

Bay of Bengal Exhibits Warming Trend During the Younger Dryas: Implications of AMOC

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Abstract

A sharp decline in temperature during the Younger Dryas (YD) preceding the current warmer Holocene is well documented in climate archives from the Northern Hemisphere high latitudes. Although the magnitude of YD cooling varied spatially, the response of YD cooling was well documented in the Atlantic and Pacific Oceans but not in the Indian Ocean. Here we investigate whether the modern remote forcing of tropical Indian Ocean sea surface temperature (SST) by Northern Hemisphere climate changes holds true for events such as the YD. Our SST reconstruction from the western Bay of Bengal exhibits an overall warming of $\sim 1.8^{\circ}\text{C}$ during the YD. We further compared our data with other existing Mg/Ca-based SST records from the Northern Indian Ocean and found no significant negative SST anomalies in both the Arabian Sea and the Bay of Bengal compared to pre and post YD, suggesting that no apparent cooling occurred during the YD in the Northern Indian Ocean. In contrast, most part of the YD exhibits positive SST anomalies in the Northern Indian Ocean that coincide with the slowdown of the Atlantic Meridional Overturning Circulation during this period.

1. Introduction

The Younger Dryas (YD), an abrupt climatic event during the last deglaciation (~12.8 to 11.5 ka), is known to have terminated the last glacial cycle following a warm interval known as the Ållerød period [e.g *Carlson*, 2013]. A sharp drop in temperature during the YD has attracted great attention among the paleoceanographic community. Several studies have suggested that a sudden release of freshwater from the Laurentide Ice Sheet into the North Atlantic weakened the Atlantic Meridional Overturning Circulation (AMOC) and triggered the YD by slowing down the thermohaline circulation [*Johnson and McClure*, 1976; *Rooth*, 1982; *McManus et al.*, 2004; *Carlson et al.*, 2007]. The slowdown of the AMOC seemed to have had far reaching impacts on global climate rather than just in the Northern Hemisphere (NH) [*Vellinga and Wood*, 2002; *Chiang and Bitz*, 2005; *Broecker et al.*, 2010]. Previously, it has been suggested that the magnitude of temperature drop during the YD varied from 3 to 10°C in the NH with a stronger cooling in higher altitudes than in lower altitudes [*Shakun and Carlson*, 2010]. In the Southern Hemisphere (SH), a slight warming was noticed instead of cooling suggesting an inter-hemispheric bipolar-seesaw nature of temperature change during the YD [*Blunier and Brook*, 2001]. However, the cooling in NH was more intense than the warming in SH during the YD interval leading to an overall net global cooling of ~0.6°C [*Shakun and Carlson*, 2010] with the least or negligible impacts in tropical regions.

The relation between cooling in the NH and the tropical Indian Ocean during the YD has been investigated by sea surface temperature (SST) reconstructions. However, the SST pattern during the YD in the Northern Indian Ocean is not consistent. For instance, surface cooling was reported from the Gulf of Aden [*Tierney et al.*, 2015], while SSTs were found to be relatively constant in the western Arabian Sea during the YD [*Saher et al.*, 2007a], and warmer with saltier sea surface conditions during the stadial periods such as Heinrich event 1 and YD, relative to the interstadials in both eastern and western Arabian Sea [*Anand et al.*, 2008]. The discrepancies in the few available records call for more studies to understand the Indian Ocean dynamics coupled with monsoon variability and global climatic teleconnections. In particular, studies from the Bay of Bengal (BoB) are less compared to the Arabian Sea regarding abrupt climatic events, and the existing SST of the YD event are still ambiguous [*Kudrass et al.*, 2001; *Govil and Naidu*, 2011; *Rashid et al.*, 2011]. In this study we investigate SST changes in the BoB during the YD and evaluate the response of tropical Indian Ocean SST to the forcing factors associated with the YD occurrence in the NH. We make use of Mg/Ca ratios of planktonic foraminifera species *Globigerinoides ruber* to derive the SST in the western BoB (Fig. 1) and compare our results with other published data from the Northern Indian Ocean region.

2. Study Area

The Indian Ocean is mainly impacted by the complex and dynamic Indian monsoon system that arises out of the temperature and pressure gradient over the Asian continent and the Indian Ocean, and is responsible for redistribution of heat and moisture over the whole densely-populated Asian region [Webster *et al.*, 1998]. During June through September, southwest monsoon winds are responsible for upwelling of cold, nutrient rich waters in western and southeastern Arabian Sea [Prell, 1984; Saher *et al.*, 2007b], and for huge amount of precipitation and river discharge to the BoB [Shetye *et al.*, 1991, 1993; Colin *et al.*, 1999; Sengupta *et al.*, 2006]. In boreal winter, the wind reverses in direction; the northeast monsoon winds induce convective mixing in the northeastern Arabian Sea and cooling of the northern BoB. The SST at the core site experiences an annual cycle peaking at >29 °C during April-June and cooling to <27 °C during November-February (Fig. 2) [Locarnini *et al.*, 2013]. Similarly, high salinity of >32.5 psu is observed through April-September, but <30.5 psu during October-December [Locarnini *et al.*, 2013]. The surface hydrography at the core location presented here is heavily influenced by the Krishna - Godavari river discharges, which depend on the strength of Indian summer monsoon. Furthermore, the seasonal reversal of monsoonal winds and precipitation, mainly responsible for the hydroclimatic changes, is also directly associated with the seasonal northward and southward shift of the Intertropical Convergence Zone (ITCZ) during summer and winter, respectively [e.g. Webster *et al.*, 1998; Gadgil, 2003].

