, Tapti, Nethravati and Indus (Subramanian, 1993). The mean $\delta^{18}O$ value of the runoff is likely to be as low as $\sim -9$‰ in some years when the summer rains are well above normal (Delaygue et al., 2001). The major runoff to the Bay derives from the rivers of Himalayan origin (Ramesh and Sarin, 1992; Karim and Veizer, 2002; Lambs et al., 2005), and increased snow-melt could also lower the mean $\delta^{18}O$ of runoff during pre-monsoon (April–May). The mean $\delta^{18}O$ value of the runoff to the Arabian Sea from the Indus is $\sim -11$‰ (Mook, 1982), while that from the Narmada and Tapti and smaller westbound rivers originating in the Western Ghats is likely to be higher (i.e., more enriched in $^{18}O$); however, in years of excess rain/floods or increased Himalayan snow-melt in spring, the mean $\delta^{18}O$ of runoff into the Arabian Sea could be significantly lower. Time series data on $\delta^{18}O$ of runoff are too sparse to refine these assertions.

The Arabian Sea and the Bay of Bengal, located at similar latitudes (Fig. 1), significantly differ in some aspects: e.g., the former loses a net ~1 m $yr^{-1}$ freshwater through evaporation, while the latter receives a net flux of ~2 m $yr^{-1}$ by intense precipitation (Prasad, 1997). River discharge and overhead precipitation together exceed evaporation in the Bay of Bengal. This flux imbalance causes the export (advection) of less saline water to the Arabian Sea (Rao and Sivakumar, 2003), e.g., the East India Coastal Current (EICC) transports low salinity water from the Bay to the southeastern Arabian Sea during winter. The West India Coastal Current (WICC) carries this water northward (Wyrtki, 1973; Han and McCreary, 2001; Schott and McCreary, 2001) and decreases the surface salinity of the eastern Arabian Sea. In contrast, during summer, the eastward flowing Indian Monsoon Current carries high-salinity water from the Arabian Sea to the Bay of Bengal (Kumar and Prasad, 1999; Vinayachandran et al., 1999). Thus both basins experience significant changes in $\delta^{18}O$ and salinity through the exchange of surface waters. Because of the geography of the Indian subcontinent, the surface water of the Bay is more diluted by runoff than that of the Arabian Sea, decreasing the surface salinity of the Bay by ~2 relative to the southern Indian Ocean (Delaygue et al., 2001). Maximum seasonal variations in salinity of 2 and 8 are observed in the Arabian Sea and the Bay of Bengal, respectively (Delaygue et al., 2001; Rao and Sivakumar, 2003).

Thus the relative contributions of the above processes to the $\delta^{18}O$–*S* relation are very different in the Arabian Sea from those in the Bay of Bengal (Delaygue et al., 2001). Higher runoff in the Bay of Bengal not only reduces the surface salinity but also stratifies the surface layer and prevents vertical mixing during summer (Rao and Sivakumar, 2003). Freshwater discharge causes less change in the sea surface salinity than does advection during winter in both the basins (Rao and Sivakumar, 2003). The northwestern Arabian Sea experiences a negative $P-E$ (excess of evaporation over precipitation), becoming less negative towards the southeastern part of the Arabian Sea, where intense rainfall leads to higher $P-E$ (Han and McCreary, 2001). River discharge into the northwestern Bay of Bengal produces minima (salinity varies from 33 to 29) during August–September, with maximum discharge in August (Han and McCreary, 2001; Rao and Sivakumar, 2003), and could significantly influence the $\delta^{18}O$–*S* relation.

Craig and Gordon (1965), in their model, used globally averaged values for hydrological fluxes to infer a $\delta^{18}O$–*S* slope; Schmidt et al. (2007) and Delaygue et al. (2001) simulated local $\delta^{18}O$ and salinity values with models covering the whole ocean, but Delaygue et al. (2001) only used the Indian part to address the problem on a regional scale for northern Indian Ocean. They inferred that, except for a few months, the same $\delta^{18}O$–*S* relation holds throughout the year, thus ruling out the effect of seasonal variations. However, regional modeling requires precise values of atmospheric and terrestrial water fluxes and values for isotopic fractionation during the phase transition of water to vapor. The relative importance of runoff and precipitation over the northern Indian Ocean is not well constrained (Delaygue et al., 2001; Rao and Sivakumar, 2003). Further, the isotopic fractionation during evaporation is a non-equilibrium process, and the isotopic composition of evaporative flux (different from the vapor flux) is estimated from mass-balance models (Craig and Gordon, 1965; Merlivat and Jouzel, 1979; Jouzel and Koster, 1996). Such uncertainties limit the validity of simple box models.

## 2. Methodology

Surface and deeper water samples were collected during eight different cruises (Tables 1–3). Sampling stations were chosen to capture the typical pattern of spatial variation of $\delta^{18}O$ and salinity (Fig. 1). All the samples were stored in tightly capped plastic bottles to prevent evaporation, and analyzed within one month of collection. Salinity measurements for all the ocean surface water samples were performed on board by Autosal with an external