



Coupling of thermocline depth and strength of the Indian, summer monsoon during deglaciation

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We investigated the variations in thermocline depth using the difference in $\delta^{18}\text{O}_{\text{c}}$ values between the two species of planktonic foraminifera, surface dwelling *Globigerinoides ruber* (s.s.) and thermocline dwelling *Neogloboquadrina dutertrei*, covering a time span of 9–23 kyr from the sediment core SK218/1 from the western Bay of Bengal (BoB). Here we show that during the strong phase of the Indian summer monsoon (12–9 kyr), a strong stratification leads to a shallow mixed layer and thermocline depth in the BoB as evident from higher $\Delta\delta^{18}\text{O}$ between the mixed layer and thermocline dwelling planktonic foraminifera species. Thus, a strong coupling between the Indian summer monsoon and thermocline depth in the BoB prevailed at a millennial time scale.

Keywords. Monsoon; the Bay of Bengal; thermocline; foraminifera; salinity; isotopes.

1. Introduction

Precipitation exceeds evaporation in the Bay of Bengal (BoB) (Cadet and Greco 1987), and a high run-off from major rivers contributes to the freshening of surface waters, mainly in the northern part of the bay which leads to a distinct salinity gradient from the north to the south (Duplessy 1982; Kudrass *et al.* 1997). The major rivers draining into the BoB are the Ganges, Brahmaputra, Mahanadi, Krishna and the Godavari. The effects of the Ganges–Brahmaputra are prominent in the northern bay, whereas those of the Krishna–Godavari are prominent in the western bay, while the Andaman Sea is mainly influenced by the outflow from the Irrawaddy, Salween and the Sitang rivers. The mixed layer depth (MLD) couples the ocean to the atmosphere and plays an important role in determining the oceanic primary productivity (Narvekar and Prasanna Kumar

2014). The MLD is known to vary throughout the BoB basin and is deeper to the north of 15°N , while it is shallower to the south of 15°N , during the summer monsoon due to strong wind mixing (Narvekar and Prasanna Kumar 2006). To the north of 15°N , along the western boundary and the east BoB, a stronger monsoon leads to intense precipitation and run-off increases due to the proximity of the rivers. A layer of fresh water forms at the surface together with increased sea surface temperatures (SSTs) which leads to stratification and decrease in the MLD (Narvekar and Prasanna Kumar 2006). Although the summer winds are strong, they are unable to break this layer and hence the MLD becomes shallow, so does the thermocline. Hence, there is a large difference in salinity between the thermocline and the MLD as a result of the stronger monsoon. This property was taken into consideration by two recent studies. Gebregiorgis *et al.*