3. Material and methods

Sediment core MD 161/17 was collected at a water depth of 790 m from the BoB ($16^{\circ}03.63'N$; $82^{\circ}01.29'E$, Fig. 1). The total length of the core is 25.08 m. The proximity of this site to the Krishna-Godavari river mouth, whose outflows are heavily influenced by the Indian Summer monsoonal strength, makes it ideal to investigate past SST changes associated with the monsoon intensity. The age model of this core was established based on ^{14}C dates, which were performed on mixed planktonic foraminifera species >150 μm fraction at the NSF-Arizona AMS Laboratory of Arizona University, USA. Radiocarbon ages were calibrated to calendar years using the CALIB 7.1 software and the Marine13 calibration curve [Stuiver and Reimer, 1993; Reimer *et al.*, 2009] with the global reservoir correction of 400 years [Southon *et al.*, 2002] (Table S1). Calibrated ages with 2σ error bars along with depth are plotted in Fig. 3. The age model utilizes the calibrated dates as tie-points and assumes linear sedimentation between the tie-points. The main focus of the present study is on the cold YD event during the last deglaciation and therefore, we used only 18.35-23.95 m section of

the core and present here the SSTs derived from Mg/Ca ratios in *G. ruber* spanning the time interval of 8-14 ka BP, which includes the YD event.

For Mg/Ca analysis, the foraminifera samples (30-40 individual specimens of >150 μm size range) were picked and cleaned by applying a slight modification of the method originally proposed in *Barker et al* [2003], and consisted of five water washes and two methanol washes followed by two oxidation steps with 1% NaOH-buffered H_2O_2 , then a weak acid leach with 0.001 M QD HNO_3 . The samples were then dissolved into 0.075 M QD HNO_3 and centrifuged for 10 minutes at 6,000 r.p.m., transferred into test tubes and diluted. Mg/Ca ratios were measured using an Agilent Technologies 700 Series ICP-OES with a CETAX ASX-520 autosampler housed at the Faculty of Geosciences, University of Bremen. Mg/Ca values are reported as mmol mol^{-1} . The instrumental precision was determined using an external, in-house standard (Mg/Ca=2.92 mmol mol^{-1}), which was run after every fifth sample. The residuals of the external standard and ECRM standard were 0.53% (0.023 mmol mol^{-1}) and 0.05% (0.002 mmol mol^{-1}) respectively. Reproducibility of the samples from replicate measurements is 0.69% (n=20), which equals 0.03 mmol mol^{-1} and accounts for an average temperature error of $\sim 0.9^\circ\text{C}$ [Mohtadi et al., 2014]. Mg/Ca values were then used to estimate SSTs using the equation $\text{Mg/Ca} = 0.449\exp(0.09 \cdot T)$ [Anand et al., 2003], where Mg/Ca is in mmol mol^{-1} and SST is in $^\circ\text{C}$. This calibration equation was selected because derived temperatures of the core tops were close to modern values at the core location, which is within error estimate of $\pm 0.3^\circ\text{C}$. Since differences in absolute temperature values during YD period, may be due to use of different calibration equations, we also calculated the SST anomalies in all records from the northern Indian Ocean. SST anomalies for each core were calculated by subtracting the averaged value of the period considered, from the individual values. The SST error estimates of the other cores (SK17, SK237 and AAS 62/1) (Fig. 1) are about $\pm 1^\circ\text{C}$ and also all these cores have good chronological control in bracketing the YD.

4. Results

In this study YD is bracketed by using 7 AMS ^{14}C dates, therefore the time span of the YD in our records is highly reliable and the interpreted SST changes remain valid regardless of the associated 2σ error with the dating. Mg/Ca ratios and estimated SST values are given in the Supplementary material as Table S2. The reconstructed SST record of core MD 161/17 varied from 24.7° to 28°C during the ~ 8 -14 ka period (Fig. 4a). Greater SST fluctuations were documented from 14 to 13.2 ka than the later period i.e 13.2 to 12.8 ka. A distinct SST warming trend (24.9 to 26.7°C) was noticed from 12.8 to 11.5 ka which exactly corresponds to the YD time period. The YD interval SSTs varied

in the lower values at the initial stage but showed an increasing trend from 12.7 to 11.5 ka (24.9 to 26.7°C), with an average of 25.9°C. The overall SST increase recorded for the YD period in MD 161/17 is 1.8°C. This shift is a striking feature in our dataset and beyond the SST error associated with the individual data points (~0.9°C). A decrease in SST of ~1.5°C from 11.5 to 11.3 ka is followed by an increase from 25.3°C at 11.3 ka to a peak of 28°C at 11 ka, and a subsequent abrupt cooling to 25.1°C at 10.5 ka, after which more or less gradual warming trend continued until 8 ka. SST anomalies were also calculated for all four cores from the northern Indian Ocean for the period between 11 to 14 ka and presented in Fig. 5. The MD 161/17 SST anomaly records negative anomalies (0 to -1°C) during 14 to 12.2 ka, but shifted to positive anomalies (~1°C) later in the YD period with some minor fluctuations until 11 ka. The major portion of YD period also document positive SST anomalies in cores SK17, AAS62/1 and SK237-GC04 from the eastern Arabian Sea (Fig. 5).

