

Ecology of foraminifera from the southwestern Bay of Bengal and its application to reconstruct monsoon seasonality during the Quaternary

A THESIS SUBMITTED IN PARTIAL FULFILLMENT FOR THE DEGREE OF

DOCTOR OF PHILOSOPHY

in Marine Science

School of Earth, Ocean and Atmospheric Sciences
GOA UNIVERSITY



By

Mohd Salman

Micropaleontology Laboratory, Geological Oceanography Division,
CSIR-National Institute of Oceanography, Goa, India

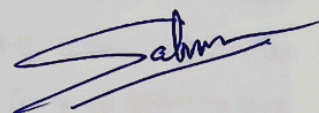
January 2026

DECLARATION

I, Mohd Salman, hereby declare that this thesis entitled "**Ecology of foraminifera from the southwestern Bay of Bengal and its application to reconstruct monsoon seasonality during the Quaternary**", represents work which has been carried out by me and that it has not been submitted, either in part or full, to any other University or Institution for the award of any research degree.

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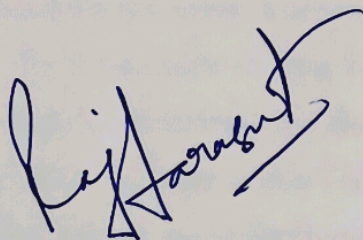
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Mohd Salman

CERTIFICATE

I hereby certify that the above declaration by the candidate, **Mohd Salman**, is true and the work was carried out under my supervision.



Dr. Rajeev Saraswat

Senior Principal Scientist

Micropaleontology Laboratory, Geological Oceanography Division,

CSIR-National Institute of Oceanography, Dona Paula, Goa.

Acknowledgment

Completing a Ph.D. is often likened to a journey, and mine has been one marked by the guidance of mentors who charted the course, the unwavering love and support of my parents, who have been my compass and anchor throughout, and the faithful support of those who walked beside me, unseen.

First and foremost, I am deeply indebted to my supervisor, [Dr. Rajeev Saraswat](#) (Senior Principal Scientist), for more than just academic guidance. His patience in the face of chaos, sharp eye for detail, and insistence on rigor helped me shape not only this thesis but also my identity as a researcher. He taught me the fundamentals of paleoceanography and paleoclimate with special reference to micropaleontology. I am thankful to him for correcting my research articles and reviewing my thesis. He always asked me to push my limits and I am grateful to work under his guidance.

To my advisory committee members [Senior Prof. Chandrashekher U. Rivonkar](#), [Prof. Vishnu Murty Matta](#), and [Assistant Prof. Niyati Kalangutkar](#), School of Earth, Ocean and Atmospheric Sciences, Goa University, Goa, thank you for your constructive critiques and unwavering support. Each of you brought a unique lens to my work, sharpening it with insight and depth I couldn't have achieved alone.

Behind every page of this thesis is the silent strength of my family. To my parents ([Mr. Asghar Husain](#) and [Mrs. Kulsum Jahan](#)), whose dreams I carry with me like heirlooms—I owe you more than words. Your sacrifices were the wind beneath my sails when waters turned rough. To my siblings, for grounding me in laughter and reminding me that life exists beyond deadlines and data.

I would like to express my sincere gratitude to [Dr. Sunil Kumar Singh](#), Director of the National Institute of Oceanography, for granting access to research facilities essential for my study. I express my sincere gratitude to [Senior Prof. Harilal B Menon](#), the Vice-Chancellor of Goa University, the Dean of the School of Earth, Ocean and Atmospheric Sciences, and the dedicated staff, particularly [Ms. Mruna](#), of the Marine Science Department at Goa University, for their invaluable assistance and administrative support. I thank the entire Human Resource Management staff of the NIO for their support during the doctoral work.

My special thanks are due to [Mr. Shashikant Velip](#) and the entire ship crew for their invaluable assistance during sample collection aboard the cruise ORV *Sindhu Sadhana* (SSD067). I also appreciate the support from [Mrs. Teja Ankush Naik](#), [Mrs.](#)

Supriya Gauresh Karapurkar, Mrs. Anita Garg, and Dr. C. Prakash Babu for their help with the CO₂ Coulometer and CN analyzer used in elemental analysis. Additionally, I am grateful to Mr. Areef A. Sardar for providing the Scanning Electron Microscope (SEM) imaging facility and support.

I express my sincere gratitude to Dr. Pankaj Kumar and Mr. Rajveer Sharma for their assistance in analyzing radiocarbon samples at the Inter-University Accelerator Centre (IUAC), New Delhi, India. I also thank Prof. Abhayanand Singh Maurya and Assistant Prof. Dharmendra Pratap Singh for providing the Isotope Ratio Mass Spectrometer facility to analyze the stable isotopic composition of foraminifera. My special thanks to Rahul Pawar for his invaluable support in stable isotope analysis. Additionally, I am grateful to Soni Rai and Nitish Kumar for their kind support during the analysis at the Indian Institute of Technology, Roorkee.

I also acknowledge the Joint Council of Scientific and Industrial Research-University Grant Commission for providing me with Junior and Senior Research Fellowships (JRF & SRF). I am also thankful to Dr. Rajeev Saraswat for providing me with financial support to continue my research work under the project FIR050301. I thank the Past Global Changes (PAGES) for providing me with a travel grant that enabled me to attend the Young Scientist Meeting 2025 and Open Science Meeting 2025.

I extend my heartfelt gratitude to Dr. Rajiv Nigam for his unwavering support and inspiring words, which served as a constant source of motivation throughout my research journey. I am thankful to Dr. Thejasino Suokhrie for always being a polite and guiding hand. I thank Dr. Shabnam Chaudhary for her kind support during the last few years of my PhD journey. A special thanks to one and only Dr. Sudhir Bhadra for his great help in identification of planktic foraminifera species and teaching the basics of Ocean Data View, Grapher and Surfer software. I thank Dr. Amrata Kaithwar for selling her scooter to me, which enabled late-night visits to work in the lab. I also extend my sincere gratitude towards Dr. Sujata Raikar for helping with foraminifera identification and giving a helping hand whenever needed. I sincerely thank Dr. Syed Mohammad Saalim and Dr. Dharmendra Pratap Singh for their invaluable support throughout my PhD journey. I am thankful to Mr. Perfectionist Shripad Bandodkar for his guidance during the early stages of this journey, especially for his precise instruction on subsampling the marine sediment cores. A special thanks to Ranjit N for his great help in sample processing and benthic foraminifera picking. My deepest gratitude goes to all my lab mates, Sudhir Bhadra, Gavendra Kumar, Deepek RP, Rinu Fathima, Anjali Kumari, Dawood Shiekh, Abhishek

Pradhan, Rajesh Kumar, Ranjit N, Suman Kumari, Manvi Tiwari, and all the dissertations and interns for their help, caffeinated debates, and shared victories. Science is often solitary, but you all made it collaborative and deeply human.

To all my closest friends in NIO, thank you for the reminders that I am more than this PhD. Thank you for answering late-night calls, believing in me when I doubted myself, and celebrating the little milestones that others may not have noticed. Many thanks to my best buddy Mr. Mohd Sadique for sharing the room, kitchen, and the best moments of my PhD journey at NIO. Sadique is the only person with whom I ever had a healthiest discussion about science. He has critical thinking about science that I always enjoyed. Once again, I thank Sadique for being a part of this journey. I am lucky and grateful to have a friend like Ahmad Shafi, who helped me a lot to clear the NET-JRF by allowing me to stay at his room for the preparation. He has a good collection of books related to Geology and Oceanography that help me a lot in qualifying the JRF.

I am thankful to my seniors at CSIR-NIO, Dr. Pankaj Kumar and Dr. Ramanand Yadav for their guidance during the initial days of this journey. I thank Dr. Firoz K. Badesab for his moral support.

Lastly, to the quiet moments that held me, the sunset and beach walks, and music that echoed in my headphones during hours of microscopy and writing, thank you. In your own way, you helped me to finish this chapter of life.

Mohd Salman

Preface

Global warming, attributed to an increase in atmospheric greenhouse gases and associated changes in the last 200 years (Feely et al., 2004), is affecting the world's climate. Therefore, the impacts of global warming on the Earth's habitats have to be assessed for the future sustenance of life. In the tropical regions, monsoon is one of the most important climatic phenomena as it sustains the livelihood of a huge population. The northern tropical Indian Ocean and adjoining landmasses are influenced by the seasonal reversal of monsoon wind. The seasonally reversing winds profoundly affect the socio-economic conditions of one of the most densely populated regions of the world (Mall et al., 2006). Monsoon in this region comprises of two distinct phases, viz. summer and winter monsoon, with over 80% of the average precipitation in the Indian subcontinent during the summer monsoon (Gadgil, 2006). Global warming will alter the land-ocean thermal contrast. Therefore, it is important to understand the possible implications of global warming on monsoon and associated processes. The past records of the monsoon intensity and other associated processes during different boundary conditions can help to predict the implications of global warming in the future. A lot of studies have reconstructed the changes in summer monsoon intensity during the past. However, little information is available about the behavior of the winter monsoon. The south-eastern coast of India receives substantial rainfall during the winter monsoon. Therefore, the southwestern Bay of Bengal is a good site to reconstruct the northeast monsoon intensity. Additionally, the region is also the conduit for the exchange of water between the southwestern Bay of Bengal and the southeastern Arabian Sea (Jensen, 2001). Therefore, this region can provide information about the winter monsoon intensity in the past and also the variation in the cross basin water exchange.

The paleoclimate studies rely on proxies, including sedimentological, geochemical, and paleontological evidence, to reconstruct the past climatic conditions. Among these, marine microfossils such as foraminifera are invaluable. These single-celled organisms, usually characterized by their protective shells, are highly sensitive to environment parameters like temperature, salinity, and oxygen levels (Boltovskoy et al., 1991; Nigam et al., 2007). Planktic and benthic foraminifera provide insights into both upper and bottom water column conditions. Additionally, benthic foraminifera serve as indicator of oxygen deficient zone (ODZ) and monsoon-driven freshwater influx, making them critical tools to reconstruct the past climate (Jorissen, 1995; Nomaki et al., 2005).

This doctoral work focused on reconstructing the past intensity of summer and winter monsoon, productivity, and ODZ during the different boundary conditions of late Quaternary, using the multiproxy approach from the southwestern Bay of Bengal. The main aim of the work was to understand the impacts of monsoon and oxygen deficiency, provide a comprehensive understanding of monsoon variability and its implications in a warming world. The work is compiled in eight chapters. The details of each chapter are given below.

Chapter 1 includes the rationale of the work as well as a review of the previous studies. The studies on ecological preferences of foraminifera and paleoclimatic reconstruction from the Bay of Bengal were reviewed to understand the current status and future directions. From the literature review, it was clear that although a good number of studies have been done on the distribution of foraminifera from the southwestern Bay of Bengal, most of the studies are confined to shallow regions like beaches and estuaries or continental shelf (Khare et al., 2007). The existence of the oxygen-deficient zone and its effect on benthic foraminifera in the southwestern Bay of Bengal has also not been addressed. The existing paleoclimatic records from this region have a relatively coarse resolution and cover only a short duration (Khare et al., 2007). The southwestern Bay of Bengal is a good site to reconstruct the northeast monsoon because it receives more fresh water flux during the northeast monsoon. Based on the above mentioned observations, the following research objectives were finalized.

- To understand the effect of ambient physico-chemical conditions, especially the oxygen deficient zone, on foraminifera.
- To reconstruct the changes in northeast monsoon intensity during the late Quaternary.
- To reconstruct the paleoproductivity changes in the southwestern Bay of Bengal.
- To reconstruct the oxygen deficient zone intensity during the late Quaternary.

The details of the modern oceanographic conditions in the Bay of Bengal, material used and the methodology followed for the doctoral work are presented in **Chapter 2**. The Bay of Bengal is characterized by strong seasonal changes driven by monsoon winds (Schott and McCreary, 2001). The major rivers like Brahmaputra, Ganga, Mahanadi, Godavari, Krishna, and Cauvery discharge a large quantity of fresh water into the bay, creating a stratified low salinity upper layer throughout the year (Krishnamurthy and Shukla, 2000). The extensive cloud cover, sediment load, and narrow shelf result in low

productivity in the Bay of Bengal (Qasim, 1977; Gomes et al., 2000). Although, coastal upwelling and winter convection are weak, but episodic events like eddies and gyres make the region locally productive (Jayalakshmi et al., 2015).

The surface and sub-surface samples collected during the 67th cruise of RV *Sindhu Sadhana* (SSD067) were used. A total of 56 multicores, 11 grabs, and one piston-core collected from the south-western Bay of Bengal (swBoB) were used. The dried coarse fraction (CF) (>63 µm) of rose-Bengal stained sediments was used to pick living and dead benthic foraminifera. The planktic foraminifera were picked from >125 µm fraction. For carbon and nitrogen analysis, the freeze-dried sediment were powdered with the help of agate mortar and pestle. The organic carbon (C_{org}) was estimated by subtracting the inorganic carbon from the total carbon. The isotopic composition of *Globigerinoides ruber*, a mixed layer and *Neogloboquadrina dutertrei*, a thermocline dwelling species was analyzed by using MAT 253 Plus Isotopic Ratio Mass Spectrometer at the Indian Institute of Technology, Roorkee. The chronology of the core was established, by using thirty-one accelerator mass spectrometer (AMS) radiocarbon dates analyzed at the Inter-University Accelerator Center, New Delhi, India.

The modern distribution and ecological preferences of planktic and benthic foraminiferal species and their contribution to the carbon burial in the southwestern Bay of Bengal are included in **Chapter 3**. The planktic foraminiferal abundance varied from 3 to 12060 specimen/g sediment, with the highest abundance (>12060 specimen/g sediment) observed along the southernmost transect. A total of 31 planktic foraminiferal species were found in the surface sediments. A very high CaCO₃ content and low planktic foraminiferal abundance on the shelf (<150 m) suggests that terrestrial input, meio and macrofauna contribute a substantial fraction of the CaCO₃ in the shelf sediments. The southernmost transect exhibited high TC, CaCO₃, and planktic foraminifera abundance, indicating their prominent role in carbon burial in this region. As compared to the northwestern and westcentral Bay of Bengal, planktic foraminiferal abundance was approximately four times higher in the swBoB.

The living benthic foraminifera (LBF) were used to understand the effect of ambient environmental parameters on various morphogroups. The highest abundance of LBF was in front of the Pennar River. The abundance of living agglutinated benthic foraminifera increased with depth but the living calcareous benthic foraminifera abundance decreased with increasing depth. The highest abundance of living calcareous angular benthic foraminifera was on the upper slope region within the zone of lowest

dissolved oxygen concentration. The living calcareous angular benthic foraminifera prefer the low dissolved oxygen, and the living calcareous rounded benthic foraminifera prefer the high dissolved oxygen concentration.

A decadal resolution marine sediment-based record of the precise onset and duration of monsoon and associated processes in the western Bay of Bengal during the 4.2 ka event is presented in **Chapter 4**. An abrupt decrease in carbon burial began at 4.65 ka and continued until 4.1 ka. The synchronous decrease in coarse fraction ($>63\ \mu\text{m}$), CaCO_3 , C_{org} and nitrogen (N) suggests a collapse in productivity. Therefore, it is clear that the changes during the 4.2 ka event lasted for 550 years and resulted in a significant decrease in carbon burial due to weaker monsoon in the swBoB. A subtle decrease in *Globigerina bulloides* abundance suggests reduced upwelling and thus the relatively weaker summer monsoon between 4.1 and 4.45 ka. An abrupt decrease in *G. ruber* at 4.45 ka suggests the cooler sea surface temperature (SST) between 4.1 and 4.45 ka.

The seasonal preferences of species were used to reconstruct the monsoon seasonality. The summer monsoon assemblage (SMA) included *G. ruber*, *G. bulloides*, *N. dutertrei*, *Trilobatus trilobus*, *T. sacculifer*, *G. siphonifera*, and *N. incompta* while winter monsoon assemblage (WMA) included *G. glutinata*, *G. falconensis*, *G. menardii* and *P. obliquiloculata*. The SMA abundance was high before 4.65 ka, decreased significantly from 4.65 to 4.1 ka, confirming the initiation of the 4.2 ka event at 4.65 ka and a weaker summer monsoon during the 4.2 ka event. However, the higher WMA abundance suggests an intensified winter monsoon, starting at 4.45 ka instead of 4.65 ka. A weaker summer monsoon and relatively stronger winter monsoon resulted in the decreased carbon burial during the 4.2 ka event.

Globigerinoides ruber and *N. dutertrei* live at different depths in the water column. Therefore, their oxygen isotopic values have been used to understand the monsoon intensity and water column stratification. The $\delta^{18}\text{O}_{G. ruber}$ was depleted from 5.1 to 4.7 ka suggesting a stronger monsoon. However, it was enriched at 4.6 ka confirming the weakening of the monsoon and initiation of the event at 4.6 ka. The most enriched $\delta^{18}\text{O}_{G. ruber}$ was between 3.95 and 4.15 ka. Interestingly, the $\delta^{18}\text{O}_{N. dutertrei}$ showed an opposite trend, unlike the $\delta^{18}\text{O}_{G. ruber}$. The difference between $\delta^{18}\text{O}_{G. ruber}$ and $\delta^{18}\text{O}_{N. dutertrei}$ was less during the 4.2 ka event, unlike before the 4.6 ka, suggesting that the sea was less stratified during the 4.2 ka event. However, the enriched value of $\Delta\delta^{18}\text{O}$ suggests highly mixed surface water conditions between 3.95 and 4.15 ka.

Based on the sediment geochemistry and foraminiferal data, a distinct change was inferred at 8.6 ka that continued until 7.8 ka, matching with the 8.2 ka event. The response of monsoon and associated processes to this 8.2 ka event is discussed in **Chapter 5**. The event lasted for 800 years in this region. The abrupt decrease in CF, CaCO₃, C_{org} and nitrogen (N) suggests a decrease in productivity during the 8.2 ka event. So far, there is no report on the initiation of 8.2 ka event at 8.6 ka and duration of 800 years from the Bay of Bengal. The decrease in *G. bulloides*/*G. ruber* ratio also suggests low productivity or oligotrophic conditions during the 8.2 ka event. The decrease in mixed layer eutrophic species (MLES), CaCO₃, C_{org}, and increase in mixed layer oligotrophic species (MLOS) also suggest low productivity during the 8.2 ka event. The summer monsoon assemblage was relatively less abundant during the 8.2 ka event suggesting a weaker summer monsoon. Although the WMA responded opposite to the SMA, it also decreased gradually from 8.7 to 6.9 ka, suggesting a weaker winter monsoon during the 8.2 ka event. However, the winter monsoon was stronger before the 8.6 ka. The difference in the stable oxygen isotopic ratio of *G. ruber* and *N. dutertrei* was also used to understand the monsoon and water column stratification during the 8.2 ka event. The $\delta^{18}\text{O}_{G. ruber}$ was depleted between 7.8 and 8.6 ka suggesting a stronger monsoon instead of weaker monsoon. However, it was enriched before and after the event confirming the initiation of the 8.2 ka event at 8.6 ka instead of 8.2 ka.

The core also contained a very well resolved section of the last glacial maximum (LGM, 19-23 kyr). It ranged from ~1250–1500 cm, with an average sedimentation rate of 55.6 cm/kyr. The resolution was ~18 yr/cm. Large scale changes were observed in all the parameters during this interval, with periodic lows and highs. The details of the climatic conditions during the last glacial maximum are included in **Chapter 6**.

The details of the variation in the intensity of oxygen deficient zone since the last glacial maximum are included in **Chapter 7**. The change in CaCO₃, C_{org}, C_{org}/N, total benthic foraminifera number (TBN/g sediment) and their morphogroups were used to investigate the variability of the oxygen deficient zone. During the LGM and Heinrich Stadial 1, reduced summer monsoon intensity and enhanced winter monsoon favoured moderately oxygenated bottom water conditions, as indicated by benthic foraminifera morphogroups and productivity markers. Despite of the reduced inflow of oxygen-rich Persian Gulf Water and Red Sea Water during LGM, weaker summer stratification, reduced productivity, influx of oxygenated water due to stronger winter monsoon, and reduced influx of oxygen-poor Arabian Sea water contributed to moderately oxygenated

conditions. With the onset of deglaciation, the intensified monsoon and increased productivity led to enhanced organic carbon flux to the seafloor, triggering bottom water deoxygenation and strengthening of the ODZ, particularly during the Bolling-Allerod, Younger Dryas and Greenlandian Age. The Meghalayan Age also experienced an intense ODZ due to high surface productivity. In contrast, periods like 8.2 ka and 4.2 ka events and the Northgrippian Age were marked by weak monsoons, reduced productivity, and thus well-oxygenated bottom waters.

The summary and future scope of the doctoral work are included in **Chapter 8**. The references cited in the thesis are listed at the end.

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List of abbreviations

Abbreviations	Full form
AABW	Antarctic Bottom Water
ABF	Agglutinated Benthic Foraminifera
AMOC	Atlantic Meridional Overturning Circulation
Angular _{BF}	Angular Asymmetrical Benthic Foraminifera
ASHSW	Arabian Sea High Salinity Water
BA	Bølling–Allerød
BoB	Bay of Bengal
CBF	Calcareous Benthic Foraminifera
CCA	Canonical Correspondence Analysis
CF	Coarse Fraction
C _{org}	Organic Carbon
DO	Dissolved Oxygen
eAS	Eastern Arabian Sea
EICC	East India Coastal Current
ENSO	El-Nino Southern Oscillation
GoM	Gulf of Mannar
GR	Grab
HS	Heinrich Stadial
IOD	Indian Ocean Dipole
ITCZ	Inter Tropical Convergence Zone
LABF	Living Angular Benthic Foraminifera
LGM	Last Glacial Maxima
LRBF	Living Rounded Benthic Foraminifera
MC	Multicore
MLES	Mixed Layer Eutrophic Species
MLOS	Mixed Layer Oligotrophic Species
N	Nitrogen
NADW	North Atlantic Deep Water
NH	Northern Hemisphere
ODZ	Oxygen Deficient Zone

OMZ	Oxygen Minimum Zone
PC	Piston Core
PGW	Persian Gulf Water
Rounded _{BF}	Rounded symmetrical Benthic Foraminifera
RSW	Red Sea Water
seAS	Southeastern Arabian Sea
SMA	Summer Monsoon Assemblage
SMC	Summer Monsoon Current
Spherical _{BF}	Spherical Benthic Foraminifera
SSS	Sea Surface Salinity
SST	Sea Surface Temperature
STIO	Southeastern Tropical Indian Ocean
swBoB	Southwestern Bay of Bengal
TBN	Total Benthic foraminifera Number
TC	Total Carbon
TIC	Total Inorganic Carbon
TPN	Total Planktic foraminifera Number
wBoB	Western Bay of Bengal
WMA	Winter Monsoon Assemblage
WMC	Winter Monsoon Current
WTIO	Western Tropical Indian Ocean
YD	Younger Dryas

Chapter 1

Introduction

1.1 Rationale

The present global warming is attributed to an increase in atmospheric greenhouse gases and associated changes in the last ~200 years (Summerhayes et al., 2024; IPCC, 2023; Feely et al., 2004). Global warming is affecting the world's climate (IPCC, 2023; Hansen et al., 2023). Therefore, the impacts of global warming on the Earth's habitats have to be assessed for the future sustenance of life. In the tropical regions, monsoon is one of the most important climatic phenomena as it sustains the livelihood of a huge population (Fiaz et al., 2025; Buckley et al., 2014). Global warming will alter the land-ocean thermal contrast. Therefore, it is important to understand the possible implications of global warming on monsoon and its associated processes. The seasonal reversal of monsoon winds influences the northern tropical Indian Ocean and adjoining landmasses (Vinayachandran et al., 2002; Schott and McCreary, 2001). The seasonally reversing winds have a profound impact on the socio-economic conditions of one of the most densely populated regions of the world (Mooley et al., 1981; Mall et al., 2006). Monsoon in this region comprises of two distinct phases, viz. summer and winter monsoon, with over 80% of the average precipitation in the Indian subcontinent during the summer monsoon (Gadgil, 2006) (**Fig. 1.1**). However, the southeastern region of the peninsular India lies behind the Western Ghats, a mountain range that acts as a barrier to the moisture-laden southwest monsoon winds, resulting in a dry rain-shadow area on its leeward side (**Fig. 1**). As a result, agricultural activity in this region is highly reliant on the winter monsoon rainfall, which contributes nearly half of the total annual rainfall for southeastern peninsular India, Sri Lanka, Southeast Asia and adjacent marine areas (Sengupta and Nigam, 2019; Kumar et al., 2007). Therefore, understanding and predicting the seasonal variability of the winter monsoon is essential for effective management of water resources and agricultural planning in this region. The seasonal response of the summer and winter monsoons to multi-decadal climatic anomalies is not well explored during the Holocene (Banerji et al., 2020). A few studies suggest that it is linear, while others indicate that it is nonlinear throughout the entire Holocene (Kaboth-Bahr et al., 2021; Resmi and Achyuthan, 2018). Therefore, the dynamic behavior of the

monsoon seasonality needs to be studied in response to Holocene abrupt climatic anomalies.

The past records of the monsoon intensity and other associated processes during different boundary conditions can help to predict the implications of global warming in the future. A lot of studies have reconstructed the changes in summer monsoon intensity from the Bay of Bengal (Suokhrie et al., 2024; 2021; Gautam et al., 2021; Ankit et al., 2017; Da Silva et al., 2017; Phillips et al., 2014; Ponton et al., 2012; Rashid et al., 2011; Govil and Naidu, 2011; Kudrass et al., 2001). Many lake sediment, speleothem and pollen records have also been used to reconstruct the summer monsoon intensity during the Holocene from the Indian subcontinent (Dixit et al., 2018; Dutt et al., 2018; Prasad et al., 2014; Berkelhammer et al., 2012). However, little information is available about the behavior of the winter monsoon. The south-eastern coast of India receives substantial rainfall during the winter monsoon (**Fig. 1.1**). Therefore, the southwestern Bay of Bengal is a good site to reconstruct the northeast monsoon intensity. Additionally, the region is also the conduit for the exchange of water between the southwestern Bay of Bengal and the Southeastern Arabian Sea (Jensen, 2001; Vinayachandran et al., 1991). Therefore, this region can provide information about the winter monsoon intensity in the past and also the variation in the cross basin water exchange.

Several proxies (sedimentological, geochemical, paleontological) are used to reconstruct the past climate. The marine microfossils foraminifera are also frequently used to reconstruct the past climate (Saraswat, 2015; Katz et al., 2010). Foraminifera are single-celled, preferentially marine microorganisms with an external protective shell

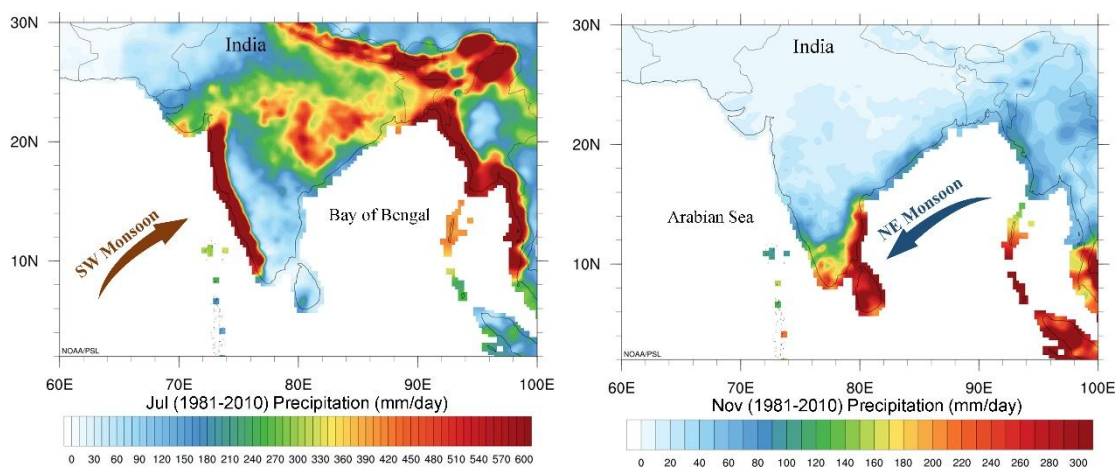


Figure 1.1: Seasonal monthly mean precipitation changes from 1981 to 2010 for July (Left) and November (Right) over the Indian subcontinent. Global Precipitation Climatology Centre (GPCC) dataset was used to create the precipitation maps (Schneider et al., 2014). Dark red arrow marks the direction of southwest monsoon winds and dark blue arrow represents northeast monsoon winds.

called a test (Saraswat, 2015, Nigam, 2005). These microorganisms are widely distributed in the world's ocean and found everywhere from the shallow coastal region to the deepest part of the ocean. They are extremely sensitive to the ambient environment (Nigam et al., 2007). Foraminifera are both benthic as well as planktic in habitat and thus provide information on both the bottom and upper water column conditions. Because of their sensitivity to environmental changes, they are widely used in paleoclimatic studies (Saraswat, 2015, Katz et al., 2010; Nigam, 2005). The abundance, diversity, and morphology of foraminifera is affected by environmental parameters such as temperature, salinity, depth, nutrients, and dissolved oxygen (Boltovskoy et al., 1991). The organic matter and dissolved oxygen are the dominant factors controlling the benthic foraminifera population, especially in the deep ocean (Singh et al., 2022; Jorissen, 1995). The different species of benthic foraminifera have different tolerance limits to oxygen. The preservation of organic matter is higher in low oxygen conditions and the organic matter uptake in such regions is dominated by benthic foraminifera (Singh et al., 2022; Nomaki et al., 2005). Therefore, benthic foraminifera can be used to reconstruct the intensity of the past oxygen minimum zones (OMZ). A particular type of benthic foraminifera are reportedly abundant in the oxygen-deficient zone of the Arabian Sea (Singh et al., 2022; Saalim et al., 2021; Suokhrie et al., 2021; Nigam et al., 2007). However, the freshwater influx due to monsoonal activity also affects the morphology of foraminifera (Nigam et al., 2009). The difference between the effects of dissolved oxygen and turbulence on benthic foraminifera morphology has to be assessed for its better applications.

1.2 Previous studies

Studies on the ecological preferences of foraminifera and paleoclimatic reconstruction from the Bay of Bengal were reviewed to understand the current status and future directions (Table 1.1).

Table 1.1: Major previous studies based on foraminifera from the southwestern Bay of Bengal and adjoining areas.

Sr. No.	Author	Year	Study Area	Major Findings
1	Cullen & Prell	1984	North Indian Ocean	Planktic foraminifera distribution and preservation in the north Indian Ocean.
2	Guptha et al.,	1997	Bay of Bengal	Documented the seasonal behavior of planktic

				foraminifera species through sediment trap study.
3	Cayre et al.,	1999	Equatorial Indian Ocean	Paleo-productivity in the equatorial Indian Ocean during the last 260,000 yr.
4	Burton & Vance	2000	Bay of Bengal	Glacial-interglacial variations in the neodymium isotope composition of seawater in the Bay of Bengal.
5	Kudrass et al.,	2001	Northern Bay of Bengal and Arabian Sea	Modulation and amplification of climatic changes in the Northern Hemisphere by Indian summer monsoon during the past 80 ky.
6	Barker et al.,	2005	World ocean data compilation	Planktic foraminiferal Mg/Ca as a proxy for past oceanic temperature and data compilation for the LGM.
7	Gandhi et al.,	2007	South east coast of India	Benthic foraminifera from the tsunamigenic sediments.
8	Rana et al.,	2007	Central east coast of India	Relict benthic foraminifera in surface sediments off central east coast of India as an indicator of sea level changes.
9	Khare et al.,	2007	Eastern coast of India, and Andaman and Nicobar Islands	An overview of foraminiferal studies in the nearshore regions off eastern coast of India, Andaman and Nicobar Islands.
10	Ahmad et al.,	2008	Bay of Bengal	Surface and deep water changes in the northeast Indian Ocean during the last 60 kyr.
11	Ota et al.,	2019	Western Bay of Bengal	Monsoon-influenced variations in plankton community structure and upper-water column stratification in the western Bay of Bengal during the past 80 kyr.
12	Rana & Nigam	2009	Western Bay of Bengal	Cyclicity in the late Holocene monsoon changes from the western Bay of Bengal.
13	Saraswat	2010	Eastern Arabian Sea and Northern Bay of Bengal	Secretion at different seawater pH controls the differential

				dissolution of biogenic carbonates.
14	Govil & Naidu	2011	Western Bay of Bengal	Variation of Indian monsoon precipitation during the last 32 kyr.
15	Rashid et al.,	2011	Bay of Bengal	Reconstructed the sea surface temperature and salinity from paired $\delta^{18}\text{O}$ and Mg/Ca measurements in planktic foraminifera <i>Globigerinoides ruber</i> and suggested that temperature and seawater $\delta^{18}\text{O}$ were $\sim 3^\circ\text{C}$ colder and $\sim 0.6\%$ depleted, respectively, during the last glacial maximum compared to the early Holocene.
16	Saraswat et al.,	2012	Southern tip of India	Linkage between seasonal insolation gradient in the Northern Hemisphere and salinity of the equatorial Indian Ocean during the last glacial period.
17	Bhonsale & Saraswat	2012	North-eastern Indian Ocean	Abundance and size variation of <i>Globorotalia menardii</i> in the northeastern Indian Ocean during the late Quaternary.
18	Ponton et al.,	2012	Bay of Bengal	Reconstructed the changes in aridification from core Indian monsoon region by using the carbon isotopes of sedimentary leaf waxes and oxygen isotopes of planktic foraminifer <i>Globigerinoides ruber</i> and suggested aridified conditions after 4000 years.
19	Satpathy et al.,	2013	Offshore of Tamil Nadu, Bay of Bengal	Severe oxygen depletion in the shallow regions of Bay of Bengal, off Tamil Nadu coast.
20	Gandhi et al.,	2014	Palk Bay, East Coast of India	The low salinity condition, less organic matter content, and fine sediment texture are the major factors controlling

				foraminiferal distribution in the Kodiyakarai region.
21	Symphonia & Nathan	2014	Southwestern Bay of Bengal	Benthic foraminifera distribution between Mahabalipuram-Marakkanam.
22	Symphonia & Nathan	2014	Southwestern Bay of Bengal	Benthic foraminifera distribution off Cuddalore.
23	Saraswat et al.,	2014	India and adjoining seas	A review of the Quaternary monsoon history.
24	Phillips et al.,	2014	Bay of Bengal	Reconstructed the variation in paleo-monsoon intensity and its effect on productivity and lithogenic fluxes during the last 110 kyr using total organic carbon, total nitrogen, CaCO ₃ , biogenic silica, $\delta^{13}\text{C}_{\text{org}}$, $\delta^{15}\text{N}$, bulk mineralogy from X-ray diffraction, bulk and lithogenic grain size distribution, magnetic susceptibility, bulk density, and Ca, Br, and Zr/Rb from X-ray fluorescence.
25	Jayalakshmi et al.,	2015	Southwestern Bay of Bengal	Response of micro and meso-zooplankton in the southwestern Bay of Bengal to a cyclonic eddy during the winter monsoon.
26	Manasa et al.,	2016	Northwestern Bay of Bengal	Assessing the suitability of benthic foraminiferal morpho-groups to reconstruct paleomonsoon from the Bay of Bengal.
27	Sijinkumar et al.,	2016	Bay of Bengal and Andaman Sea	$\delta^{18}\text{O}$ and salinity variability from the last glacial maximum to Recent.
28	Suokhrie et al.,	2017	Central-western Bay of Bengal	Indian monsoon variability during the last 2000 yr.
29	Azhar et al.,	2017	Arabian Sea and Bay of Bengal	OMZ contrasts between the Arabian Sea and Bay of Bengal implied by the difference in remineralization depth.
30	Gandhi et al.,	2017	Southeastern coast of India	Distribution of benthic foraminifera, sediment

				characteristics, and environmental conditions.
31	Saraswat et al.,	2017	Western Bay of Bengal	The highest abundance of <i>Asterorotalia trispinosa</i> is close to the outfall region of Ganga-Brahmaputra rivers and decreases away from the river mouths. <i>A. trispinosa</i> is rare in front of both the Cauvery and Pennar river outfall regions. The ecological preference of <i>A. trispinosa</i> suggests it as a potential paleomonsoon proxy, the restricted distribution implies limited application.
32	Singh et al.,	2017	Southeastern Arabian Sea	The vertical distribution of living benthic foraminifera to understand the relationship between ecological parameters and foraminiferal abundance.
33	Panmei et al.,	2017	Bay of Bengal	This record of sea surface temperature exhibits an overall warming of 1.8°C during the Younger Dryas and slowdown of Atlantic Meridional Overturning Circulation.
34	Saolim et al.,	2018	Mahanadi Basin, western Bay of Bengal	Ecological preferences of agglutinated benthic foraminiferal morphogroups.
35	Nagarajan & Nathan	2018	Southwestern Bay of Bengal	Benthic foraminiferal distribution and biofacies in the shelf part of the Bay of Bengal.
36	Ma et al.,	2019	Bay of Bengal	Changes in intermediate circulation in the Bay of Bengal since the last glacial maximum.
37	Panmei et al.,	2018	Bay of Bengal	Reconstructed the variability of terrigenous input to the Bay of Bengal for the last ~80 kyr using the variations in calcium carbonate and magnetic susceptibility, along with Ba, Ti, and Al.

38	Symphonia & Nathan	2019	Continental shelf-slope, southwestern Bay of Bengal	Taxonomy of recent benthic foraminifera.
39	Gautam et al.,	2019	Southwestern Bay of Bengal	Late Quaternary Indian monsoon variability record from the southwestern Bay of Bengal.
40	Loveson & Nigam	2019	East coast of India	Reconstructed late Pleistocene and Holocene sea level curve.
41	Saraswat et al.,	2019	Southeastern Arabian Sea	Indonesian throughflow controlled the westward extent of the Indo-Pacific Warm Pool during the glacial-interglacial intervals.
42	Devendra et al.,	2019	Northeastern Indian Ocean	Paleoproductivity and associated changes in the north-eastern Indian ocean since the last glacial.
43	Ma et al.,	2019	Bay of Bengal	Reconstructed the changes in intermediate circulation in the Bay of Bengal since the last glacial maximum as inferred from benthic foraminiferal assemblages and geochemical proxies.
44	Suokhrie et al.,	2020	Bay of Bengal and the Arabian Sea	Lack of denitrification causes a difference in benthic foraminifera living in the oxygen deficient zones of the Bay of Bengal and the Arabian Sea.
45	Kaithwar et al.,	2020	Southeastern Arabian Sea	The abundant presence of <i>Bolivina currai</i> , <i>Hopkinsinella glabra</i> , and <i>Bulimina elegans</i> was correlated positively /negatively with the dissolved oxygen. Besides this, five other assemblages of living benthic foraminifera were also correlated with dissolved oxygen and organic carbon.
46	Liu et al.,	2020	Central Bay of Bengal	Reconstructed millennial-scale variability of Indian summer monsoon during the last 42 kyr

				by using foraminiferal Mg/Ca and oxygen isotope records from the Central Bay of Bengal.
47	Paketi et al.,	2020	Western Bay of Bengal	Used Sr-Nd isotope ratio coupled with major elemental compositions to understand the variation in lithological composition and contrasting rainfall distribution patterns during last 60 kyr.
48	Anbuselvan & Nathan	2021	Southwestern Bay of Bengal	Documented the distribution of planktic foraminiferal assemblages and their relationship with the environmental variables such as water depth, temperature and salinity.
49	Bhadra & Saraswat	2021	Western Bay of Bengal	Reported the surface distribution and ecology of planktic foraminifera in the riverine influx dominated western Bay of Bengal.
50	Suokhrie et al.,	2021	Westcentral Bay of Bengal	Reported the spatial distribution of living benthic foraminifera and its comparison with the ambient parameters to utilize them as proxy in paleo-ecology and paleoclimatic changes.
51	Gautam et al.,	2021	Southwestern Bay of Bengal	Reconstructed the changes in Indian monsoon variability during the last 46 kyr using the stable isotopic record of planktic foraminifera.
52	Verma et al.,	2021	Western Bay of Bengal	Documented the benthic foraminiferal response to the millennial-scale variations in monsoon-driven productivity and deep-water oxygenation in the western Bay of Bengal during the last 45 ka.
53	Sijinkumar et al.,	2021	Bay of Bengal and Andaman Sea	A record of Indian summer monsoon variability and its

				associated water column stratification and productivity changes during the late Quaternary.
54	Liu et al.,	2021	Bay of Bengal and Andaman Sea	Used trace elements (Mg/Ca and Ba/Ca) and stable isotopes of planktonic foraminifera (<i>Globigerinoides ruber</i>) and suggested that the last glacial maximum was ~3°C colder than the present sea surface temperature.
55	Bhadra & Saraswat	2022	Western Bay of Bengal	Documented exceptionally high foraminiferal dissolution in the western Bay of Bengal at depths deeper than 1000 m by using the fragmentation indices and the ratio between dissolution resistant and dissolution susceptible assemblage.
56	Bhadra & Saraswat	2022	Western Bay of Bengal	Reconstructed the influence of the mid-Pleistocene transition (MPT) on the monsoon and associated productivity in the Indian Ocean by using the planktic foraminifera species and their summer and winter monsoon assemblages.
57	Verma et al.,	2022	Western Bay of Bengal	Reconstructed the monsoon-related changes in surface hydrography and productivity during the last 45 kyr BP by using the planktic foraminiferal species and their assemblages.
58	Bhadra et al.,	2023	Western Bay of Bengal	Reported the first occurrence of <i>G. ruber</i> (pink) in the Bay of Bengal at 399 ka and the last occurrence datums at 128 ka and 123 ka in the Bay of Bengal and Arabian Sea, respectively.