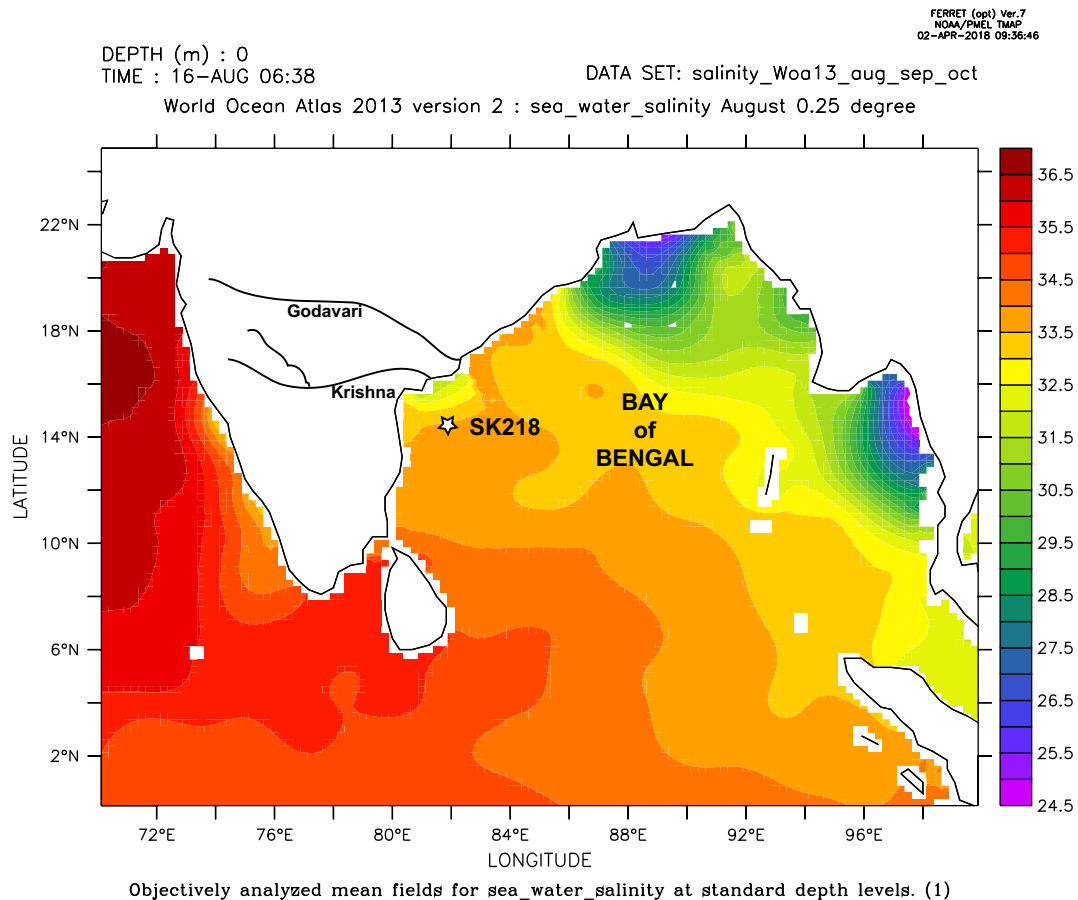


Figure 1. Core location of SK218/1 in the western BoB overlaid on the surface salinities for the month of August from World Ocean Atlas 13. The proximity of the rivers, Krishna and Godavari, lowers the salinity at the core location during the SW monsoon.

(2016) have discussed the surface hydrology of the BoB during the last glacial maximum (LGM) to the present using a sediment core from the Andaman Sea. They used thermocline and surface dwelling foraminifera species to reconstruct freshwater-induced stratification and showed that the LGM was characterised by relatively weaker South Asian summer monsoon (SAM) and reduced run-off. Sijinkumar *et al.* (2016) used a suite of 11 sediment cores from the BoB and the Andaman Sea to understand the north–south salinity gradient and showed that during the mid-Holocene, the northern and southern BoB and the Andaman Sea were fresher relative to modern times, whereas the BoB as a whole was more saline relative to modern times during the LGM. The Andaman Sea however remained unchanged during this time. There are other studies that reconstruct salinity in the northern BoB (Kudrass *et al.* 1997; Rashid *et al.* 2011) but only one study exists in the western BoB, a region influenced by the Krishna–Godavari river outflow (Govil and Naidu 2011; Core SK218/1).

They observed that the southwest (SW) monsoon was weak during the LGM and its strength increased beginning from the Bølling–Ållerød (B–A). It weakened during the Younger Dryas (YD) and thereafter it strengthened during the Holocene. These observations are consistent with all BoB studies. However, it is not known if the summer monsoon strength was coupled to the thermocline and the MLD variations back in time. We therefore use the sediment core, SK218/1 (figure 1) from the western BoB, influenced by the riverine run-off, to understand the changes in thermocline depth (see figure 2) from the LGM to the early Holocene when the strength of the SW monsoon is known to have gradually increased.

2. Materials and methods

Core SK218/1 was collected at a water depth of 3307 m from the BoB (14°02′06″N; 82°00′12″E)