5. Discussion

5.1. SST variability during the YD in the northern Indian Ocean

To understand the Northern Indian Ocean SSTs response regionally during the YD, we have compared our results with published SST records derived from Mg/Ca of *G. ruber* with a sufficient temporal resolution and a reliable chronology from other regions of the Northern Indian Ocean (Fig. 4). The reconstructed SST record of the core MD 161/17 from the BoB clearly show a warming trend and positive SST anomalies during the YD. Similarly, the eastern Arabian Sea SST records of Core SK17 [Anand *et al.*, 2008], Core AAS62/1 [Kessarkar *et al.*, 2013], Core SK237-GC04 [Saraswat *et al.*, 2013] also document distinct increasing SST during YD (Fig. 4) and also major portion of YD in all these records show positive SST anomalies (Fig. 5). Therefore, apparently northern Indian Ocean document a warming trend during the YD.

On the contrary, surface cooling of ~0.5°C during the YD was reported from the Gulf of Aden in the westernmost Arabian Sea [Tierney *et al.*, 2015]. The authors suggested that the Indian Ocean SST cooling is the link between deglacial Indian monsoon failure and the North Atlantic stadials, implying a direct in-phase connection between the Indian Ocean SST variability and the remote North Atlantic cooling episodes such as Heinrich Event 1 and the YD. However, within the western Arabian Sea a warming SST trend during YD was documented in Core NIOP 929 [Saher *et al.*, 2007] and Core NIOP 905 [Anand *et al.*, 2008], which indicates inconsistency of SST during the YD modulated through the spatial variability of upwelling in western Arabian Sea. Therefore, the existing contrast of the SST pattern between the eastern Arabian Sea and the western Arabian Sea

during the YD is attributed to the monsoon upwelling, which is more intense in the western Arabian Sea. Hence the seasonal SST difference was larger during the last glacial period when upwelling was weak [Naidu and Malmgren, 2005; Saher et al., 2007]. The upwelling proxy records from the western Arabian Sea do not show a reduced upwelling strength during the YD [Naidu and Malmgren, 1995; Overpeck et al., 1996], hence Gulf of Aden documented 0.5°C cooling during the YD [Tierney et al., 2015]. Moreover, it has been argued that the southwest monsoon strength and the associated upwelling in the western Arabian Sea do not necessarily correspond to the monsoon rainfall over the Indian subcontinent [Govil and Naidu, 2011], casting doubt on a straightforward relationship between the western Arabian Sea cooling and the North Atlantic climate. Therefore, we argue that there is no coherent regional response of cooling in the northern Indian Ocean during the YD rather a striking warming trend is noticed during this period.

The YD warming of the BoB was accompanied by an increase of $\delta^{18}\text{O}_{\text{sw}}$ in the western BoB [Govil and Naidu, 2011], Andaman Sea [Rashid et al., 2007], the northern BoB [Rashid et al., 2011], and also in the eastern Indian Ocean [Mohtadi et al., 2014] indicating that the Indian summer monsoon rainfall reduced during the YD. In addition, reduction of Indian summer monsoon during the YD is also confirmed by the Speleothem records from the northern Indian region [Sinha et al., 2005] (Fig. 5). Decreased intensity of Indian summer monsoon regionally during the YD in both continent [Sinha et al., 2005] and oceanic regions [Mohtadi et al., 2014] was associated with the southward displacement of the ITCZ forced by the north Atlantic cooling through atmospheric teleconnection, which were also imprinted in the denitrification [Altabet et al., 2002] and tropical convection processes [Ivanochko et al., 2005] records of the western Arabian Sea.

5.2. Global Teleconnection

The YD event signatures are globally imprinted, generally with widespread cooling in NH and warming in southern hemisphere (SH), the so-called bi-polar seesaw [EPICA, 2006; Barker et al., 2009; Landais et al., 2015]. In the Atlantic ocean, 1-3°C SST cooling was documented in the North Atlantic region [e.g. Bard et al., 2000; Carlson et al., 2008], 1-7 °C SST cooling in Norwegian Sea [Benway et al., 2010], 0.5-1°C cooling in north African SSTs [Zhao et al., 1995; deMenocal et al., 2000], 3-4°C cooling in the Cariaco Basin [Lea et al., 2003], and slight cooling (~1°C) near the end of the YD period in the Gulf of Mexico [Flower et al., 2004], except for near the southeastern United States where SST increased by ~1.5°C owing to trapping of heat resulted from the AMOC slowdown [Grimm et al., 2006; Carlson et al., 2008a]. The Pacific coast of the United States and Canada recorded 2–3°C of YD cooling [Barron et al., 2003; Vacco et al., 2005]. Likewise, SST records from

China Sea suggest 0.5-1°C cooling [Sun *et al.*, 2005; Kubota *et al.*, 2010]. The SH, on the other hand, generally warmed during the YD [Shakun and Carlson, 2010; Clark *et al.*, 2012; Shakun *et al.*, 2012]. A warming of 0.3-1.9°C was observed from the SST records of southeast Atlantic to New Zealand [Lamy *et al.*, 2004; Pahnke and Sachs, 2006; Carlson *et al.*, 2008]. The warming trend of the BoB record during YD mimimics the Antarctica warming documented in the EPICA ice core record (EPICA, 2006) and also the onset of deglaciation timing in the northern Indian Ocean coincides with Antarctica warming [Naidu and Govil, 2010] suggest a coherent deglaciation response exist between the northern Indian Ocean and Antarctica. However, because of the limitations of chronological constrains and sampling resolution of the SST data presented in this study we are not able to discuss the lead/lag of the BoB SST variations with either Greenland and Antarctic Ice core records.