59	Bhadra et al.,	2023	Western Bay of Bengal	Reconstructed the upper water column stratification during pre MPT, MPT and post MPT by using mixed layer and thermocline assemblages and oxygen isotopic ratio of planktic foraminifera.
60	Saraswat et al.,	2023	Northern Indian Ocean	Estimated the effect of freshwater induced salinity and diagenetic changes on the stable oxygen isotopic variation in <i>Globigerinoides ruber</i> ; also revised the salinity and temperature equations for Arabian Sea and Bay of Bengal.
61	Saraswat et al.,	2023	Gulf of Mannar	Reconstructed the carbon burial changes during the Meghalayan Age by using the elemental composition of bulk sediments and the relative abundance of <i>Globigerina bulloides</i> .
62	Suokhrie & Saraswat	2024	Westcentral Bay of Bengal	Reconstructed the variation in monsoon induced upwelling, mixed layer and thermocline during the last glacial-interglacial transition, by using the mixed layer and thermocline dwelling planktic foraminifera, <i>Globigerina bulloides</i> and the stable oxygen isotopic ratio of surface dwelling <i>Globigerinoides ruber</i> .
63	Thirumalai et al.,	2025	Bay of Bengal	Presented a century-scale record of summer monsoon runoff variability and marine biogeochemical impacts in the Bay of Bengal since the Last Glacial Maximum. Also reported a productivity collapse during both the extreme states of peak

				monsoon excess and deficits— both due to upper-ocean stratification.
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From the literature review, it is clear that, although, a good number of studies have been done on the distribution of foraminifera from the Bay of Bengal, most of the studies are confined to shallow regions like beaches and estuaries or continental shelf (Anbuselvan & Nathan, 2021; Symphonia & Nathan, 2019; Nagarajan & Nathan, 2018; Gandhi et al., 2017, 2014; Manasa et al., 2016; Symphonia & Nathan, 2014; Rana et al., 2007; Gandhi et al., 2007; Khare et al., 2007; Cullen & Prell, 1984). Additionally, only a few studies addressed the relationship between ambient environmental conditions and foraminifera from a shallow (shelf) to deep (abyssal) environment (Suokhrie et al., 2021, 2020; Bhadra & Saraswat, 2021; Saalim et al., 2018; Saraswat et al., 2017). A detailed study on the ecology of planktic and benthic foraminifera from the shelf to the deeper regions of the southwestern Bay of Bengal is non-exist. The existence of the OMZ and its effect on benthic foraminifera in the southwestern Bay of Bengal has also not been addressed. The knowledge of the assemblage thriving in the OMZ will help to reconstruct the intensity and extent of the OMZ in the past.

The existing paleoclimatic records from this region have a relatively coarse resolution and cover only a short duration (Verma et al., 2022; Liu et al., 2021; Gautam et al., 2021; Paketi et al., 2021; Liu et al., 2020; Ma et al., 2018; Ankit et al., 2017; Da Silva et al., 2017; Sijinkumar et al., 2016; Saraswat et al., 2014; Phillips et al., 2014; Ponton et al., 2012; Rashid et al., 2011; Govil and Naidu, 2011; Ota et al., 2009; Khare et al., 2007). Therefore, high resolution records are required to understand both the short-term and long-term past monsoon and associated changes. Several workers have reconstructed paleoclimate/monsoon from the Bay of Bengal, but most of them have focused on southwest monsoon (Ding et al., 2004) and the work on northeast monsoon is limited. Therefore, a detailed work on northeast monsoon intensity in the past is required to understand its role in the Indian economy and its response to global warming. The southwestern Bay of Bengal is a good site to reconstruct the northeast monsoon because it receives more freshwater flux during the northeast monsoon (**Fig. 1.1**).

1.3 Objectives:

Based on the above mentioned observations, the following research objectives were finalized.

- To understand the effect of ambient physico-chemical conditions, especially the oxygen deficient zone, on foraminifera.
- To reconstruct the changes in northeast monsoon intensity during the late Quaternary.
- To reconstruct the paleoproductivity changes in the southwestern Bay of Bengal.
- To reconstruct the oxygen deficient zone intensity during the late Quaternary.

Chapter 2

Study Area and Methodology

2.1 Hydrography of the Bay of Bengal

The Bay of Bengal (BoB) is a unique tropical basin, situated in the eastern part of the northern Indian Ocean. The northern Indian Ocean experiences a semi-annually reversing monsoon wind pattern (Vinayachandran et al., 2002; Schott and McCreary, 2001). The winds generally blow from the southwest during the summer monsoon (May-September) and from the northeast during the winter monsoon (November-February). The seasonal winds induce a reversal of ocean circulation along the east and west coast of India (Shankar et al., 2002). The western Bay of Bengal also experiences seasonally reversing monsoon winds, which alter the direction of the East India Coastal Current (EICC) (Schott and McCreary, 2001; Shetye et al., 1991) (**Fig. 2.1**). At the onset of the summer season, the summer monsoon current (SMC) transports high salinity water from the southeastern Arabian Sea (seAS) into the southwestern Bay of Bengal (swBoB) (**Fig. 2.1**). It joins the EICC along the eastern margin of India and flows towards the north (**Fig. 2.1**). From June

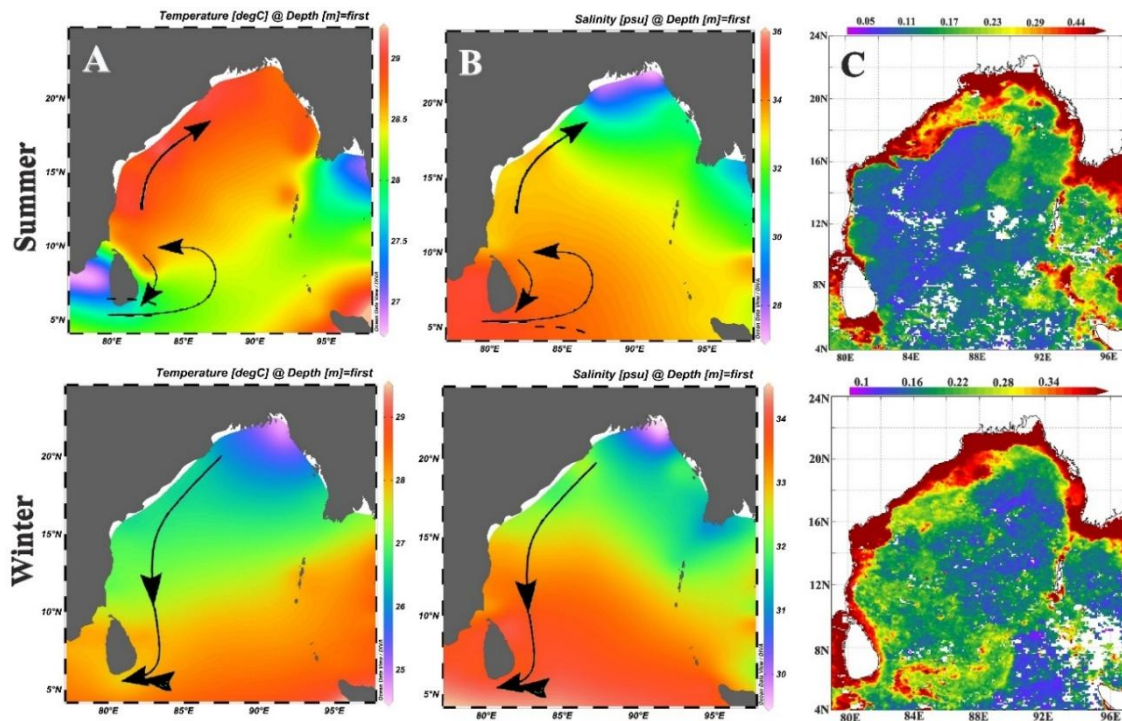


Figure 2.1: The modern seasonal (summer and winter) sea surface temperature (A), salinity (B) and chlorophyll-a (C) concentration in the western Bay of Bengal (Melin, 2013; Locarnini et al., 2018; Zweng et al., 2019). The arrows indicate seasonal surface currents (after Schott et al., 2009).

to September, SMC transports approximately 10 Sv (Sverdrups) of water into the BoB (Vinayachandran et al., 1999). Simultaneously, there is an influx of high salinity and carbonate-enriched water from the Gulf of Mannar to the Palk Bay, estimated at $\sim 16.301 \text{ m}^3/\text{s}$ (Jagadeesan et al., 2013). This high salinity carbonate-enriched water subsequently flows into the swBoB and enhances carbon burial in this region (Salman and Saraswat, 2024; Jagadeesan et al., 2013). The southwest monsoon also induces weak upwelling near the eastern margin of India during the summer (Shetye et al., 1996), leading to increased surface chlorophyll-a content (Shetye et al., 1991) (**Fig. 2.1**).

Brahmaputra, Ganga, Mahanadi, Godavari, Krishna, and Cauvery are the major rivers discharging a large quantity of fresh water into the Bay of Bengal (Jana et al., 2018; Unger et al., 2003). Amongst these, Krishna, Pennar, and Cauvery rivers discharge a large amount of freshwater into the southwestern Bay of Bengal during both summer as well as winter monsoon (Krishnamurthy and Shukla, 2000). The dispersal of freshwater over the basin leads to a low-saline upper water column with strong stratification throughout the year. The strong stratification is one of the major causes of the low primary productivity in the bay than that in the Arabian Sea (Prasanna Kumar et al., 2002). The physical processes like coastal upwelling and winter convection are not very active in the Bay of Bengal (Jayalakshmi et al., 2015). The intense stratification inhibits vertical mixing, causing intense evaporation on the surface which eventually leads to the cyclone. The conditions reverse in winter and the decrease in surface temperature leads to the formation of an “inversion layer” unlike the “barrier layer” in summer. The mixed layer depth remains shallow ($\sim 30 \text{ m}$) due to weak winds and strong stratification (Prasanna Kumar et al., 2002).

The stratification, along with primary productivity, drives the strength of the OMZ in the Bay of Bengal (Udaya Bhaskar et al., 2021). The dissolved oxygen availability is higher in the southwestern Bay of Bengal than in the northwestern part (Sarma et al., 2013). The northwestern region has higher productivity and low dissolved oxygen due to lower suspended particulate matter concentration compared to the southwestern region (Sarma et al., 2013) (**Fig. 2.1**). Although the northwestern Bay of Bengal receives massive freshwater input from the Ganga-Brahmaputra river system, the suspended particulate matter (SPM) is surprisingly lower in this region compared to the southwestern Bay. The study by Sarma et al. (2013) shows that SPM in the coastal Bay ranged between 9.1 and 33.7 mg/L, with higher values in the southwest (30-34 mg/L) and lower values in the northwest (9.2-12.4 mg/L). This contrary pattern occurs because most of the sediments

discharged by the Ganga-Brahmaputra are transported eastward and offshore, rather than accumulating along the northwestern coast, as indicated by regional clay mineralogy (Rao et al., 1988). Moreover, strong stratification caused by the large freshwater influx in the NW region limits vertical mixing and resuspension of bottom sediments, further reducing SPM in surface waters. The lower SPM enhances light penetration, promoting higher phytoplankton growth and chlorophyll-a concentrations (3.6-5.3 mg/m³), despite relatively low nutrient levels resulting from biological uptake (Sarma et al., 2013). In contrast, the southwestern Bay of Bengal, although less influenced by river discharge, shows higher SPM, which reduces light availability (and hence lower chlorophyll-a, 0.8-1.2 mg/m³). Thus, the low SPM in the northwestern Bay of Bengal results from the eastward transport of riverine sediments and suppressed vertical mixing due to stratification, not from a lack of sediment supply (Sarma et al., 2013). The oxygen minimum zone (OMZ) in the Bay of Bengal is not as strong as that in the Arabian Sea (Al Azhar et al., 2017; McCreary et al., 2013). The denitrification does not occur in the Bay of Bengal (Howell et al., 1997). The higher settling rate of particulate organic matter in the bay than in the Arabian Sea is one of the factors suggested for weaker OMZ and lack of denitrification in the Bay of Bengal (Gauns et al., 2005; Ittekkot et al., 1991). However, the remineralization depth may decrease due to the continuous warming of the Indian Ocean (Lee et al., 2015; Known et al., 2009). Therefore, the region may witness denitrification (Bristow et al., 2016).

2.2 Geology and hydrography of the southwestern Bay of Bengal

The Cauvery basin is one of the major peri-cratonic rift basins that developed during the rifting processes associated with the breakup of the Gondwana supercontinent during the Mesozoic Era. As the Indian Plate separated from Madagascar, rift valleys formed, leading to the development of the Cauvery basin (Twinkle et al., 2016). The Cauvery River deposits a huge amount of sediment resulting in a very high sedimentation rate in the basin. Along the entire continental shelf of western Bay of Bengal (wBoB), the highest sand (%) was observed near the Cauvery River mouth at 11°N transect (Jacob et al., 2008). The river discharge and sediment load in the Cauvery River peaks during the month of November (Vaithiyanathan et al., 1988). The topography and gradient of the land around the Cauvery River basin plays a crucial role in the high sedimentation rate. The river originates from the Western Ghats, with steep slopes, leading to more physical weathering. As the river flows toward the coast, it cuts through various types of terrains, including rocks, soils, and

sediments, carrying more sediment along the way (Vybhav et al., 2020; Vaithiyanathan et al., 1988; Subramanian, 1978). The Cauvery River also forms a delta, near the river mouth, where the sediments are deposited. The delta acts as a natural sediment trap, slowing down the flow and allowing the sediments to settle on the seafloor. This process contributes to

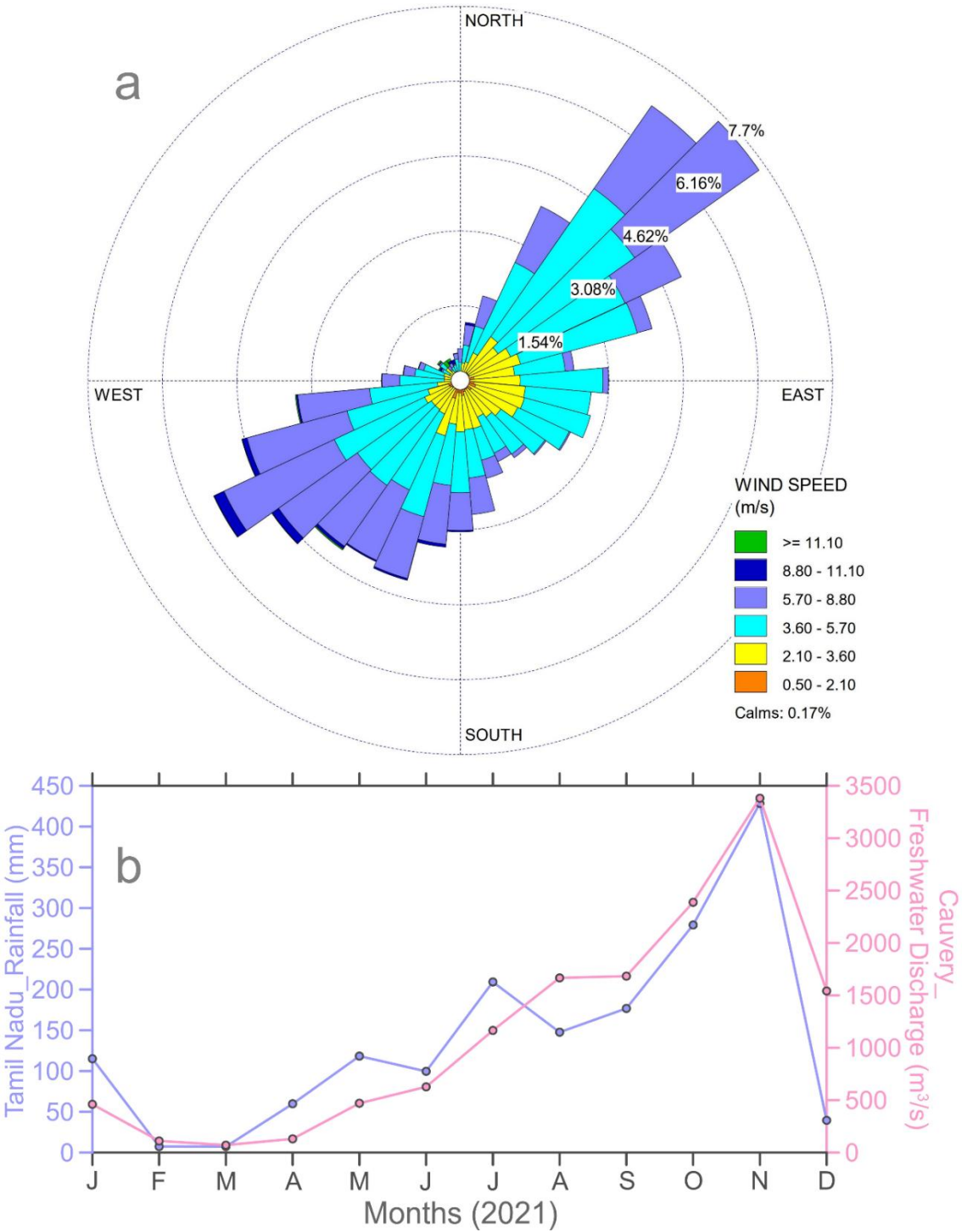


Figure 2.2: The wind rose diagram (a) shows the dominant wind pattern at the core location and the line graph (b) shows the monthly mean rainfall in the Tamil Nadu state and freshwater discharge of Cauvery River.

offshore sediment accumulation (Alappat et al., 2010; Singh and Rajamani, 2001).

The land and ocean thermal contrast results in moisture-laden winds blowing over Indian subcontinent from June to August. The intertropical convergence zone reaches its northernmost position by August resulting in ~70% of the annual rainfall over India (Dixit and Tandon, 2016). Later, as the ITCZ recedes southward, reaching its southernmost position in January, a reversal in pressure gradient triggers the onset of winter monsoon, mainly over southeast India (Dixit and Tandon, 2016). From the wind rose diagram, it is evident that the highest frequency of winds in the Cauvery basin is from the northeast during the winter season (**Fig. 2.2a**). As a result of these intense northeast winds, Tamil Nadu receives the maximum precipitation during the winter season (October to December) (Kripalani and Kumar, 2004; Khole and De, 2003) (**Fig. 2.2b**). Thus, the winter monsoon is the predominant rainy season in Tamil Nadu accounting for approximately ~60% of the rainfall in coastal areas and nearly 40–50% in the interior districts (Khole and De, 2003). During both the summer and winter monsoon seasons, the Krishna, Pennar, and Cauvery rivers discharge substantial amounts of freshwater into the southwestern Bay of Bengal (swBoB) (Khole and De, 2003; Krishnamurthy and Shukla, 2000). However, the freshwater discharge of Cauvery River dominates from October to December (**Fig. 2.2b**). In the swBoB, the seasonal variation in sea surface salinity is around 1 psu (**Fig. 2.3**), with the minimum during early winter (November – December). This decrease in salinity is attributed to the precipitation associated with the winter monsoon over southeast India (Dimri et al., 2016).

The Bay of Bengal exhibits relatively lower primary productivity as compared to the Arabian Sea. The diminished productivity in the Bay of Bengal is primarily attributed to the extensive cloud cover, sediment load, stratification, and a narrow shelf (Sarma et al., 2016; Prasanna Kumar et al., 2010; 2004; 2002; Madhupratap et al., 2003; Gomes et al., 2000). The Bay of Bengal experiences weak coastal upwelling and winter convective mixing as compared to the Arabian Sea (Prasanna Kumar et al., 2002; Shetye et al., 1991). However, the eddies and gyres are common in the Bay of Bengal and enhance the productivity in the area (Jayalakshmi et al., 2015; Sabu et al., 2015; Prasanna Kumar et al., 2010; 2004; 2002; Muraleedharan et al., 2007; Vinayachandran et al., 2005). During the winter season, cyclonic eddies that bring nutrient-rich and cold water to the surface are common in the swBoB resulting in a cooler sea surface temperature while in the summer season, the anticyclonic eddies are more prevalent leading to a warmer sea surface temperature (Jayalakshmi et al., 2015; Chen et al., 2013).

The southwestern Bay of Bengal experiences warmer sea surface temperature (SST) during the summer season ranging from 28.5 to 29.5°C (**Fig. 2.3**). The sea surface salinity (SSS) is however low during this season ranging from 32.5 to 34.0 psu. A relatively, cooler SST is observed during the winter season ranging from 26.0 to 27.5°C and SSS is also low in the swBoB ranging from 32.0 to 33.5 psu (**Fig. 2.3**). The Chlorophyll-a, except in very shallow marginal areas, is low in swBoB during the month of July (**Fig. 2.3**). The Chlorophyll-a content is high in the swBoB during January suggesting relatively higher marine productivity during the winter season (**Fig. 2.3**).

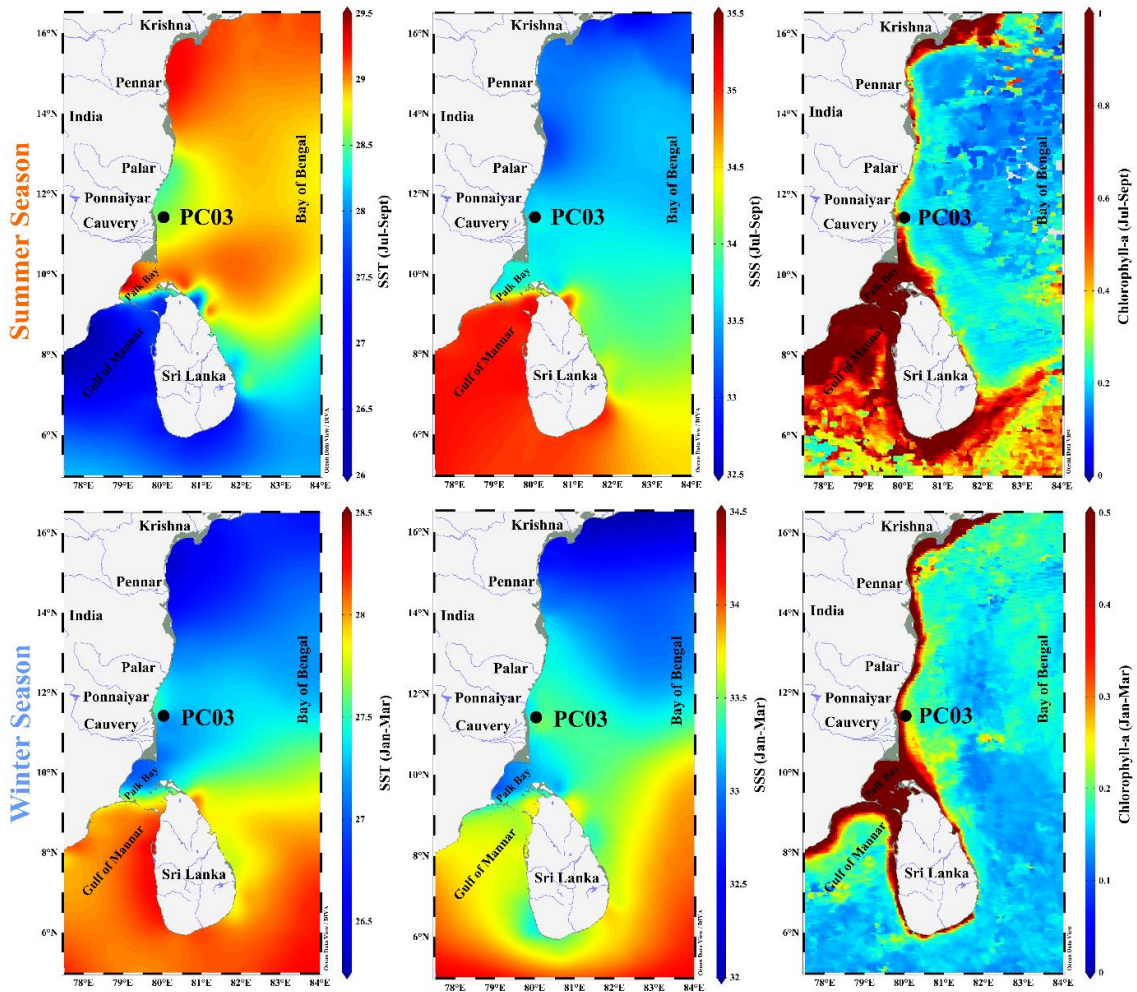


Figure 2.3: Seasonal changes in temperature, salinity and chlorophyll-a in the southwestern Bay of Bengal. The black dot is the core location (SSD067 PC03). The blue lines mark the major rivers.

2.3 Materials and methodology

2.3.1 Sample collection

During the 67th cruise of RV *Sindhu Sadhana* (SSD067) in November 2019, a series of surface sediment samples and a piston core (PC03) were collected from the swBoB (**Fig. 2.4; Table 2.1**). A total of 67 surface samples (56 multicores and 11 grabs) were collected along coast perpendicular transects, at depths varying from shallow (24 m) to deep (3505 m) (**Fig. 2.4; Table 2.1**). The sandy substrate hindered the operation of multicorer on the inner shelf. The sediments at such locations were sampled by using a grab sampler. The multicores were carefully sub-sampled onboard at 1 cm interval. The bottom water, from the undisturbed sediment-water interface in the multicorer tubes, was also collected and its temperature, salinity, pH and dissolved oxygen (DO) were measured. One-half of each sub-sample was stored in 100 mL plastic containers. The rose-Bengal solution was added to the containers and shaken gently to mix well with the sediments to stain and preserve living benthic foraminifera. The other half was stored in zip-lock plastic bags and subsequently used for microfossil study and bulk sediment geochemistry. The piston-core, SSD067 PC03 (Longitude 80.0323°E, Latitude 11.4202°N and water depth 509 m) was collected for paleoclimatic reconstruction (**Fig. 2.4**). The total length/recovery of the core was 15.72 m. The core was sub-sampled into 1.0 to 1.5 m long sections onboard and stored at 4°C. Subsequently, all sections were sub-sampled at 1 cm intervals in the laboratory.

Table 2.1. Sample details with latitude, longitude, and water depth. MC, GR, and PC are the notations for Multicore, Grab, and Piston core locations, respectively.

Sr. No.	Station	Latitude (°N)	Longitude (°E)	Water Depth (m)
1	MC01	14.5556	80.4010	318
2	MC02	14.5538	80.4159	598
3	MC03	14.5569	80.4214	721
4	MC04	14.5560	80.4385	1117
5	MC05	14.5401	81.3057	2994
6	MC06	14.5553	80.4418	1249
7	MC07	14.5557	80.4695	1556
8	MC08	14.5555	80.4913	1762
9	MC09	14.5532	80.5462	2018
10	MC10	14.5520	80.7283	2526
11	MC11	13.5501	81.3001	3289

12	MC12	13.5486	80.7996	2964
13	MC13	13.5462	80.7742	2735
14	MC14	13.5505	80.6819	2225
15	MC15	13.5486	80.6429	1725
16	MC16	12.4500	80.6833	1106
17	MC17	12.4524	80.6270	731
18	MC18	13.5493	80.6075	1522
19	MC19	13.5449	80.5945	1124
20	MC20	13.5502	80.5784	895
21	MC21	13.5507	80.5708	620
22	MC22	13.5488	80.5687	480
23	MC23	13.5500	80.5390	256
24	MC24	13.5498	80.5779	878
25	MC25	13.5483	80.5080	154
26	MC26	12.4643	80.5957	168
27	MC27	12.4626	80.5952	286
28	MC28	12.4726	80.5980	470
29	MC29	12.4594	80.7205	1437
30	MC30	12.2700	81.3000	3429
31	MC31	12.4560	80.8670	3254
32	MC32	12.4534	80.7527	2760
33	MC33	11.4244	80.0682	738
34	MC34	11.4202	80.0323	509
35	MC35	11.4193	79.9768	277
36	MC36	11.4189	79.9658	128
37	MC37	11.4143	79.9572	91
38	MC38	11.4178	79.9135	52
39	MC39	11.4188	79.8634	28
40	MC40	11.4203	80.1159	1036
41	MC41	11.4236	80.2714	1480
42	MC42	11.4193	80.4839	2007
43	MC43	11.4132	80.5663	2570
44	MC44	11.4162	80.7722	3112
45	MC45	10.7480	81.0000	3505
46	MC46	10.6535	80.8314	2982
47	MC47	10.5996	80.7096	1695

48	MC48	10.5567	80.6732	1240
49	MC49	10.5376	80.6521	1009
50	MC50	10.5122	80.6229	760
51	MC51	10.4979	80.6060	510
52	MC52	10.3785	80.1340	24
53	MC53	10.4048	80.2012	52
54	MC54	10.4125	80.2445	73
55	MC55	10.4489	80.3770	157
56	MC56	10.4812	80.5168	249
57	GR01	14.5688	80.2672	26
58	GR02	14.5473	80.3332	52
59	GR03	14.5562	80.3539	66
60	GR04	14.5537	80.3737	104
61	GR05	13.5479	80.4845	78
62	GR06	13.5523	80.4081	29
63	GR07	12.4480	80.2168	29
64	GR08	12.4425	80.2735	78
65	GR09	12.4434	80.2775	46
66	GR10	12.4538	80.5223	76
67	GR11	12.4479	80.5794	129
68	PC03	11.4202	80.0323	509

2.3.2 Sample processing

The surface and subsurface sediment samples for the foraminiferal studies were processed by following the established standard procedure (Naik et al., 2017; Saraswat et al., 2017). For the living benthic foraminifera, rose-Bengal-stained sediments were transferred into a measuring cylinder and kept for 24 hours to determine the bulk sediment volume. The sediments were wet sieved with a 63 μ m mesh. For dead benthic and planktic foraminifera, an appropriate amount (15–20 g) of sediment was collected in a pre-weighed and properly labeled petri-dish. The petri-dish was covered with a cling film and deep frozen for 24 hours at -30 °C. The frozen sediment samples were then freeze-dried. Freeze-drying helps in avoiding the hard lumps that usually form if the clay and organic matter-rich sediments are oven-dried. A portion (1–2 g) of the freeze-dried sediments was separated for geochemical analysis. An appropriate amount (15–20 g) of dried bulk sediments was weighed and wet sieved by using a 63 μ m mesh with very low water pressure to prevent

foraminiferal test breakage. Both the stained and non-stained fractions $>63\ \mu\text{m}$, retained on the sieve, were then transferred to a glass beaker and dried overnight. The dried coarse fractions were subsequently weighed and stored in plastic vials for further analysis. A total of 123 surface (0–1 and 1–2 cm) and 1572 subsurface (SSD067 PC03) samples were processed.

2.3.3 Picking and identification of foraminifera

A weighed aliquot of the dried $>63\ \mu\text{m}$ fraction (non-stained) was dry sieved at $125\ \mu\text{m}$, and the resulting fractions were weighed. A representative weighed aliquot of fraction $>125\ \mu\text{m}$ was used to pick the planktic foraminifera. A minimum of 300 intact planktic foraminifera shells were picked from each sample by using the Olympus stereozoom microscope (SZX16) and mounted on a micropaleontological slide. In case of sandy coarse fraction, where sufficient number of planktic foraminifera were not available, a minimum 1 g of coarse fraction was used. The planktic foraminifera were identified by following Schiebel and Hemleben (2017), Shrivastav et al.

(2016), Ovechkina et al. (2010), Hemleben et al. (1989), Bé (1967), and Parker (1962).

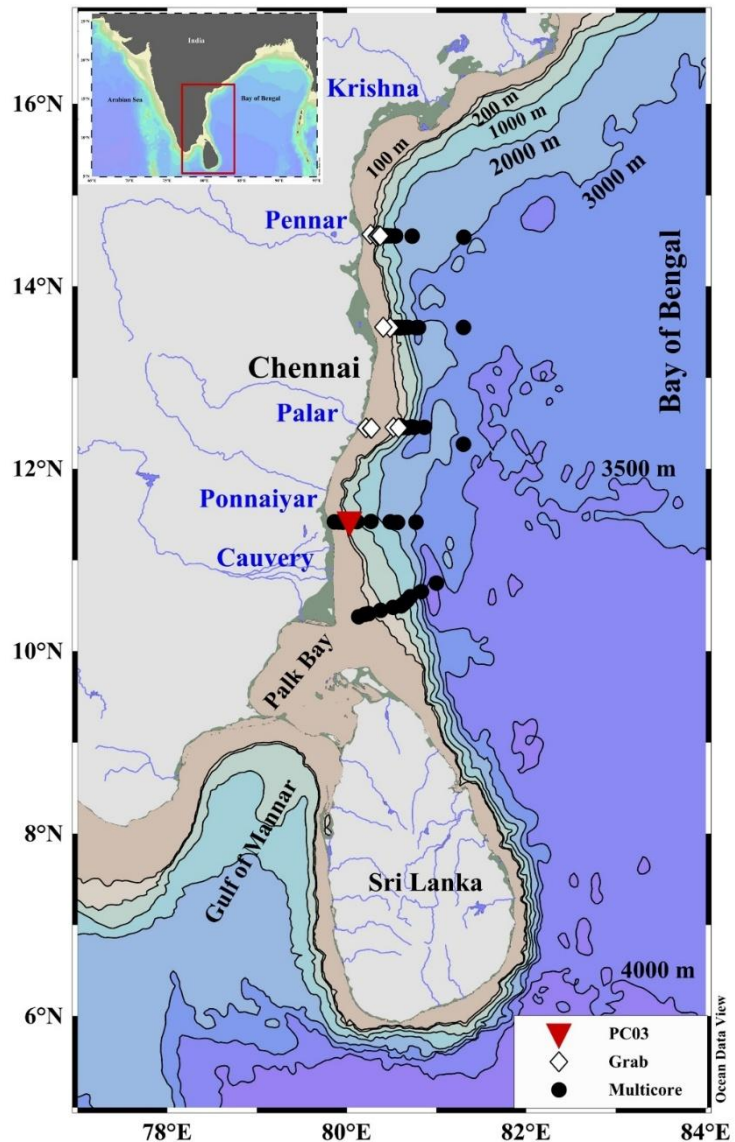


Figure 2.4: The sample locations in the southwestern Bay of Bengal. The black dots are the multicore stations, the white diamonds are grab samples, and the red inverted triangle is the location of piston core SSD067 PC03. The rivers draining into the region are marked by blue lines. The background colour and contours are bathymetry.

Globigerinoides ruber (white) and its morphotypes (*sensu stricto* and *sensu lato*) live throughout the year and thrive in the upper part of the mixed layer. Both morphotypes represent annual surface water conditions (Thirumalai et al., 2014). Therefore, *sensu stricto* and *sensu lato* morphotypes were counted together as *G. ruber*. The subspecies *Globorotalia menardii neoflexuosa* was also considered under *G. menardii* (Siccha and Kucera, 2017; Cullen and Prell, 1984). Spezzaferri et al., (2015) suggested *Trilobatus* as a new genus to distinguish the *trilobus* and *ruber* clades. They further suggested to use the term *Trilobatus trilobus* for specimens without sac and *Trilobatus sacculifer* for specimens with sac, in view of their different ecological preferences. Recently, a similar approach was also used by Bhadra and Saraswat (2021) and Anbuselvan and Nathan (2021). The absolute abundance was estimated by using the following equation.

$$\text{Absolute Abundance} = \frac{\text{Number of Planktic Foraminifera}}{\text{Wt. CF (>125 } \mu\text{m) used for picking}} \times \frac{\text{Wt. CF (>125 } \mu\text{m) after dry sieving}}{\text{Wt. CF (>63 } \mu\text{m) taken for dry sieving}} \times \frac{\text{Wt. CF (>63 } \mu\text{m)}}{\text{Wt. Dry Sediment}}$$

The number of specimens of different planktic foraminiferal species were counted to calculate their relative abundance. Before processing, the volume of wet sediment was measured to calculate the flux of planktic foraminifera and calcium carbonate. The total planktic foraminiferal number per gram sediment (TPN/g sediment) and the relative abundance (%) of different species were plotted by using Ocean Data View software (Schlitzer, 2021).

The dried CF (>63 μm) of rose-Bengal (stained) sediment samples was used to pick living benthic foraminifera. The aliquot of coarse fraction required to pick the living benthic foraminifera was obtained by coning and quartering. A minimum of 300 living (rose-Bengal stained) shells were picked from the two core top sections (0-1 and 1-2 cm) for all the surface samples. The absolute abundance of living benthic foraminifera (TBF/g dry sediment) was estimated for both 0–1 cm and 1–2 cm intervals. The total population of living benthic foraminifera was categorized into two morphogroups, namely calcareous and agglutinated foraminifera. The living calcareous benthic foraminifera population was further divided into two distinct morphogroups, namely angular and rounded benthic foraminifera by following the published literature (Saalim et al., 2018; Nigam et al., 1992; 2007).

For paleomonsoon reconstruction, one-fourth of >63 μm fraction from each section

of the core SSD067 PC03 was dry sieved with 150 μm mesh to pick the planktic foraminifera. A representative aliquot of the sieved coarse fraction obtained after coning and quartering was weighed and spread in a picking tray to pick planktic foraminifera. A minimum of 300 specimens were picked from the alternate intervals for 8.2 and 4.2 ka events and identified by following the published literature mentioned above. Similarly, planktic foraminifera were also picked from 172 samples for the entire sediment core at an interval of 10 cm. The total planktic foraminifera number per gram of dry sediment (TPN/g sediment) was calculated. The relative abundance of *Globigerinoides ruber*, *Globigerina bulloides*, *Neogloboquadrina dutertrei*, *Trilobatus trilobus*, *Trilobatus sacculifer*, *Globigerinella siphonifera*, and *Neogloboquadrina incompta* was used for the summer monsoon assemblage (SMA), as these species proliferate during the summer monsoon season in the southwestern and western Bay of Bengal (Salman and Saraswat, 2024; Maeda et al., 2022; Stoll et al., 2007; Guptha et al., 1997). In contrast, *Globigerinita glutinata*, *G. falconensis*, *Globorotalia menardii* and *Pulleniatina obliquiloculata* were dominant during the winter monsoon (Salman and Saraswat, 2024; Bhadra and Saraswat, 2021; Maeda et al., 2021; Stoll et al., 2007; Guptha et al., 1997) and thus were considered for the winter monsoon assemblage (WMA). The studies based on sediment traps and surface sediments have indicated that *G. ruber* and *T. sacculifer* prefer warmer, low-nutrient oligotrophic conditions (Salman and Saraswat, 2024; Guptha et al., 1997). Therefore, these species were categorized as mixed-layer oligotrophic species (MLOS). In contrast, *G. bulloides*, *G. glutinata*, and *G. falconensis* flourish in cooler, nutrient-rich eutrophic environments, classifying them as mixed-layer eutrophic species (MLES).