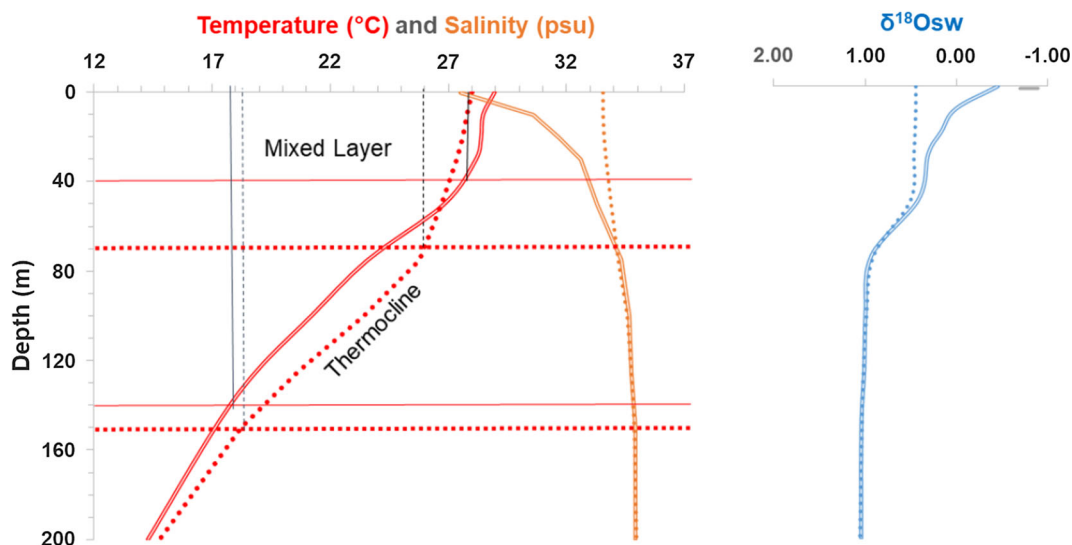


Figure 2. Temperature and salinity climatology profiles up to 200 m water depth obtained from the LEVITUS dataset at a grid of 14.5°N and 82.5°E, for the seasons February–April (dotted red and orange) and August–October (double line: red and orange) near the core location. The mixed layer and thermocline are marked by the horizontal lines (dotted red for February–April and solid red for August–October). The $\delta^{18}\text{O}_{\text{sw}}$ was calculated from the salinity data and equation by Sengupta *et al.* (2013) and the profiles are plotted in blue (dotted: February–April and solid: August–October).

(figure 1). The age model of this core was established based on eight accelerator mass spectrometry ^{14}C dates providing an age between 0.6 and 36.8 kyr for this entire core (see Govil and Naidu 2011). We have used the core section from ~ 23 to 9.3 kyr as we were particularly interested in analysing the strength of the SW monsoon and its relation to changes in thermocline depth. The $\delta^{18}\text{O}$ analyses were performed on *Neogloboquadrina dutertrei* (250–350 mm size range) and *Globigerinoides ruber* (*sensu stricto* morphotype; s.s.). Both species were analysed using a Thermo Delta V Plus mass spectrometer equipped with an automatic carbonate preparation device (Kiel IV) at the National Institute of Oceanography, Goa. The mean external error and reproducibility (1σ) of the carbonate standard is better than 0.07‰ and 0.05‰ for $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$, respectively. Oxygen and carbon isotopic values are calibrated against the NBS 19 standards and are expressed as per mil (‰) vs. the Vienna Pee Dee belemnite (VPDB) standard.

3. Results and discussion

$\delta^{18}\text{O}$ values of *N. dutertrei* range from -1.87‰ to 0.36‰ throughout the entire record of ~ 23 –9.3 kyr and a shift of $\sim 2\text{‰}$ between the glacial and interglacial (figure 3a) periods. The $\delta^{18}\text{O}$ values of *G. ruber* (s.s.) range from -3.5‰

to 0.54‰ with a shift of $\sim 1.5\text{‰}$ between the glacial and interglacial (figure 3a) periods. The difference in $\delta^{18}\text{O}$ of these two species of planktonic foraminifera, i.e., the shallower dwelling *G. ruber* (s.s.) and thermocline dweller *N. dutertrei* (denoted as $\Delta\delta^{18}\text{O}$), ranges from -0.14‰ to -1.63‰ (figure 3b). Temperature and salinity data were obtained from the LEVITUS dataset at a grid close to our sediment core location (14.5°N and 82.5°S; Levitus and Boyer 1994) at a depth of 200 m and are plotted for two different seasons, February–April and August–October (figure 2). The SST difference between these two seasons is only 1°C . The surface salinity, however, shows a large difference (six salinity units) between the two seasons. The August–October season encompasses the peak SW monsoon. It is during this time that the *G. ruber* fluxes are slightly higher in the central BoB sediment trap (CBBT; 13°09'N and 84°21'E), which lies further east of our sediment core location (Guptha *et al.* 1997).

