

Coupling of thermocline depth and strength of the Indian Summer Monsoon during the deglaciation

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Running Title:

Thermocline depth and Indian Summer Monsoon

Abstract

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

Keywords: monsoon, Bay of Bengal, thermocline, foraminifera, salinity, isotopes

Introduction:

Precipitation exceeds evaporation in the Bay of Bengal (BoB) (Cadet and Greco, 1987), and high runoff from major rivers contributes to freshening of surface waters mainly in the northern part of the Bay which leads to a distinct salinity gradient from north to south (Duplessy, 1982; Kudrass *et al.*, 2001). The major rivers draining into the BoB are the Ganges, Brahmaputra, Mahanadi, Krishna, and Godavari. The effects of Ganges-Brahmaputra are prominent in the northern Bay whereas Krishna-Godavari in the western Bay, while the Andaman Sea is mainly influenced by the outflow from Irrawaddy, Salween and 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 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, stronger monsoon leads to more intense precipitation and runoff increases due to proximity of rivers. A layer of fresh water forms at the surface together with increased sea surface temperatures (SSTs) which leads to stratification and a decrease in MLD (Narvekar and Prasanna Kumar, 2006). Though the summer winds are strong, they are unable to break this layer and hence MLD shallows and so does the thermocline. Hence there is a large difference in salinity between the thermocline and mixed layer as a result of stronger monsoon. This property was taken into consideration by two recent studies. Gebregeorgis *et al.* (2016) have discussed the surface hydrology of the BoB during the LGM to 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 runoff. Sijinkumar *et al.*, (2016) have used a suite of 11 sediment cores from the BoB and Andaman Sea to understand the north-south salinity gradient and shown that during the mid-Holocene the northern and southern BoB and the Andaman Sea were fresher relative to modern whereas the BoB as a whole was more saline relative to modern during LGM. The Andaman Sea however remained unchanged during this time. There are other studies that reconstruct salinity in the northern BoB (Kudrass *et al.*, 2001; 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 last glacial maximum (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 the BoB studies. However it is not known if the summer monsoon strength was coupled to the

thermocline and MLD variations back in time. We therefore use the sediment core, SK 218/1 (Fig. 1) from the western BoB, influenced by riverine runoff, to understand the changes in thermocline depth (See Fig. 2) from the LGM to the Early Holocene when the strength of the southwest (SW) monsoon is known to have gradually increased.

1. Materials and methods:

Core SK218/1 was collected at a water depth of 3307m from the BoB (14°02'06" N; 82°00'12" E) (Fig. 1). The age model of this core was established based on 8 AMS ^{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 strengthening SW monsoon and its relation to changes in thermocline depth. The $\delta^{18}\text{O}$ analyses were performed on *Neogloboquadrina dutertrei* (250 to 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 NIO, Goa. The mean external error and reproducibility (1σ) of 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 (‰) versus the Vienna Pee Dee belemnite (VPDB) standard.

2. Results and discussion:

$\delta^{18}\text{O}$ values of *N. dutertrei* range from -1.87 to 0.36 ‰ throughout the entire record of ~ 23 to 9.3 kyr and a shift of ~ 2‰ between glacial and interglacial (Fig. 3a). The $\delta^{18}\text{O}$ of *G. ruber* (s.s.) range from -3.5 to 0.54 ‰ with a shift of ~ 1.5‰ between glacial and interglacial (Fig. 3a). 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 $\Delta \delta^{18}\text{O}$), range from -0.14 to -1.63 ‰ (Fig. 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) to depth of 200 m and are plotted for two different seasons, February to April and August to October (Fig. 2). The SST difference between these two seasons is only 1°C. The surface salinity however shows a large difference (6 salinity units) between the two seasons. The August to October season encompasses the peak southwest monsoon. It is during this time that the *G. ruber* fluxes are slightly higher in the central Bay of Bengal sediment trap (CBBT; 13°09'N and 84°21'E), which is further east of our sediment core location (Guptha *et al.*, 1997).

The MLD and thermocline depth were approximately demarcated using the temperature profile near core location (Fig. 2). We chose the seasons of February to April (pre-monsoon) and August to

October (peak SW monsoon and end of) so as to get a picture of large contrast in salinity as an effect of change in southwest monsoon intensity and to observe significant differences in thermocline depth. It is observed that the MLD and thermocline shallows during the August-October season and deepens during the February-April season. The shallow MLD and thermocline occurs due to stratification as a result of freshwater input from Krishna-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 using equations from Sengupta *et al.*, 2013. Sengupta *et al.*, 2013 have identified water masses based on temperature-salinity relationship and also measured the $\delta^{18}\text{O}$ of the 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; 1. Bay of Bengal Water (BoBW; 0 to 50 m) and 2. Mixed Zone waters (MZ 60 to 200 m). We use the two different $\delta^{18}\text{O}$ -salinity calibrations for these water masses; $S = (\delta^{18}\text{O}_{\text{sw}} + 4.58) / 0.15$ for BoBW waters and $S = (\delta^{18}\text{O}_{\text{sw}} + 5.52) / 0.188$ for Mixed Zone water (Sengupta *et al.*, 2013). The $\delta^{18}\text{O}_{\text{sw}}$ thus obtained are shown in Fig. 2. The $\delta^{18}\text{O}$ -salinity slope of 0.15‰/psu determined by Sengupta *et al.*, 2013 is similar to that of 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 from the BoB to establish $\delta^{18}\text{O}$ -salinity calibration wherein they calculate a similar slope (0.15‰/psu). We therefore feel confident to use this calibration by Sengupta *et al.*, 2013 for recalculating the downcore salinity from SK 218/1 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 lowest during the monsoon season and highest during pre-monsoon. We use this understanding and apply it to the down core $\Delta \delta^{18}\text{O}$ data to deduce 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 (Fig. 2). However the surface salinity difference is large (Fig. 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 (Fig. 3b).