To understand the paleo-oxygenation changes in the western Bay of Bengal, a total of 172 samples from the entire PC03, at an interval of 10 cm, were used to pick benthic foraminifera. The dried CF ($>63\ \mu\text{m}$) was used to pick a minimum of 300 benthic foraminifera shells. The absolute abundance of benthic foraminifera (TBF/g dry sediment) was estimated for the entire core. Interestingly, we got only calcareous benthic foraminifera throughout the core except a few intervals. The agglutinated benthic foraminifera were rare throughout the core with a maximum of upto 5 specimens out of 300 specimens, only in the topmost five section of the core. The total calcareous benthic foraminifera population was divided into three distinct morphogroups, namely angular, rounded, and spherical benthic foraminifera by following the published literature (Singh et al., 2022; Nigam et al., 1992; 2007).

2.3.4 Elemental analysis of bulk sediment

The freeze-dried sediment was powdered with the help of agate mortar and pestle for geochemical analysis. A total of 123 surface (0–1 and 1–2 cm) and 1572 piston core (SSD067 PC03) samples at 1 cm interval from the entire core, were powdered and analyzed. The CN Elemental Analyzer (Flash2000NC Thermo) was used to analyze the total carbon (TC) and nitrogen (N) content of the sediments. The NC soil was used as the standard for TC and N with the values of $2.29 \pm 0.07\%$ and $0.21 \pm 0.01\%$, respectively. The standard deviation and associated error for TC were $\pm 0.05\%$ and 2.24% and for N $\pm 0.01\%$ and 4.61% , respectively. The total inorganic carbon (TIC) was analyzed by using a CO_2 Coulometer (CM50170 UIC). The calcium carbonate (CaCO_3) standard, with inorganic carbon of 12%, was used for TIC analysis. The standard deviation was $\pm 0.070\%$ and the associated error was 0.58% . The organic carbon (C_{org}) was obtained by subtracting the inorganic carbon from the total carbon. The CaCO_3 percentage in the bulk sediment was calculated from TIC and the atomic weight of carbon in CaCO_3 by using the following equation

$$\text{CaCO}_3 (\%) = \text{TIC} (\%) \times \frac{100}{12}$$

2.3.5 Stable isotopic analysis

One-fourth of $>63 \mu\text{m}$ fraction was dry sieved with 250 and $350 \mu\text{m}$ mesh to pick the planktic foraminifera species, namely *G. ruber* (*sensu stricto*) and *N. dutertrei* for isotopic analysis. The clean and well-preserved specimens were picked and analyzed for isotopic analysis. The analysis was done at the Stable Isotope Laboratory, Department of Earth Sciences, Indian Institute of Technology, Roorkee, using a Thermo Fisher Scientific MAT 253 Plus 10-kVa isotope ratio mass spectrometer, coupled with Kiel-IV carbonate device. The foraminifera shells were treated with 105% orthophosphoric acid (H_3PO_4) within the Kiel carbonate system at 72°C to produce the CO_2 for isotopic analysis. The secondary lab standard [Indian Ocean Coral-(IOC) from the Maldives] was run after every 5 samples to check the precision and accuracy of the instrument with a deviation of $\pm 0.22\text{‰}$ for $\delta^{18}\text{O}$ and $\pm 0.17\text{‰}$ for $\delta^{13}\text{C}$. The oxygen isotopic ratio in foraminifera were reported relative to the reference standard (VPDB) using the delta notation.

The $\delta^{18}\text{O}_{G. ruber}$ includes the contribution of both the seawater $\delta^{18}\text{O}$ ($\delta^{18}\text{O}_{\text{sw}}$) and temperature of the ambient environment. The measured $\delta^{18}\text{O}_{G. ruber}$ was used to estimate the $\delta^{18}\text{O}_{\text{sw}}$ by using the temperature equation derived for *Orbulina universa* (Bemis et al.,

1998). As the sea surface temperature (SST) data for the same core is not available, the *G. ruber* Mg/Ca based SST record from the nearby southeastern Arabian Sea was used to correct for temperature contribution (Saraswat et al., 2013). The global ice volume contribution was taken from Bintanja and Van de Wal, (2008). The ice volume correction (ivc) values were subtracted from the $\delta^{18}\text{O}_{\text{sw}}$ to obtain the local $\delta^{18}\text{O}_{\text{sw}}$ ($\delta^{18}\text{O}_{\text{sw-ivc}}$). A conversion factor of 0.27 was applied to convert the $\delta^{18}\text{O}_{\text{sw-ivc}}$ from VPDV to Vienna Standard Mean Ocean Water (VSMOW). The sea surface salinity was estimated using the relation between salinity and $\delta^{18}\text{O}_{\text{sw}}$, given by Saraswat et al. (2023), for the Bay of Bengal.

$$\delta^{18}\text{O}_{\text{sw}} = 0.13 * \text{salinity} - 4.23$$

where 0.13 is the slope of salinity, which means the rate of change in $\delta^{18}\text{O}_{G. ruber}$ for unit change in salinity, and 4.23 is the constant.

The thermocline paleotemperature records from the Bay of Bengal are not available. Therefore, the ice volume contribution was directly subtracted from $\delta^{18}\text{O}_{G. ruber}$ and $\delta^{18}\text{O}_{N. dutertrei}$. After ice volume correction, the $\delta^{18}\text{O}_{N. dutertrei}$ was subtracted from $\delta^{18}\text{O}_{G. ruber}$ to calculate the difference of $\delta^{18}\text{O}$ between mixed layer and thermocline, represented as following.

$$\Delta\delta^{18}\text{O} = \delta^{18}\text{O}_{G. ruber} - \delta^{18}\text{O}_{N. dutertrei}$$

2.3.6 Environmental parameters

The annual average temperature and salinity data for all the multicore and grab stations were downloaded from the World Ocean Atlas 2018 (Locarnini et al., 2018; Zweng et al., 2018) via the Live Access Server of CSIR-National Institute of Oceanography Goa. Both the average temperature and salinity profiles of the mixed layer and thermocline were acquired. The mixed layer was considered as 30 m, while the thermocline depth was set from 30 to 150 m, following Muraleedharan et al. (2007). The annual average Chlorophyll-a data was obtained from the “Chlorophyll-a, NASA VIIRS Ocean color (2012 onward)” database through Live Access Server of CSIR-National Institute of Oceanography, Goa and used as a productivity indicator. The ocean currents data with ‘u’ and ‘v’ vectors (GOFS 3.1: 41-layer HYCOM + NCODA Global 1/12° Reanalysis) were obtained from HYCOM data server. The average surface currents data, for July and January of 2015, was taken with temporal resolution of 3 hours.

The hourly wind speed and direction data for 2021 at a height of 10 m were

obtained from the NASA POWER Data Access Viewer (<https://power.larc.nasa.gov/data-access-viewer/>) and utilized to generate a wind rose diagram using WRPLOT view software. The monthly mean rainfall data for Tamil Nadu was sourced from the Centre for Hydrometeorology and Remote Sensing (CHRS) RainSphere platform (<https://rainsphere.eng.uci.edu/>), while freshwater discharge data for the Cauvery River was acquired through the Group on Earth Observations (GEO) HydroViewer, a platform dedicated to global water sustainability (<https://hydroviewer.geogloss.org/>). The dissolved oxygen was measured onboard by using titration following Winkler's method.

Table 2.2: Details of the radiocarbon dates with lab code, associated errors, 1 sigma age range, calibrated ages, and the sedimentation rate.

Sample Interval	Lab ID	Age (BP)	Error (+/-)	Age range (1sigma)	Calib. Age (ka)	S. R. (cm/kyr)
PCO3 (49-50)	IUACD#21C4067	886	23	230 - 551	0.382	128.27
PCO3 (99-100)	IUACD#21C4068	1481	24	743 - 1101	0.932	90.91
PCO3 (149-150)	IUACD#21C4069	1995	29	1278 - 1657	1.466	93.63
PCO3 (199-200)	IUACD#21C4070	2530	30	1877 - 2312	2.094	79.62
PCO3 (249-250)	IUACD#21C4071	2932	29	2352 - 2771	2.584	102.04
PCO3 (299-300)	IUACD#22C5196	3292	29	2809 - 3230	3.029	112.36
PCO3 (349-350)	IUACD#22C5197	3819	29	3446 - 3888	3.677	77.16
PCO3 (399-400)	IUACD#22C5198	4251	30	3988 - 4457	4.237	89.29
PCO3 (449-450)	IUACD#22C5199	4772	36	4691 - 5163	4.914	73.86
PCO3 (499-500)	IUACD#22C5200	5433	32	5490 - 5875	5.679	65.36
PCO3 (549-550)	IUACD#22C5201	6299	34	6398 - 6801	6.612	53.59
PCO3 (599-600)	IUACD#22C5202	6788	39	6954 - 7335	7.144	93.98
PCO3 (649-650)	IUACD#22C5203	7470	37	7622 - 7974	7.808	75.30
PCO3 (696-697)	IUACD#22C5204	8267	37	8459 - 8911	8.691	53.23
PCO3 (749-750)	IUACD#23C5788	9160	62	9548 - 10029	9.811	47.32
PCO3 (899-900)	IUACD#23C5791	10720	76	11729 - 12320	12	68.52
PCO3 (949-950)	IUACD#23C5792	12594	58	13831 - 14434	14.164	23.11
PCO3 (996-997)	IUACD#23C5793	13004	53	14499 - 15102	14.792	74.84
PCO3 (1049-1050)	IUACD#24C7149	13579	49	15348 - 15852	15.608	64.95
PCO3 (1099-1100)	IUACD#24C7150	14716	56	16828 - 17302	17.057	34.51
PCO3 (1149-1150)	IUACD#24C7158	14962	52	17096 - 17600	17.363	163.40
PCO3 (1199-1200)	IUACD#24C7159	15575	64	17842 - 18293	18.088	68.97

PC03 (1249-1250)	IUACD#24C7160	16633	60	19000 - 19463	19.238	43.48
PC03 (1299-1300)	IUACD#24C7161	17110	57	19570 - 20058	19.823	85.47
PC03 (1382-1383)	IUACD#24C7162	18405	63	21153 - 21709	21.422	51.91
PC03 (1429-1430)	IUACD#24C7163	19452	88	22395 - 22827	22.605	39.73
PC03 (1462-1463)	IUACD#24C7164	19985	86	22933 - 23412	23.191	56.31
PC03 (1499-1500)	IUACD#24C7165	20578	71	23650 - 24090	23.842	56.84
PC03 (1571-1572)	IUACD#24C7166	22076	80	25312 - 25725	25.516	43.01

2.3.7 Chronology of PC03

The radiocarbon (^{14}C) dating is commonly used to estimate the age of carbon-bearing materials (e.g. foraminifera, corals, stalagmites, trees, sediments) deposited in natural environment during the last ~55 kyr. The mixed planktic foraminiferal shells were picked to develop the chronology of PC03. The dried coarse fraction ($>63\ \mu\text{m}$) was sieved with $125\ \mu\text{m}$ mesh to pick the planktic foraminifera. The chronology of the core was established, by using 29 accelerator mass spectrometer (AMS) radiocarbon dates analyzed at the Inter-University Accelerator Center, New Delhi, India. Most of the dates were spaced at an interval of 50 cm throughout the entire length of the core (**Fig. 2.5, Table 2.2**), if there was no change in sediment texture. If there was any change in sediment texture, intervals were selected accordingly. All the ^{14}C ages were calibrated by using Marine20 calibration curve in the CALIB 8.2 radiocarbon calibration program (Heaton et al., 2020) with a reservoir correction (ΔR) and uncertainty of -57 ± 165 yr for the swBoB (Southon et al., 2002; Dutta et al., 2001a; 2001b; 2001c). The average sedimentation rate for the entire core was 72.79 cm/kyr, which gives an average resolution of ~14 yrs/cm (**Fig. 2.5, Table 2.2**).

2.3.8 Statistical analysis

All the planktic foraminifera species except those with relative abundance $<0.2\%$, were used along with different environmental parameters for Canonical Correspondence Analysis (CCA) and Pearson Correlation Coefficients (R^2). The relative abundance of all species was used to calculate the Dominance, Simpson (1-D) and Shannon (H) diversity indices by using the Paleontological Statistics software, 'PAST' version 4.14 (Hammer et al., 2001). Similarly, CCA was also run with an absolute abundance of benthic foraminifera and their morphogroups to understand their correlation with ambient environmental parameters for section 3.2.

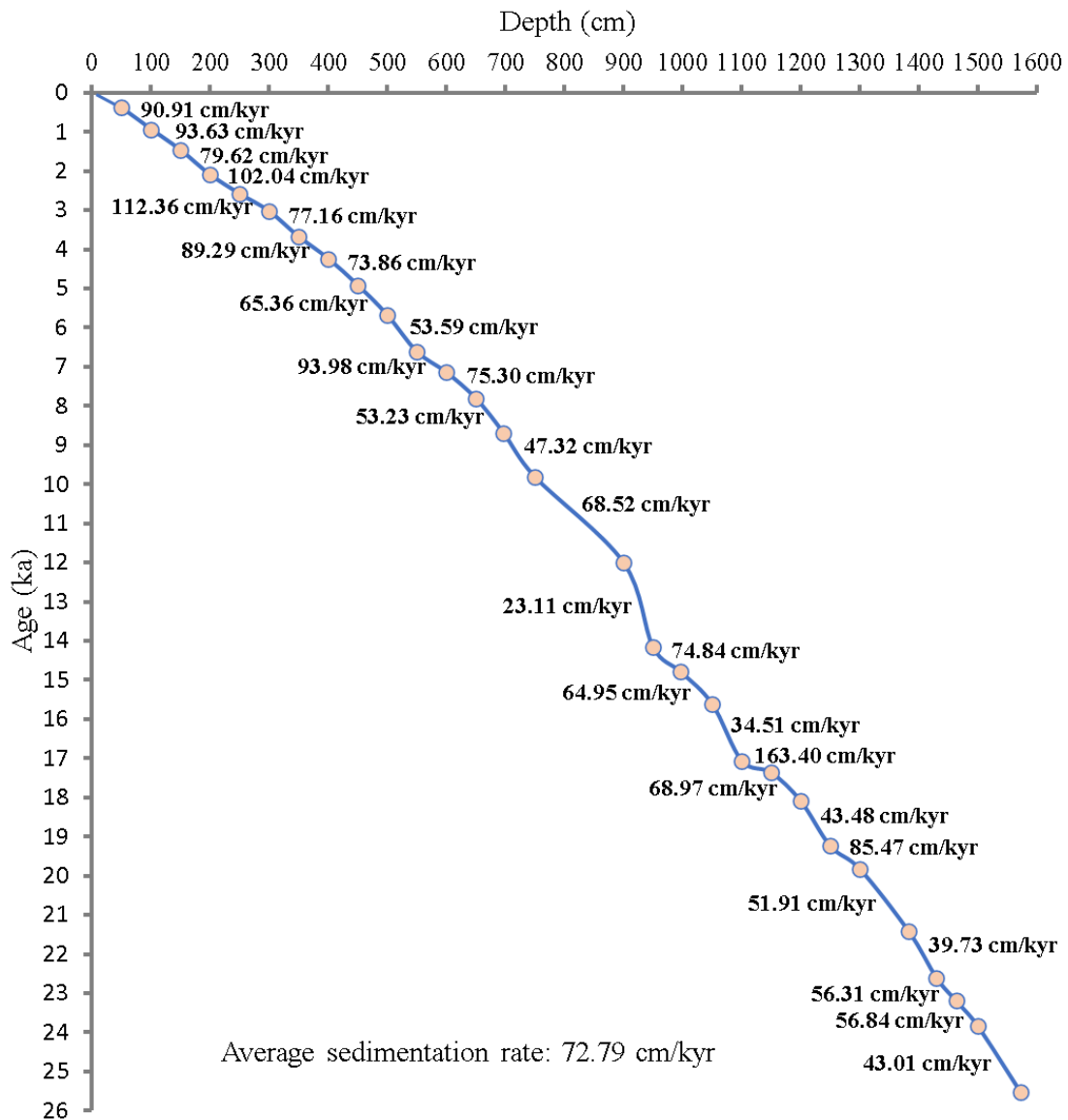


Figure 2.5: Age-depth profile and sedimentation rate of the sediment core (SSD067 PC03) used for this study.

2.3.9 Spectral and wavelet analysis

In order to determine the cyclicity and frequency of transient events during the last glacial maximum, spectral and wavelet analysis was done for CaCO_3 (%) and C_{org} (%). The analysis was done with PAST' version 4.14 (Hammer et al., 2001) by using simple periodogram algorithm for spectral analysis and wavelets for unequal spacing algorithm for wavelet analysis. In spectral analysis the significant periodicities were considered with the significance level of > 99 and 95 %.

Chapter 3

Ecology and Distribution of Planktic and Benthic Foraminifera

3.1 Planktic foraminiferal ecology and carbon burial in the southwestern Bay of Bengal

3.1.1 Introduction

Oceans absorb a large fraction of anthropogenic carbon dioxide (CO₂), thus restricting its build up in the atmosphere (Heinze et al., 2015). Once dissolved, CO₂ either reacts with water molecules to form carbonate ions and hydrogen ions or is sequestered by the phytoplankton to form the organic matter (Oschlies et al., 2025; Fig. 3.1). The weathering of silicate rocks releases the carbonate and calcium ions and reaches the sea through riverine influx, which also carries a lot of nutrients and terrestrial organic matter supporting the phytoplankton bloom in the upper water column (Penman et al., 2025; Fig. 3.1). The marine organisms use these carbonate ions to precipitate calcareous shells (Schiebel, 2002). The terrestrial and marine organic carbon and inorganic calcareous shells sink down to the ocean floor and get buried for a long time (Fig. 3.1). The single celled free-floating planktic foraminifera are amongst the major marine calcareous organisms inhabiting the upper ocean waters (Hemleben et al., 1989; Bé, 1977). They sequester a large amount of CO₂ and bury it in the ocean sediments for millions of years (Saraswat and Nigam, 2013). Around 0.36-0.88 Gt/yr CaCO₃ is the total contribution of planktic foraminifera to the surface sediments of the world ocean, which is ~32-80% of the total deep-ocean calcite budget (Schiebel, 2002). The ambient ocean conditions, including temperature, salinity, pH, productivity, and currents, strongly affect planktic foraminifera and their diversity (Lombard et al., 2011; Fraile et al., 2008; Zaric et al., 2005; Watkins et al., 1996). The sea surface temperature is considered as the most important factor that controls the abundance and diversity of planktic foraminifera (Rutherford et al., 1999), shell size (Schmidt et al., 2004a) and composition (Morey et al., 2005). Therefore, changes in the seawater properties due to the global warming, will affect planktic foraminifera and in turn carbon burial capacity of the oceans.

Additionally, being extremely sensitive to minor changes in the ambient

environment, planktic foraminifera are extensively used in paleoclimatic reconstruction (Nigam, 2005; Katz et al., 2010; Saraswat, 2015). The past environmental conditions and the composition of ambient seawater can be reconstructed by using faunal abundance as well as through the chemical composition of their hard shells and changes in the stable isotopic and elemental ratio (Schiebel and Hemleben, 2017). The comparison of surface distribution of recent planktic foraminifera with the ambient parameters provides information about the ecological preferences of the species and the factors controlling the geochemical signatures of the tests. The knowledge of the environmental control on the distribution, morphology, species abundance, and geochemical composition of the foraminiferal shells, is used to reconstruct the paleoceanographic conditions (Kucera, 2007).

The factors affecting planktic foraminiferal diversity and abundance, as well as the chemical characteristics of their shells can be studied by using living specimen collected by using plankton net (Iwasaki et al., 2017). However, one-time snapshot sampling does not address the resultant planktic foraminiferal abundance in marine sediments representing annual conditions. The sediment traps are deployed to understand the year round abundance and diversity of planktic foraminifera as well as their contribution to the carbon burial (Maeda et al., 2022; Kiss et al., 2021). However, such studies are mostly confined to the open ocean and are very limited (Bhadra and Saraswat, 2021). Although, the foraminiferal assemblage in the surface sediments is partially modulated by factors other than the ambient conditions prevailing at the time of living population, the ambient conditions still strongly modulate the resultant foraminiferal population in the sediments. That is why, this sedimentary foraminiferal assemblage is the basis of all the paleoclimatic and paleoceanographic reconstruction. Therefore, an improved paleoclimatic reconstruction by using foraminiferal characteristics requires a comparison of the foraminiferal assemblages in the sediments with the ambient conditions (Saraswat et al., 2023; Bhadra and Saraswat, 2022; Wang et al., 2021; de Villiers, 2005; Schiebel, 2002; Peterson and Prell, 1985).

The northern Indian Ocean is strongly influenced by the seasonal reversal of monsoon and thus provides a unique environmental setting for planktic foraminiferal studies. Therefore, multiple efforts were made to investigate the planktic foraminiferal distribution in the northern Indian Ocean (Stainbank et al., 2019; Munir and Sun, 2018; Darling et al., 2017; Symphonia and Senthil Nathan, 2014; Bhonsale and Saraswat, 2012; Saraswat, 2010; Chaudhary et al., 2003; Schulz et al., 2002; Naidu, 1993; Duplessy et al.,

1981; Cullen and Prell, 1984; Bé and Hutson, 1977; Bé and Tolderlund, 1971; Zobel, 1971). Both the eastern Arabian Sea (eAS) and Bay of Bengal (BoB), receive a large amount of monsoon influx and direct precipitation (Behara et al., 2019; Vinayachandran et al., 2013; Rao and Sivakumar, 2003). The huge freshwater influx, seasonal reversal of monsoon, surface water stratification and mesoscale eddies create distinct environmental conditions in the Bay of Bengal (Jana et al., 2018; Li et al., 2017) to understand the response of planktic foraminifera to monsoon and associated processes. Therefore, a few

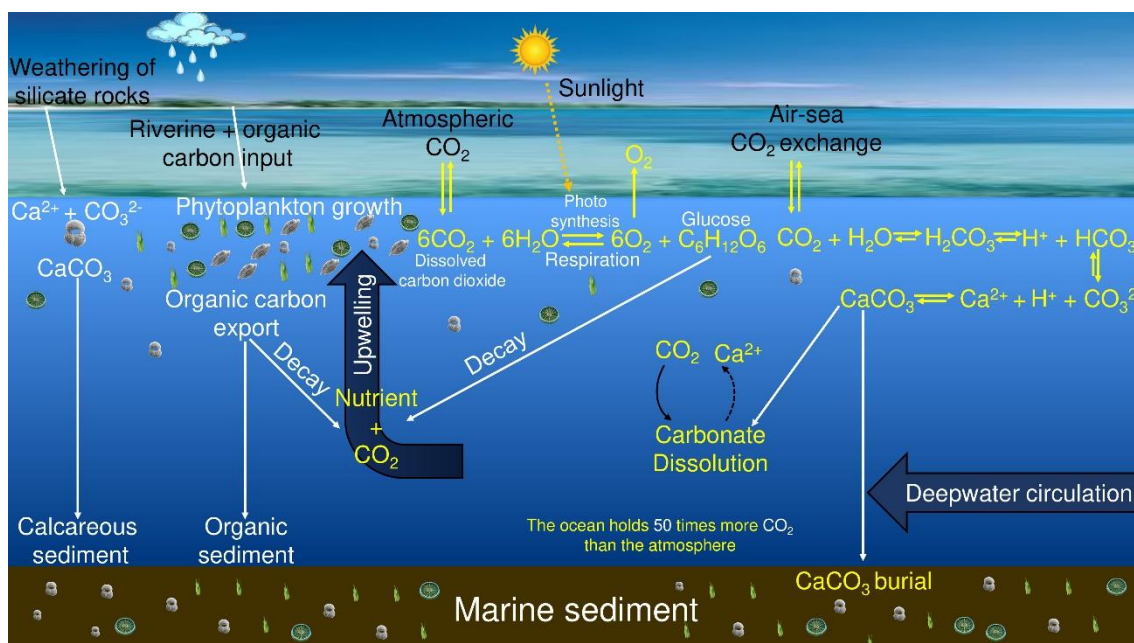


Figure 3.1: Graphical visualization of different sources of organic and inorganic carbon in marine environment and their long-term burial in marine sediment.

attempts were made to document the planktic foraminiferal distribution from the Bay of Bengal (Maeda et al., 2022; Bhadra and Saraswat, 2021; Anbuselvan and Senthil Nathan, 2021; Symphonia and Senthil Nathan, 2014; Chaudhary et al., 2003; Guptha et al., 1997; Cullen and Prell, 1984). Guptha et al., (1997) and Maeda et al., (2022) reported an increase in the planktic foraminiferal flux in the central Bay of Bengal during both the summer and winter monsoon and the lowest flux from March to May, by using the sediment trap data. Recently, Bhadra and Saraswat (2021) studied foraminiferal distribution in the western Bay of Bengal (wBoB) to understand the response of planktic foraminifera to the huge fresh water influx. However, such information is missing from the southwestern Bay of Bengal (swBoB) as highlighted by Anbuselvan and Senthil Nathan (2021) and Symphonia and Senthil Nathan (2014). Anbuselvan and Senthil Nathan (2021) investigated a small part of the swBoB, but their work was limited to the continental shelf. Therefore, it is necessary to understand the distribution and ecology of

planktic foraminifera from the swBoB as the region is influenced by both the summer as well as winter monsoon.

The regional physical processes like mesoscale eddies, cross-basin water exchange, seasonal surface circulation pattern, and the monsoonal flux directly affect the hydrology of the swBoB (Jayalakshmi et al., 2015; Vinayachandran, 2009; Vinayachandran et al., 1999). The effect of these processes on the planktic foraminiferal population in the swBoB is still not clear. The region also has very high productivity resulting in a large organic matter flux. The organic matter availability not only strongly affects living foraminiferal population but also the preservation of the shells in the sediments. The utility of the organic matter as food for marine organisms depends on its origin and quality. The organic matter can be of marine or terrestrial origin. The rivers

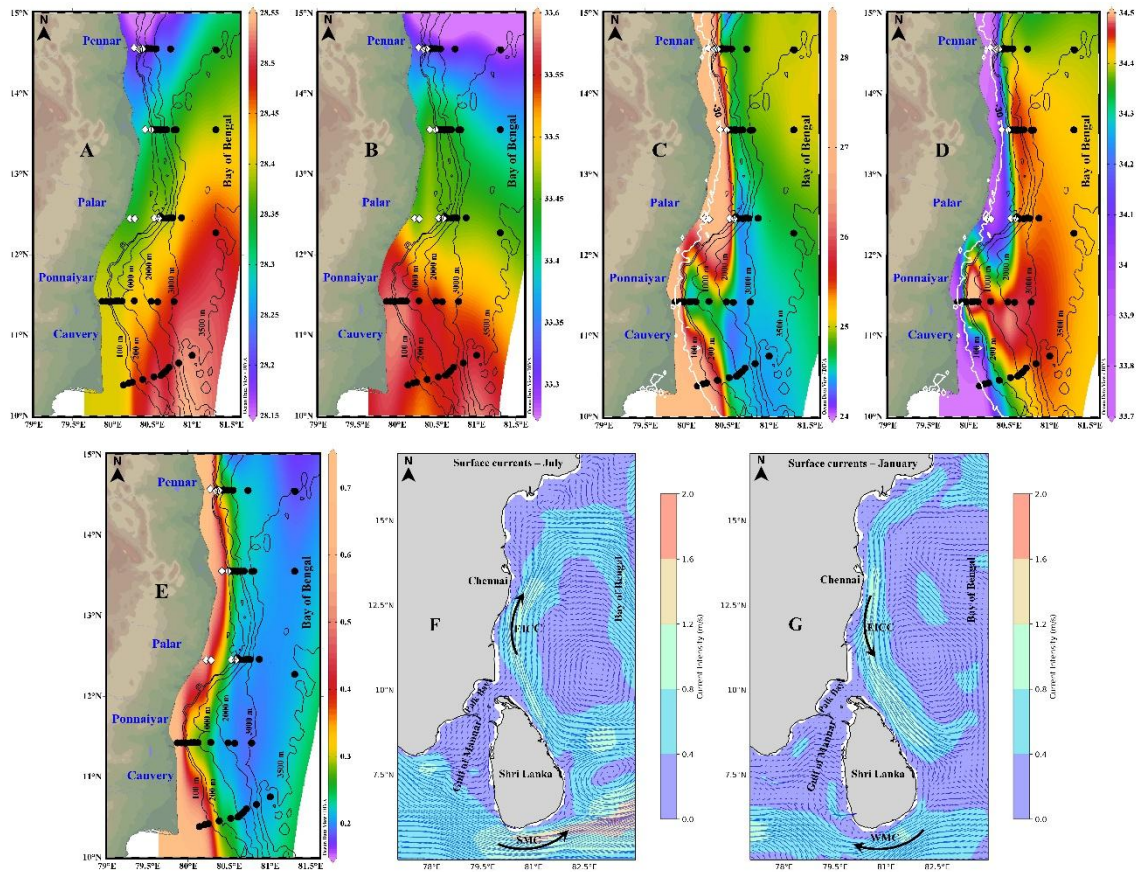


Figure 3.2: The annual average mixed layer temperature (A), mixed layer salinity (B), thermocline temperature (C), thermocline salinity (D), surface Chlorophyll-a (E) and seasonal (July representing southwest monsoon, and January representing northeast monsoon) surface currents (F & G), in the southwestern Bay of Bengal (Locarnini et al., 2018; Zweng et al., 2018). The background colour and length of the arrows indicate the intensity of the currents. The major currents in the region are marked as East India Coastal Current (EICC), Summer Monsoon Current (SMC) and Winter Monsoon Current (WMC). The water depth between mixed layer and thermocline is marked by white contour line at 30 m. The black dots and white diamonds are the sample locations and major rivers marked by blue lines. The black contour lines are bathymetry.

transport terrestrial organic matter to the sea (Gao et al., 2022). The C_{org}/N ratio can be used to differentiate the sources of organic matter. The low C_{org}/N ratio (<7) indicates nitrogen rich marine produced organic matter (Saraswat et al., 2023; Ramaswamy et al., 2008; Suokhrie et al., 2022). However, post-depositional diagenetic changes modulate the C_{org}/N ratio (Waples & Sloan, 1980; Freudenthal et al., 2001).

The correlation of the surface distribution of recent planktic foraminifera with the ambient parameters will help to understand the factors affecting the planktic foraminiferal population in this region. The objective of this work was to understand the planktic

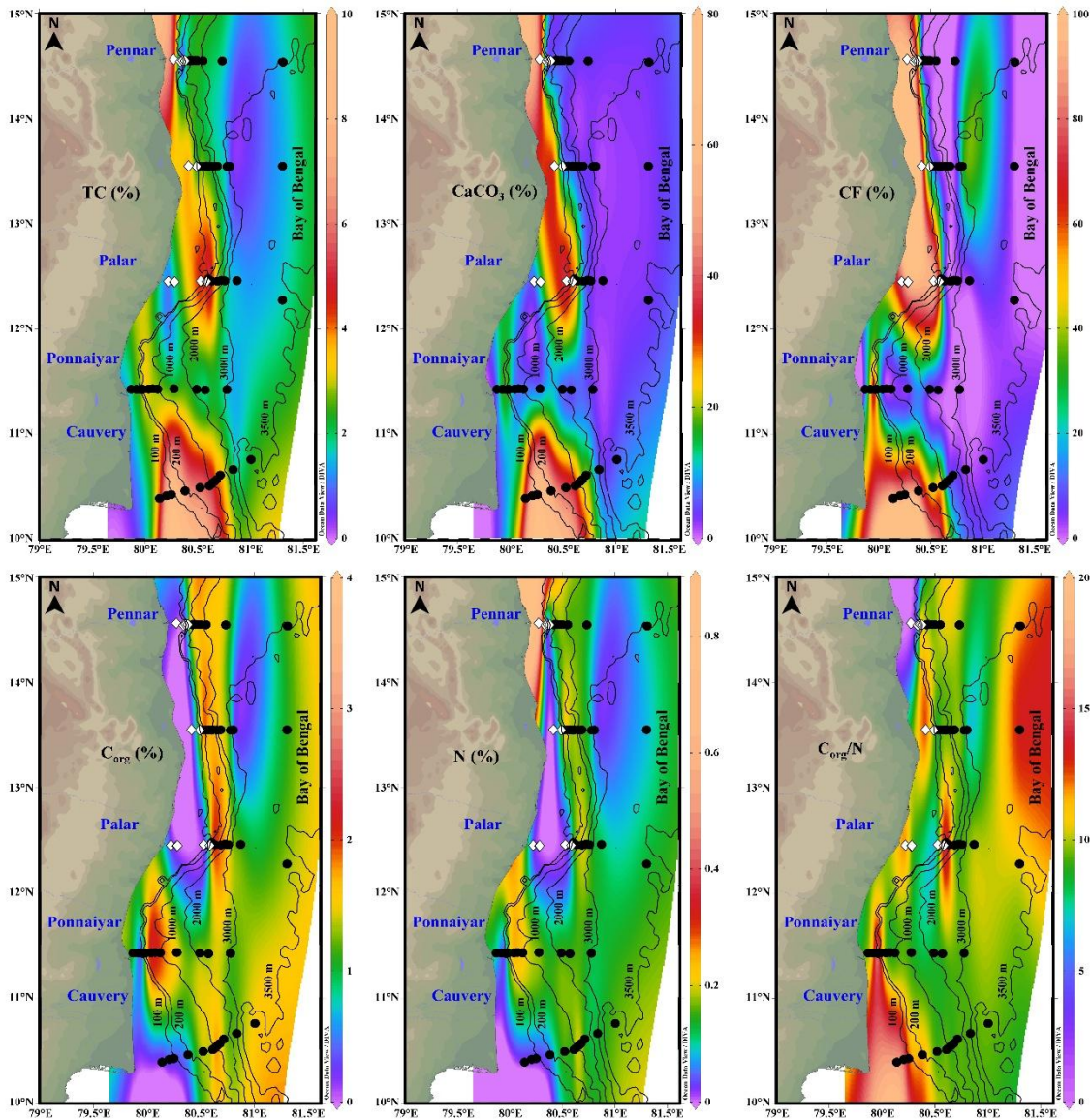


Figure 3.3: The total carbon (TC), calcium carbonate ($CaCO_3$), coarse fraction (CF), organic carbon (C_{org}), Nitrogen (N) percentage and the ratio of organic carbon to nitrogen (C_{org}/N) in the surface sediments of the southwestern Bay of Bengal. The black dots and white diamonds are the sample locations and major rivers are marked by blue lines. The labelled contour lines are bathymetry.

foraminiferal contribution to the carbon burial in the swBoB. Additionally, the factors affecting the planktic foraminiferal abundance and diversity in this region have also been studied to understand the fate of planktic foraminifera in the present climate change scenario.

3.1.2 Results

Ambient parameters

Both, the mixed layer temperature and salinity were highest in the southernmost region and decreased from south to north (**Fig. 3.2**). However, the thermocline temperature increased and salinity decreased from south to north. The thermocline temperature decreased and salinity increased towards the open ocean (**Fig. 3.2**). The lowest mixed layer temperature and salinity were 28.18°C, 33.28 psu and the highest were 28.53°C, 33.58 psu, respectively. For thermocline, the lowest temperature and salinity were 24.42°C, 33.73 psu, and the highest were 28.20°C, 34.50 psu. The annual average Chlorophyll-a concentration was high only in the shelf region throughout the swBoB (**Fig. 3.2**).

The organic carbon varied from the lowest of 0.01% to the highest of 3.61%. The highest C_{org} was on the continental slope, all throughout the swBoB. Incidentally, C_{org} was low along the southernmost transect but the total carbon (TC) was high (**Fig. 3.3**). The rest of the transects had high TC and $CaCO_3$ content only at shallow stations (shelf). $CaCO_3$ abundance varied from 4.17 to 774.53 mg/cm³ (**Table 3.1**), with an average of 102.90 mg/cm³. Interestingly, the highest C_{org}/N ratio was also in the southernmost transect and it decreased with depth. However, an opposite trend was observed in the northern part of the study area. Here, C_{org}/N increased with depth (**Fig. 3.3**). The coarse fraction (>63 µm) percentage was also high along the southernmost transect and for the rest of the transects it was high only at shallow stations (**Fig. 3.3**).

Table 3.1: Sample details with latitude, longitude, depth, absolute abundance (TPN, #/g sediment), shell flux (PF, shells/cm³) of planktic foraminifera, and $CaCO_3$ flux (mg/cm³) in the southwestern Bay of Bengal.