As evident from the northern Indian Ocean records in the present study, there is a distinct increasing trend of SST from 12.3 to 11.5 ka in the BoB (Fig. 4a). Similar warming during YD were documented from Caribbean Sea and off Brazil [0.25-1.2°C: Jaeschke *et al.*, 2007], off tropical West Africa [-0.2°C: Weldeab *et al.*, 2007] and eastern tropical Pacific [e.g., Benway *et al.*, 2006; Lea *et al.*, 2006]. Scientific consensus exist that the AMOC slowed down due to sudden influx of fresh water to the North Atlantic, and ultimately triggered the YD with significant cooling in the North Atlantic realm [e.g. McManus *et al.*, 2004; Broecker, 2006a; Carlson *et al.*, 2007]. The cooling associated with the YD caused an extensive winter sea ice cover, which increased the albedo and prevented the release of ocean heat causing the westerly winds to shift to a more southern path [Brauer *et al.*, 2008]. As a consequence, Siberian-like winter conditions prevailed over the Northern Atlantic and adjacent landmasses [Denton *et al.*, 2005]. As Northern Indian Ocean did not cool during the YD, we hypothesise that the AMOC slowdown caused a warming in the Northern Indian Ocean.

Community Climate System Model (CCSM3) North Atlantic water hosing experiments have shown that a drastic slowed-down AMOC caused abrupt cooling in the North Atlantic during YD and Heinrich Stadials [Otto-Bliesner and Brady, 2010; Kageyama *et al.*, 2013], due to the reduced northward Atlantic heat transport. The cooling caused due to slowdown of AMOC quickly propagates zonally in the northern hemisphere through atmospheric advection of by the westerly winds [Clement and Peterson, 2008]. The imbalance due to the northern hemisphere cooling (southern hemisphere warming) is compensated by generating anomalous energy transport from southern hemisphere to northern hemisphere causing reorganisation of Hadley circulation that involves northward cross-equatorial flow in the upper branch accompanied by southward flow in the

lower branch. The reorganization of the Hadley circulation is associated with westerly, low level wind anomaly over the south equatorial Indian Ocean weakens the southeast trade winds crossing the Equator turns into weaker south westerly causes reduction of upwelling in the western Arabian Sea. Whereas the strengthening of eastward winds over the equator, a unique feature in the Indian Ocean responsible for warm equatorial, also contributes to the warming in the eastern tropical Indian Ocean including BoB. Thus, the reorganisation of Hadley circulation as a result of the slowdown of AMOC caused warming in the northern Indian Ocean during the YD.

A number of studies, particularly from the Arabian Sea, link the cold episodes in the North Atlantic to weakened Indian and Asian summer monsoons at millennial time scales during the last glacial period [e.g. *Schulz et al.*, 1998; *Burns et al.*, 2003]. Although a wide range of mechanisms have been offered to explain the teleconnections between the North Atlantic cooling and tropical Indian Ocean hydroclimate, e.g. monsoon weakening in response to regional sea surface cooling [*Tierney et al.*, 2008; *Stager et al.*, 2011], changes in the monsoon intensity [*Griffiths et al.*, 2009; *Mohtadi et al.*, 2011] associated with a southward shift in the mean [*Lewis et al.*, 2011] or winter position of the ITCZ [*Muller et al.*, 2012; *Mohtadi et al.*, 2011]. Integrated paleoclimate data and model study revealed that drastic changes in the tropical Indian Ocean climate occurred as a response to the AMOC slowdown during Heinrich stadials and the YD, which involves a reorganization of the Hadley circulation with a southward shift of the ITCZ across the entire equatorial Indian Ocean [*Mohtadi et al.*, 2014].

Recently, it has been proposed that SST cooling of the Indian Ocean is the link between the Indian monsoon and North Atlantic cold climate intervals (such as the YD) that causes the reduction in the monsoon strength [*Tierney et al.*, 2015]. Although Indian summer monsoon variability is remotely forced by the north Atlantic temperature changes mediated through the meridional troposphere temperature gradient [*Goswami et al.*, 2006], our results show that the Northern Indian Ocean SST in general, and BoB SST in particular did not cool during the YD rather showing warming which is attributed to the slowdown of AMOC. Furthermore, recently it has been suggested that Indian Ocean warming weakens the strength of the Indian Summer Monsoon [*Roxy et al.*, 2015]. Reduction of Indian summer monsoon and warming BoB during the YD supports the hypothesis of *Roxy et al* [2015].

6. Conclusions

We reconstructed SST records based on shell Mg/Ca of *G. ruber* from the BoB sediment core, and compared it with other published Mg/Ca-derived SST records from the Northern Indian Ocean. Our

data show that the BoB warmed up to $\sim 1.8^{\circ}\text{C}$ during the YD period which is consistent with the other records. We hypothesise that the AMOC slowdown due to fresh water influx to the North Atlantic during the YD caused an overall cooling in the North Atlantic and a coeval warming in the tropical Indian Ocean. The warming of the BoB during the YD coincides with a southward shift of ITCZ that result in a reduction of the Indian Summer Monsoon, and suggests that recent evidences of Indian Ocean warming may have a profound negative influence on the Indian summer monsoon rainfall.