The MLD and thermocline depth were approximately demarcated using the temperature profile near the core location (figure 2). We chose the seasons of February–April (pre-monsoon) and August–October (peak SW monsoon and the end of it) so as to get a picture of a large contrast in salinity as an effect of change in the SW monsoon intensity and to observe significant differences in thermocline depth. It is observed that the MLD and thermocline become shallow

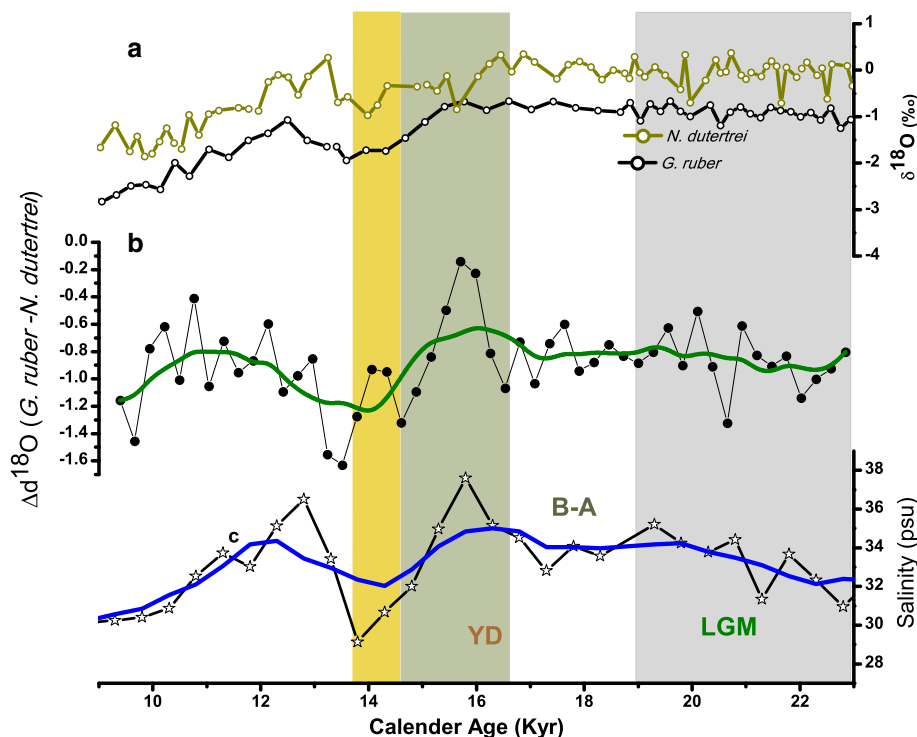


Figure 3. (a) $\delta^{18}\text{O}$ (‰) measured from *G. ruber* (black line) and *N. dutertrei* from Core SK218/1 (green line); (b) the difference between $\delta^{18}\text{O}$ of *G. ruber* and *N. dutertrei* is denoted as $\Delta\delta^{18}\text{O}$, and (c) the salinity record from SK218/1, recalculated from Govil and Naidu (2011) using the $\delta^{18}\text{O}_{\text{sw}}$ –salinity calibration from Sengupta et al. (2013). The thick lines in (b) and (c) are five-point smoothing of the data.

during the August–October season and deepen during the February–April season. The shallow MLD and thermocline occur due to stratification as a result of the freshwater input from the Krishna to the Godavari rivers during the SW monsoon. The modern seawater $\delta^{18}\text{O}$ ($\delta^{18}\text{O}_{\text{sw}}$) values were calculated using the salinity data and equations of Sengupta et al. (2013) who identified water masses based on the temperature–salinity relationship and also measured the $\delta^{18}\text{O}$ of different water masses. The water column depth of 200 m that we have selected to demarcate the thermocline and the MLD is presently influenced by two water masses: (i) the BoB waters (BoBW, 0–50 m) and (ii) mixed zone waters (MZ, 60–200 m). We use two different $\delta^{18}\text{O}$ –salinity calibrations for these water masses: $S = (\delta^{18}\text{O}_{\text{sw}} + 4.58)/0.15$ for the BoBW waters and $S = (\delta^{18}\text{O}_{\text{sw}} + 5.52)/0.188$ for MZ waters (Sengupta et al. 2013). The $\delta^{18}\text{O}_{\text{sw}}$ values thus obtained are shown in figure 2. The $\delta^{18}\text{O}$ –salinity slope of $0.15\text{‰}/\text{psu}$ determined by Sengupta et al. (2013) is similar to that by Delaygue et al. (2001), Singh et al. (2010) and Achyuthan et al. (2013). Sijinkumar et al. (2016) have put together all available $\delta^{18}\text{O}_{\text{sw}}$ data of the BoB to establish $\delta^{18}\text{O}$ –salinity calibration wherein they calculate