Station	Latitude (°N)	Longitude (°E)	Depth (m)	TPN (#/g sediment)	PF Flux (shells/cm ³)	$CaCO_3$ Flux (mg/cm ³)
MC01	14.5556	80.4010	318	342	98	14.59
MC02	14.5538	80.4159	598	296	64	12.49
MC03	14.5569	80.4214	721	332	68	11.08
MC04	14.5560	80.4385	1117	81	27	16.95

MC05	14.5401	81.3057	2994	132	45	6.25
MC06	14.5553	80.4418	1249	147	26	10.20
MC07	14.5557	80.4695	1556	229	35	10.04
MC08	14.5555	80.4913	1762	379	81	13.69
MC09	14.5532	80.5462	2018	427	70	12.93
MC10	14.5520	80.7283	2526	137	46	6.93
MC11	13.5501	81.3001	3289	147	11	4.17
MC12	13.5486	80.7996	2964	64	48	7.79
MC13	13.5462	80.7742	2735	299	55	9.16
MC14	13.5505	80.6819	2225	141	31	5.62
MC15	13.5486	80.6429	1725	46	11	5.53
MC16	12.4500	80.6833	1106	440	118	16.56
MC17	12.4524	80.6270	731	536	91	
MC18	13.5493	80.6075	1522	83	18	9.01
MC19	13.5449	80.5945	1124	196	59	12.51
MC20	13.5502	80.5784	895	107	54	12.10
MC21	13.5507	80.5708	620	490	153	18.66
MC22	13.5488	80.5687	480	5769	130	19.09
MC23	13.5500	80.5390	256	428	172	14.54
MC24	13.5498	80.5779	878	140	43	12.68
MC25	13.5483	80.5080	154	689	495	30.80
MC26	12.4643	80.5957	168	2237	2428	670.19
MC27	12.4626	80.5952	286	1030	179	36.19
MC28	12.4726	80.5980	470	557	144	32.63
MC29	12.4594	80.7205	1437	282	51	10.34
MC30	12.2700	81.3000	3429	107	36	6.23
MC31	12.4560	80.8670	3254	79	17	5.75
MC32	12.4534	80.7527	2760	81	21	6.99
MC33	11.4244	80.0682	738	231	69	11.04
MC34	11.4202	80.0323	509	323	78	12.28
MC35	11.4193	79.9768	277	430	89	12.22
MC36	11.4189	79.9658	128	227	270	450.14
MC37	11.4143	79.9572	91	55	58	51.60
MC38	11.4178	79.9135	52	281	318	127.83
MC39	11.4188	79.8634	28	10	2	9.10
MC40	11.4203	80.1159	1036	393	114	13.87
MC41	11.4236	80.2714	1480	217	59	9.08
MC42	11.4193	80.4839	2007	450	83	9.42
MC43	11.4132	80.5663	2570	258	81	6.49
MC44	11.4162	80.7722	3112	248	51	6.77
MC45	10.7480	81.0000	3505	426	56	8.77
MC46	10.6535	80.8314	2982	296	54	9.01
MC47	10.5996	80.7096	1695	2170	652	44.17
MC48	10.5567	80.6732	1240	32	129	22.04

MC49	10.5376	80.6521	1009	6193	2509	72.84
MC50	10.5122	80.6229	760	7034	1845	55.25
MC51	10.4979	80.6060	510	12060	6329	172.21
MC52	10.3785	80.1340	24	7	10	470.66
MC53	10.4048	80.2012	52	900	862	540.35
MC54	10.4125	80.2445	73	1111	1134	774.53
MC55	10.4489	80.3770	157	4260	3354	397.48
MC56	10.4812	80.5168	249	5374	4069	181.86
GR01	14.5688	80.2672	26	4	6	668.68
GR02	14.5473	80.3332	52	27	36	45.30
GR03	14.5562	80.3539	66	90	70	24.70
GR04	14.5537	80.3737	104	549	322	35.84
GR05	13.5479	80.4845	78	231	81	13.84
GR06	13.5523	80.4081	29	7	10	421.07
GR07	12.4480	80.2168	29	3	3	54.68
GR08	12.4425	80.2735	78	10	12	154.84
GR09	12.4434	80.2775	46	34	44	183.43
GR10	12.4538	80.5223	76	315	342	429.16
GR11	12.4479	80.5794	129	916	978	229.12

Absolute abundance of planktic foraminifera

The total number of planktic foraminifera varied from 3 to 12060 specimen/g dry sediment. The highest absolute abundance was in the southernmost transect but the rest of the transects had a very low abundance of planktic foraminifera (**Fig. 3.4**). The total planktic foraminifera number was >2000 up to the maximum of 12060 specimen/g dry sediment in the southernmost transect. However, in other transects, the maximum absolute abundance of planktic foraminifera was <2000 specimen/g dry sediment. A majority of the stations had <500 planktic foraminifera/g dry sediment. Additionally, planktic foraminifera were abundant throughout the slope, only along the southernmost transect. In case of all other transects, planktic foraminifera abundance was slightly high only on the upper slope region. The absolute abundance of planktic foraminifera increased up to a depth of 1000 m and decreased after 1000 m. The range of planktic foraminiferal flux varied between 2 and 6329 shells/cm³ (**Table 3.1**), with a mean value of 434 shells/cm³. The highest planktic foraminifera shell flux was along the southernmost transect (**Fig. 3.5**).

Species abundance

A total of 31 planktic foraminifera species belonging to fifteen genera were found in the surface sediments of the swBoB (**Fig. 3.6**). The dominant species (relative abundance

>5%) include *Globigerinita glutinata*, *Globigerina bulloides*, *Globigerinoides ruber*, *Neogloboquadrina dutertrei*, *Trilobatus trilobus*, and *Trilobatus sacculifer* (Fig. 3.7). The dominant species constitute ~76% of the total planktic foraminiferal population (Fig. 3.6). The moderately abundant species (with average relative abundance >1% and <5%) were *Neogloboquadrina incompta*, *Globorotalia menardii*, *Globigerina falconensis*, *Globigerinella siphonifera*, *Globoturborotalita rubescens*, *Globoquadrina conglomerata*, *Globigerinoides tenellus*, and *Pulleniatina obliquiloculata* (Fig. 3.6, 3.8). The rare species with average relative abundance

<1% include *Globorotaloides hexagonus*, *Globigerinella calida*, *Globorotalia unguata*, *Globigerinella radians*, *Turborotalita quinqueloba*, *Orbulina universa*, *Globigerinoides conglobatus*, *Hastigerina pelagica*, *Neogloboquadrina pachyderma* (Fig. 3.6, 3.9). The relative abundance of *Turborotalita humilis*, *Sphaeroidinella dehiscens*, *Beella digitata*, *Globorotalia eastropacia*, *Globorotalia tumida*, *Globigerinella adamsi*, *Neogallitellia vivans*, and *Globigerinita uvula* was <0.2% (Fig. 3.6).

Dominant species

Globigerinita glutinata was predominant in the swBoB (Fig. 3.6). The relative abundance of *G. glutinata* increased with depth except for a few stations. *Globigerinita glutinata* was abundant at the deepest (>3000 m) stations of the two southern transects. However,

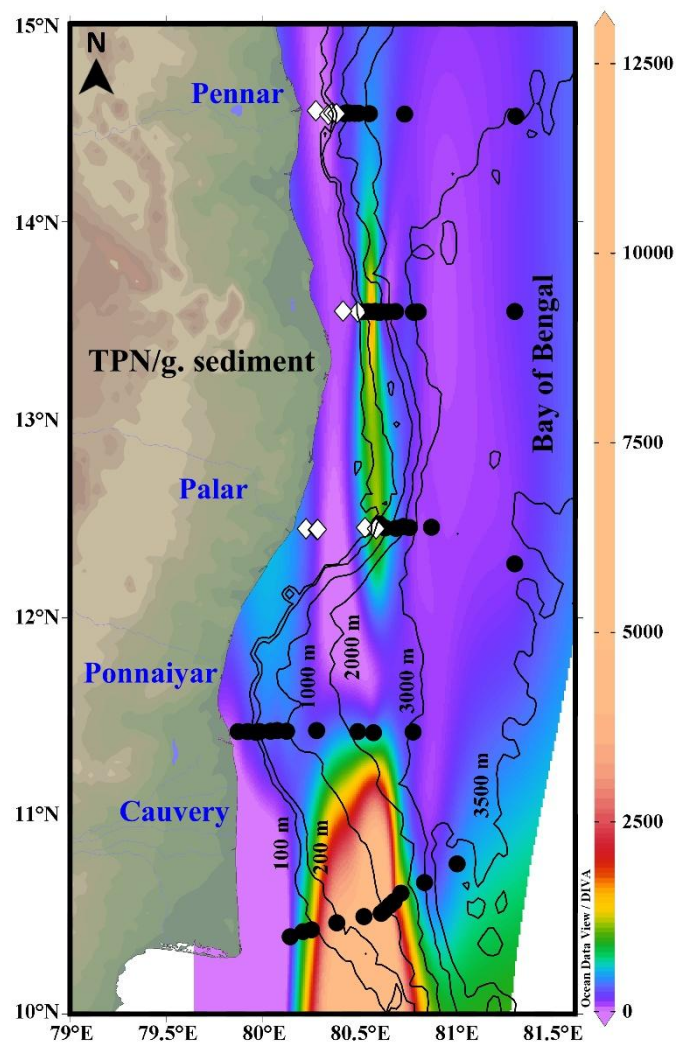


Figure 3.4: The absolute abundance of planktic foraminifera (TPN/g sediment) in the swBoB. The black dots and white diamonds are the sample locations and major rivers are marked by blue lines. The labelled contour lines are bathymetry.

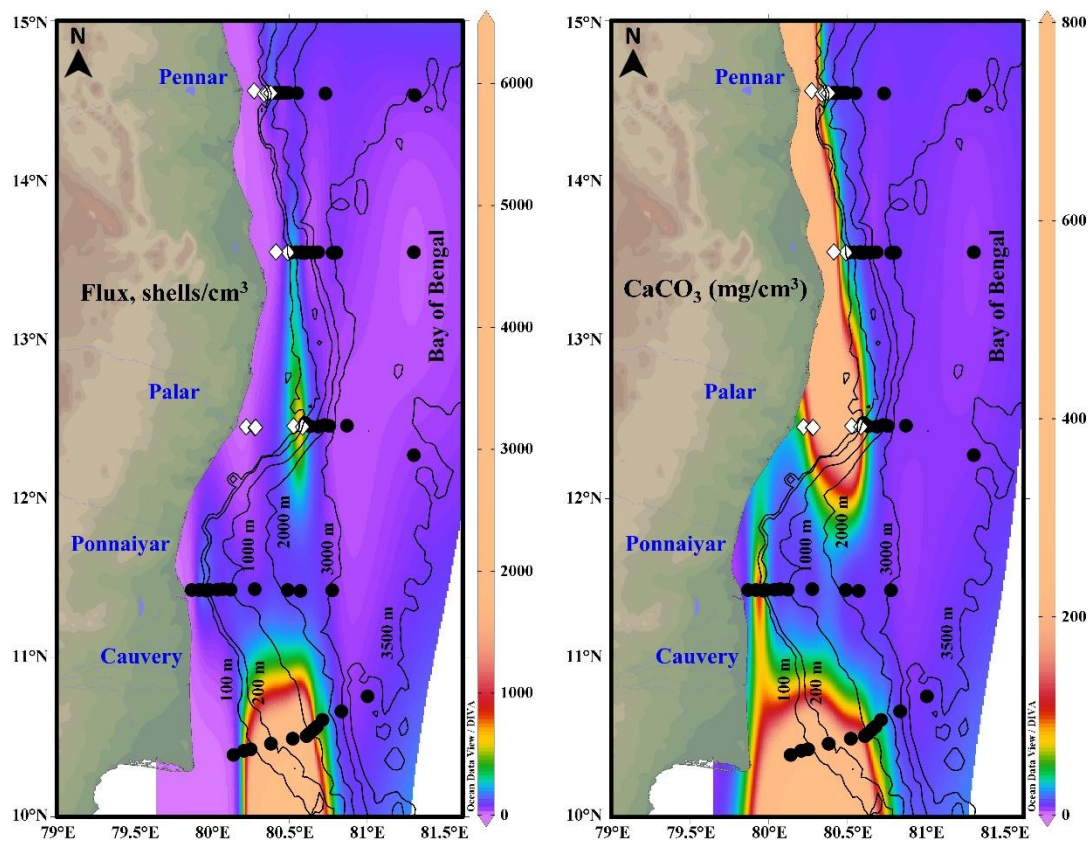


Figure 3.5: The flux rate of total planktic foraminifera (shells/cm³) and calcium carbonate (mg/cm³) in the southwestern Bay of Bengal.

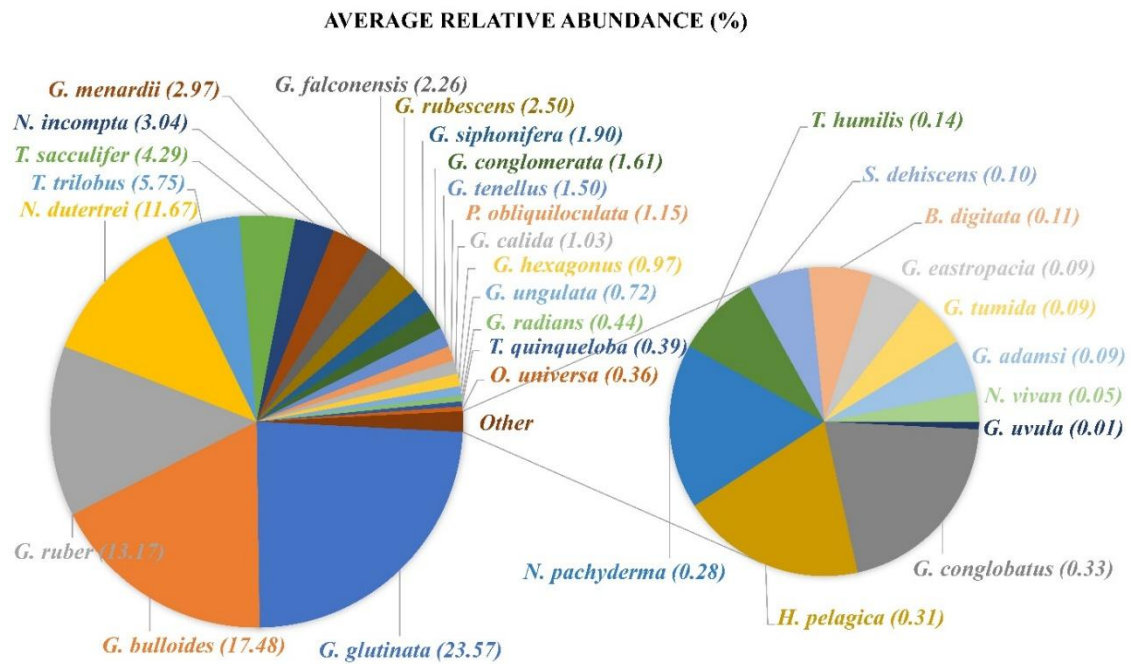


Figure 3.6: Pie diagram showing the average relative abundance of planktic foraminifera species at all the stations in the southwestern Bay of Bengal. The numbers with species names are their average relative abundance.

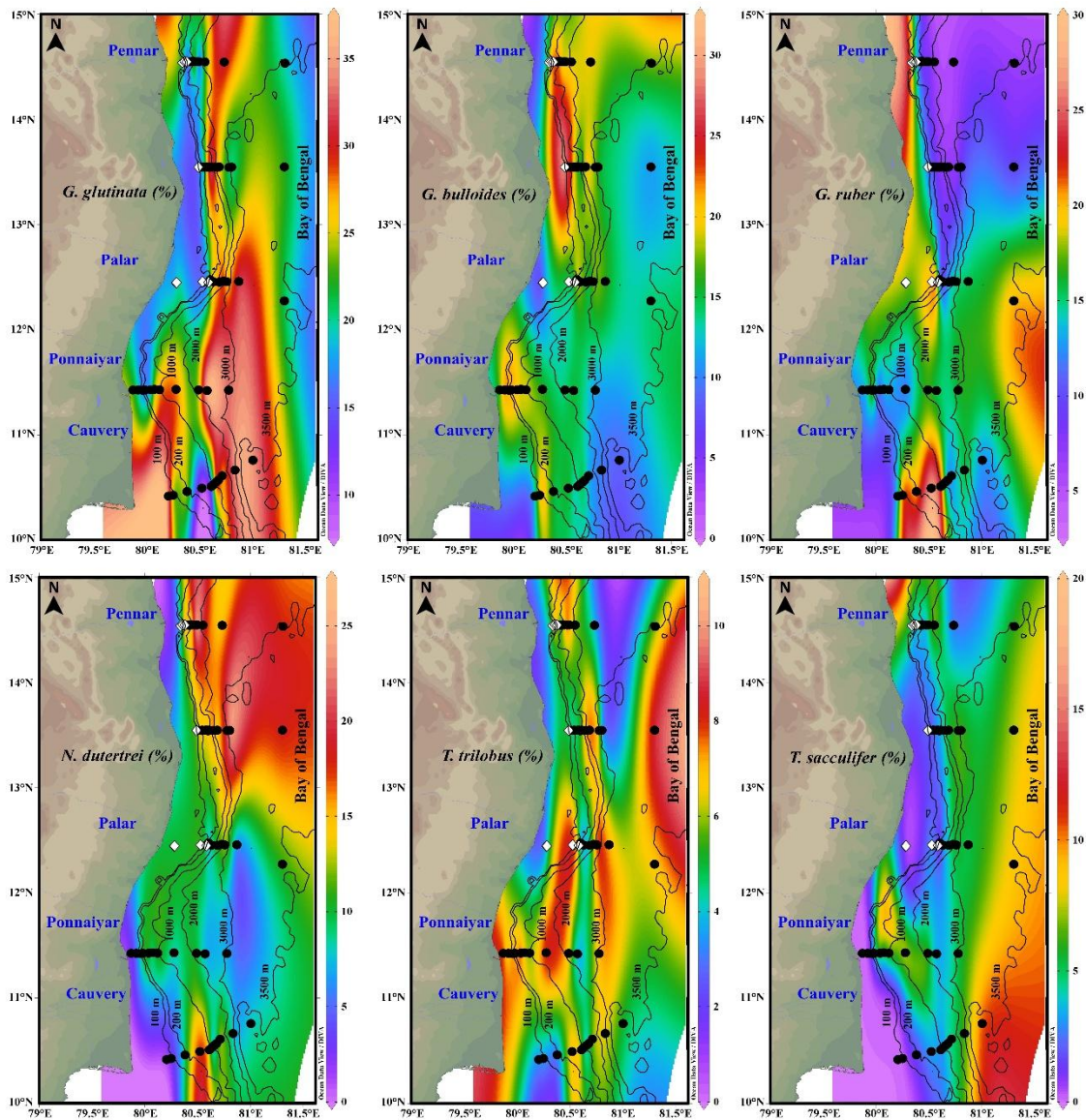


Figure 3.7: The relative abundance (%) of dominant planktic foraminifera species in the southwester Bay of Bengal. The black dots and white diamonds are the sample locations and major rivers are marked by blue lines. The labelled contour lines are bathymetry.

the deepest (>3000 m) stations along the upper 3 transects, had a relatively low abundance of *G. glutinata*. In general, *G. glutinata* was relatively more abundant in the southern part of the study area (**Fig. 3.7**). The second most abundant species was *G. bulloides* (**Fig. 3.7**). The percentage of *G. bulloides* was high on the outer shelf and upper slope region except very shallow stations and decreased with increasing depth. The relative abundance of *G. bulloides* was low off Cauvery River but it was high off Chennai and in the northern part of swBoB (**Fig. 3.7**). The highest relative abundance of *G. ruber* was in the southernmost transect and decreased towards north. Its abundance was high at very shallow stations off Pennar River, while in the southernmost transect, *G. ruber* was abundant on the slope (**Fig. 3.7**). *Neogloboquadrina dutertrei* was more abundant along

the uppermost two transects and its abundance increased with depth. The percentage of *N. dutertrei* was low in the southern part of the study area, except the upper slope region (Fig. 3.7). *Trilobatus trilobus* was abundant on the northern side of the Cauvery Basin.

Moderately abundant species

Neogloboquadrina incompta was moderately abundant on the shelf in front of the Pennar and Cauvery rivers. However, it was more abundant on the upper slope in the middle part of the swBoB. The percentage of *G. menardii* increased with increasing water depth in

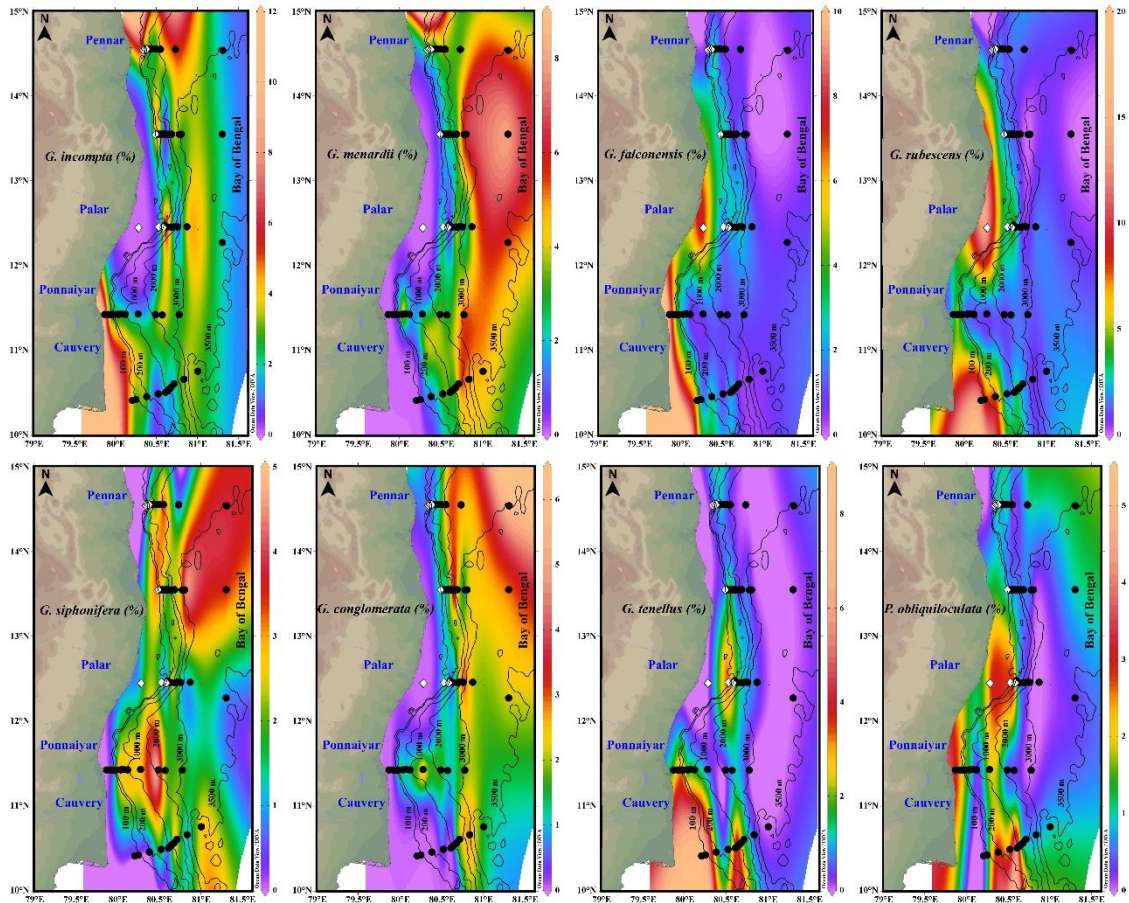


Figure 3.8: The relative abundance (%) of moderately abundant planktic foraminifera species in the southwestern Bay of Bengal. The black dots and white diamonds are the sample locations and major rivers are marked by blue lines. The labelled contour lines are bathymetry.

all the transects (Fig. 3.8). It was rarely present on the entire continental shelf. *Globigerina falconensis* was more abundant only on the shallow shelf region especially near Cauvery River (Fig. 3.8). It was rarely present at all other stations on the outer shelf and further deeper regions. The abundance of *G. bulloides* was low in front of the Cauvery River whereas *G. falconensis* had a high relative abundance in this region (Fig. 3.8). *Globigerinella siphonifera* was less abundant on the shelf and highly abundant at deeper

stations except the southern part (**Fig. 3.8**). The abundance of *G. rubescens* was similar to that of *G. falconensis*. It was abundant at only a very few shallow water stations on the continental shelf. The highest abundance of *G. rubescens* was in the southernmost transect (**Fig. 3.8**). The relative abundance of *G. conglomerata* increased with increasing depth and it was relatively more abundant in the northern part of the study area (**Fig. 3.8**). The surface distribution of *G. conglomerata* was similar to that of *G. menardii* (**Fig. 3.8**). The relative abundance of *G. tenellus* was similar to that of *G. falconensis*, with the zones of high relative abundance confined to the continental shelf, especially in the southern part of the study area. However, it was also abundant at a few stations on the upper slope, unlike *G. falconensis*. The relative abundance of *P. obliquiloculata* was high in the southern part as compared to the northern part of the study area. *Pulleniatina obliquiloculata* did not have any systematic distribution pattern with depth (**Fig. 3.8**).

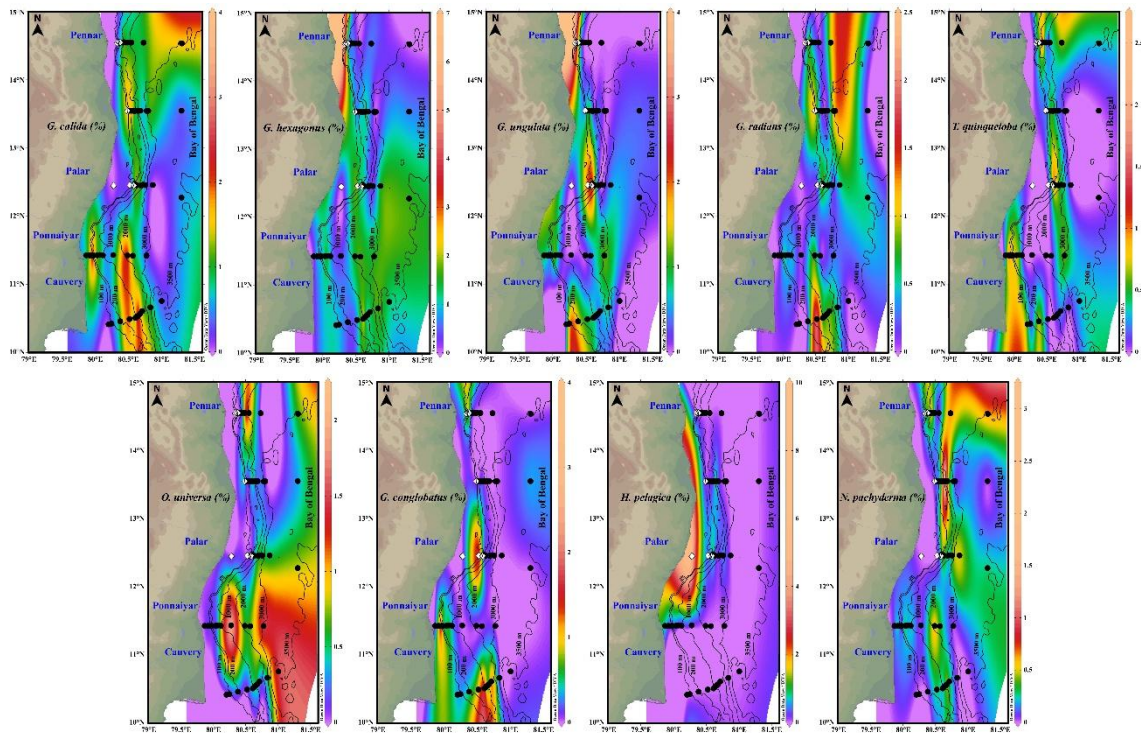


Figure 3.9: The relative abundance (%) of rare planktic foraminifera species in the southwestern Bay of Bengal. The black dots and white diamonds are the sample locations and major rivers are marked by blue lines. The labelled contour lines are bathymetry.

Rare species

The average relative abundance of *G. hexagonus* was 0.97 (%) in the swBoB. Only, one shallowest station in front of the Pennar River, had a high relative abundance of *G. hexagonus*. The abundance of *G. calida* was comparatively high on the slope and it was rare on the shelf, except in the northern Cauvery Basin (**Fig. 3.9**). The surface distribution

of *T. quinqueloba* was similar to that of *G. calida*. *Globorotalia unguolata* was relatively more abundant on the shelf as well as upper slope and was almost absent at deeper stations. Thus, the relative abundance of *G. unguolata* decreased with increasing depth (Fig. 3.9). *Orbulina universa* was absent at most of the stations on the continental shelf. It was more abundant in the southern part as compared to northern part of study area. The relative abundance of *G. conglobatus* was high at a few stations on the shelf and it was absent at the slope except the southern transect. Here, it was more abundant on the slope (Fig. 3.9). *Hastigerina pelagica* was found only at a few stations on the shelf.

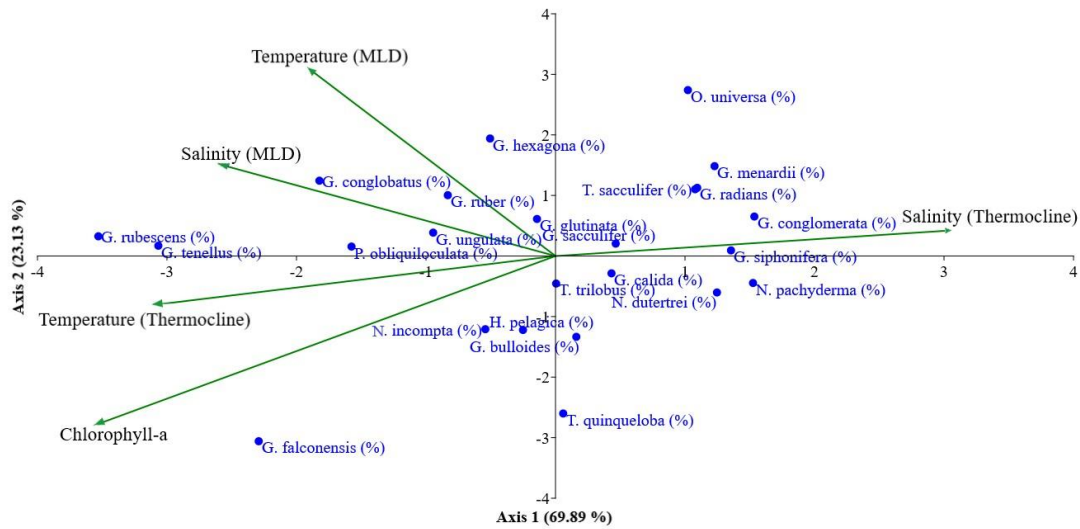


Figure 3.10: Canonical Correspondence Analysis (CCA) plot between ambient parameters and the relative abundance of planktic foraminifera species. The ambient parameters are marked by solid green lines.

Interestingly, the relative abundance of *H. pelagica* was very high (~10%) at station GR08. *Neogloboquadrina pachyderma* was abundant on the upper slope in the northern part of the study area.

Canonical correspondence analysis

The Canonical Correspondence Analysis helped to understand the relation between environmental parameters and planktic foraminiferal species. Axis 1 and Axis 2 represent 69.89% and 23.13% of the variance for the selected species explained with CCA (Fig. 3.10). The permutation (p) values for both axes were the same as <0.01 and the Eigen values were 0.0481 and 0.0159, respectively. The relative abundance of *G. ruber* was positively correlated with the temperature and salinity of the mixed layer (Fig. 3.10).

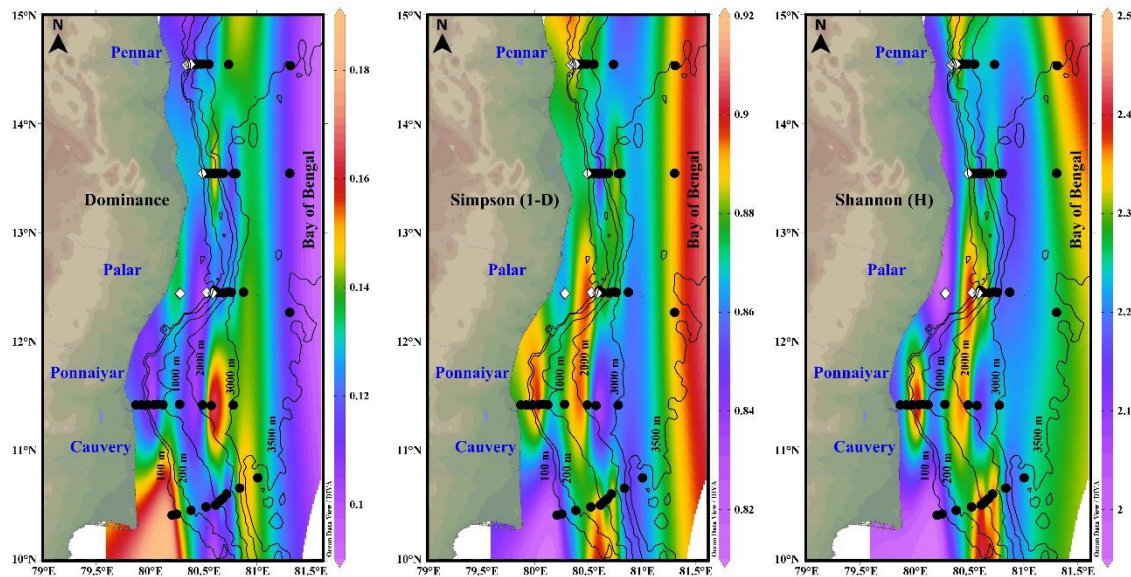


Figure 3.11: The Dominance, Shannon (H) and Simpson (1-D) diversity indices of planktic foraminifera species from the southwestern Bay of Bengal. The black dots and white diamonds are the sample locations and major rivers are marked by blue lines. The labelled contour lines are bathymetry.

A similar preference for warmer temperature and high salinity in the mixed layer was also observed for *G. conglobatus*. The abundance of *G. menardii* was positively correlated with thermocline salinity and negatively correlated with thermocline temperature. However, *G. unguolata* showed an opposite preference, suggesting that *G. menardii* is a deep water and dissolution resistance species, and *G. unguolata* is shallow warm water and dissolution susceptible species (**Fig. 3.10**). Similarly, *N. dutertrei* and *G. conglomerata* were negatively correlated with mixed layer temperature and salinity, as well as thermocline temperature and positively correlated with thermocline salinity. *Neogloboquadrina pachyderma* was also positively correlated with thermocline salinity. Several species (*G. falconensis*, *G. incompta*, *H. pelagica*, *G. bulloides*, and *T. quinqueloba*) were positively correlated with chlorophyll-a, clearly indicating that food availability controls their abundance (**Fig. 3.10**). However, *O. universa*, *G. menardii*, *T. sacculifer* were negatively correlated with chlorophyll-a, suggesting their preference for low productivity oligotrophic waters.

Diversity indices

The stations on the continental shelf and lower slope had low Shannon and Simpson diversity index but dominance was high (**Fig. 3.11**), while on the upper slope, it was opposite. The deepest stations of the three northern transects had a relatively higher planktic foraminiferal diversity. Overall, the planktic diversity was more in the southern

part except at the deepest stations of the three northern transects. A relatively high diversity was also observed on the slope (Fig. 3.11).

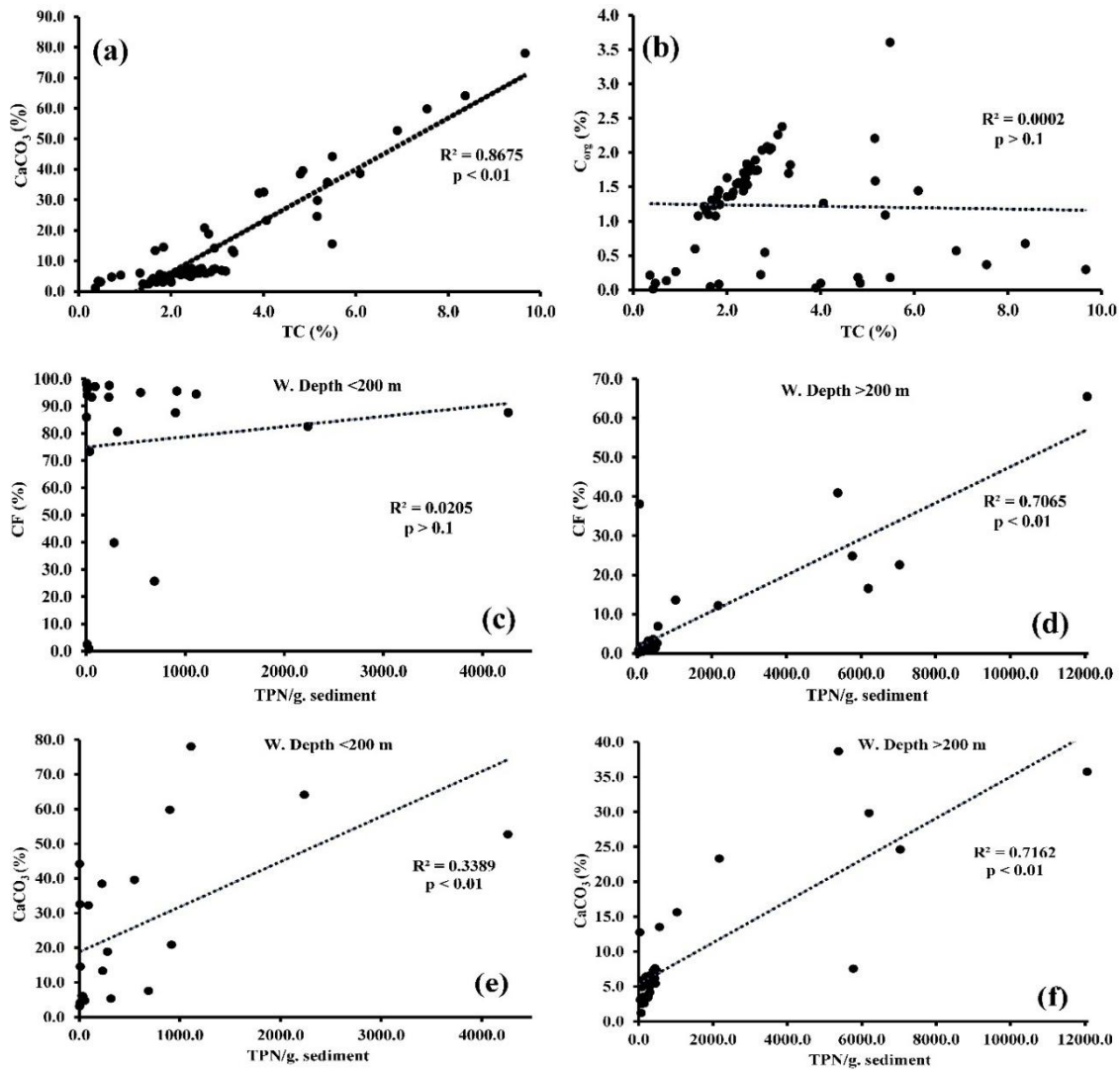


Figure 3.12: The correlation between TC and CaCO_3 (a), TC and C_{org} (b), TPN/g sediment and CF at <200 m w. depth (c), TPN/g sediment and CF at >200 m w. depth (d), TPN/g sediment and CaCO_3 at <200 m w. depth (e), and TPN/g sediment and CaCO_3 at >200 m water depth (f) from the southwestern Bay of Bengal.

Pearson correlation coefficient

The Pearson correlation values were used to understand the positive and negative correlation between the relative abundance of planktic foraminiferal species and different environmental parameters (Table 3.2). The probability values were calculated to test the null hypothesis and to know the significance of the correlation (Table 3.2). Only those correlation values that had p-values 0.01, 0.05, and 0.1 were considered statistically very significant, significant, and insignificant, respectively. Surprisingly, *Globigerinita glutinata* was not strongly correlated with any environmental parameter. However, *G. bulloides* exhibited a robust negative correlation ($R^2 = -0.5064$; $p < 0.01$) with the

temperature of mixed layer (**Table 3.2**). On the contrary, *G. ruber* had a strong positive correlation with both the temperature ($R^2 = 0.4469$; $p < 0.01$) and salinity ($R^2 = 0.4055$; $p < 0.01$) of the mixed layer (**Table 3.2**). *Neogloboquadrina dutertrei* was negatively correlated with temperature ($R^2 = -0.4146$; $p < 0.01$) and salinity ($R^2 = -0.5136$; $p < 0.01$) of the mixed layer and thermocline temperature ($R^2 = -0.2868$; $p < 0.05$) and positively correlated ($R^2 = 0.2799$; $p < 0.05$) with thermocline salinity. *Globorotalia menardii* exhibited a negative correlation ($R^2 = -0.3649$; $p < 0.01$) with thermocline temperature and a positive correlation ($R^2 = 0.3285$; $p < 0.05$) with thermocline salinity (**Table 3.2**). *Globorotalia unguolata* was positively correlated with thermocline temperature ($R^2 = 0.2959$; $p < 0.05$) and negatively with thermocline salinity ($R^2 = -0.3250$; $p < 0.05$). *Globigerinoides conglobatus* was positively correlated with mixed layer salinity ($R^2 = 0.2479$; $p > 0.05$). *Neogloboquadrina incompta* ($R^2 = 0.2610$; $p < 0.05$), *G. falconensis* ($R^2 = 0.7021$; $p < 0.01$), *G. rubescens* ($R^2 = 0.4840$; $p < 0.01$), *G. tenellus* ($R^2 = 0.4235$; $p < 0.01$) were positively correlated with Chlorophyll-a while *N. dutertrei* ($R^2 = -0.3621$; $p < 0.01$), *T. trilobus* ($R^2 = -0.3046$; $p < 0.05$), *T. sacculifer* ($R^2 = -0.4824$; $p < 0.01$), *G. menardii* ($R^2 = -0.5498$; $p < 0.01$), *G. siphonifera* ($R^2 = -0.3411$; $p < 0.01$), *G. conglomerata* ($R^2 = -0.4059$; $p < 0.01$) and *O. universa* ($R^2 = -0.3070$; $p < 0.05$) were negatively correlated with Chlorophyll-a (**Table 3.2**).