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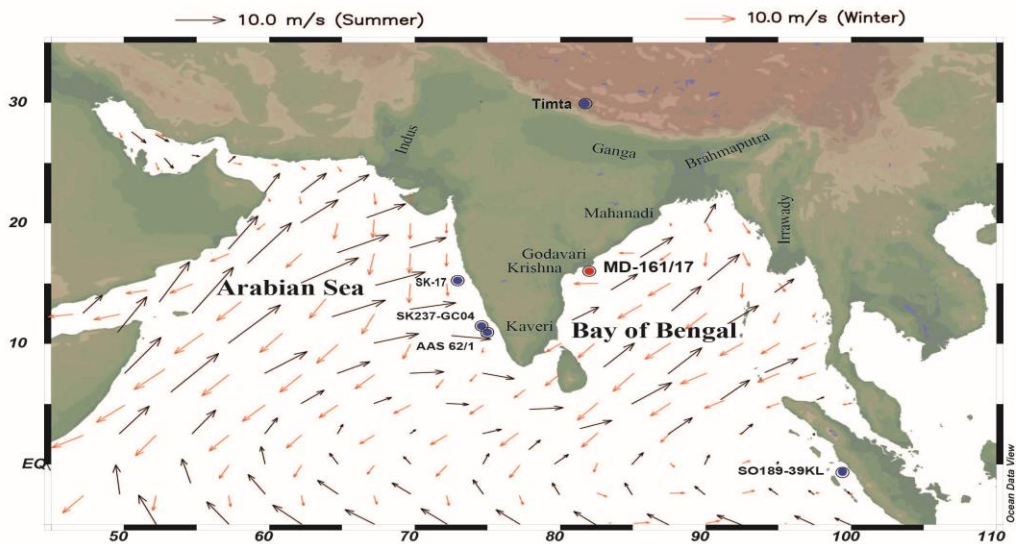


Figure 1. A schematic map of surface winds during summer and winter seasons in the northern Indian Ocean and the location of core MD 161/17 (red dot) along with other cores and caves locations (blue dots) compared in the study. Base map is generated using Ocean Data View [Schlitzer, 2016].

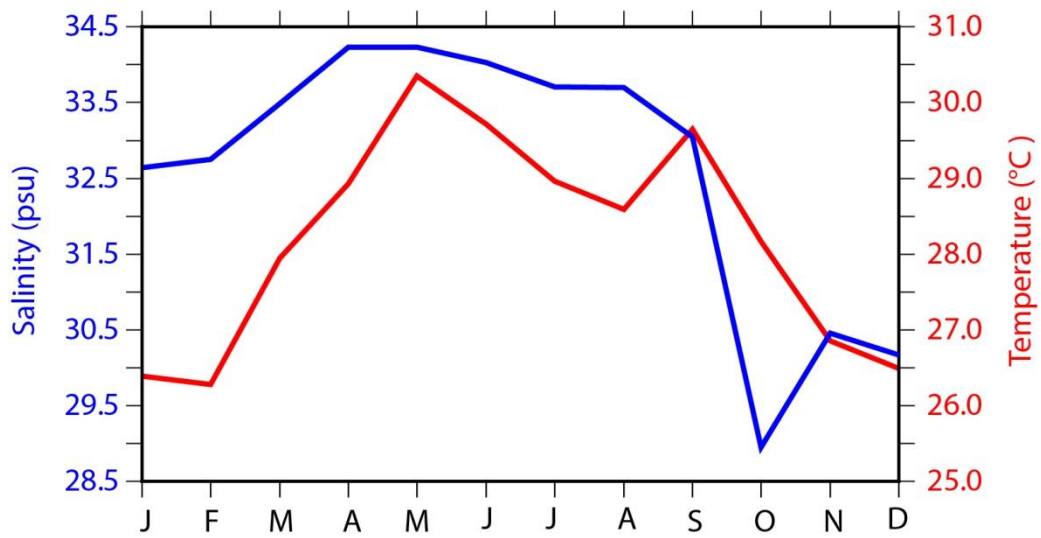


Figure 2. Monthly changes in the modern sea surface salinity (psu) and temperature at the MD 161/17 core location in the Bay of Bengal [*Locarnini et al.*, 2013].