The $\Delta \delta^{18}\text{O}$ record of the sediment core was compared to reconstructed salinity and both the profiles show a good resemblance (Fig. 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 (Fig. 3b and c). Down core $\Delta \delta^{18}\text{O}$ suggests the following: the lowest $\Delta \delta^{18}\text{O}$ values signify shallowing

of the MLD and thermocline due to stronger monsoon and hence freshening and stratification whereas higher $\Delta \delta^{18}\text{O}$ signify deepening of MLD and a deeper thermocline due to weakening of monsoon (Fig. 3b and c). The salinity record shows an overall seawater freshening from 23kyr to 10kyr (Fig. 3c). The monsoon as shown by the salinity record, was weak during the Last Glacial Maximum (LGM). A prominent peak in $\Delta \delta^{18}\text{O}$ and salinity is noticed at ~16kyr which signifies weak monsoon. This weakening in monsoon is supported by peat records from Nilgiri hills, southern India, which show that moist conditions started after ~16kyr and monsoon strengthened (Sukumar *et al.*, 1993). Furthermore, $\Delta \delta^{18}\text{O}$ and salinity from SK218/1 shows that the monsoon strengthened during the (B-A) and again weakened during the Younger Dryas (YD) which is in line with previous studies in this region (Kudrass *et al.*, 2001; Govil and Naidu, 2011; Rashid *et al.*, 2011). In general the MLD is shallow north of 15°N and deeper to south of 15°N (Narvekar and Prasanna Kumar, 2006) due to stratification as a result of precipitation and runoff in the North. In the southern region, the fresh water influence is less and the mixed layer is deepened due to higher wind speed and also the presence of Arabian Sea high salinity waters. Our core location is slightly southern to 15°N but is influenced by the Krishna-Godavari (K-G) river runoff and hence we see fresh water stratification during SW monsoon.

3. Conclusions:

The difference in $\delta^{18}\text{O}$ of surface and thermocline dwelling foraminifera, *G. ruber* and *N. dutertrei* clearly demonstrated the coupling between the stratification driven by the monsoon rainfall and its influence on the shallowing of thermocline in the western Bay of Bengal. Strong stratification and shallowing of thermocline from 12 to 9 kyr than the last glacial period was noticed which signifies a stronger monsoon during 12 to 9 kyr.

Acknowledgements: We thank the Director, NIO for his encouragement and support. Anita Garg was responsible for the measurements on IRMS. Jayu Narvekar helped with the figures. This is NIO contribution NO. XXXX.

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Figure Legends:

Fig. 1. Core location of SK218/1 in the western Bay of Bengal overlaid on surface salinities for the month of August from World Ocean Atlas 13. Proximity of rivers, Krishna and Godavari lowers salinity at core location during southwest monsoon.

Fig. 2. Temperature and Salinity climatology profiles up to 200m water depth obtained from LEVITUS dataset at a grid of 14.5°N and 82.5°E, for the seasons of February to April (dotted red and orange) and August to October (double line; red and orange) near to core location. The mixed layer and thermocline are marked by the horizontal lines (dotted red for Feb-Apr and solid red for Aug-Oct). The $\delta^{18}\text{O}_{\text{sw}}$ was calculated from the salinity data and equation by Sengupta *et al.*, 1993 and the profiles are plotted in blue (dotted: Feb-Apr and solid: Aug-Oct).

Fig. 3. a) $\delta^{18}\text{O}$ (‰) measured from *Globigerinoides ruber* (black line) and *Neogloboquadrina dutertrei* from Core SK 218/1 (green line); b) The difference between $\delta^{18}\text{O}$ of *G. ruber* and *N. dutertrei* denoted as $\Delta\delta^{18}\text{O}$; and c) salinity record from SK218/1, recalculated from Govil and Naidu, 2011 using the $\delta^{18}\text{O}_{\text{sw}}$ -salinity calibration form Sengupta *et al.*, 2013 . The thick lines in (b) and (c) are 5 point smoothing of the data.