3.1.3 Discussion

Carbon burial in the southwestern Bay of Bengal

The mean CaCO_3 (14.19%) was much higher in the swBoB than the north-western and west-central BoB. It was 7.60% in the west-central BoB (Soukhrie et al., 2021) and for the entire north-western and west-central BoB, it was 5.69% including a few data points from swBoB (Saraswat et al., 2017). These values suggest the trend of increasing CaCO_3 content in the surface sediments towards the south in western BoB. The maximum carbon burial was along the southernmost transect (**Fig. 3.3**) suggesting the influence of Gulf of Mannar water during the summer monsoon season. The carbon burial in marine realm occurs by two different pathways (Cartapanis et al., 2018), namely transformation of dissolved inorganic carbon (DIC) into C_{org} through the utilization of DIC by photosynthetic organisms, and extraction of DIC from the surrounding water by calcareous organisms to produce carbonate minerals (CaCO_3) and their burial at the seafloor. The C_{org} burial in marine sediments is modulated by the primary productivity in

Table 3.2: The positive and negative Pearson correlation (R^2) values are given in green and blue colours, respectively. The significant probability (p) values are highlighted in orange colour.

Species	Temperature (MLD) R^2	Temperature (MLD) p	Salinity (MLD) R^2	Salinity (MLD) p	Temperature (Thermocline) R^2	Temperature (Thermocline) p	Salinity (Thermocline) R^2	Salinity (Thermocline) p	Chl-a R^2	Chl-a p
<i>G. glutinata</i> (%)	0.2210	0.0844	0.1638	0.2069	-0.0965	0.4544	0.0169	0.9003	-0.1273	0.3307
<i>G. bulloides</i> (%)	-0.5064	0.0001	-0.2964	0.0233	0.1149	0.3761	-0.0052	0.9672	0.2383	0.0657
<i>G. ruber</i> (%)	0.4469	0.0005	0.4055	0.0009	0.2109	0.1038	-0.1945	0.1351	0.1037	0.4200
<i>N. dutertrei</i> (%)	-0.4146	0.0010	-0.5136	0.0001	-0.2868	0.0267	0.2799	0.0280	-0.3621	0.0035
<i>G. sacculifer</i> (%)	-0.0231	0.8597	-0.1109	0.3929	-0.2238	0.0832	0.2480	0.0549	-0.3046	0.0149
<i>T. trilobus</i> (%)	-0.0618	0.6356	-0.0061	0.9623	0.0532	0.6927	0.0303	0.8261	0.0611	0.6446
<i>T. sacculifer</i> (%)	0.0344	0.7874	-0.1215	0.3535	-0.3735	0.0028	0.3490	0.0054	-0.4824	0.0001
<i>N. incompta</i> (%)	-0.1188	0.3646	-0.0591	0.6519	0.0375	0.7793	-0.1145	0.3582	0.2610	0.0447
<i>G. menardii</i> (%)	0.0347	0.7906	-0.2268	0.0796	-0.3649	0.0043	0.3285	0.0122	-0.5498	0.0001
<i>G. falconensis</i> (%)	-0.0307	0.8206	0.2078	0.1110	0.3888	0.0018	-0.3725	0.0045	0.7021	0.0001
<i>G. siphonifera</i> (%)	-0.1061	0.4134	-0.1509	0.2466	-0.2836	0.0268	0.3309	0.0076	-0.3411	0.0069
<i>G. rubescens</i> (%)	0.2115	0.1049	0.2895	0.0237	0.5500	0.0002	-0.5607	0.0005	0.4840	0.0020
<i>G. conglomerata</i> (%)	-0.1375	0.2895	-0.3176	0.0116	-0.2246	0.0814	0.2249	0.0800	-0.4059	0.0026
<i>G. tenellus</i> (%)	0.2897	0.0204	0.4032	0.0009	0.3366	0.0117	-0.3319	0.0146	0.4235	0.0035
<i>P. obliquiloculata</i> (%)	0.2059	0.1138	0.2363	0.0640	0.2380	0.0626	-0.1774	0.1570	0.2249	0.0795
<i>G. hexagona</i> (%)	0.0655	0.6227	0.0622	0.6535	0.1850	0.1451	-0.2636	0.0501	-0.1145	0.3251
<i>G. calida</i> (%)	0.1112	0.3963	0.1810	0.1677	-0.2866	0.0279	0.3383	0.0085	-0.0890	0.5017
<i>G. unguolata</i> (%)	-0.0659	0.6127	-0.0418	0.7484	0.2959	0.0203	-0.3250	0.0145	0.0925	0.4864
<i>G. radians</i> (%)	0.0501	0.7036	-0.0350	0.7932	-0.2903	0.0254	0.2419	0.0568	-0.2407	0.0634
<i>T. quinqueloba</i> (%)	-0.2277	0.0762	-0.1113	0.4047	0.0464	0.7322	-0.0618	0.6371	0.1272	0.3173
<i>O. universa</i> (%)	0.2133	0.1021	0.1053	0.4207	-0.2892	0.0225	0.2719	0.0359	-0.3070	0.0218
<i>G. conglobatus</i> (%)	0.1848	0.1499	0.2479	0.0509	0.2154	0.0815	-0.1331	0.2820	0.0879	0.4761
<i>H. pelagica</i> (%)	0.0563	0.6664	0.1088	0.4060	-0.1437	0.2585	0.1323	0.2833	0.0482	0.7018
<i>N. pachyderma</i> (%)	-0.1665	0.2141	-0.1812	0.1645	-0.1291	0.2744	0.0884	0.4175	-0.1503	0.1754

the surface water (Middelburg, 2019; Thunell et al., 2000; Calvert and Pedersen, 1992; Pedersen et al., 1992; Calvert et al., 1991), oxygen level in subsurface water (Middelburg, 2019; Paropkari et al., 1993), sedimentation rate/mineral ballast (Hedges and Keil, 1995; Ittekkot et al., 1991), texture of sediments (Calvert et al., 1995), and dilution by terrigenous material (Kandasamy and Nath, 2016; Rullkotter, 2006). However, the CaCO_3 content of the sediments is substantially modified by biological productivity in the overlying water column, dissolution in the water column as well as in sediments (Bhadra and Saraswat, 2022) and dilution by terrigenous matter (Yadav et al., 2022; Govil et al., 2004).

The C_{org} content in the sediments was low on the shelf while the surface water Chlorophyll-a was high. The low sedimentary organic matter despite high water column productivity means poor organic matter preservation in the sediments. The poor C_{org} preservation was because of the coarse-grained texture of the sediments and the influence of oxygenated water (Hedges and Keil, 1995). The C_{org} was also likely to be transported to the deeper depth with finer fraction of sediment due to the influence of riverine influx. The C_{org} content was maximum at intermediate depth (slope) suggesting the high C_{org} burial rate on the slope (Fig. 3.3). But the TC and CaCO_3 percentage were high on the shelf (Fig. 3.3), suggesting that a large part of the TC on the shelf was contributed by the calcareous organisms (foraminifers, pteropods, ostracods, gastropods and others). The strong positive correlation ($R^2 = 0.8675$; $p < 0.01$) between TC and CaCO_3 (Fig. 3.12a)

confirms that a major part of TC was contributed by calcareous shells in the form of calcite mineral (CaCO_3) throughout the swBoB. Incidentally, there was no correlation ($R^2 = 0.0002$; $p > 0.1$) between TC and C_{org} (**Fig. 3.12b**). The low C_{org} and N content along the southernmost transect and high C_{org}/N ratio suggests a high burial rate of terrestrially derived C_{org} . The high TC, CaCO_3 and CF percentages along this transect suggest the maximum burial and sequestration of carbon through calcareous shell flux (**Fig. 3.5**). The quick burial due to higher sedimentation rate decreases the remineralization during the downward flux of the planktic foraminiferal shells in the water column as well as after its settlement in the sediments. The higher burial rates, therefore, are mainly responsible for the better carbonate preservation in this part. Additionally, the higher overall planktic population is also attributed to the influx of more saline and carbonate saturated water from the Arabian Sea. However, the high C_{org} and N in front of the Pennnar River and the low C_{org}/N ratio (**Fig. 3.3**) suggest relatively, higher marine produced C_{org} burial.

The mean C_{org} content and C_{org}/N ratio of sediments were 1.23% and 10.29, respectively, at a mean water depth of 1080 m in the swBoB. The mean C_{org} and C_{org}/N , in the entire western BoB, were 1.19% and 11.16 at a mean water depth of 564 m (Saalim et al., 2021; Suokhrie et al., 2020; Saraswat et al., 2017). The high C_{org} and low C_{org}/N suggests slightly higher C_{org} burial in the swBoB and less influence of terrigenous input. Nevertheless, the C_{org} content in the sediments was relatively higher in the west-central BoB (Saraswat et al., 2017). The high C_{org} in the west-central BoB was attributed to the upwelling in this region (Shetye et al., 1991). The low C_{org} and high C_{org}/N in the extreme northwest BoB (Saraswat et al., 2017) and in front of the Cauvery River suggests high terrigenous input and less C_{org} burial. A similar mean C_{org} (1.2%), but much higher mean C_{org}/N (17.43) were reported at a mean depth of 281 m in the western BoB from the Krishna River mouth to the Mahanadi River mouth (Krishna et al., 2013). However, here the sampling was limited to a maximum water depth of 1209 m. Therefore, the dominance of terrestrially derived organic matter at the shallow depth stations results in higher C_{org}/N . The mean C_{org} (2.43%) at a mean water depth of 984 m in the eastern Arabian Sea along the western margin of India (Paropkari et al., 1993) suggests almost two fold C_{org} burial as compared to the western BoB. The highest carbon burial was in the Gulf of Mannar where the mean C_{org} and C_{org}/N at a mean water depth of 1059 m were 4.2% and 9.47, respectively (Singh et al., 2022).

Absolute abundance of planktic foraminifera

We observed the highest planktic foraminiferal abundance reported so far from the entire western Bay of Bengal (wBoB). In wBoB, the highest abundance, previously reported by Bhadra and Saraswat (2021) was 5086 specimen/g dry sediment. But, in swBoB, the highest planktic foraminiferal abundance was an order of magnitude higher (12000 specimen/g dry sediment) than so far reported from the wBoB (**Fig. 3.4**). The swBoB is a unique basin and is influenced by seasonally reversing monsoon circulation (Shankar et al., 2002; Schott and McCreary, 2001). During the southwest monsoon, summer monsoon current transports 10 Sv of high salinity (0.9 psu higher) water from the seAS into the BoB (Vinayachandran et al., 1999). At the same time, the influx of carbonate-saturated water from the Gulf of Mannar to the Palk Bay is $\sim 16.301 \text{ m}^3/\text{s}$ (Jagadeesan et al., 2013). This high salinity carbonate saturated water further moves into the swBoB (**Fig. 3.2F**) and facilitates the abundance of planktic foraminifera in the southern part of the Cauvery Basin. The temperature, salinity, food availability (Lombard et al., 2011; Fraile et al., 2008; Zaric et al., 2005; Watkins et al., 1996) and dissolution (Wang et al., 2021; de Villiers, 2005; Schiebel, 2002; Peterson and Prell, 1985) are the key parameters affecting planktic foraminiferal abundance. The warmer, saline and carbonate-saturated waters sustain abundant presence of planktic foraminifera in the swBoB. Additionally, planktic foraminifera have been reported to secrete thicker and heavier shells at high salinity (Zarkogiannis et al., 2022). As the thicker and heavier shells sink faster, the exposure time to dissolution is less. Ultimately, the flux of calcareous shells increases.

The absolute abundance of planktic foraminifera increased with increasing water depth up to the upper slope. The highest planktic foraminiferal abundance was observed on the upper slope but it decreased below 1000 m water depth (**Fig. 3.4**). The decrease in planktic foraminiferal abundance at deeper stations suggests the dissolution below the foraminiferal lysocline (Bhadra and Saraswat, 2021). As we move from south to north, the foraminiferal lysocline shoals from 2600 m to 2000 m in the BoB (Cullen and Prell, 1984). The absolute abundance of planktic foraminifera also shows a similar shallowing trend of foraminiferal lysocline from south to north in the swBoB (**Fig. 3.4**).

The planktic foraminiferal population was low throughout the continental shelf as compared to the slope. Additionally, the abundance of planktic foraminifera was very low on the inner shelf (**Fig. 3.4**). The low planktic foraminiferal abundance on the shallow shelf is attributed to several factors. The planktic foraminifera require sufficient water column depth to thrive. A large number of planktic species migrate to deeper depths for

reproduction (Schiebel and Hemleben, 2017). The lack of sufficient water column depth on the shelf restricts the migration of several species and results in low planktic foraminifera abundance. The terrigenous flux is also high on the continental shelf. The low salinity and high turbidity due to terrigenous influx is detrimental to planktic foraminifera (Kucera, 2007). Additionally, the dilution due to terrigenous influx also lowers the planktic foraminiferal abundance on the shelf (Bhadra and Saraswat, 2021; Cullen and Prell, 1984).

The low C_{org} content and high C_{org}/N ratio in surface sediments suggest low productivity along the southernmost transect. The high C_{org}/N ratio in front of the Cauvery River mouth (**Fig. 3.3**), suggests high terrestrial organic matter input. Therefore, the high abundance of planktic foraminifera in the southern part was attributed to the influx of seAS's high salinity and carbonate-saturated water (Trask, 1936). The high terrestrial input means a high energy condition that results in the deposition of coarse-grained sediments (Bejugam and Nayak, 2017). The coarse-grained sediments and the influence of oxygenated water led to the poor preservation of organic matter (Suokhrie et al., 2022; Saalim et al., 2021; Bejugam and Nayak, 2017; Calvert et al., 1995; Calvert and Pedersen, 1992). Despite the riverine influx, salinity was high in this region because the high salinity seAS water mixes with less saline BoB water and thus maintains the high salinity in the southern part during the summer monsoon (**Fig. 3.2B, 3.2F**). However, the C_{org}/N was low in front of the Pennar River mouth and increased with depth (**Fig. 3.3**). This trend in C_{org}/N is because most of the riverine influx is from the northern side.

Planktic foraminiferal contribution to the carbon burial

The high $CaCO_3$ content and relatively low absolute abundance of planktic foraminifera (TPN/g sediment) on the shelf (<150 m) suggests that the planktic foraminifera are not a major contributor of $CaCO_3$ to the surface sediments on the continental shelf of swBoB (**Fig. 3.3, 3.4**). The very weak positive correlation ($R^2 = 0.0205$; $p > 0.1$) between CF and TPN/g sediment (**Fig. 3.12c**) also confirms that the large part of CF on shallow shelf was contributed either by sand fraction or other meio and macro fauna. However, planktic foraminifera mainly contributed to the $CaCO_3$ content in the surface sediment at depths >200 m (**Fig. 3.12f**), all throughout the swBoB, as evident by the strong positive correlation ($R^2 = 0.7162$; $p < 0.01$). The strong positive correlation ($R^2 = 0.7065$; $p < 0.01$) between CF and TPN/g sediment (**Fig. 3.12d**) also confirms that the large part of CF is planktic foraminifera shells. Along the southernmost transect, the high TC, $CaCO_3$,

TPN/g sediment and low C_{org} (**Fig. 3.3, 3.4**) as well as the strong positive correlation between $CaCO_3$ and TPN/g sediment (**Fig. 3.12f**), suggests that the substantial part of TC is of planktic foraminiferal origin in this region.

The average absolute abundance of planktic foraminifera was 257 specimen/g sediment in the north-west and west-central BoB (Bhadra and Saraswat, 2021). However, the average absolute abundance was around four times higher (920 specimen/g sediment) in the swBoB. Similarly, the $CaCO_3$ content was also high in the swBoB relative to other parts of western BoB. The absolute abundance of planktic foraminifera also was relatively higher off Mahanadi River (Bhadra and Saraswat, 2021) where the TIC ($CaCO_3$) percentage was also high (Saraswat et al., 2017). Therefore, planktic foraminiferal shells constituted a large fraction of the $CaCO_3$ on the slope and deeper regions of the wBoB.

Ecological preferences of the dominant planktic foraminiferal species

Globigerinita glutinata, was the most abundant within the swBoB region, with an average relative abundance of 23.55% (**Fig. 3.6, 3.7**). This species dominated the planktic foraminiferal assemblages, indicating its ecological success and adaptability within the local marine environment. Earlier, the dominant presence of *G. glutinata* was also reported from the wBoB (Bhadra and Saraswat, 2021). Interestingly, *G. glutinata* was not the most abundant species in the sediment traps deployed in the Bay of Bengal (Maeda et al., 2022; Gupta et al., 1997). The difference was because the sediment trap studies used $>150\ \mu m$ fraction to pick planktic foraminifera, but we have used a relatively finer fraction ($>125\ \mu m$). Therefore, the size variation of the CF used for picking may create a difference in its relative abundance. The maximum abundance of *G. glutinata* is commonly reported in subtropical to temperate waters (Schmuker and Schiebel, 2002; Schiebel and Hemleben, 2000). But, in nutrient-rich conditions, it is more prominent in tropical waters as well (Mazumder et al., 2009; Schiebel et al., 2001). *Globigerinita glutinata* is a cosmopolitan species with a wide geographical distribution and can tolerate a large range of temperature and salinity (Mazumder et al., 2009; Hemleben et al., 1989). In the swBoB, *G. glutinata* displayed only a subtle positive correlation with R^2 of 0.2210 ($p > 0.05$) for mixed layer temperature and 0.1638 ($p > 0.1$) for mixed layer salinity, not suggesting its strong preference for warmer and saline waters (**Fig. 3.10, Table 3.2**). The higher relative abundance of *G. glutinata* was observed in the southern part as compared to the northern part of swBoB (**Fig. 3.7**), which might suggest the influence of high salinity and nutrient rich water influx from the seAS and Palk Bay. However, in wBoB,

the relative abundance of *G. glutinata* was negatively correlated with mixed layer temperature and suggested to prefer relatively cooler water (Bhadra and Saraswat, 2021). In the northeast Atlantic, food and nutrient availability control the relative abundance of *G. glutinata* (Schiebel et al., 2017; Chapman, 2010; Schiebel et al., 2001; Schiebel and Hemleben, 2000). In the Gulf of Aden, *G. glutinata* was associated with the entrainment of nutrients into the surface waters during the winter monsoon (Ivanova et al., 2003). However, in the Caribbean Sea, *G. glutinata* was associated with eddy-induced productivity (Schmuker and Schiebel, 2002). In the swBoB also, productivity is controlled by the entrainment of eddy-driven nutrients into the surface water during the winter monsoon (Jayalakshmi et al., 2015). Therefore, we conclude that the presence of eddies controls the distribution of *G. glutinata*, rather than any specific change in temperature and salinity in the swBoB.

Globigerina bulloides was the second most abundant species in the swBoB (Fig. 3.7). *Globigerina bulloides* is considered a non-symbiotic species usually found within the upper 60 m of the water column and prominently associated with temperate to sub-polar water masses, as well as upwelling settings in the low latitudes (Schiebel and Hemleben, 2017; Peterson et al., 1991). *Globigerina bulloides* prefers cold and nutrient rich water (Reynolds and Thunnell, 1985; Thompson, 1981). The relative abundance of *G. bulloides* was high on the shelf and upper slope (Fig. 3.7) where the concentration of chlorophyll-a was also high (Fig. 3.2E). A positive correlation with Chlorophyll-a and a negative correlation with mixed layer temperature in the CCA plot as well as a significant negative correlation coefficient with mixed layer temperature ($R^2 = -0.5064$; $p < 0.01$) further confirms its preference for relatively cooler and nutrient rich waters (Table 3.2). During the southwest monsoon, coastal upwelling brings the deeper cold and nutrient-rich water to the surface and results in increased phytoplankton production on the shelf. Although, the upwelling in the swBoB is not as strong as in the northwestern and central western Bay of Bengal, prominent upwelling is reported till the Chennai coast (Shetye et al., 1991). Therefore, the high relative abundance of *G. bulloides* in the northern part of the swBoB along the Chennai coast was ascribed to upwelling during the summer season.

The high relative abundance of *G. ruber* was observed in the southern part of swBoB (Fig. 3.7) where the temperature and salinity both were high. *Globigerinoides ruber* is a symbiont-bearing species that resides within the mixed layer, most frequently found in the tropical to subtropical waters of the global ocean (Bijma et al., 1990a; Hemleben et al., 1989; Bé, 1977). *Globigerinoides ruber* and *G. bulloides* show an

opposite trend which means *G. ruber* prefers high salinity and warm water but *G. bulloides* prefers relatively cooler water. Both, the CCA plot (**Fig. 3.10**) and Pearson correlation values (**Table 3.2**) also confirm that *G. ruber* prefers high salinity and warm water as compared to *G. bulloides*. A similar ecological preference of these species was also reported from the wBoB (Bhadra and Saraswat, 2021). *Globigerinoides ruber* doesn't show any significant correlation with Chlorophyll-a (**Table 3.2**), implying that the food availability is not a major controlling factor for *G. ruber*. It has been reported that *G. ruber* prefers oligotrophic conditions (Ravelo and Andreasen, 1999; Ravelo et al., 1990). The relative abundance of *G. ruber* was high only on the continental shelf. According to Berger (1968), *G. ruber* is a dissolution susceptible species. The low abundance of *G. ruber* at deeper depths suggests higher dissolution in the northern part as compared to the southern part of swBoB. The high relative abundance of *G. ruber* at deeper depth along the southernmost transect (**Fig. 3.7**), suggests better preservation of planktic foraminifera in the southern part of swBoB.

The relative abundance of *N. dutertrei* was negatively correlated with thermocline temperature and positively correlated with thermocline salinity suggesting its cold-water preference (**Fig. 3.10, Table 3.2**). *Neogloboquadrina dutertrei* is frequently found in the tropical to subtropical waters (Schiebel and Hemleben, 2000; Bé, 1977). It was previously reported as a thermocline dweller species (Billups and Spero, 1996; Thunell and Reynolds, 1984; Fairbanks et al., 1982). But, as the thermocline salinity decreased towards the north in the swBoB (**Fig. 3.2D**), the relative abundance of *N. dutertrei* increased (**Fig. 3.7**). A similar high relative abundance of *N. dutertrei* was also reported in the low salinity waters of the northern and western BoB (Bhadra and Saraswat, 2021; Cullen and Prell, 1984; Cullen and Duplessy, 1979). Therefore, *N. dutertrei* can be used as a proxy for paleosalinity reconstruction in the BoB. As the relative abundance of *N. dutertrei* increased with increasing water depth, the high percentage of *N. dutertrei* at deeper depths is ascribed to its dissolution resistant nature (Bhadra & Saraswat, 2022).

Trilobatus sacculifer was negatively correlated with chlorophyll-a suggesting its preference for oligotrophic waters (**Fig. 3.10**). Previously, several authors have reported its preference for oligotrophic surface waters, like *G. ruber* (Schiebel et al., 2004; Conan and Brummer, 2000; Naidu and Malmgren, 1996a). *Trilobatus sacculifer* is a symbiont-bearing mixed layer species and reproduces at ~80 m water depth, during the full moon, probably within the chlorophyll maxima (Bijma and Hemleben, 1994; Bijma et al., 1990a; Hemleben et al., 1989). It was considered as a euryhaline species and can tolerate the

salinity range from 24 to 47 psu, and temperatures from 14 to 32°C (Bijma et al., 1990b). Recently, *T. sacculifer* was assigned to two separate species, namely *Trilobatus trilobus* (without sac) and *T. sacculifer* (with sac). Interestingly, *Trilobatus trilobus* and *T. sacculifer* show an opposite trend. The relative abundance of *T. trilobus* was more in the shallow regions, except a few stations, as compared to *T. sacculifer* and negatively correlated with temperature and salinity of the mixed layer. The relative abundance of *T. sacculifer* was also negatively correlated with temperature and salinity of the mixed layer but positively correlated with thermocline salinity (**Fig. 3.10**). The positive correlation ($R^2 = 0.3490$; $p < 0.01$) with thermocline salinity and negative correlation ($R^2 -0.3735$; $p < 0.01$) with thermocline temperature suggests that *T. sacculifer* is a deep-cold water preferring species that migrates to deeper and cold water during reproduction. The high abundance at deeper depth also indicates that *T. sacculifer* is less prone to dissolution as compared to *T. trilobus*.

Factors affecting moderately abundant species

Neogloboquadrina incompta was more abundant in the swBoB as compared to wBoB (**Fig. 3.8**). The highest relative abundance of *N. incompta* in the wBoB was only 1.38% (Bhadra and Saraswat, 2021). However, in the swBoB, its highest relative abundance was 25%. The average abundance of *N. incompta* was also ~3% in the swBoB (**Fig. 3.6**). Additionally, *N. incompta* showed a weak positive correlation with annual average Chlorophyll-a concentration (**Table 3.2**). In the northeastern Atlantic Ocean also, *N. incompta* was abundant during the low productive summer season (Schiebel and Hemleben, 2000). It has also been reported that at the time of highest food availability, *N. incompta* is replaced by *G. bulloides* and *T. quinqueloba* (Schiebel et al., 2001). Therefore, *N. incompta* prefers oligotrophic conditions in swBoB as compared to wBoB, during the summer.

The average relative abundance of *G. menardii* in swBoB was 2.79% but in wBoB it was 1.15% (Bhadra and Saraswat, 2021). It was also reported that as the temperature and salinity increased from north to south in BoB, the relative abundance of *G. menardii* also increased (Bhonsale and Saraswat, 2012; Cullen and Prell, 1984). However, in swBoB the relative abundance of *G. menardii* decreased towards the south and was higher at the deeper stations of the northern part of the study area (**Fig. 3.9**). In swBoB, both the temperature and salinity of the mixed layer increased towards the south, but the thermocline temperature decreased and salinity increased (**Fig. 3.2**). From the CCA, we

report a positive correlation of *G. menardii* with thermocline salinity and a negative correlation with thermocline temperature. The thermocline water was less saline and warmer at shallow stations (<150 m, water depth). The thermocline salinity increased and temperature decreased with depth creating a large gradient. This gradient does not vary significantly from south to north (**Fig. 3.2**). Therefore, from both the CCA and correlation analysis, it is evident that *Globorotalia menardii* exhibits a positive correlation with thermocline salinity and a negative correlation with temperature (**Fig. 3.10, Table 3.2**). The high abundance of *G. menardii* at deeper stations also suggests that *G. menardii* is a dissolution resistant species. The decreasing abundance of *G. unguolata* with depth was attributed to its dissolution susceptible nature and preference for less saline and warm water of thermocline. The apparent calcification depth for *G. unguolata* is 81 ± 6 m positioned in the upper thermocline, whereas for *G. menardii*, it is 109 ± 8 m situated in the lower thermocline (Stainbank et al., 2019). *Globorotalia unguolata* has a lighter and thinner shell as compared to *G. menardii* (Lamb and Beard, 1972). Therefore, *G. unguolata* is comparatively more prone to dissolution, unlike *G. menardii*. A similar dissolution ranking of these species was also reported by Bhadra and Saraswat (2022) from the north-west Bay of Bengal.

Globigerina falconensis was more abundant on the inner shelf and its abundance decreased towards the open ocean. The distribution pattern of *G. falconensis* is similar to that of *G. bulloides*. The relative presence of *G. falconensis* was high near the Cauvery River mouth (**Fig. 3.8**) where the upwelling was not as prominent as above the Chennai coast (Shetye et al., 1991). However, the winter monsoon influx is high in the southern part of the swBoB, indicating that *G. falconensis* can be used as a winter monsoon species. In the Arabian Sea also, *G. falconensis* was more abundant during the winter monsoon in the zone of vertical mixing and *G. bulloides* was more abundant during the summer season in the regions of strong upwelling (Schulz et al., 2002). Therefore, the change in the occurrence of these two species may be a good proxy to reconstruct the monsoon seasonality. *Globigerina falconensis* showed a very strong positive correlation ($R^2 = 0.7021$; $p < 0.01$) with Chlorophyll-a (**Fig. 3.10, Table 3.2**) suggesting food availability as a major factor controlling its abundance.

The relative abundance of *G. siphonifera* increased towards the north as the salinity decreased (**Fig. 3.2, 3.8**). It means that *G. siphonifera* prefers low salinity waters. *Globigerinella siphonifera* was the most abundant species in the tropical to subtropical region (Bé, 1977). Experimentally, it is determined that the temperature and salinity

tolerance range for *G. siphonifera* is 10-30°C and 27-45 psu, respectively (Bijma et al., 1990b). However, *G. siphonifera* abundance was negatively correlated with thermocline temperature and positively with thermocline salinity in the swBoB (Table 3.2). Bhadra and Saraswat (2021) considered it a dominant species in the wBoB where the freshwater influx was more as compared to the swBoB. Here, *G. siphonifera* was moderately abundant. Therefore, its low abundance in swBoB was attributed to the low freshwater influx. Another moderately abundant species, *Globoturborotalita rubescens* was more abundant on the shallow inner shelf particularly close to Cauvery River mouth (Fig. 3.8). The surface distribution of *G. rubescens* was similar to that of *G. falconensis*. Interestingly, the C_{org}/N ratio was also high in the regions having higher abundance of these species (Fig. 3.3), suggesting their preference for riverine influx derived terrestrial organic matter. The high abundance of *G. rubescens* along the continental margin was also observed by Bhadra and Saraswat (2021) and Bé and Hutson (1977).

The maximum relative abundance of *Globoquadrina conglomerata* was ~6% with an average abundance of 1.53%. *Globoquadrina conglomerata* was negatively correlated with mixed layer salinity ($R^2 = 0.3176$; $p < 0.05$), as well as thermocline temperature and positively correlated with thermocline salinity (Fig. 3.10, Table 3.2). Its abundance increased from south to north (Fig. 3.8), like *N. dutertrei*, *G. menardii*, and *G. siphonifera* (Fig. 3.7) as the thermocline temperature increased and salinity decreased (Fig. 3.2C, 3.2D). Its relative abundance also increased with increasing water depth (Fig. 3.8). *Globoquadrina conglomerata* is absent in the Atlantic Ocean and rare in the oligotrophic regions of the Indian and Pacific Oceans (Schiebel et al., 2004; Bé, 1977; Parker, 1962). But, in swBoB, *G. conglomerata* was moderately abundant. The higher abundance of *G. conglomerata* at deeper depth also suggests its dissolution resistant nature like *N. dutertrei*, *G. menardii*, and *G. siphonifera*. The relative abundance of *G. tenellus* was comparatively high on the outer shelf and upper slope and it was almost absent at lower slope and at further deeper depths. But, in the Arabian Sea, a relatively high number of *G. tenellus* was observed at the lower slope (Naidu, 1993). *Globigerinoides tenellus* was positively correlated with the mixed layer salinity as compared to mixed layer temperature, suggesting its high salinity preference. Pearson coefficient also indicates its relatively strong positive correlation with mixed layer salinity and Chlorophyll-a (Table 3.2).

Pulleniatina obliquiloculata was abundant on the shelf and upper slope of the southern part of the study area. It is a cosmopolitan species (Li et al., 1997; Bé, 1977)

reproducing at a monthly cycle and undergoing gametogenic (Hemleben et al., 1989). *Pulleniatina obliquiloculata* showed a weak positive correlation with mixed layer salinity and temperature as well as thermocline temperature (**Fig. 3.8**). The warm water on the shallow shelf and southern part of the study area facilitates relatively high abundance of *P. obliquiloculata*. Bhadra and Saraswat (2021) also reported that the higher abundance of *P. obliquiloculata* in the wBoB is sustained by the warm water conditions on the inner shelf.

Planktic foraminiferal diversity

The planktic foraminifera were less diverse on the shelf and the diversity increased with depth (**Fig. 3.11**). The low diversity on the continental shelf may be because of the shallow water depth and associated small range of seawater temperature but large salinity fluctuations not being favourable for several species. Additionally, many species migrate to deeper depths to reproduce. The shallow depth restricts the migration (Schiebel and Hemleben, 2017). The turbulence on the shelf is also high, resulting in low diversity. Beyond the shelf, the diversity was high till 2000 m water depth (**Fig. 3.11**). However, planktic foraminifera were more diverse on the shelf in wBoB (Bhadra and Saraswat, 2021). It may be because of the shallow lysocline in northwestern BoB, restricting the diversity on the upper slope. The high diversity on the upper slope region in the present study suggests a deeper lysocline in swBoB. The low species diversity at >2000 m was due to lysoclinical dissolution (**Fig. 3.11**). The surface mixed layer dwelling species like *G. ruber*, *G. bulloides*, *G. falconensis*, *G. rubescens* and *T. trilobus* are more prone to dissolution (Bhadra & Saraswat, 2022). As the tests of dissolution prone species dissolve preferentially, the diversity decreases. The obvious question is why are these surface dwelling species more prone to dissolution? The surface dwelling species live in a warm and less dense water of the euphotic zone and form relatively light, porous, and open-structured shells. The dissolution resistant species such as *G. menardii*, *P. obliquiloculata*, and *N. dutertrei* live below the photic zone where the growth rate is also slow and thus they form relatively thicker and dense shells (Berger and Parker, 1970). The high diversity at the deepest stations of the three northern transects may be due to the presence of more thermocline species. The high salinity and carbonate saturated waters of the Gulf of Mannar and Palk Bay (**Fig. 3.2F**), coupled with the terrestrial influx, strongly dominate in the southern part of the swBoB. Such waters support a large diversity of planktic foraminiferal species in this part of the Bay of Bengal.

3.1.4 Conclusions

The highest carbon burial was observed in the southernmost part of the western BoB. We did not find large variation in C_{org} burial throughout the western BoB. However, the $CaCO_3$ burial was much higher in the southern part of swBoB where the planktic foraminiferal flux was also very high. Therefore, it is evident that the planktic foraminifera contribute significantly to the total carbon burial in the surface sediments of the swBoB. The very high planktic foraminiferal number in the swBoB suggests the biannual influence of the Palk Bay and Gulf of Mannar waters during the summer monsoon season as well as the higher productivity during winter season. Therefore, we conclude that planktic foraminifera can be effectively used to understand the change in the influence of the southeastern AS water in the swBoB. *Globigerinoides ruber* showed a preference for high salinity and warm waters. We infer that the relative abundance of *G. bulloides* and *G. falconensis* responds to the upwelling during summer and conditions during the winter monsoon, respectively. The chaotic distribution pattern of *G. glutinata* may be the result of cyclonic eddies during the winter season. *Neogloboquadrina dutertrei*, *G. menardii*, *G. siphonifera* and *G. conglomerata* were negatively correlated with mixed layer temperature and mixed layer salinity, as well as thermocline temperature and positively correlated with thermocline salinity. However, their spatial distribution from south to north suggests a preference for warm and less saline thermocline water. *Globorotalia menardii*, *P. obliquiloculata*, *N. dutertrei*, *G. conglomerata*, *G. glutinata* and *T. sacculifer* were considered dissolution resistant species due to their high abundance at deeper depths while the surface dwelling species like *G. ruber*, *G. bulloides*, *G. falconensis*, *G. rubescens* and *T. trilobus* were more prone to dissolution. The high abundance of *T. sacculifer* at deeper depths suggests that *T. sacculifer* was less prone to dissolution as compared to *T. trilobus*. The positive correlation of *G. falconensis*, *G. incompta*, *G. bulloides*, *H. pelagica* and *T. quinqueloba* with Chlorophyll-a suggests that the food availability also strongly controlled the planktic foraminiferal abundance. These findings will help to document the factors affecting the chemical composition of planktic foraminifera test and will also help to reconstruct the paleoclimatic and paleoceanographic conditions from the Bay of Bengal.

3.2 Ecology of living benthic foraminifera and lysoclinical dissolution

3.2.1 Introduction

In the northern Indian Ocean, Arabian Sea and Bay of Bengal have a strong oxygen-minimum zone (OMZ) occurring at intermediate depths (Azhar et al., 2017; Sarma et al., 2013; Bristow et al., 2016; Naqvi et al., 1996). Although, both the basins are located at similar latitudes, the OMZ in the Bay of Bengal differs from that in the Arabian Sea. The low bacterial respiration and faster sinking rate of organic matter, due to high riverine influx, do not lead to strong denitrification (Azhar et al., 2017; Naqvi et al., 1996). The Bay of Bengal OMZ is the fourth most significant denitrification centre, but remains incomprehensible. The biogeochemical estimates don't show evidence of a very strong denitrification in the Bay of Bengal. However, it is opined that the Bay of Bengal OMZ is on the verge of functional anoxia (Azhar et al., 2017; Bristow et al., 2016; Naqvi et al., 1996). Therefore, its impact on benthic community has to be assessed in the present global warming scenario.

Benthic foraminifera, single-celled, preferentially marine microorganisms with an external protective shell called a test, dominate the meiobenthic community on the shelf and slope of the marginal marine regions (Lei et al., 2014; Saraswat and Nigam, 2013). Compositionally, their hard exoskeletons are made up either of calcium carbonate or sediment particles (Saraswat and Nigam, 2013). These microorganisms are widely distributed in the world's ocean and found everywhere from the shallow coastal regions to the deepest part of the ocean. They are extremely sensitive to the ambient environment (Nigam et al., 2007). Because of their sensitivity to environmental changes, they are widely used in paleoclimatic studies, including paleo-oxygenation. The abundance, diversity, and morphology of foraminifera are affected by environmental parameters such as temperature, salinity, pH, water depth, dissolved oxygen, and food availability (Suokhrie et al., 2021a; Nigam et al., 2007; Boltovskoy et al., 1991). Therefore, the factors affecting benthic foraminiferal abundance, diversity and composition have to be assessed for their better application in environmental impact assessment and paleoclimatic studies.

The organic matter content and dissolved oxygen concentration are the dominant factors controlling the benthic foraminiferal population, especially in the deep ocean

(Singh et al., 2021; Suokhrie et al., 2021; Saalim et al., 2021; Jorissen, 1995). The organic matter flux and the dissolved oxygen concentration at the bottom and in the sediment pore-water, modulate benthic foraminiferal composition, as well as spatial and vertical distribution in the sediment (Singh et al., 2018; Jorissen et al., 1995, Fontanier et al., 2002). The oxygen tolerance limits of benthic foraminifera are species specific. The preservation of organic matter is higher in low oxygen conditions and the organic matter uptake in such regions is dominated by benthic foraminifera (Nomaki et al., 2005). Therefore, benthic foraminifera can be used to reconstruct the intensity of the past OMZ.

A particular type of benthic foraminifera is reportedly abundant in the OMZ of the Arabian Sea (Nigam et al., 2007). However, the freshwater influx due to monsoonal activity also affects the morphology of foraminifera. If the monsoon intensity is high, the riverine discharge will be more into the sea, thus creating turbulent conditions in the shallow regions. The abundance of rounded symmetrical benthic foraminifera increases in the high energy (shallow) and oxygenated conditions, whereas angular asymmetric benthic foraminifera dominate the calm and oxygen-deficient environment (Nigam et al., 2009). The differentiation between the effect of dissolved oxygen and turbulence on benthic foraminifera morphology has to be assessed for its better application.