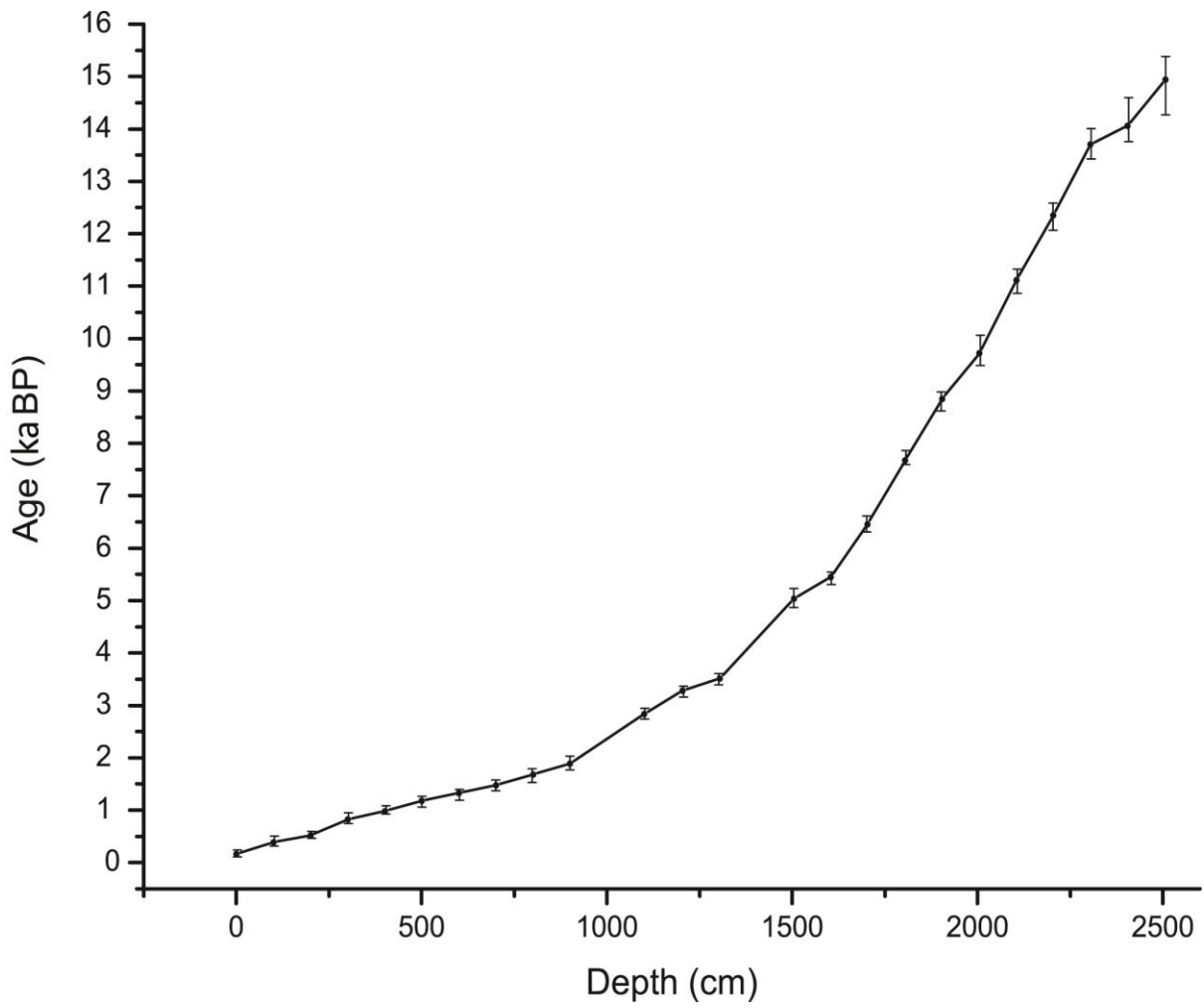


Figure 3. Chronology of the MD161/17 core based on AMS ^{14}C dates obtained on mixed planktonic foraminifera species. ^{14}C dates are calibrated to calendar years and plotted against depth of the core with 2σ error