a similar slope ($0.15\text{‰}/\text{psu}$). We therefore feel confident about using this calibration by Sengupta et al. (2013) for recalculating the downcore salinity from SK218/1 from the $\delta^{18}\text{O}_{\text{sw}}$ data by Govil and Naidu (2011). Furthermore, we calculated the difference in $\delta^{18}\text{O}$ at the base of the mixed layer and the thermocline depth, denoted as $\Delta\delta^{18}\text{O}_{\text{sw}}$. The $\Delta\delta^{18}\text{O}_{\text{sw}}$ is observed to be the lowest during the monsoon season and the highest during the pre-monsoon. We use this understanding and apply it to the downcore $\Delta\delta^{18}\text{O}$ data to deduce the variation in thermocline depth and in turn changes in the monsoon intensity from LGM to early Holocene.

The $\delta^{18}\text{O}_{\text{c}}$ is influenced by the $\delta^{18}\text{O}$ of sea water and water temperature. $\delta^{18}\text{O}$ seawater is controlled by the evaporation and precipitation and also global ice volume. As shown earlier, the seasonal SSTs do not differ much from each other and the water column temperatures between seasons do not have drastic differences (figure 2). However, the surface salinity difference is large (figure 2) and since the ice volume change will affect the mixed layer and thermocline in a similar manner, when one dataset is subtracted from the other, the ice volume effect will be cancelled out. Hence,

the downcore $\Delta\delta^{18}\text{O}$ calculated here should mainly result from changing salinity (figure 3b).

The $\Delta\delta^{18}\text{O}$ record of the sediment core was compared with reconstructed salinity and both the profiles show a good resemblance (figure 3b and c) which further indicates that $\Delta\delta^{18}\text{O}$ can be used as a measure of the thermocline depth and mixed layer salinity changes driven by monsoonal precipitation (figure 3b and c). Downcore $\Delta\delta^{18}\text{O}$ suggests the following: the lowest $\Delta\delta^{18}\text{O}$ values signify the shallowing of the MLD and thermocline due to the stronger monsoon and hence freshening and stratification, whereas a higher $\Delta\delta^{18}\text{O}$ signifies the deepening of MLD and a deeper thermocline due to the weakening of the monsoon (figure 3b and c). The salinity record shows an overall seawater freshening from 23 to 10 kyr (figure 3c). The monsoon, as shown by the salinity record, was weak during the LGM. A prominent peak in $\Delta\delta^{18}\text{O}$ and salinity is noticed at ~ 16 kyr which signifies a weak monsoon. This weakening in monsoon is supported by peat records from the Nilgiri hills, southern India, which shows that moist conditions started after ~ 16 kyr and the monsoon strengthened (Sukumar *et al.* 1993). Furthermore, $\Delta\delta^{18}\text{O}$ and salinity from SK218/1 show that the monsoon strengthened during the B–A and again weakened during the YD which is in line with previous studies in this region (Kudrass *et al.* 1997; Govil and Naidu 2011; Rashid *et al.* 2011). In general, the MLD is shallow to the north of 15°N and deeper to the south of 15°N (Narvekar and Prasanna Kumar 2006) due to the stratification as a result of the precipitation and run-off in the north. In the southern region, the freshwater influence is less and the mixed layer is deepened due to the higher wind speed and also the presence of the high salinity waters of the Arabian Sea. Our core location is slightly southern to 15°N , but is influenced by the Krishna–Godavari (K–G) river run-off and hence we observe freshwater stratification during the SW monsoon.

4. Conclusions

The difference in $\delta^{18}\text{O}$ of surface and thermocline dwelling foraminifera, *G. ruber* and *N. dutertrei*, clearly demonstrates the coupling between the stratification driven by the monsoon rainfall and its influence on the shallowing of thermocline in the western BoB. Strong stratification and shallowing of thermocline from 12 to 9 kyr compared with

the last glacial period was noticed which signifies a stronger monsoon during the 12–9 kyr.

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