The main objective of this work was to assess the factors affecting the ecology and spatial distribution of living (rose-Bengal stained) benthic foraminifera from shallow

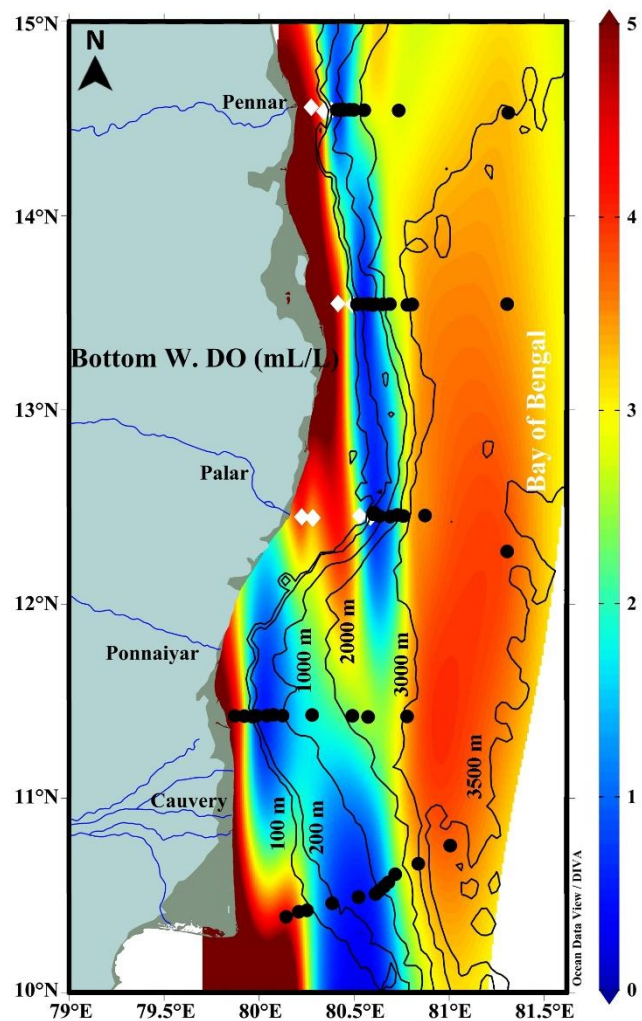


Figure 3.13: Spatial variation in bottom water dissolved oxygen in the southwestern Bay of Bengal. The black dots and white diamonds are the sample locations, and major rivers are marked by blue lines. The labelled contour lines are bathymetry.

(Shelf) to deep ocean floor and their microhabitats. A special emphasis was given to understand the effect of OMZ on the relative abundance of benthic foraminiferal morphogroups. The changes in the relative abundance of calcareous and agglutinated benthic foraminifera has also been used to delineate the benthic foraminiferal lysocline depth.

3.2.2 Results

Dissolved oxygen

The bottom water dissolved oxygen (DO) varied from a minimum of 0.31 mL/L to the

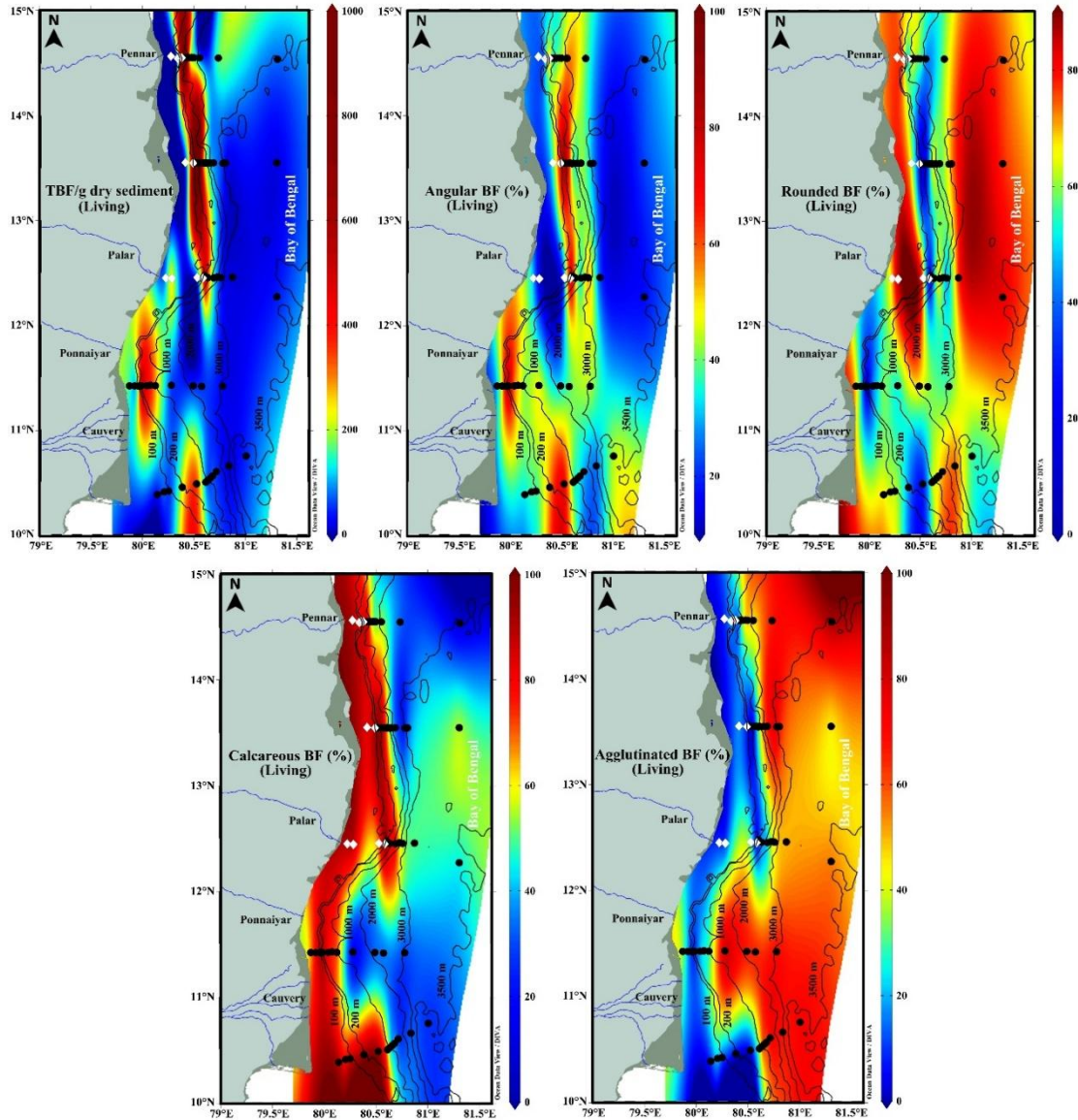


Figure 3.14: Variation in absolute abundance of total living benthic foraminifera (TBF/g dry sediment), relative abundance of living angular benthic foraminifera (LABF %), living rounded benthic foraminifera (LRBF %), living calcareous benthic foraminifera (Calcareous BF %) and living agglutinated benthic foraminifera (Agglutinated BF %) in the southwestern Bay of Bengal. The black dots and white diamonds are the sample locations, and major rivers are marked by blue lines. The labelled contour lines are bathymetry.

maximum of 4.60 mL/L (**Fig. 3.13 & 3.15**). Between a depth range of 150 and 720 m, the dissolved oxygen concentration was ≤ 0.50 mL/L, and it varied from 0.5 to 1 mL/L between 720 and 1125 m. From 1125 to 1800 m, the oxygen concentration ranged between 1 to 2 mL/L, and below 1800 m, it was higher than 2 mL/L (**Fig. 3.13 & 3.15**).

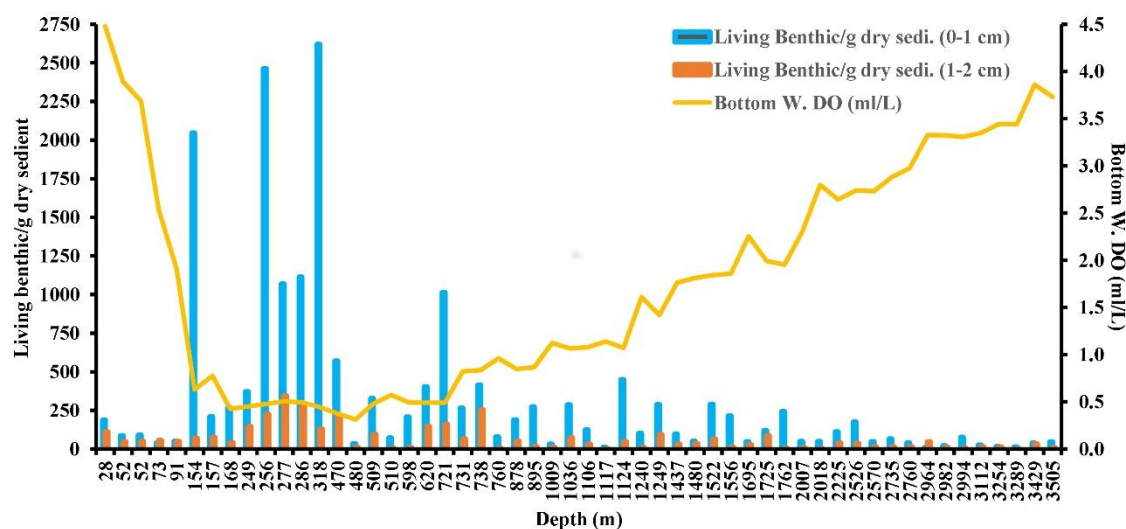


Figure 3.15: The comparison of absolute abundance of living benthic foraminifera in the top 0–1 and 1–2 cm intervals from the southwestern Bay of Bengal. The yellow line is the variation in bottom water dissolved oxygen.

Absolute abundance of living benthic foraminifera in 0–1 and 1–2 cm intervals

The absolute abundance of living benthic foraminifera (TBF/g dry sediment) was estimated for both 0–1 cm and 1–2 cm intervals. The living benthic foraminiferal abundance varied between 4 and 2620 living specimen/g dry sediment (**Fig. 3.14**). The highest abundance of living benthic foraminifera in the 0–1 cm interval was observed in front of the Pennar River (**Fig. 3.14**). The sediments in this area exhibited low C_{org}/N , TC, TIC, and CF, while N levels were high (**Fig. 3.3**).

The lowest abundance of living benthic foraminifera was recorded in front of

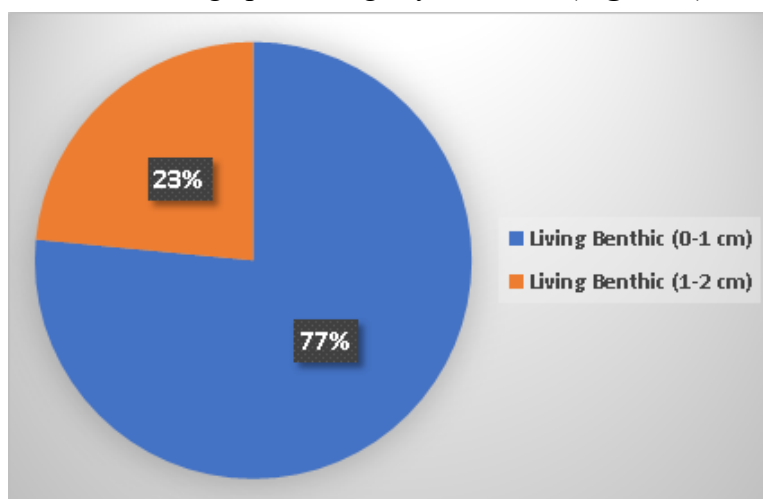


Figure 3.16: The percentage of total living benthic foraminifera population in 0–1 and 1–2 cm intervals from the southwestern Bay of Bengal.

the Cauvery River (**Fig. 3.14**), where the sediments had high C_{org}/N , TC, TIC, and CF, but very low TN. Notably, $\%C_{org}$ was also relatively low in front of the Cauvery River (**Fig. 3.3**). The highest abundance of living benthic foraminifera was found at intermediate depths, where C_{org} content was high, but dissolved oxygen was low (**Fig. 3.13, 3.14, 3.3**). The living benthic foraminiferal (TBF) abundance showed a positive correlation with C_{org} and N, and was negatively correlated with DO (**Fig. 3.18**). In the 1–2 cm interval as well, the peak abundance was at the intermediate water depth (150 to 750 m), well within the OMZ (**Fig. 3.15**). The increased food availability contributes to the higher abundance of living benthic foraminifera at these intermediate depths. The absolute abundances in both 0–1 cm and 1–2 cm intervals were compared (**Fig. 3.15**). Interestingly, the majority of living benthic foraminifera inhabit the top 0–1 cm of sediments, particularly in OMZ, while a significant number also resides in the 1–2 cm intervals in well-oxygenated waters (**Fig. 3.15**). Approximately 77% of the total living benthic foraminifera were found in 0–1 cm interval, with 23% in 1–2 cm interval (**Fig. 3.16**).

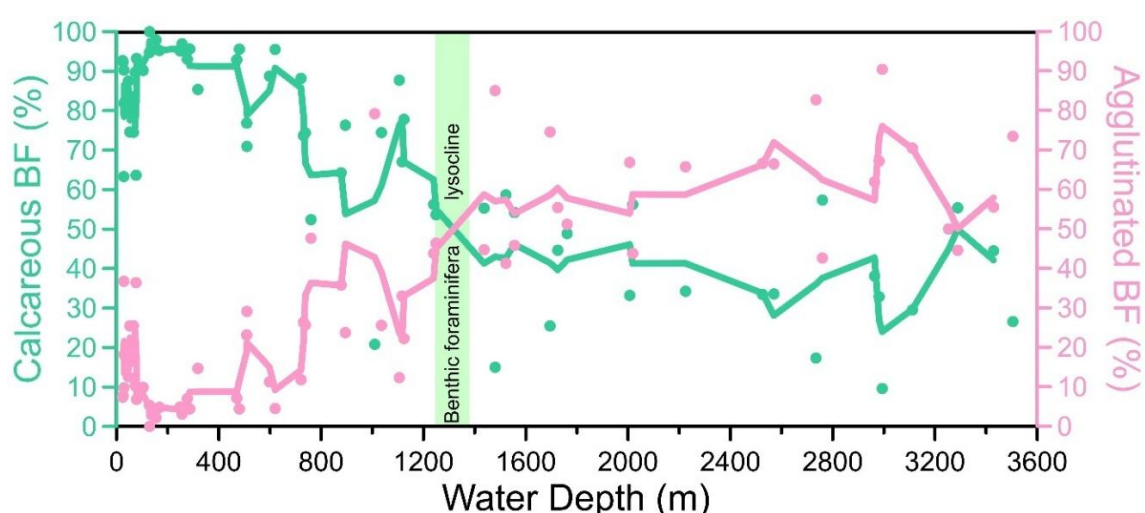


Figure 3.17: Depth-wise variation in relative abundance of calcareous and agglutinated benthic foraminifera from the southwestern Bay of Bengal. The benthic foraminiferal lysocline is marked by a light green colour bar.

Living benthic foraminifera morphogroups

The living benthic foraminifera were categorized into two broad groups, namely calcareous and agglutinated foraminifera. Furthermore, the living calcareous benthic foraminiferal population was further divided into two distinct morphogroups, namely angular and rounded benthic foraminifera. The highest relative abundance of living angular benthic foraminifera (LABF %) was at intermediate depths where $\%C_{org}$ was high and dissolved oxygen was low. The living rounded benthic foraminifera (LRBF %) were

less abundant at intermediate depths but more prevalent on the shallow shelf and deeper regions (**Fig. 3.14**). LRBF showed a very strong positive correlation with DO (**Fig. 3.18**). The relative abundance of living calcareous benthic foraminifera (CBF %) was high on the shelf and upper slope regions shallower than 1250 m, whereas living agglutinated benthic foraminifera (ABF %) were more abundant at depths greater than 1250 m (**Fig. 3.17**). The opposite trend in calcareous and agglutinated benthic foraminifera was attributed to the lysocline dissolution at 1250 m (**Fig. 3.17**). The living agglutinated benthic foraminifera were positively correlated with depth and negatively correlated with pH. Interestingly, it also showed a negative correlation with bottom water temperature and salinity (**Fig. 3.18**). The abundance of CBF was positively correlated with pH (**Fig. 3.18**).

3.2.3 Discussion

Response of benthic foraminifera and their morphogroups to ODZ

The dissolved oxygen abruptly decreased at 150 m and remained low (≤ 0.5 mL/L) until 720 m, which was at its lowest level in the southwestern Bay of Bengal (**Fig. 3.13, 3.15**). Even up to 1800 m, it remained ≤ 2 mL/L. Interestingly, %C_{org} was highest at this depth (**Fig. 3.13**). The high living benthic foraminiferal abundance within the OMZ suggests that the higher food availability with low oxygen level and less predatory pressure leads to an increase in living benthic foraminiferal abundance. Because most of the macrofauna remain under stress in oxygen-deficient environments, that's why the predatory pressure is less. In case of meiofauna, including benthic foraminifera, only opportunistic species thrive in low oxygen environments. The high food availability for these opportunistic species of benthic foraminifera causes a high reproduction rate, resulting in higher total living benthic foraminiferal abundance at intermediate depths (**Fig. 3.14**). Therefore, although, benthic foraminifera are less diverse in oxygen-deficient environments, they are highly abundant (Suokhrie et al., 2021a; Saalim et al., 2021; Singh et al., 2021; Kaithwar et al., 2020).

As the dissolved oxygen concentration increased at depths below lower slope, C_{org} and N content of the sediments decreased, the living benthic foraminifera abundance also decreased (**Fig. 3.14, 3.15, 3.3**). The living benthic foraminiferal abundance was low below 1800 m and on the shelf at depths shallower than 150 m. At these depths, the DO was high and C_{org} content was low (**Fig. 3.15**). The benthic foraminiferal abundance at these depths showed a very strong correlation with C_{org} and N content while negative

correlation with DO (**Fig. 3.18**). This positive response of living benthic foraminifera to organic matter and negative impact of DO is in good agreement with other records from the Bay of Bengal (Suokhrie et al., 2021a, 2020; Saalim et al., 2021) and Arabian Sea (Singh et al., 2021, 2018; Kaithwar et al., 2020; Schumacher et al., 2007; Gooday et al., 2000; Jannink et al., 1998).

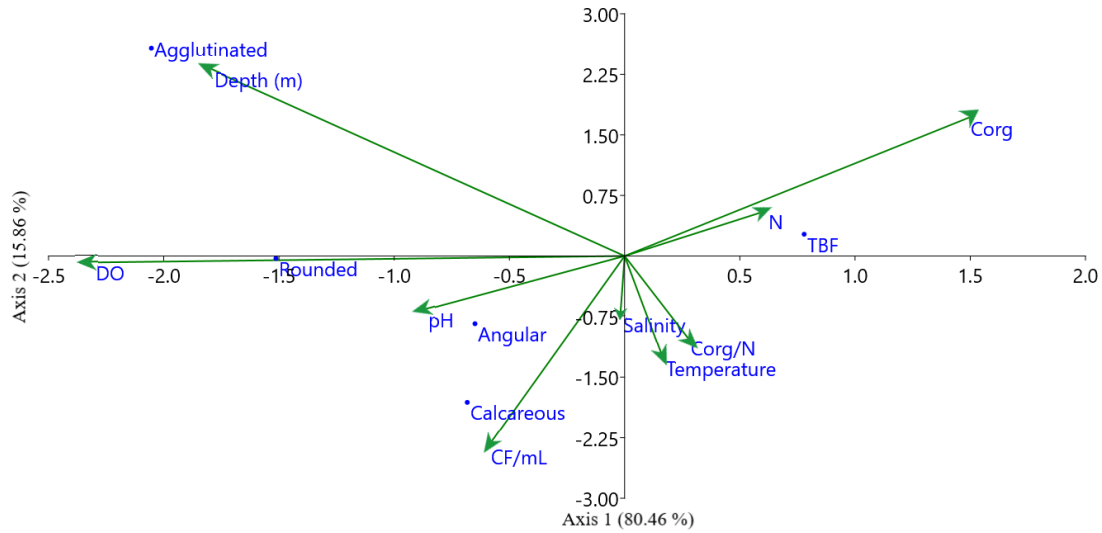


Figure 3.18: Canonical correspondence analysis (CCA) plot between the ambient parameters and the absolute abundance of living benthic foraminifera and the relative abundance of their morphogroups. The solid green lines mark the ambient parameters. Axis 1 and Axis 2 represent 80.46 % and 15.86 % of the variance for the selected morphogroups. The permutation (p) values for both axes were the same as <0.01 and the eigenvalues were 0.248 and 0.049, respectively.

Monsoon is the main driver that controls the marine productivity and the organic matter flux from the land through riverine input. The finer sediments at the intermediate depth preserve more C_{org} (Meyers, 1994). Therefore, the high C_{org} content in marine sediments between 150 and 1800 m is attributed to finer sediments. Additionally, C_{org} content in marine sediments is strongly modulated by dissolved oxygen concentration (Meyers, 1994). The lower dissolved oxygen concentration at intermediate depths provides better preservation of organic matter, which can lead to higher C_{org} content at these depths. However, the higher Chlorophyll-a content suggested higher surface productivity along the continental margin (**Fig. 3.3**). The low C_{org} content on the shallow shelf (>150 m), despite of the high productivity was attributed to coarse-grained sediment and the influence of oxygenated water, which leads to the degradation of organic matter. The high turbulence or disturbance due to riverine influx also lead to lower benthic foraminifera abundance on the shelf (**Fig. 3.14**). The living benthic foraminifera were less abundant along the southern transect in front of Cauvery River where C_{org} content was

less. The DO remained similar as at other comparable depths (Fig. 1), clearly suggesting that the food availability remains the primary factor controlling the benthic foraminiferal abundance. DO comes after it as the second most important factor modulating benthic foraminiferal abundance. The high C_{org}/N ratio along the southern transect also suggests more terrigenous input and less availability of marine-produced organic matter in the marine sediment (**Fig. 3.3**).

The total living benthic foraminifera abundance in both 0–1 and 1–2 cm intervals followed the same bathymetric trend from shallow to deep regions (**Fig. 3.15**). The difference in benthic foraminiferal population between 0–1 and 1–2 cm intervals increased in low oxygen environments (**Fig. 3.15**). Out of the total living benthic foraminifera in the top 2 cm of the sediments, round 90% population lived in the top 0–1 cm in OMZ same as in case of the southeastern Arabian Sea (Singh et al., 2018). However, in shallow oxygenated environments (>150 m), only 50 % population lived in 0–1 cm interval and 50% in 1–2 cm interval (**Fig. 3.15**). Therefore, the bottom water DO plays a crucial role in controlling the average living depth of benthic foraminifera within the marine sediments. Overall, from shallow to deep marine environments, around 77% benthic foraminifera lived in the top 0–1 cm and 23% lived in 1–2 cm (**Fig. 3.16**). Therefore, for studies on living benthic foraminifera, it is essential to consider the top 2 cm of the sediments. Relying solely on the 0–1 cm interval may result in the loss of a substantial fraction (~23%) of the living benthic foraminifera.

The samples collected along coast perpendicular latitudinal transects spanning from shallow continental shelf to the deeper regions between Pennar to Cauvery River also helped to understand the relative effect of turbulence/disturbance due to high energy conditions in shallow environment and bottom water DO on angular and rounded benthic foraminifera. Initially, the idea of change in the relative abundance of angular-asymmetrical and rounded-symmetrical benthic foraminifera to reconstruct the monsoonal conditions was proposed by Nigam et al. (1992). Later, they proposed the rectilinear or angular benthic foraminifera as a proxy for OMZ (Nigam et al., 2007). Therefore, the relative effect of turbulence and dissolved oxygen on these morphogroups is not clear. What exactly controls their abundance? The high CF and C_{org}/N ratio and low C_{org} and N content in front of Cauvery suggests high energy conditions or turbulence, and for Pennar River, the situation was opposite (**Fig. 3.3**). Interestingly, from north to south, there was no major change in either LABF abundance or bottom water DO. In fact, LABF were relatively less abundant in front of the Pennar River where C_{org} content was

the highest and energy conditions were relatively calm, as inferred from low C_{org}/N ratio (**Fig. 3.14, 3.3**). Similarly, LRBF relative abundance should have been more abundant in high energy conditions in front of the Cauvery River, but, it was not. Therefore, it is clear that at intermediate depths DO is the primary factor and food availability becomes the second most important factor modulating LABF and LRBF abundance. In the bathymetric zonation of these two morphogroups, of course, LRBF are more abundant on the high-energy and well oxygenated shallow shelf environments, while LARF are less abundant at these depths (**Fig. 3.14**). However, LABF are highly abundant in relatively calmer conditions on the upper slope and close to shelf break regions where the DO concentration is low and food availability is high. Therefore, within the OMZ region, LABF and LRBF primarily respond to the DO changes and not to the energy conditions. Overall, in the case of angular and rounded BF morphogroups, DO is the primary factor, and then food availability comes second, but for total living BF population, food availability is the primary factor and DO ranks second.

Benthic foraminiferal lysocline

A prominent bathymetric shift in the relative abundance of calcareous and agglutinated benthic foraminifera was observed on the upper slope (**Fig. 3.14**). The abundance of calcareous benthic foraminifera was negatively correlated with depth and positively with pH, while agglutinated benthic foraminifera were positively correlated with depth and negatively with pH (**Fig. 3.18**). The higher abundance of calcareous benthic foraminifera on the continental shelf and upper slope regions and an abrupt decrease at 1250 m suggest a major shift in dissolution. The agglutinated benthic foraminiferal abundance increased to >50% of the total benthic foraminiferal population beyond this depth (**Fig. 3.17**). A similar exceptionally high agglutinated benthic foraminiferal abundance was also reported on the slope of the western Bay of Bengal (Saalim et al., 2018). Similarly, planktic foraminifera and their fragmentation index suggested a significant increase in dissolution after 1000 m depth in the western Bay of Bengal. The shallowest lysocline depth was reported in the western Bay of Bengal compared to other parts of the world's ocean (Bhadra and Saraswat, 2022b). However, the lysocline depth was slightly deeper in the southwestern Bay of Bengal as inferred from the relative abundance of calcareous and agglutinated benthic foraminifera (**Fig. 3.14**). The primary reason for the shallow lysocline in the western Bay of Bengal was an increase in carbonate solubility due to increase in pCO_2 and a decrease in pH of the ambient bottom water and pore water within

the sediments. The calcareous benthic foraminiferal shells start to dissolve below 7.5 pH (Nigam et al., 2008; Saraswat et al., 2015).

The high nutrient input through riverine influx, eddies and coastal upwelling makes the region locally productive (Sarma et al., 2020; Jayalakshmi et al., 2015). Both, the marine-produced and terrestrial-derived organic matter flux contributes to the C_{org} in the marine sediments. A higher DO availability leads to the degradation of organic matter, which releases enough CO_2 to increase the carbonate solubility. In the western Bay of Bengal, there is enough DO available below 1000 m, which increases the degradation of organic matter and supports higher respiration rate (Bhadra and Saraswat, 2022b, Bristow et al., 2017; Suokhrie et al., 2020). The positive correlation of agglutinated benthic foraminifera with DO and negative correlation with C_{org} confirms the dissolution with degradation of organic matter below 1250 m (Fig. 3.3). In the southwestern Bay of Bengal, DO concentrations were always higher than 1.5 mL/L below 1250 m water depth (Fig. 3.15). However, the DO level was relatively low above 1250 m water depth, which may have supported a slightly deeper lysocline at 1250 m in the southwestern Bay of Bengal (Fig. 3.17). The agglutinated benthic foraminifera were negatively correlated with bottom water temperature (Fig. 3.17). The cold water at deeper depths can hold more gases (CO_2), which can promote the carbonate dissolution. The minimum and maximum pH range varied between 7.5 and 8.1 in the western Bay of Bengal (Saalim et al., 2018). However, the pH range was relatively high between 7.6 and 8.3 in the southwestern Bay of Bengal, supporting a slightly deeper lysocline.

3.2.4 Conclusions

From the surface distribution of living benthic foraminifera, it is inferred that low DO levels, particularly at intermediate depth, correlate with high living benthic foraminifera abundance due to increased food availability and reduced predatory pressure. However, as dissolved oxygen concentration increases, a decline in benthic foraminifera abundance is observed, particularly on the shelf and deeper regions where DO is high and C_{org} content is low. The food availability is the primary factor controlling the total living benthic foraminiferal abundance in the southwestern Bay of Bengal. The study also highlights the importance of considering the top 2 cm of sediments to assess the living benthic foraminiferal population, as significant fractions may be overlooked if only the top 1 cm is analyzed. Furthermore, the research underscores the role of turbulence and energy conditions in shaping the abundance of angular and rounded morphogroups of

benthic foraminifera, with DO emerging as the primary factor influencing their distribution. The bathymetric shift in calcareous and agglutinated foraminifera suggested that the foraminiferal lysocline depth lie at 1250 m in the southwestern Bay of Bengal. Overall, the findings contribute to a deeper understanding of benthic foraminifera dynamics in response to environmental changes, particularly in the context of ongoing climate change and its impact on marine ecosystems.

Chapter 4

Precise Timing, Duration, and Impact of 4.2 ka Event on Monsoon Seasonality and Productivity

4.1 Introduction

A large part of the agriculture depends on monsoon for irrigation. Even a minor deviation in monsoon precipitation strongly impacts food grain production (Prasanna, 2014; Gagdil and Gagdil, 2006; Kumar et al., 2004). Therefore, huge efforts are made to predict monsoons. Although several factors are found to affect monsoon, a complete understanding of the monsoon dynamics is still lacking, resulting in uncertain monsoon prediction. The knowledge of the monsoon intensity and its seasonal behavior during the contrasting boundary conditions during the Holocene can help to better understand monsoon dynamics (Banerji et al., 2020). Initially, the Holocene was believed to be climatologically stable. However, the high-resolution records revealed the prevalence of extreme climatic events during the Holocene (Fleitmann et al., 2003; Gupta et al., 2003; Wang et al., 2005). A prominent climatic anomaly that occurred at the boundary between the Northgrippian and Meghalayan ages of the Holocene Epoch is defined as 4.2 ka event, Holocene Event 3, or 4000 yr BP event (Perry and Hsu, 2000). Globally, the 4.2 ka event has been reported as a cold and dry event in the mid to low-latitude regions of the Northern Hemisphere (Ran and Chen, 2019; Banerji et al., 2020; Giesche et al., 2023). The climatic anomaly at 4.2 ka, a megadrought event, led to the collapse of many civilizations including the Egyptian Old Kingdom in the Nile Valley, the Akkadian Empire in Mesopotamia, and the Harappan civilization of the Indus Valley (Railsback et al., 2018; Stanley et al., 2003, Staubwasser et al., 2003; Cullen et al., 2000). The solar activity, volcanic forcing, El-Nino Southern Oscillation (ENSO), Indian Ocean Dipole (IOD) and shift in Inter-Tropical Convergence Zone (ITCZ) are suggested as the possible causes that triggered the 4.2 ka event (Banerji et al., 2020; Toth and Arosen, 2019). The 4.2 ka event doesn't have any associated evidence of meltwater flux from the high-latitude ice sheets unlike the 8.2 ka event (Mayeski et al., 2004). Although, a majority of the studies suggest it as a climatic event, its beginning and end are not properly defined (Sharifi et al., 2015; Berkelhammer et al., 2012; Cullen et al., 2000). The transition regionally differs in its climatological and chronological signatures (Helama et al., 2024).

Indian monsoon, a vital component of the world's climate system, also experienced a disruption during the 4.2 ka event (Saraswat et al., 2023; Sengupta et al., 2020; Saraswat et al., 2016; Prasad et al., 2014; Dixit et al., 2014; Staubwasser et al., 2003; Kudrass et al., 2001). However, the implications of this event are mixed, with some areas experiencing droughts and others receiving extensive rainfall (Cullen et al., 2000; Staubwasser et al., 2003; Booth et al., 2005). These erratic weather patterns likely harmed agriculture and water resource management. The mismatch in climatic implications of 4.2 ka event is likely because of the difference in the monsoon seasonality in different parts of the Indian subcontinent. The Indian subcontinent is at the heart of the Asian monsoon system where it receives a huge summer as well as winter monsoon precipitation (**Fig. 1.1**; Dai et al., 2023). The majority of the rivers drain into the Bay of Bengal. The southwestern Bay of Bengal (swBoB) receives an enormous amount of riverine influx during both the summer and winter monsoon seasons (Sandeep and Pant, 2019; Vinayachandran et al., 2013). The swBoB is also influenced by the seasonal reversal of east India coastal current (EICC) (**Fig. 4.1**). Therefore, the swBoB can be a unique site to reconstruct the monsoon seasonality during the extreme climatic events of the Holocene.

Several workers have reconstructed the summer monsoon intensity in the past, from the western Bay of Bengal (Suokhrie et al., 2024; 2021; Gautam et al., 2021; Ankit et al., 2017; Da Silva et al., 2017; Phillips et al., 2014; Ponton et al., 2012; Rashid et al., 2011; Govil and Naidu, 2011; Kudrass et al., 2001). Many lake sediment, speleothem and pollen records have also been used to reconstruct the summer monsoon intensity during the Holocene from the Indian subcontinent (Dixit et al., 2018; Dutt et al., 2018; Prasad et al., 2014; Berkelhammer et al., 2012). However, the records of winter monsoon intensity during the past, especially during the 4.2 ka event, are limited (Verma et al., 2022). Additionally, most of the marine records from the Bay of Bengal don't have the temporal resolution to precisely define sub-centennial monsoon intensity changes during the Holocene (Verma et al., 2022; Liu et al., 2021; Gautam et al., 2021; Peketi et al., 2021; Ankit et al., 2017; Da Silva et al., 2017; Sijinkumar et al., 2016; Phillips et al., 2014; Ponton et al., 2012; Rashid et al., 2011; Govil and Naidu, 2011; Kudrass et al., 2001). Therefore, high-resolution marine records are required to understand the monsoon seasonality, productivity and carbon burial response to major as well as short-term climatic events of the Holocene.

Planktic foraminifera, owing to their high sensitivity to even subtle environmental fluctuations, serve as invaluable proxies in paleoclimate reconstructions (Salman and

Saraswat, 2024; Saraswat, 2015, Katz et al., 2010; Nigam, 2005). Their fossilized remains provide insights into past oceanographic conditions through species abundance and geochemical signatures preserved in their calcite shells (Schiebel and Hemleben, 2017). Total planktic foraminifera, species abundance, and their assemblages allow for precise reconstructions of past oceanographic conditions and climate dynamics (Kucera, 2007). By integrating faunal assemblages and bulk sediment geochemistry data, planktic foraminifera proxies offer a robust framework for understanding long-term ocean-climate interactions.

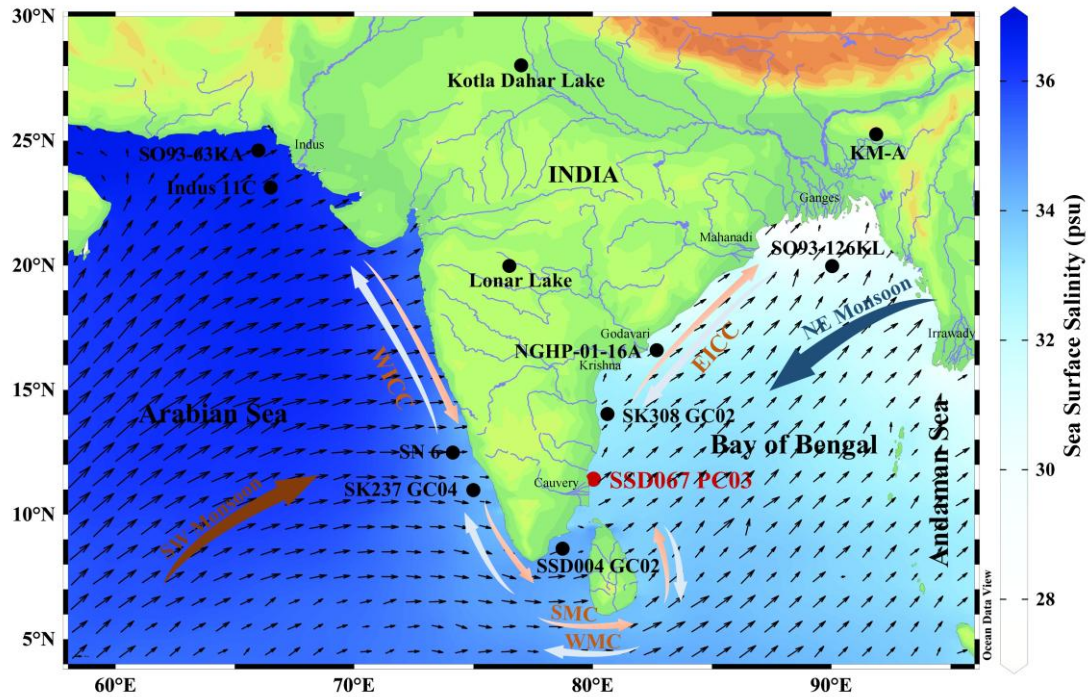


Figure 4.1: The core location used for this study (SSD67 PC03) is marked by red dot and other marine and terrestrial proxy records discussed in the paper are marked by black dots. SK308 GC02 (Suokhrie and Saraswat, 2024), NGHP-01-16A (Ponton et al., 2012), SO93–126KL (Kudrass et al., 2001), SSD004 GC02 (Saraswat et al., 2023), SK237 GC04 (Saraswat et al., 2013), SN 6 (Tiwari et al., 2015), Indus 11C (Giosan et al., 2018), SO90-63KA (Burdanowitz et al., 2020), Loner Lake (Prasad et al., 2014), Kotla Dahar Lake (Dixit et al., 2014b), KM-A (Berkelhammer et al., 2012). The small black arrows show the wind pattern during July month and the bigger orange and blue arrows represent the summer and winter monsoon patterns, respectively. The light orange and white arrows represent surface ocean currents during the summer and winter season along the east and west coast of India. The background is the annual average salinity. The blue lines mark the major rivers.

This chapter includes the first ever decadal resolution multi-proxy marine sediment record of the monsoon seasonality and associated parameters during the 4.2 ka event from the swBoB, a region strongly influenced by winter monsoon. The work is based on the use of planktic foraminiferal assemblages to differentiate between the summer and winter monsoon seasonality.

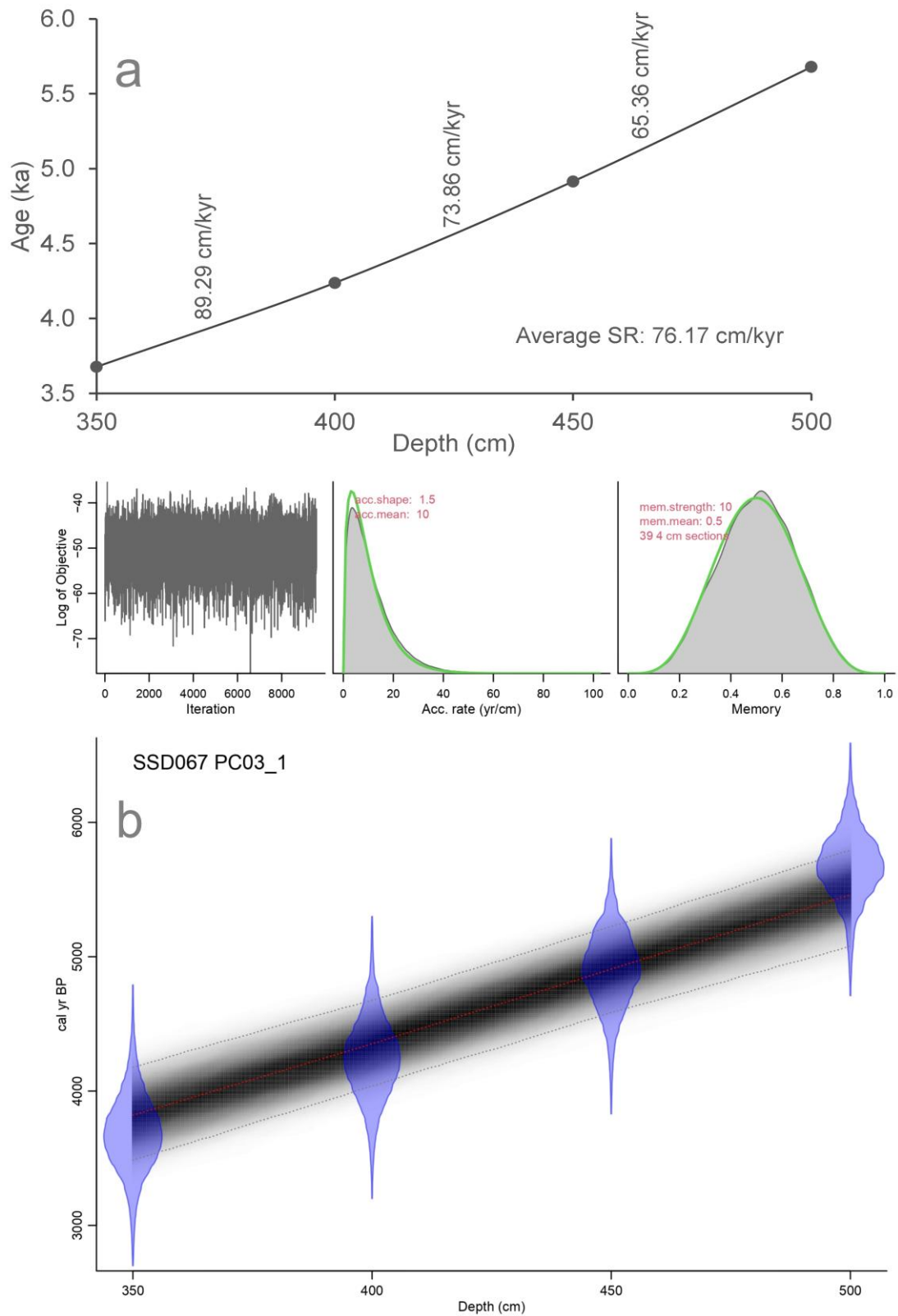


Figure 4.2: Age-depth profile and sedimentation rate in the studied section of core SSD67 PC03. The black and gray gradient represents the modelled uncertainty for different depths. Blue violin plots indicate the posterior probability distributions for age at specific depths. The red dashed line is the median age-depth relationship, while dotted lines represent confidence intervals.