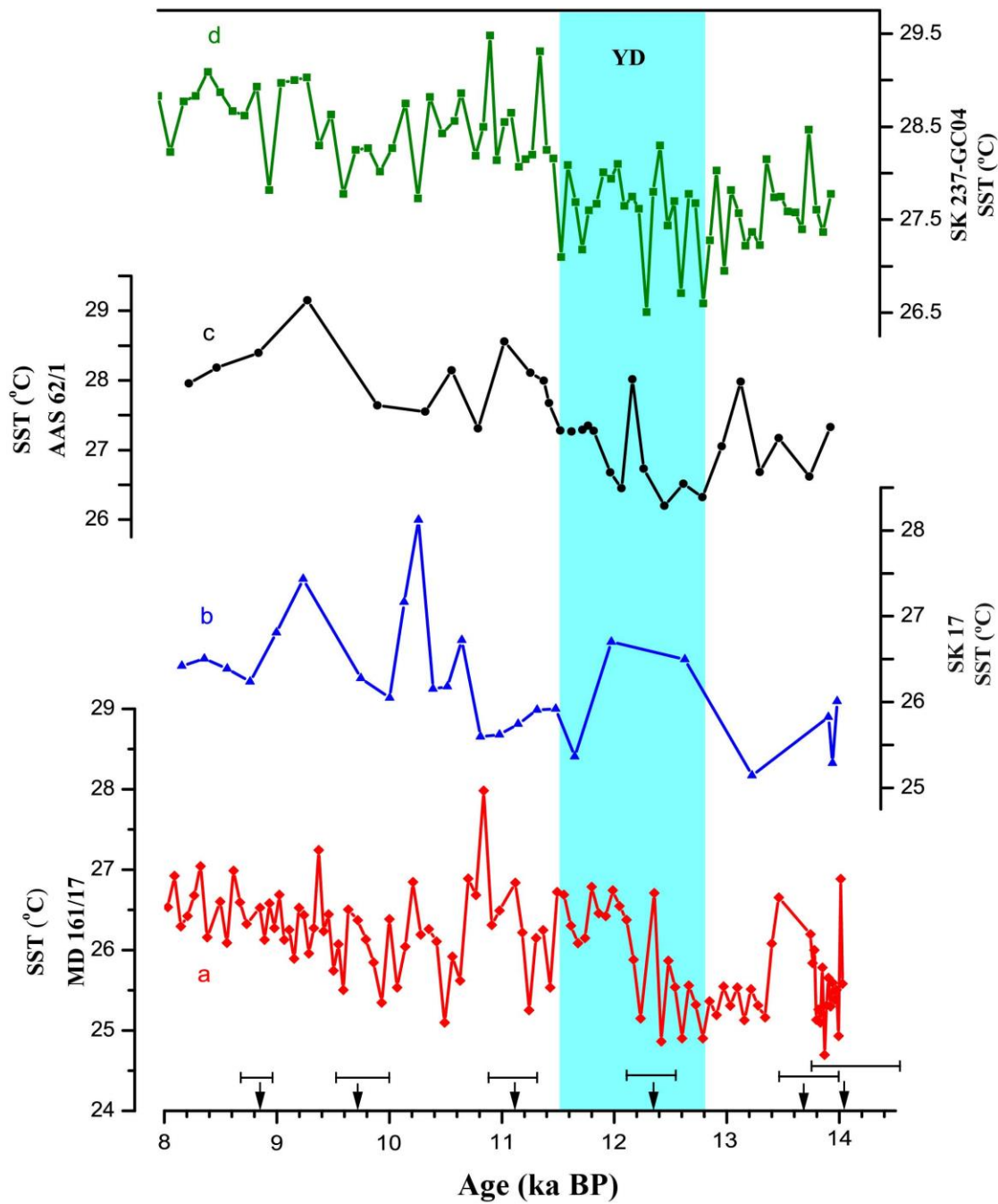


Figure 4. Variations in sea surface temperature (SST) records: (a) MD 161/17 (present study); (b) SK 17 [Anand *et al.*, 2008]; (c) AAS 61/1 [Kessarkar *et al.*, 2013]; and (d) SK237-GC04 [Saraswat *et al.*, 2013]. Light greenish band denotes the YD interval (12.8 - 11.5 ka). AMS ^{14}C date controls points and 2σ error bars are shown with arrows pointed to X axis.