Figures

Fig.1

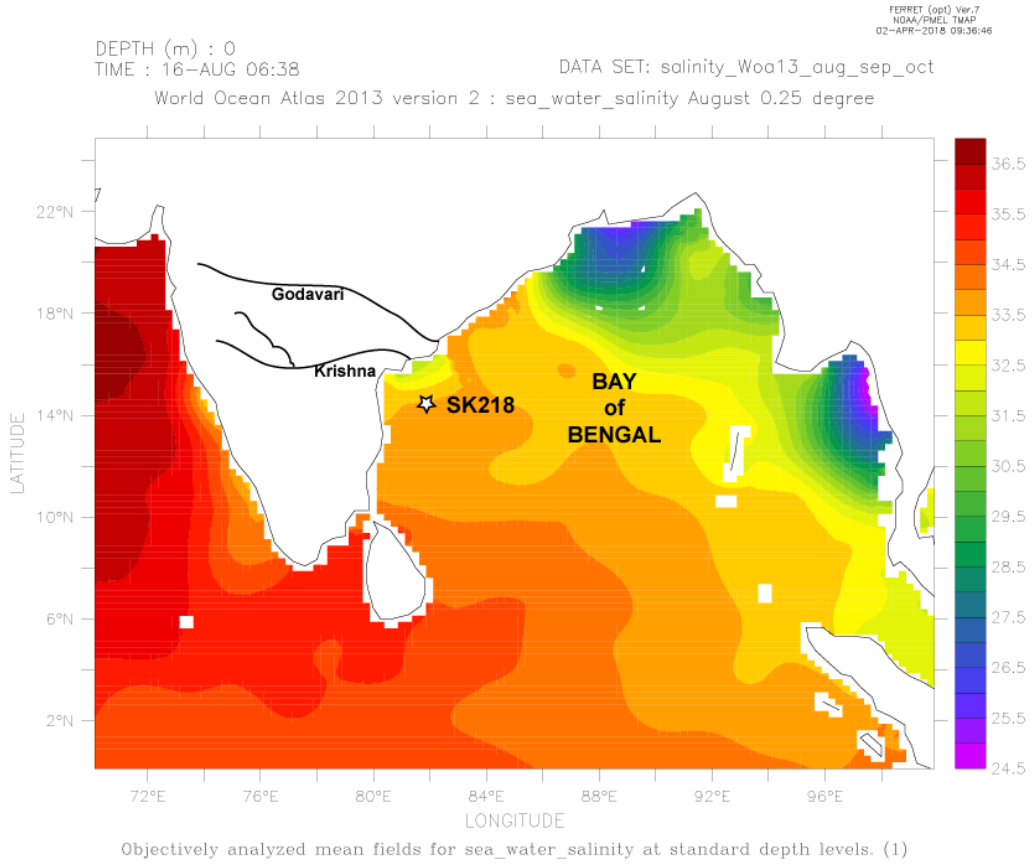


Fig. 2

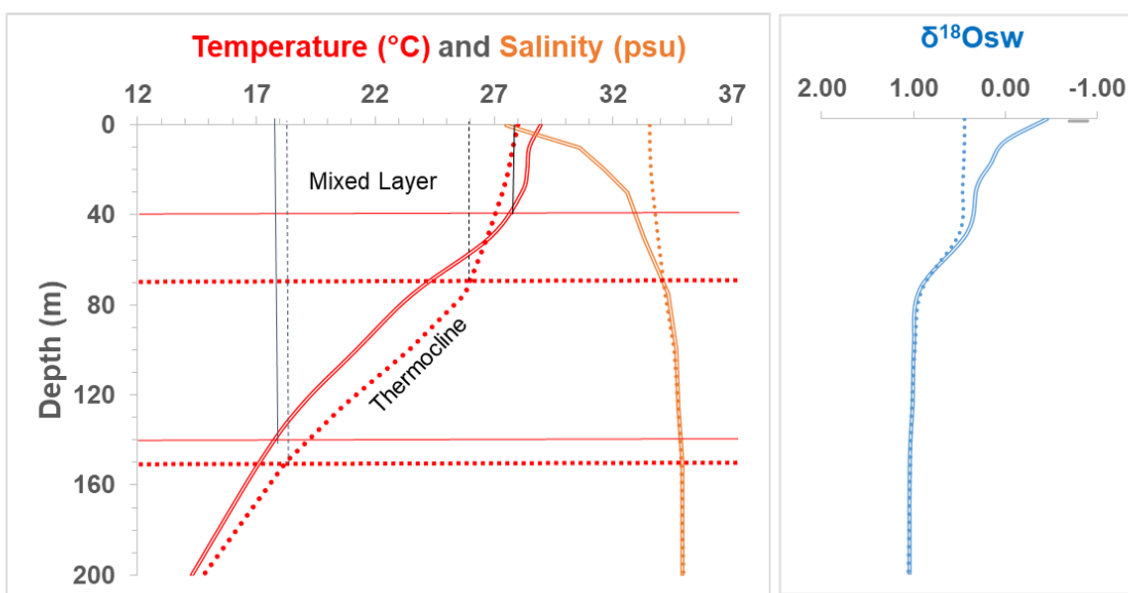


Fig. 3