4.2 Objective

Based on the above mentioned observations, the given research objective was finalized.

- To reconstruct the impact of 4.2 ka event on monsoon seasonality and productivity on the Indian subcontinent and adjoining areas.

4.3 Results

This chapter includes the detailed foraminiferal and sediment characteristics of the section of the core covering the time interval between 3.68 ka and 5.68 ka. The lowest sedimentation rate (65.36 cm/ka) in this section was towards the bottom part and the highest (89.29 cm/ka) was in the top part (**Fig. 4.2**). The average sedimentation rate was 76.17 cm/kyr, with one cm representing ~13 years, providing a decadal resolution.

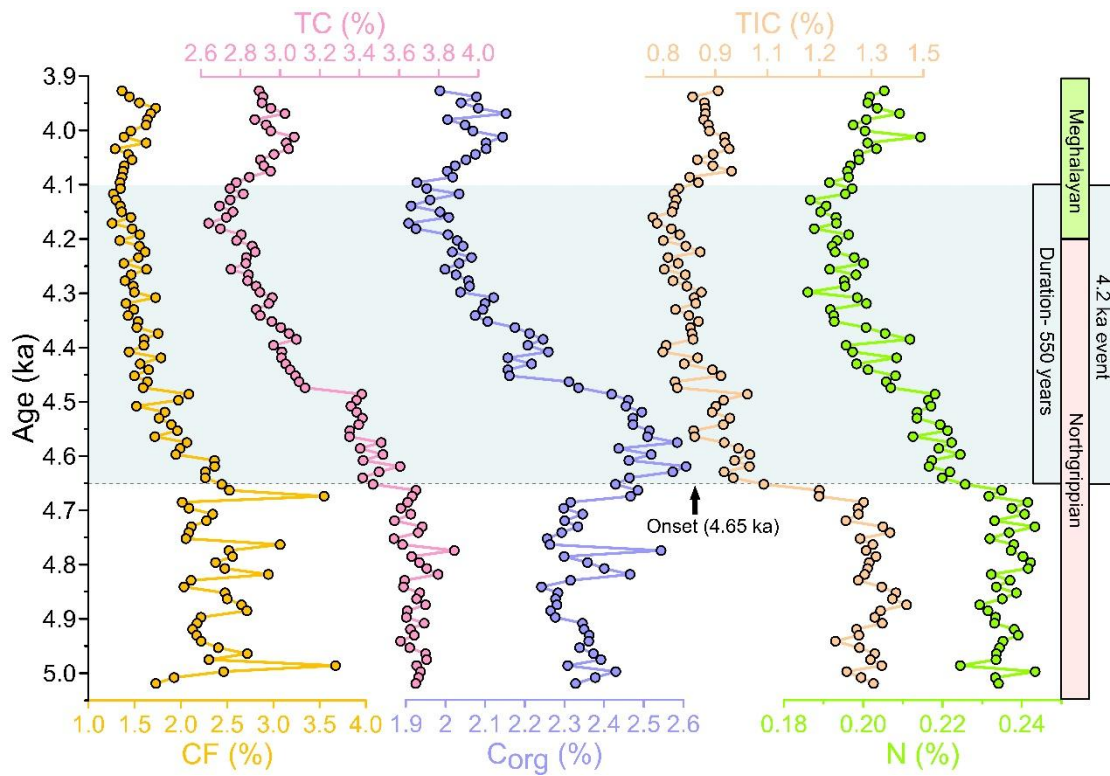


Figure 4.3: Variation in coarse fraction (CF), total carbon (TC), organic carbon (C_{org}), total inorganic carbon (TIC), and nitrogen (N) percentage during the 4.2 ka event in the southwestern Bay of Bengal. The 4.2 ka event is the boundary between Northgrippian and Meghalayan and is marked by the light blue-green bar.

4.3.1 Geochemical characteristics

A very prominent shift was observed in all the geochemical proxies during the Northgrippian-Meghalayan transition, coinciding with 4.2 ka event. Both, the CF and TC began to decrease at 4.65 ka and continued to decrease until 4.1 ka (**Fig. 4.3**). A similar decrease was also observed in C_{org}. Although, the decrease in C_{org} began at 4.5 ka, it also

continued to decrease up to 4.1 ka. The decrease in C_{org} during this interval was $\sim 0.7\%$. However, prior to the decrease, C_{org} abruptly increased by $\sim 0.3\%$, in the latter part of the Northgrippian Age. A similar trend of an abrupt increase in the latter part of Northgrippian Age, followed by a gradual decrease beginning at 4.65 ka and continuing up to 4.1 ka was also observed in C_{org}/N (**Fig. 4.3, 4.5**). C_{org}/N varied between a narrow range of ~ 9.5 and 12, throughout the studied section. Prior to 4.65 ka, the TIC was $\sim 1.4\%$ and it decreased abruptly to 0.9% at 4.65 ka. The decrease in TIC began earlier at 4.68 ka and continued

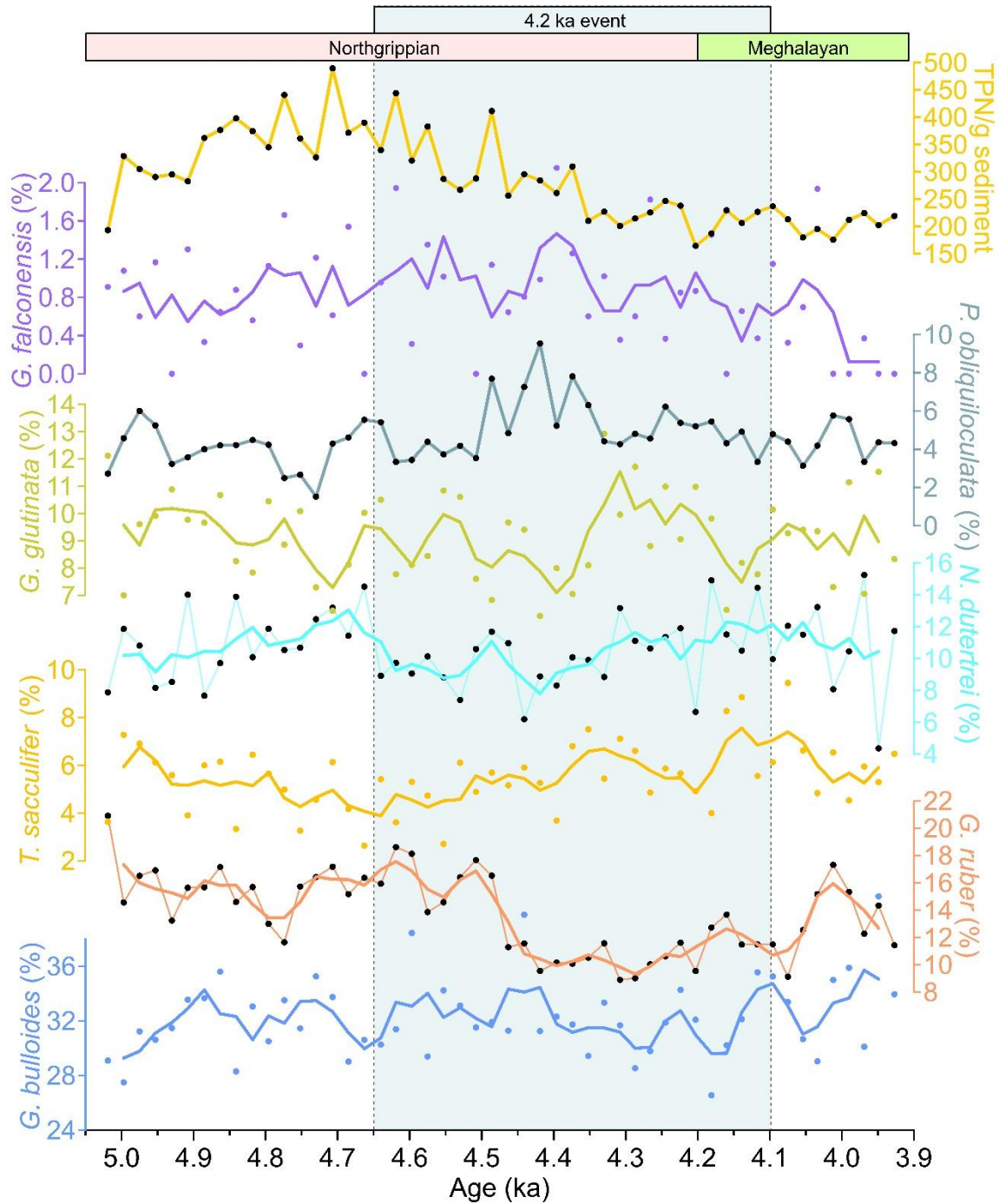


Figure 4.4: Variation in the relative abundance of planktic foraminifera species during the 4.2 ka event in the southwestern Bay of Bengal. The 4.2 ka event is marked by the light blue-green bar.

up to 4.1 ka, like C_{org} . Similarly, N and $CaCO_3$ percentages also decreased abruptly at 4.65 ka and continued until 4.1 ka suggesting the collapse in productivity from 4.65 to 4.1 ka (Fig. 4.3, 4.5).

4.3.2 Relative abundance of planktic foraminifera species and their assemblages

Globigerina bulloides was the most abundant species. The relative abundance of *G. bulloides* increased until 4.45 ka, followed by a subtle decrease between 4.45 to 4.15 ka (Fig. 4.4). A distinct increase in its relative abundance was observed after 4.15 ka. A comparatively more distinct difference was observed in the relative abundance of *Globigerinoides ruber* during the 4.2 ka event. Its relative abundance decreased at 4.46 ka and remained low until 4.05 ka. Prior to 4.46 ka, the relative abundance of *G. ruber* was ~17% and suddenly decreased to 10% after 4.46 ka (Fig. 4.4). The response of *G. bulloides* and *G. ruber* was delayed by 200 years as compared to the change in geochemical characteristics of the sediments. The relative abundance

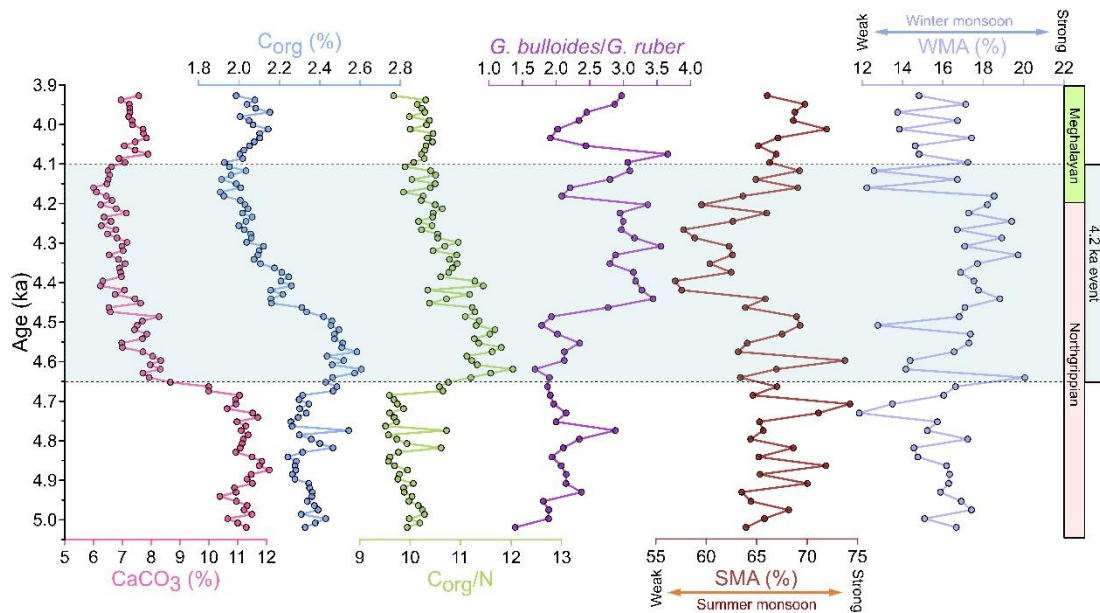


Figure 4.5: Variation in $CaCO_3$, C_{org} , C_{org}/N , *G. bulloides*/*G. ruber*, summer and winter monsoon assemblages (SMA and WMA) during the 4.2 ka event in the southwestern Bay of Bengal. The 4.2 ka event is marked by the light blue-green bar.

of *Neogloboquadrina dutertrei*, a thermocline dwelling species, abruptly decreased at 4.65 ka (Fig. 4.4). An abrupt increase in *Pulleniatina obliquiloculata* was observed at 4.45 ka. The relative abundance of *G. falconensis* was high between 4.65 ka and 4.1 ka (Fig. 4.4). The highest total planktic foraminifera number (TPN) was at 4.7 ka and it gradually declined until 3.9 ka (Fig. 4.4). A distinct decrease in SMA and increase in WMA was observed during the 4.2 ka event suggesting a weaker summer monsoon and

a stronger winter monsoon (**Fig. 4.5**). A very prominent increase in *G. bulloides*/*G. ruber* ratio was observed at 4.45 ka and continued until 4.1 ka (**Fig. 4.5**).

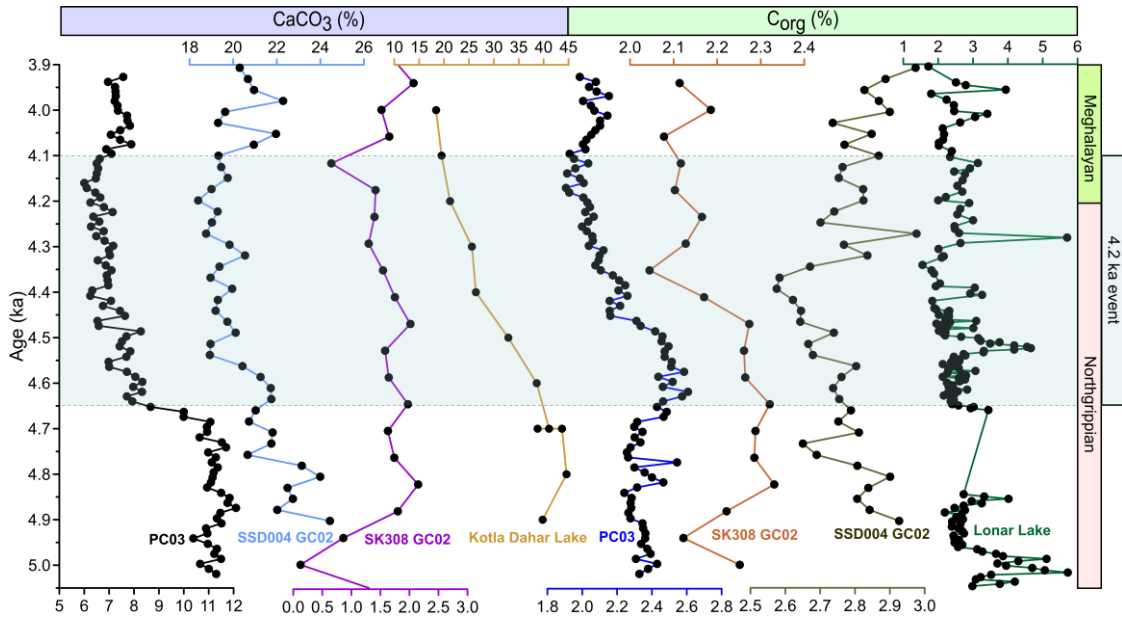


Figure 4.6: Comparison of CaCO_3 and C_{org} in SSD067 PC03 with other marine records from the northern Indian Ocean [SSD004 GC02 (Saraswat et al., 2023); SK308 GC02 (Suokhrrie et al., 2021)], and terrestrial records from India [Kotla Dahar Lake (Dixit et al., 2014b); Lonar Lake (Prasad et al., 2014)].

4.4 Discussion

4.4.1 Onset, duration, and carbon burial response to the 4.2 ka event

Based on the sediment geochemistry and foraminiferal data, we report that the onset of 4.2 ka event was at 4.65 ka and it continued until 4.1 ka (**Fig. 4.3, 4.4, 4.5**). Therefore, the event lasted for close to 550 years. The onset of the event is not uniform in all the records, so much so that in some records, it began at the end of other records (Cullen et al., 2000; Berkelhammer et al., 2012). The duration of the event also varies between different records. In some records, the event lasted for one or two centuries while in others, the changes were observed for a longer duration. The maximum reported duration (550 years) of this event in our record is similar to that reported earlier as well (Railsback et al., 2018). The earliest beginning of the event was reported at 4.7 ka from the Thar desert (Enzel et al., 1999) and the latest termination of the event was reported at 3.41 ka (Psomiadis et al., 2018). Globally, the response to the event may differ in a few years to decades due to the movement of climatic signals and regional response to the event (Railsback et al., 2018). So far, there is no report of the initiation of 4.2 ka event at 4.65 ka and duration of 550 years from the BoB. The precise beginning, termination and the duration of the event could not be ascertained at many a places as most of the records had

poor millennial to centennial or even lower resolution. But in the Gulf of Mannar, Saraswat et al., (2023) used a multi-decadal resolution sediment core and reported that there was a sudden decrease in carbon burial between 4.6 and 4.1 ka. The changes in CaCO_3 and C_{org} were very prominent. Suokhrie et al., (2021) reported an abrupt decrease in C_{org} from the Pennar offshore region. Similarly, Dixit et al., (2014b) also reported an abrupt decrease in CaCO_3 from the Kotla Dahar Lake (Fig. 4.6) that confirm the extent of this event from 4.65 to 4.05 ka in the northern Indian Ocean as well as in the Indian subcontinent. We also report an abrupt decrease in carbon burial beginning at 4.65 ka and continuing until 4.1 ka.

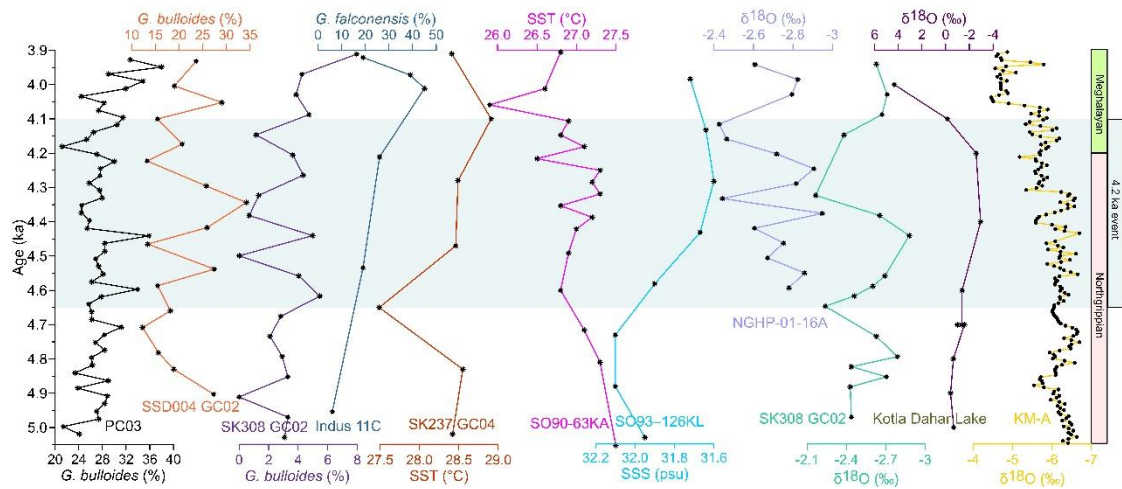


Figure 4.7: The records of the monsoon variability during the 4.2 ka event from the northern Indian Ocean and Indian subcontinent. SSD067 PC03 (This study); SSD004 GC02 (Saraswat et al., 2023); SK308 GC02 (Suokhrie and Saraswat, 2024); Indus 11C (Giosan et al., 2018); SK237 GC04 (Saraswat et al., 2013); SO90-63KA (Burdanowitz et al., 2020); SO93-126KL (Kudrass et al., 2001); NGHP-01-16A (Ponton et al., 2012); terrestrial record from India, Kotla Dahar Lake (Dixit et al., 2014b) and a speleothem records from Northeast India, KM-A (Berkelhammer et al., 2012).

The reduced carbon burial was driven by a decrease in both the organic (C_{org}) and inorganic carbon (CaCO_3). The synchronous decrease in coarse fraction ($>63 \mu\text{m}$), CaCO_3 , C_{org} and nitrogen (N) suggests a collapse in productivity (Fig. 4.3). The decrease in carbon burial in this region can be either due to weaker summer or winter monsoon (Fig. 4.5). However, the winter monsoon was relatively stronger as evident from the higher relative abundance of WMA (Fig. 4.5). The productivity should be high if the winter monsoon was stronger because the productivity in swBoB is high during winter season (Fig. 2.3). The high C_{org}/N ratio suggests increased terrestrial input or the degradation of organic matter. The increased abundance of coarse grained sediments due to high terrestrial input and oxygenated water leads to the poor preservation of organic

matter (Kessarkar and Rao, 2007; Suokhrie et al., 2021; Saalim et al., 2021). The poorly preserved organic matter has low N and thus higher C_{org}/N ratio (Fig. 4.3, 4.5). The synchronous decrease in TPN and $CaCO_3$ is attributed to the low salinity due to the strong winter monsoon. As the summer monsoon weakens, SMC slows down thus reducing the influx of high salinity Arabian Sea water to the study area (Dimri et al., 2016; Vinayachandran et al., 1999). The weaker summer monsoon also means reduced influx of the Gulf of Mannar and Palk Bay's high salinity water in to the swBoB. The calcareous organisms produce lighter shells in less saline water (Zarkogiannis et al., 2022; Trask, 1938), thereby resulting in low $CaCO_3$ flux. The decrease in TPN and $CaCO_3$ was not due to dissolution because the core was retrieved from 509 m water depth, well above the lysocline (1000 m) in the wBoB (Salman and Saraswat, 2024; Bhadra and Saraswat, 2023). Therefore, we conclude that the net carbon burial decreased during the 4.2 ka event as the summer monsoon weakened and the winter monsoon intensified.

4.4.2 Planktic foraminiferal response to the 4.2 ka event

A subtle decrease in *Globigerina bulloides* abundance suggests reduced upwelling and thus the relatively weaker summer monsoon between 4.45 and 4.1 ka. *Globigerina bulloides* is a eutrophic species that prefers cold and nutrient-rich waters. In low-latitude regions, it is associated with upwelling-induced high productivity (Schulz et al., 2002; Naidu et al., 1999; Reynolds and Thunell, 1985; Cullen and Prell, 1984). The response of *G. bulloides* was delayed by 200 years like *G. ruber*, as compared to the carbon burial (Fig. 4.4). *Globigerinoides ruber* prefers well-mixed warmer and oligotrophic conditions (Salman and Saraswat, 2024; Bhadra and Saraswat, 2021; Guptha et al., 1997). An abrupt decrease in *G. ruber* at 4.45 ka suggests the cooler sea surface temperature (SST) between 4.45 and 4.1 ka. Interestingly, the response of *Neogloboquadrina dutertrei* was synchronous with carbon burial and was very distinct. The relative abundance of *N. dutertrei* decreased from the high of ~14% at 4.65 ka to the lowest of ~6% during the 4.2 ka event. *Neogloboquadrina dutertrei* is a thermocline species that prefers warmer and less saline water (Guptha et al., 1997). The high abundance of *N. dutertrei* was also reported in low salinity regions from the northern and western BoB (Salman and Saraswat, 2024; Bhadra and Saraswat, 2021). From an abrupt decrease in the relative abundance of *N. dutertrei* at 4.65 ka BP, we infer a drop in temperature and an increase in salinity of the thermocline. *Pulleniatina obliquiloculata* thrives in sub-surface waters at the lower limit of thermocline. Therefore, it can be used as a proxy for the changes in

thermocline depth (Ravelo et al., 1990; Baohua et al., 1997). Based on the plankton tow studies, *P. obliquiloculata* was suggested to be abundant during the winter season (Bé and Tolderlund 1971; Ravelo and Fairbanks, 1992). The high abundance of *Pulleniatina obliquiloculata* during 4.2 ka event suggests the deepening of thermocline due to stronger winter monsoon and weaker summer monsoon. Interestingly, the increase in *Globigerina falconensis* also suggests a stronger winter monsoon between 4.65 to 4.1 ka (Fig. 4.4). *Globigerina falconensis* is also a eutrophic mixed layer species that prefers high food availability. It was abundant in the winter monsoon-dominated regions of the northern Indian Ocean (Salman and Saraswat, 2024; Schulz et al., 2002). Therefore, the high abundance of *G. falconensis* suggests the stronger winter monsoon during the 4.2 ka event (Fig. 4.4). The highest planktic foraminiferal abundance was at 4.65 ka and gradually declined until 4.1 ka suggesting the highest productivity at the onset and lowest at the end of the 4.2 ka event (Fig. 4.4).

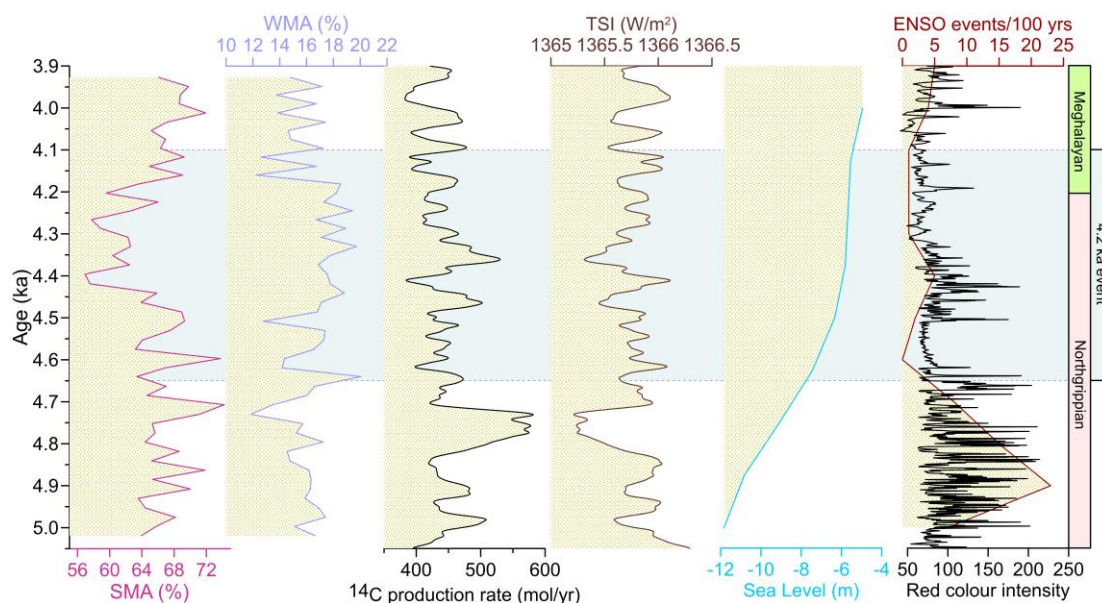


Figure 4.8: The changes in summer and winter monsoon (SMA and WMA), ¹⁴C production rate (mol/yr), total solar irradiance (TSI, w/m²), sea level (m), and ENSO activity during the 4.2 ka event.

4.4.3 Monsoon seasonality and productivity changes during the 4.2 ka event

The southwestern Bay of Bengal is strongly influenced by summer and winter monsoon seasonality (Schott and McCreary, 2001). The planktic foraminifera species flourish in different seasons (Salman and Saraswat, 2024; Maeda et al., 2021; Stoll et al., 2007; Gupta et al., 1997; Schulz et al., 2002). Therefore, the summer and winter monsoon assemblages were used to understand the monsoon seasonality. The relative abundance of SMA was high before the 4.65 (Fig. 4.5). It gradually decreased after the 4.65 ka and

was lowest at 4.4 ka confirming the initiation of the 4.2 ka event at 4.65 ka. After 4.4 ka, it started increasing until 4.1 ka. Overall, it was low from 4.65 to 4.1 ka which also confirmed the weakening of the summer monsoon during this period. On the other hand, the WMA abundance was high from 4.45 to 4.2 ka (**Fig. 4.5**) suggesting that the winter monsoon was intense during this interval. However, the initiation of winter monsoon intensification occurred at 4.45 ka BP instead of 4.65 ka BP.

Globigerina bulloides/*G. ruber* ratio did not change at 4.65 ka and abruptly increased at 4.45 ka (**Fig. 4.5**) suggesting high productivity/eutrophic conditions or less stratified upper water column conditions from the 4.45 to 4.1 ka BP. However, the decrease in CaCO_3 and C_{org} suggests low productivity and a high $\text{C}_{\text{org}}/\text{N}$ ratio also suggests lesser marine productivity (**Fig. 4.5**). Before the 4.65 ka, the $\text{C}_{\text{org}}/\text{N}$ ratio was low. Therefore, the high ratio of *G. bulloides*/*G. ruber* may not be the result of high productivity. *Globigerina bulloides* does not show a major shift, and was less affected by this cooling event. Therefore, the increase in *G. bulloides*/*G. ruber* ratio was the result of cooling. As the sea surface temperature decreases, the abundance of *G. ruber* also decreases (Salman and Saraswat, 2024) resulting in the high *G. bulloides*/*G. ruber* ratio.

4.4.4 Regional variability of 4.2 ka event

The 4.2 ka event has been reported from various parts of the Indian subcontinent and northern Indian Ocean by using different proxy records (**Fig. 4.1**). A majority of the records suggest a shift in monsoon intensity during the 4.2 ka event (Saraswat et al., 2023; 2016; Shah et al., 2022; Laskar and Bohra, 2021; Giosan et al., 2018; Bhushan et al., 2018; Sandeep et al., 2017; Kathayat et al., 2017; Berkelhammer et al., 2012; Dixit et al., 2014b, 2018; Prasad et al., 2014; Staubwasser et al., 2003; Kudrass et al., 2001). Saraswat et al. (2023) also reported a decrease in C_{org} and CaCO_3 (%) at ~4.6 ka in the Gulf of Mannar indicating a weak summer monsoon until 4.1 ka (**Fig. 4.6**). A similar study carried out from the offshore region of Pennar River also reported a decline in C_{org} and CaCO_3 (**Fig. 4.6**), confirming the onset (at 4.65 ka), duration (550 years), and the decrease in carbon burial as a result of the 4.2 ka event in wBoB and GoM. However, the decrease in C_{org} (%) at 4.5 ka (**Fig. 4.6**) and low $\text{C}_{\text{org}}/\text{N}$ ratio between 4.5-4.2 ka from the Lonar Lake suggests weakening of summer monsoon at 4.5 ka instead of 4.2 ka in central India (Prasad et al., 2014). The decrease in CaCO_3 (%) at 4.65 ka in the Dahar Lake (**Fig. 4.6**), northwestern India, suggests a gradual change in precipitation while the enriched $\delta^{18}\text{O}$ of gastropods implies an abrupt weakening of the Indian summer monsoon at 4.1 ka (Dixit

et al., 2014b). Hence, the concurrent response in both the marine and terrestrial regions confirms the widespread impact of the event across the Indian subcontinent.

The change in the relative abundance of *G. bulloides* was not very prominent in our record (Fig. 4.7). *Globigerina bulloides* exhibited a similar less discernible response to the summer monsoon in swBoB and GoM also during the 4.2 ka event (Fig. 4.7). The subdued response of *G. bulloides* in this region is because of its overall low abundance. The weaker upwelling in this region results in low abundance of *G. bulloides*. However, a decrease in *G. bulloides* offshore Pennar River suggests a weaker summer monsoon between 4.6 and 4.1 ka (Fig. 4.7). We report a high *G. falconensis* abundance thus suggesting a stronger winter monsoon between 4.65 to 4.1 ka (Fig. 4.4). The high abundance of *G. falconensis* in the northeastern Arabian Sea (neAS) also suggests a stronger winter monsoon between 4.5–3.95 ka (Fig. 4.7). Schulz et al., (2002) linked the higher abundance of *G. falconensis* to strong vertical mixing due to a strong winter monsoon during the 4.2 ka event. However, there was a gradual decrease in winter monsoon also towards the 4.1 ka (Fig. 4.4). Therefore, both summer and winter monsoon were weaker towards the end of event at ~4.1 ka. This transition coincides well with the termination of the mature phase of the Indus civilization (Khan et al., 2024; Berkelhammer et al., 2012; Dixit et al., 2014b; Staubwasser et al., 2003).

Interestingly, an abrupt decrease in SST was observed at 4.65 ka in the seAS (Fig. 4.7). SST exhibited a minor increase between 4.5–4.1 ka, followed by another decrease after 4.1 ka (Saraswat et al., 2013). Similarly, alkenone-based SST reconstruction (Burdanowitz et al., 2020) from the northern Arabian Sea (nAS) also suggests a gradual decrease in temperature from 4.65 to 4.0 ka (Fig. 4.7), confirming the widespread cooling throughout the Indian subcontinent at the boundary between the Northgrippian and Meghalayan Ages. A centennially resolved record (Kudrass et al., 2001) from the northern Bay of Bengal (nBoB) suggested low salinity between 4.6 and 4.0 ka (Fig. 4.7). The low salinity implies increased precipitation or riverine influx due to intense monsoon. However, the weaker summer monsoon during the 4.2 ka event, has been reported from various parts of the northern Indian Ocean (Laskar and Bohra, 2021; Staubwasser et al., 2003). The northern Bay of Bengal also receives huge summer monsoon flux. Therefore, the salinity should be high if the summer monsoon was weaker. The low salinity is thus attributed to the advection of low salinity water from the swBoB to the nwBoB, as the winter monsoon was intense during the 4.2 ka event. The changes in salinity confirm the onset and duration of the 4.2 ka event as observed in this study. From the west-central

Bay of Bengal, Ponton et al. (2012) reported enriched $\delta^{18}\text{O}_{G. ruber}$ at 4.25 to 4.0 ka (**Fig. 4.7**) while from swBoB, Suokhrie and Saraswat (2024) also reported enriched $\delta^{18}\text{O}_{G. ruber}$ at 4.45 to 4.1 ka suggesting a weaker summer monsoon (**Fig. 4.7**). The higher $\delta^{18}\text{O}$ of a gastropod species at 4.2 to 4.0 ka from the Kotla Dahar Lake also suggests a weaker summer monsoon after 4.2 ka onward (**Fig. 4.7**). Furthermore, a speleothem record from the Mawmluh Cave northeast India suggested the initiation of the event at 4.3 ka, with extreme drought between 4.05 to 3.9 ka (**Fig. 4.7**). The discrepancy in timing is attributed to the difference in resolution and chronological diachrony.

4.4.5 Possible causes of 4.2 ka event

The solar forcing has been linked with the variation in Indian monsoon intensity on a millennial to multi-centennial scale. The small insolation changes caused large variation in precipitation throughout the Holocene (Thamban et al., 2007). The strength of the solar activity is closely linked with the production rate of radionuclides. During the high solar irradiance, the increased solar winds and stronger solar magnetic field provide better shielding to the Earth from the cosmic rays, leading to a lower production rate of ^{14}C . On the contrary, the ^{14}C production rate is high when the solar wind and magnetic field are weaker during low solar irradiance (Roth and Joos, 2013). The global ^{14}C production rate decreased at 4.7 ka due to the increased solar irradiance (Roth and Joos, 2013; **Fig. 4.8**). However, the SMA of planktic foraminifera suggests the initiation of a weaker summer monsoon at 4.65 ka (**Fig. 4.8**). A similar dry condition was also reported from the Thar desert matching with the reduced ^{14}C production rate at 4.7 ka (Enzel et al., 1999). Staubwasser et al. (2003) also reported that the weaker monsoon in the neAS during the 4.2 ka event coincided with a reduction in ^{14}C production rate and enhanced solar radiation. Interestingly, the seasonal insolation changes (0-30°N) in the Northern Hemisphere (NH) were different. On a millennial scale the summer insolation gradually decreased and winter insolation increased during the 4.2 ka event (Burdanowicz et al., 2020). SMA and WMA of planktic foraminifera suggest a weaker summer monsoon and stronger winter monsoon between 4.65 and 4.1 ka, respectively (**Fig. 4.8**). Therefore, the seasonal insolation changes played a crucial role in both the summer and winter monsoon changes on a millennial scale. However, on a decadal scale, the changes in solar insolation are not comparable.

The rate of sea-level rise was higher before the event but decreased at 4.5 ka and almost stabilized between 4.5 and 4.1 ka (Grant et al., 2014; **Fig. 4.8**). However, the sea

level was 3 m higher than the present sea level between 4.35 and 4.10 ka. Subsequently, it was 2 m below the present sea level after 4.10 ka along the east coast of India (Loveson and Nigam, 2019). Das et al. (2022) also reported a low sea level stand between 4.15 and 3.625 ka along the southwest margin of Gujarat.

The internal and short-term climatic phenomena like ENSO, IOD, and ITCZ shifting could have also caused the abrupt climatic shift during the 4.2 ka event (Booth et al., 2005; Staubwasser et al., 2003). The seasonal migration of ITCZ controls the duration, termination, and amount of precipitation during the summer and winter monsoon seasons in the tropical and subtropical regions (Geen et al., 2020). Generally, the ITCZ migrates towards the warm hemisphere on seasonal and longer timescale (Schneider et al., 2014). The summer ITCZ shifted towards the south since the last 4.5 ka in response to the decreasing summer insolation in the Northern Hemisphere (Fleitmann et al., 2007). The summer ITCZ shift coincides with the onset of the weaker summer monsoon phase (Fleitmann et al., 2007; Fig. 4.8). Therefore, the ITCZ shift due to the decreasing insolation in the Northern Hemisphere affected the summer monsoon precipitation over the Indian subcontinent. However, there was no abrupt shift in ITCZ at 4.65 ka. The insolation decreased progressively from 6.3 ka onward (Fleitmann et al., 2007). Therefore, the shift in ITCZ may not be the sole factor responsible for the weak summer monsoon during the 4.2 ka event.