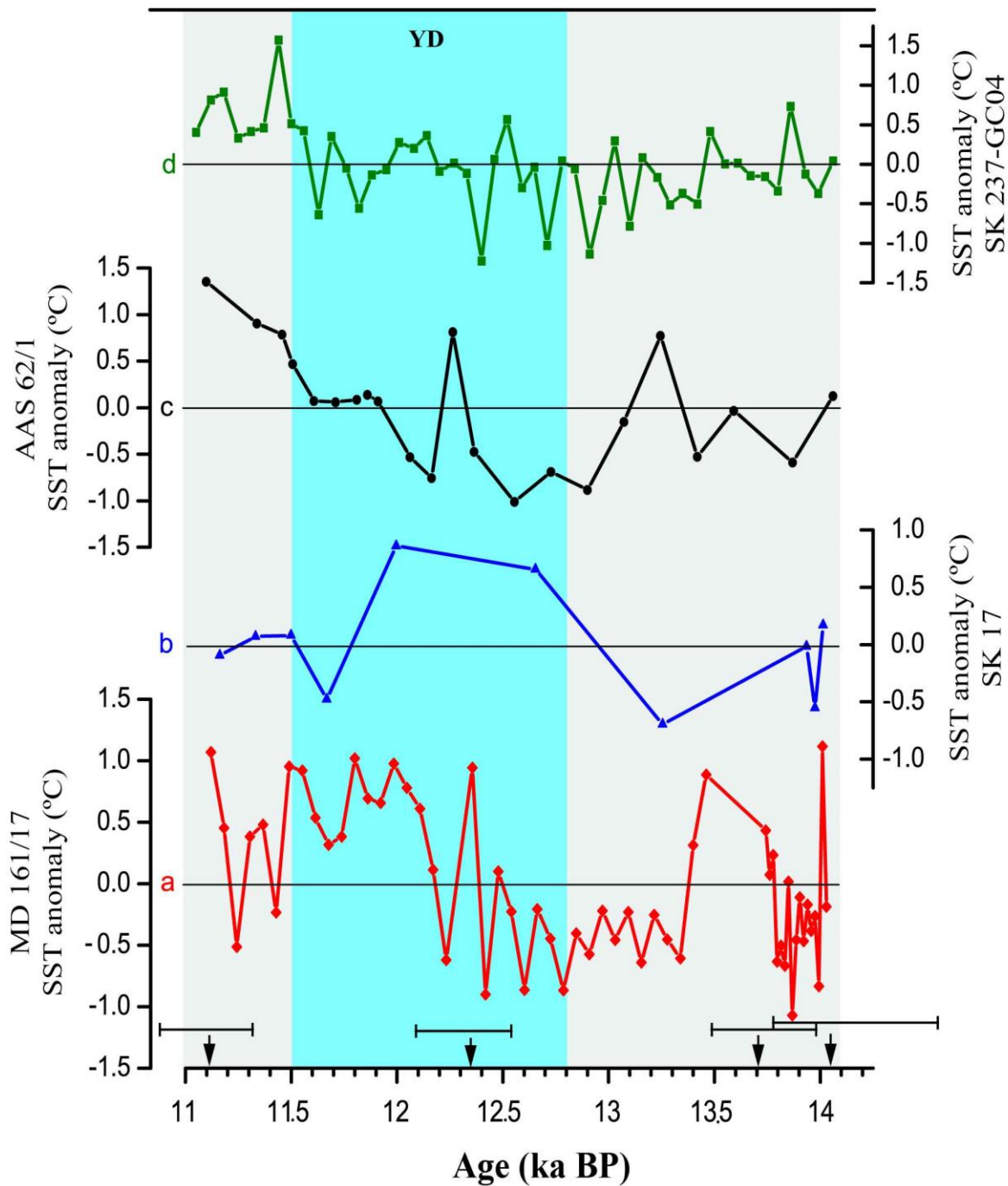


Figure 5. Variations in the SST anomalies observed for the time period 11-14 ka in the northern Indian Ocean cores: (a) MD 161/17 (present study), (b) SK 17 [Anand *et al.*, 2008], (c) AAS 62/1 [Kessarkar *et al.*, 2013], and (d) SK 237-GC04 [Saraswat *et al.*, 2013]. Light greenish band represents the YD interval. AMS ¹⁴C date controls points along with 2 σ error bars are shown with arrows pointed to X axis.

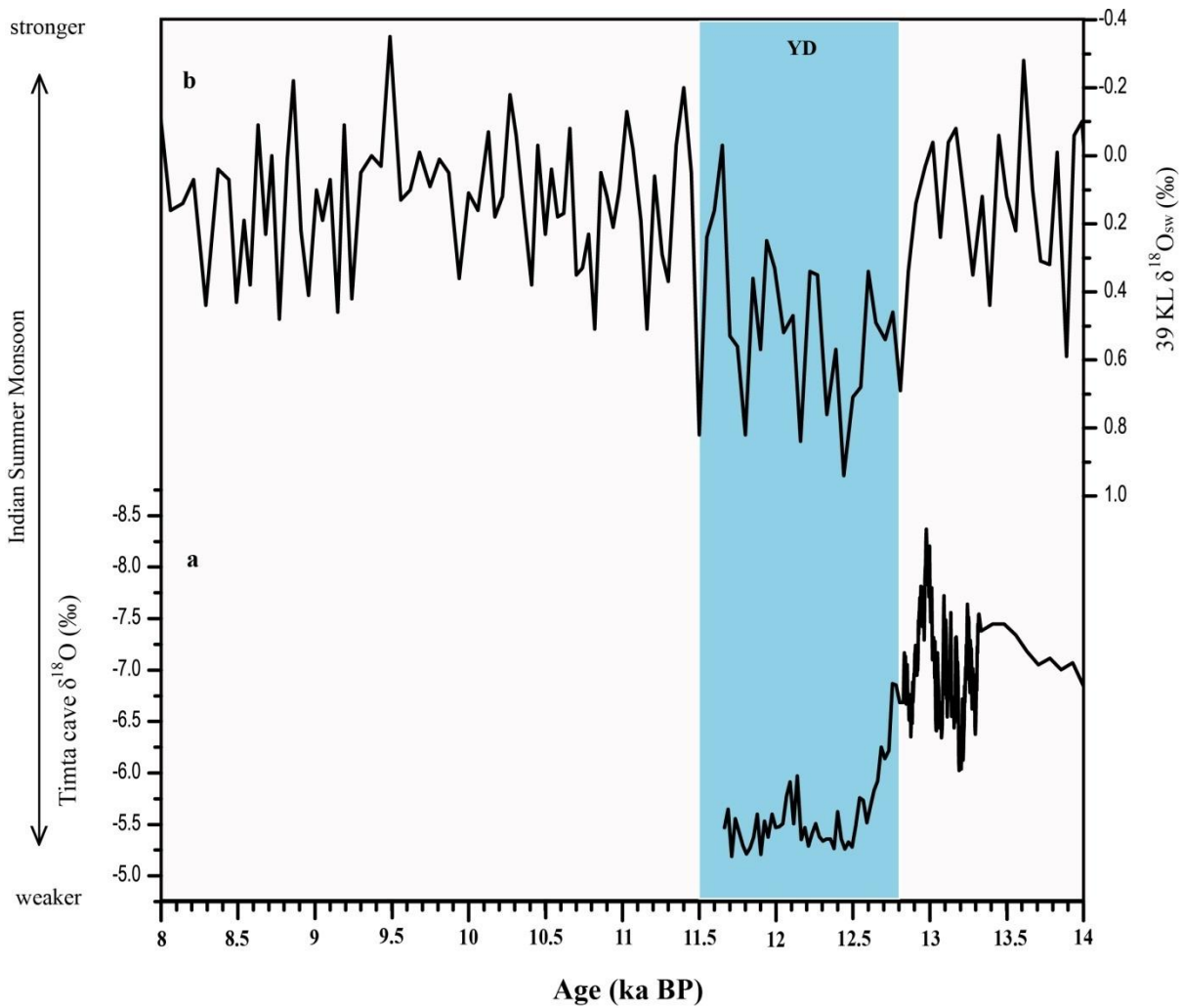


Figure 6. (a) $\delta^{18}\text{O}$ records of speleotherm from the Timta Cave from northern Indian region [Sinha *et al.*, 2005]; (b) $\delta^{18}\text{O}_{\text{sw}}$ record from the eastern Indian Ocean [Mohtadi *et al.*, 2014]. Reduction of Indian summer monsoon during the YD was documented in both terrestrial and marine records highlighted in light blue.