In the eastern Pacific Ocean, both the intensity and frequency of ENSO events per 100 years decreased abruptly at 4.65 ka and continued until 4.04 ka (Moy et al., 2002; Fig. 4.8). El Nino, La Nina and associated Southern Oscillation in the equatorial Pacific Ocean strongly affect the Indian monsoon (Ratna et al., 2024). During the El Nino (warm phase), Indian subcontinent experiences drought conditions while during the La Nina (cool phase), it receives extensive rainfall (Ratna et al., 2024). The ENSO activity increases with the warming (Latif and Keenlyside, 2009). The 4.2 ka event was a cold phase in the low to mid-latitude regions (Railsback et al., 2018). Therefore, the low ENSO activity due to the colder interval in the low latitude regions resulted in a weaker summer monsoon during the 4.2 ka event. A study from the South China Sea, a region influenced by both the summer and winter monsoons, also suggested that the warm phase of ENSO was weaker while the cold phase was stronger during the 4.2 ka event (Dang et al., 2020). Therefore, the cool (La Nina) phase of ENSO during the winter season resulted in a stronger winter monsoon.

The positive IOD phase starts with the cooling in the southeastern tropical Indian Ocean (STIO) and warming of the western tropical Indian Ocean (WTIO) in May – June while negative IOD phase starts with warm STIO and cool WTIO in December – January (Brown et al., 2009; Saji et al., 1999). A strong positive IOD phase causes heavy rainfall over east Africa and Indian subcontinent (Saji et al., 1999). However, at the same time, the strong El Nino phase may cause drought conditions over the Indian subcontinent (Brown et al., 2009). During the early-middle Holocene, the weak IOD variability was associated with strengthened monsoon when the ENSO activity was suppressed while strong IOD variability was associated with weaker monsoon during the late Holocene when ENSO events were more frequent (Brown et al., 2009; Abram et al., 2007). The records of IOD variability during the 4.2 ka event are unavailable. Therefore, from the eastern and western TIO, high-resolution paleo records are required to understand the seasonal behavior of IOD events during the Holocene (Brown et al., 2009).

4.5 Conclusions

We provide the first ever insights into the complex interaction between climatic factors and seasonal monsoon dynamics on decadal resolution during the 4.2 ka event. The 4.2 ka event began at 4.65 ka and persisted until 4.1 ka, with a duration of 550 years in the Bay of Bengal. The event exhibited diachroneity in its onset across different records but resulted in a significant decrease in carbon burial. The weakening of summer monsoon and the strengthening of winter monsoon were key drivers reducing the carbon burial during the 4.2 ka event. The synchronous decrease in coarse fraction, CaCO_3 and nitrogen indicated a collapse in productivity. However, the winter monsoon assemblage of planktic foraminifera species and high $\text{C}_{\text{org}}/\text{N}$ ratio suggested a stronger winter monsoon leading to degradation of marine organic matter and deposition of terrestrial-derived organic carbon. The seasonal insolation changes, ENSO activity and ITCZ shifting were the major factors driving the 4.2 ka event. The findings from this study and comparison with previous research on 4.2 ka event, confirm its occurrence and climatic impacts across various regions of the Indian subcontinent and the northern Indian Ocean.

Chapter 5

Response of Monsoon Seasonality, Productivity, and Upper Water Column Stratification to the 8.2 ka Event

5.1 Introduction

The ongoing Holocene interglacial epoch started post the deglaciation, and has experienced several abrupt climatic events (Bond et al., 2001). Among them, the 8.2 ka event, which has been reported in different proxy records between 8.9 and 8.0 ka, is the most pronounced short-term cooling event in the Northern Hemisphere, within the Holocene (Dixit et al., 2024; Banerji et al., 2022; He et al., 2022; Mayewski et al., 2004). It is characterized by significant north Atlantic cooling, likely driven by freshwater influx from the proglacial Lake Agassiz-Ojibway and Hudson Bay Ice Saddle (Rush et al., 2023; Matero et al., 2017). The ice-melt freshened the north Atlantic Ocean, where the deep water forms, causing a decrease in the density of seawater, reducing the deep water formation, and resulting in the slowdown of the Atlantic Meridional Overturning Circulation (AMOC) (Kleiven et al., 2008). The Gulf Stream, a major heat transporter from the tropics to high latitude regions of the north Atlantic Ocean, slowed down and thus triggered the Northern Hemisphere cooling during the 8.2 ka event (Dixit et al., 2024; Matero et al., 2017; Wang et al., 2005). Both, the proxy data and model simulations suggest a cooling of 3°C mainly concentrated in the north Atlantic region (McKay et al., 2024; Wierma and Renssen, 2006). The hydrological shifts associated with the 8.2 ka event were widespread, leading to reduced precipitation (12.4%) in the Northern Hemisphere and increased precipitation (4.2%) in the Southern Hemisphere (He et al., 2022). Wiersma et al. (2011) simulated a cool response in the Northern Hemisphere and a warm response in the Southern Hemisphere. However, the link between north Atlantic cooling during the 8.2 ka event and the regional climatic phenomena like Intertropical Convergence Zone (ITCZ) migration, El Nino-Southern Oscillation (ENSO), and Indian Ocean Dipole (IOD) affecting the Asian monsoon system have not been well documented from the northern Indian Ocean. With accelerated global warming, ice sheet melting will increase and that may trigger similar abrupt cooling events (Aguiar et al., 2021; Park et al., 2019). A key question is how the global monsoon will respond if greenhouse gases continue to rise alongside enhanced meltwater fluxes. Therefore, these events are of

particular interest due to concerns that similar rapid shifts can occur in future.

A significant reduction in precipitation is recorded in marine and lake sediments, and speleothems from the Asian monsoon regions during the 8.2 ka event (McKay et al., 2024; Dixit et al., 2024, 2014; Tan et al., 2020; Cheng et al., 2009; Wang et al., 2005). The limited marine records from the Arabian Sea exhibit the well-preserved signatures of the 8.2 ka event (Staubwasser et al., 2002; Altabet et al., 2002), whereas records from the Bay of Bengal suggest a potential temporal lag in its manifestation (Ponton et al., 2012; Rashid et al., 2011; Kudrass et al., 2001). However, the evidence of 8.2 ka event in the northern equatorial Indian Ocean remains limited, likely due to low sampling resolution (Banerji et al., 2022). Additionally, a majority of the marine records covering the entire Holocene have focused on the summer monsoon variations (Dixit et al., 2024; Banerji et al., 2022). Therefore, the southeast margin of India, a winter monsoon dominated region, needs to be studied to understand the behaviour of both the summer and winter monsoons to climatic anomalies like 8.2 ka event. Understanding the response of upper water column stratification to such short-term cooling climatic anomalies is also important as it controls the marine productivity, air-sea gaseous exchange, dissolved oxygen, and heat flux (Rai et al., 2024; Bhadra et al., 2023; Tarique et al., 2023; Shanthi et al., 2016).

The dynamic structure of the upper water column in the southwestern Bay of Bengal is shaped by a complex interplay of seasonal wind intensity, intense rainfall, river runoff, formation of cyclonic and anticyclonic eddies, and advection of Arabian Sea high salinity water during the summer season (Jayalakshmi et al., 2015; Narvekar and Prasanna Kumar, 2014; Patnaik et al., 2014; Muraleedharan et al., 2007). The mixed-layer and thermocline dwelling planktic foraminifera species, having different depth habitats, record different stable isotopic signatures to reveal the temperature and salinity gradient of the upper water column (Rai et al., 2024; Bhadra et al., 2023; Saraswat et al., 2023). Therefore, the oxygen isotopic difference between the mixed-layer and thermocline dwelling planktic foraminifera species can be used as a proxy to trace the upper water column stratification during the geologic past (Rai et al., 2024; Bhadra et al., 2023).

As such, there is no record of the monsoon seasonality and its associated upper water column stratification in response to the 8.2 ka vent from the northern Indian Ocean. Dixit et al. (2024) pointed towards the need for more high-resolution proxy records with precise age control and to refine the lead-lag relationship in climate events. The decadal scale marine sediment record presented here, provides the first well-structured, strong evidence of the response of regional climate change in the northern Indian Ocean to the north Atlantic climatic forcing that occurred around 8.2 ka event. This study aims to

understand the impact of 8.2 ka event on monsoon seasonality, productivity, and upper water column stratification by using bulk sediment geochemistry, planktic foraminifera species abundance, and their stable isotope signatures from the south-western Bay of Bengal, a region highly influenced by both the summer and winter monsoons.

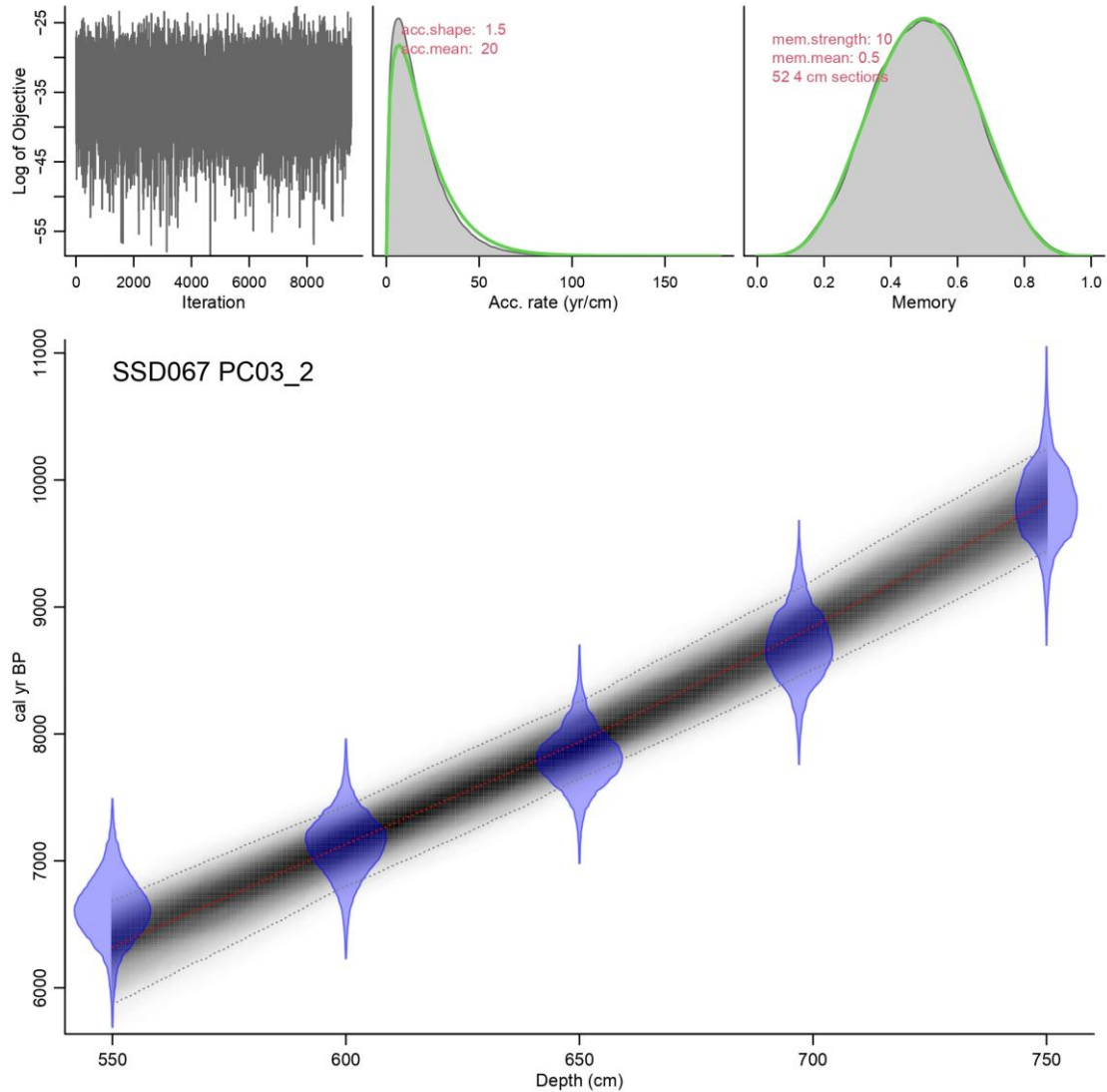


Figure 5.1: The age-depth profile and sedimentation rate in the studied section of core SSD067 PC03. The black and grey gradient represents the modelled uncertainty at different depths. Blue violin plots indicate the posterior probability distributions for age at specific depths. The red dashed line is the median age-depth relationship, while dotted lines represent confidence intervals.

5.2 Objective

Based on the above mentioned observations, the given research objective was finalized.

- To reconstruct the response of monsoon seasonality, productivity, and upper water column stratification to the short-term climatic anomalies around the 8.2 ka event.

5.3 Results

The average sedimentation rate in the studied section (550-750 cm) was 64.69 cm/kyr having an average resolution of 15.46 yr/cm (**Fig. 5.1; Table 5.1**). Bacon age model was run to establish the chronology and to avoid the sudden change in sedimentation rate (**Fig. 5.1; Table 5.1**). From the Bacon age model based chronology, the average resolution was 20 yr/cm (**Fig. 5.1**).

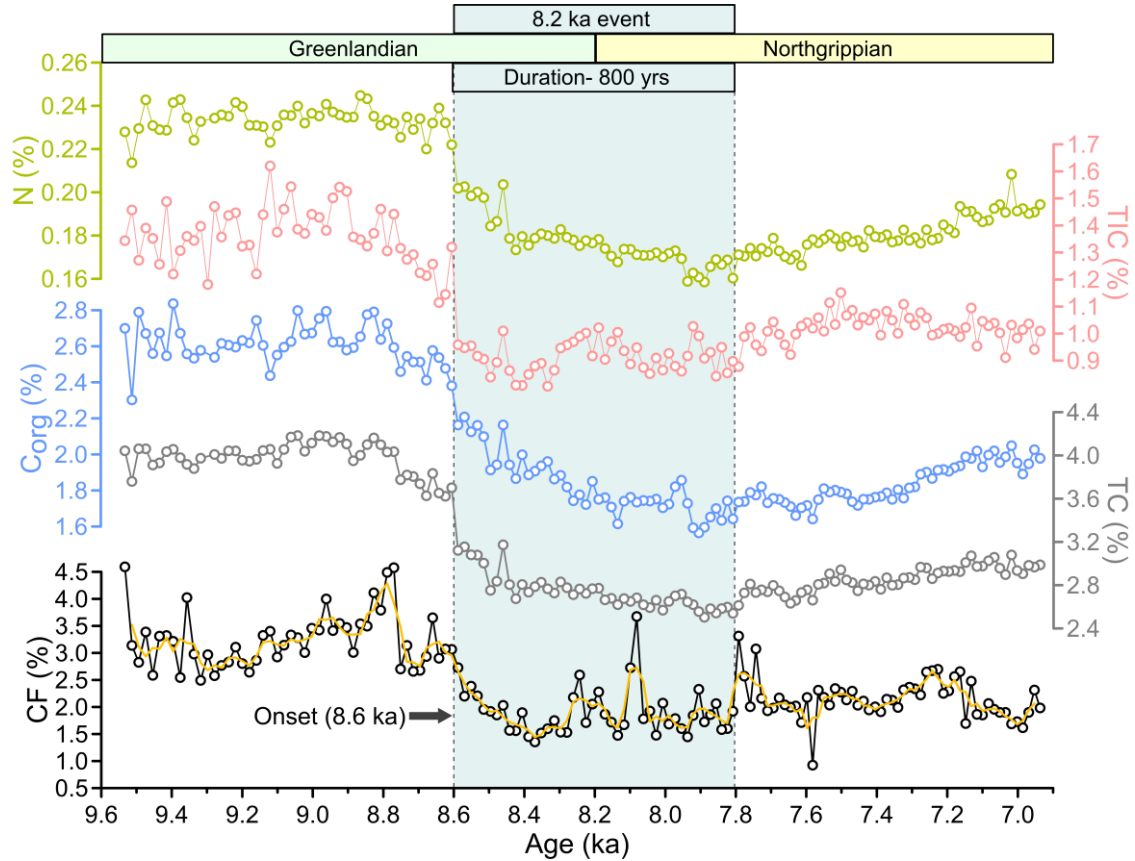


Figure 5.2: Variation in coarse fraction (CF), total carbon (TC), organic carbon (C_{org}), total inorganic carbon (TIC), and nitrogen (N) percentage during the 8.2 ka event in the southwestern Bay of Bengal. The 8.2 ka event is the boundary between Greenlandian and Northgrippian and is marked by the light blue-green bar.

5.3.1 Changes in CF, TC, TIC, C_{org} , N and C_{org}/N during the 8.2 ka event

A significant decrease in coarse fraction ($>63 \mu m$), which includes shells and may also contain sandy grains excluding silt and clay, occurred at 8.6 ka (**Fig. 5.2**). A similar decline in TC, C_{org} , TIC and N was observed at 8.6 ka, continuing until 7.8 ka, coinciding with the boundary between Greenlandian and Northgrippian ages, also known as 8.2 ka event (**Fig. 5.2**). Before the event, TC was 4.2%, and it dropped to 2.8% during the 8.2 ka event. All the variables - CF, TC, C_{org} , TIC, and N were at their maximum level during the Greenlandian Age but decreased to their lowest value by the end of the event. The recovery began after the event, though levels remained lower than that prior to the event

(Fig. 5.2). Notably, both CaCO_3 and $\text{C}_{\text{org}}/\text{N}$ ratio also decreased during the 8.2 ka event (Fig. 5.4). These changes suggest a reduction in marine productivity and terrestrial input during the event, supporting the idea that the onset of the event occurred at 8.6 ka rather than 8.2 ka and lasted for 800 years.

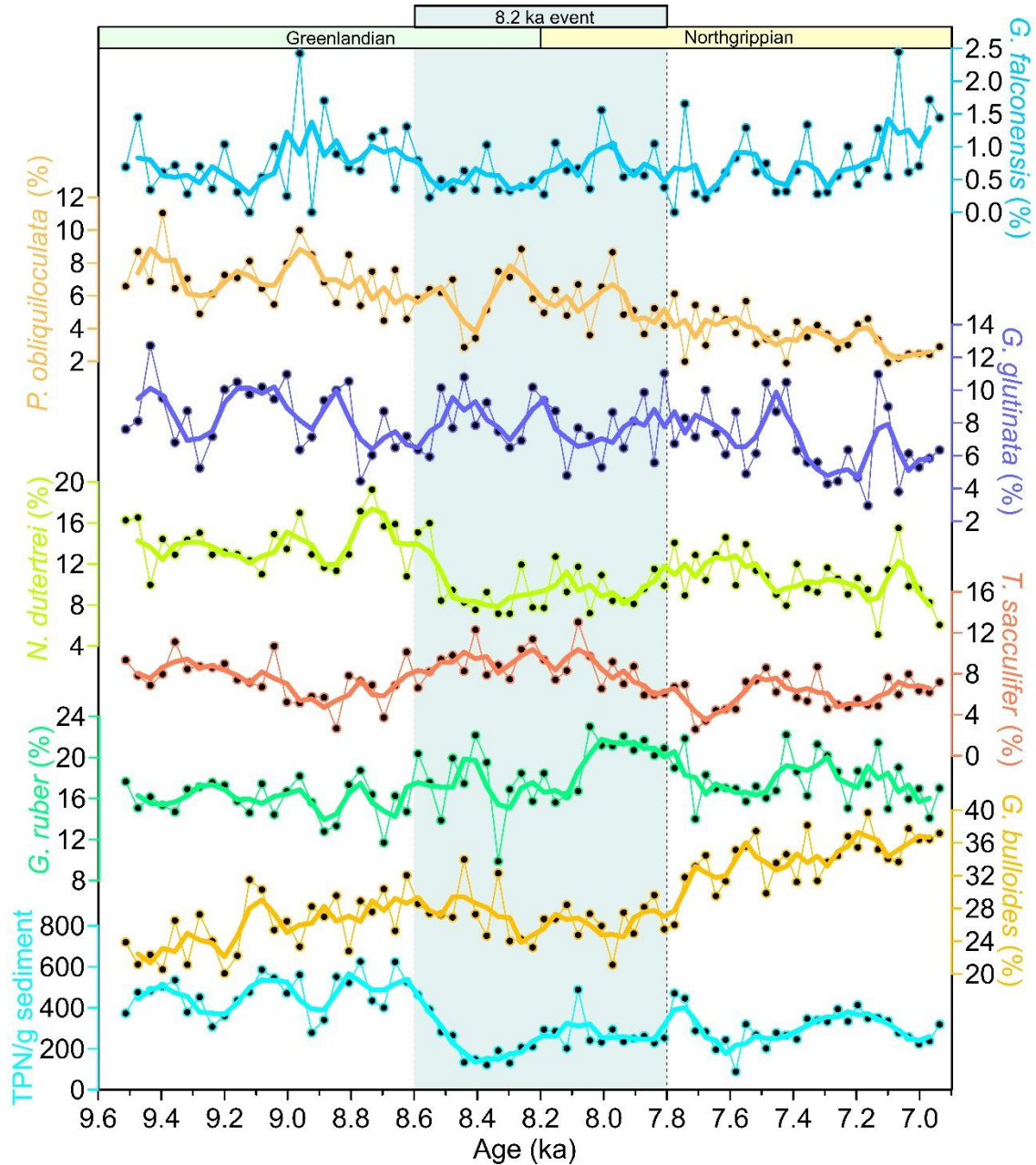


Figure 5.3: The variation in the relative abundance of planktic foraminifera species during the 8.2 ka event in the southwestern Bay of Bengal. The 8.2 ka event is marked by the light blue-green bar.

5.3.2 Planktic foraminifera species and their assemblages

A sudden decrease was observed in total planktic foraminiferal number (TPN/g sediment) at 8.6 ka and continued until 7.8 ka confirming the onset at 8.6 ka and duration of 800 years. The planktic foraminiferal abundance decreased from 500 to just 100 during the

event (**Fig. 5.3**). The lowest *G. bulloides* (%) abundance was noticed before the event but there was a gradual increase, prior to 8.2 ka event. However, the relative abundance of *G. bulloides* decreased during the event. The highest *G. bulloides* (%) abundance was after the event (**Fig. 5.3**). Interestingly, the relative abundance of mixed layer species *G. ruber* (%) and *T. sacculifer* (%) was high during the event, whereas it was low before and after the event suggesting that oligotrophic conditions prevailed during the event (**Fig. 5.3**). The relative abundance of *Neogloboquadrina dutertrei*, a thermocline species, decreased from 16 to 10% during the event indicating the decrease in subsurface chlorophyll maxima (**Fig. 5.3**). *Globigerinita glutinata* doesn't show any major shift during the event

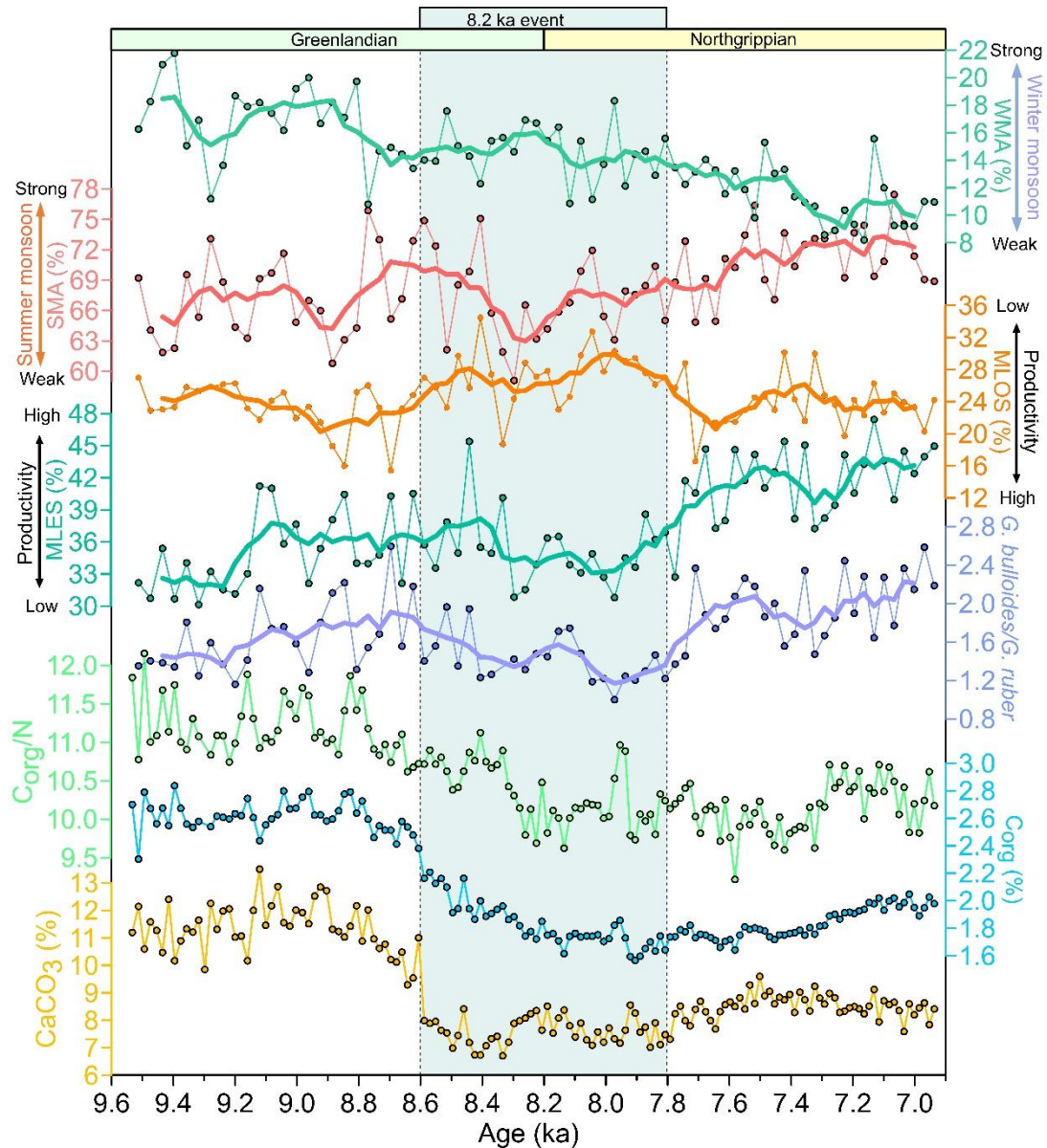


Figure 5.4: The variation in $CaCO_3$, C_{org} , C_{org}/N , *G. bulloides*/*G. ruber*, mixed layer eutrophic species (MLES), mixed layer oligotrophic species (MLOS), summer and winter monsoon assemblages (SMA and WMA) during the 8.2 ka event in the southwestern Bay of Bengal. The 8.2 ka event is marked by the light blue-green bar.

while *P. obliquiloculata* abundance decreased gradually since starting to the end of the studied section showing a trend opposite to that of *G. bulloides*. Similarly, *G. falconensis* (%) also doesn't show a major shift during the event suggesting that the stable winter monsoon conditions prevailed in the swBoB (**Fig. 5.3**). A sharp decline in *G. bulloides*/*G. ruber* ratio was observed suggesting low marine productivity and relatively warmer sea surface temperature during the event (**Fig. 5.4**). However, the decrease in MLES and increase in MLOS was observed during the event (**Fig. 5.4**). A prominent decrease was noticed in SMA indicating a weaker summer monsoon while the response of WMA was opposite to that of summer monsoon. There was a gradual decrease in WMA starting from 9.6 ka and continuing up to 6.9 ka (**Fig. 5.4**). The seasonal preferences of foraminifera are established based on their modern distribution in surface sediments, plankton tows and sediment traps. The sediment trap and core-top studies from the northern Indian Ocean revealed large seasonal variations in shell fluxes and spatial distribution of planktic foraminiferal species (Salman and Saraswat, 2024; Bhadra and Saraswat, 2021; Maeda et al., 2021; Stoll et al., 2007; Schulz et al., 2002; Conan and Brummer, 2000; Guptha et al., 1997). The southwestern Bay of Bengal has pronounced seasonality characterized by distinct summer and winter phases (Schott and McCreary, 2001). These seasonal phases influence the sea surface temperature, salinity, and nutrient availability, which govern the planktic foraminiferal abundance. Therefore, different planktic foraminifera species thrive in specific monsoon conditions, making them effective indicators of seasonal hydrographic changes. The relative abundance of *Globigerinoides ruber*, *Globigerina bulloides*, *Neogloboquadrina dutertrei*, *Trilobatus trilobus*, *Trilobatus sacculifer*, *Globigerinella siphonifera*, and *Neogloboquadrina incompta* peaks during the summer monsoon season in the southwestern and western Bay of Bengal. Therefore, they have been considered as a summer monsoon indicator. However, the relative abundance of *Globigerinita glutinata*, *Globigerina falconensis*, *Globorotalia menardii* and *Pulleniatina obliquiloculata* dominates during the winter monsoon (Salman and Saraswat, 2024; Bhadra and Saraswat, 2021; Maeda et al., 2021; Stoll et al., 2007; Guptha et al., 1997). The studies based on sediment traps and surface sediments have indicated that *G. ruber* and *T. sacculifer* prefer warmer, low-nutrient oligotrophic conditions (Salman and Saraswat, 2024; Bhadra and Saraswat, 2021; Guptha et al., 1997), categorizing them as mixed-layer oligotrophic species (MLOS). In contrast, *G. bulloides*, *G. glutinata*, and *G. falconensis* flourish in cooler, nutrient-rich eutrophic environments, classifying them as mixed-layer eutrophic species (MLES). *Globigerinoides ruber* has a preference for warmer, oligotrophic, and stratified sea surface conditions, while *G. bulloides* prefers

cooler, eutrophic, and upwelling/vertical mixing (Salman and Sarasawat, 2024; Bhadra and Saraswat, 2021; Stoll et al., 2007). Therefore, the *G. bulloides*/*G. ruber* ratio can be used as an SST, productivity, upwelling, and stratification proxy.

5.3.3 Stable isotope signatures during the 8.2 ka event

The $\delta^{18}\text{O}$ of *G. ruber*, a mixed layer species, and *N. dutertrei*, a thermocline species, was used to reconstruct the monsoon and water column stratification changes during the 8.2 ka event. The most depleted $\delta^{18}\text{O}_{G. ruber}$ was during the event while it was more enriched before and after the event (Fig. 5.5). However, $\delta^{18}\text{O}_{N. dutertrei}$ was relatively enriched during the event. Therefore, the difference ($\Delta\delta^{18}\text{O} \text{‰}$) of $\delta^{18}\text{O}_{G. ruber}$ and $\delta^{18}\text{O}_{N. dutertrei}$ was large during the 8.2 ka event (Fig. 5.5). The decrease in sea surface salinity also suggests that the well-stratified water column conditions prevailed during the 8.2 ka event (Fig. 5.5).

5.4. Discussion

5.4.1 Onset, duration, and impact of 8.2 ka event

The monsoon and associated changes control the CF, TIC, C_{org} , N, and C_{org}/N of marine sediments. A prominent decrease in CF, majorly containing the biogenic shells, suggests a drop in marine productivity (Fig. 5.2). The low productivity reduces the biogenic shell flux to the sediment (Saraswat et al., 2023). A major shift in TC at 8.6 ka indicates the decrease in both the marine and land-derived organic and inorganic carbon (Fig. 5.2). A minor increase in C_{org}/N ratio suggests increased contribution of land derived organic matter or decrease in marine productivity due to weaker monsoons (Fig. 5.4). Interestingly, an abrupt decrease in C_{org} , TIC, and N at 8.6 ka confirms the onset of the event at 8.6 ka rather than 8.2 ka and the termination at 7.8 ka (Fig. 5.2). Therefore, the event lasted for 800 years. The onset of the event is not uniform in all the records, so much so that in some records, it began at the end of other records. The duration of the event also varies between different records. In some records, the event lasted for one or two centuries, while in others, the changes were observed for a longer duration.

Globally, the response to the event may differ from a few years to decades due to the movement of climatic signals and regional responses to the event (Dixit et al., 2024; McKay et al., 2024). So far, there is no report of the initiation of 8.2 ka event at 8.6 ka and duration of 800 years from the BoB. The precise beginning, termination, and duration of the event could not be ascertained at many places as most of the records have poor

millennial to centennial or even lower resolution (Banerji et al., 2022). A sub-centennial record from the Indo-Pacific warm pool region suggested that the event lasted for 200

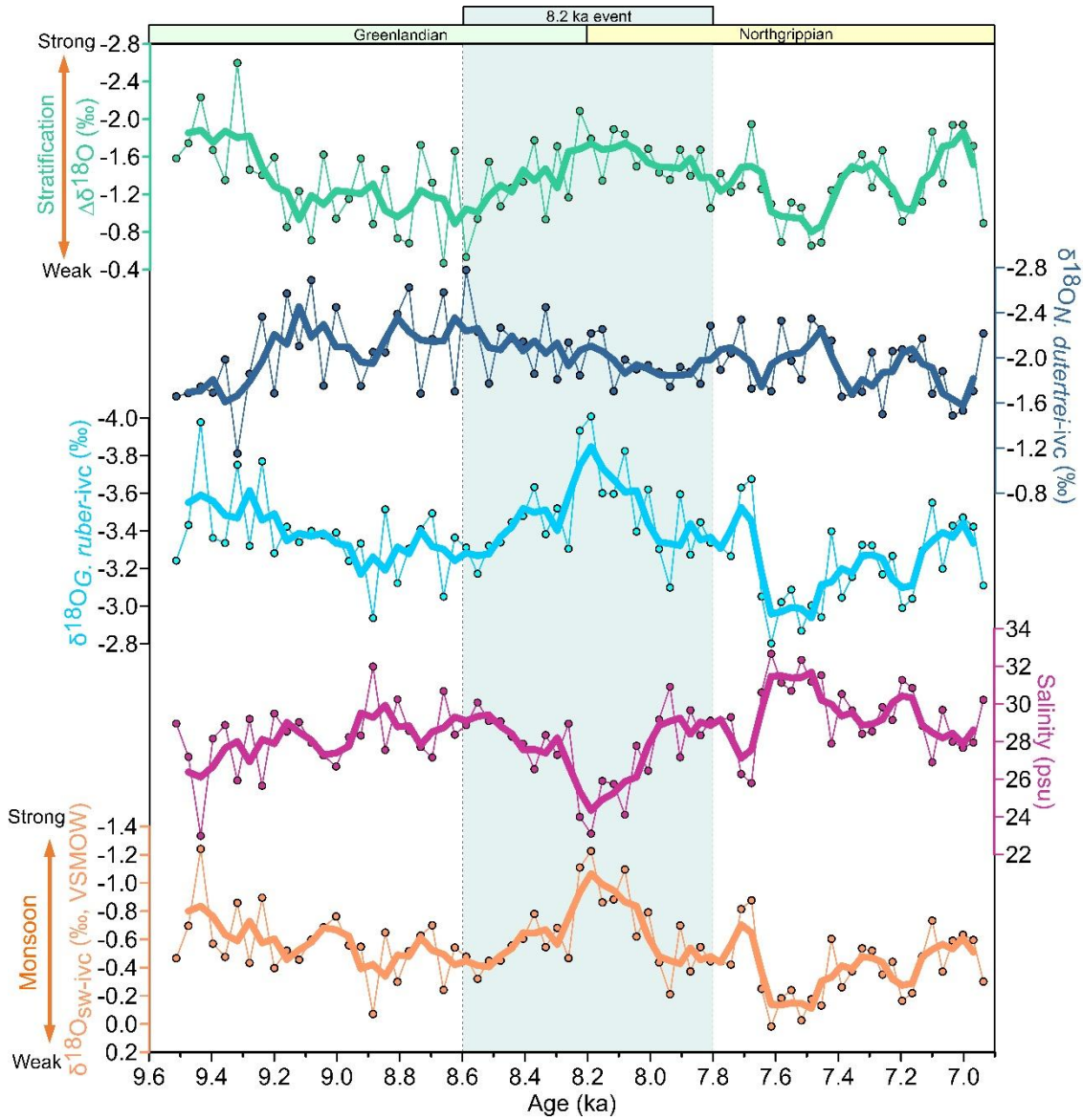


Figure 5.5: The changes in $\delta^{18}\text{O}$ of *G. ruber* and *N. dutertrei*, the difference between $\delta^{18}\text{O}$ of *G. ruber* and *N. dutertrei* ($\Delta\delta^{18}\text{O}$), and sea surface salinity (SSS) during the 8.2 ka event.

years from 8.15–7.96 ka and there was a lag of 100 years (Dixit et al., 2024). But we report that there was no lag in the Bay of Bengal, and the onset at 8.6 ka, matching well with the north Atlantic cooling suggests fast propagation of climatic signals through atmospheric connections rather than oceanic circulation. Due to the lack of high-resolution records, most of the studies suggested the transmission of climatic signals through oceanic circulation. However, our decadal to multi-decadal resolution record suggests propagation of the signal through the atmospheric teleconnections rather than oceanic processes in case of short-term climatic anomalies.

The extreme sensitivity of planktic foraminifera to climate change can reveal the

history of monsoon and its associated processes from the Bay of Bengal which is highly influenced by monsoonal activity. The high planktic foraminiferal abundance before 8.6 ka suggests increased productivity before the event. An abrupt decrease in the abundance at 8.6 ka continuing until 7.8 ka indicates low productivity from 8.6 to 7.8 ka confirming the impact of 8.2 ka event in the swBoB (**Fig. 5.3**). In tropical regions, *G. bulloides* is associated with upwelling-induced high productivity regions (Schulz et al., 2002; Naidu et al., 1999; Reynolds and Thunell, 1985; Cullen and Prell, 1984). Therefore, the relatively low abundance of *G. bulloides* suggests low productivity and weaker summer monsoon between 8.6 to 7.8 ka (**Fig. 5.3**). *Globigerinoides ruber* and *T. sacculifer* proliferate in warmer and oligotrophic conditions (Salman and Saraswat, 2024; Bhadra and Saraswat, 2021; Guptha et al., 1997). The abundance of *G. ruber* and *T. sacculifer* was high during the event (**Fig. 5.3**). Although, both species don't suggest a sea surface cooling during this interval, their high abundance confirms that oligotrophic conditions prevailed during the 8.2 ka event in the swBoB. The relative abundance of *Neoglobobulimina dutertrei*, a thermocline species that prefers eutrophic and warmer waters of thermocline (Salman and Saraswat, 2024; Bhadra and Saraswat, 2021; Gupta et al., 1997), abruptly decreased at 8.6 ka (**Fig. 5.3**), suggesting a decline in productivity and thermocline temperature. *Pulleniatina obliquiloculata*, a thermocline species, indicates winter season (Sijinkumar et al., 2011; Bé and Tolderlund, 1971; Ravelo and Fairbanks, 1992). *Pulleniatina obliquiloculata* abundance gradually decreased from 9.6 to 6.9 ka, showing a trend opposite to the *G. bulloides* (**Fig. 5.3**). In contrast, a relatively high abundance of *P. obliquiloculata* suggests a relatively stronger winter monsoon during the 8.2 ka event. However, it was highest before the event and lowest after the event. *Globigerina falconensis* is a eutrophic mixed-layer species that exhibits a preference for environments with high food availability (Salman and Saraswat, 2024), suggesting a subtle decrease in winter productivity during the 8.2 ka event (**Fig. 5.3**). No major shift in *G. glutinata* suggests a similar winter monsoon, as the high abundance of *G. glutinata* is often associated with winter season (Guptha et al., 1997).

5.4.2 Monsoon seasonality and productivity changes during the 8.2 ka event

The summer monsoon assemblage was relatively less abundant during the 8.2 ka event, suggesting a weaker summer monsoon. However, the WMA responded opposite to the SMA. The winter monsoon assemblage was highest before the event, moderate during the event, and lowest after the event, whereas SMA was high after the event, suggesting a moderately intensified winter monsoon during the 8.2 ka event. There was a gradual

strengthening of summer and weakening of winter monsoons until 8.6 ka (**Fig. 5.4**). After 8.6 ka, the winter monsoon further intensified while summer monsoon weakened until 7.8 ka. If we combine both SMA and WMA, there was a net reduction in Indian monsoon

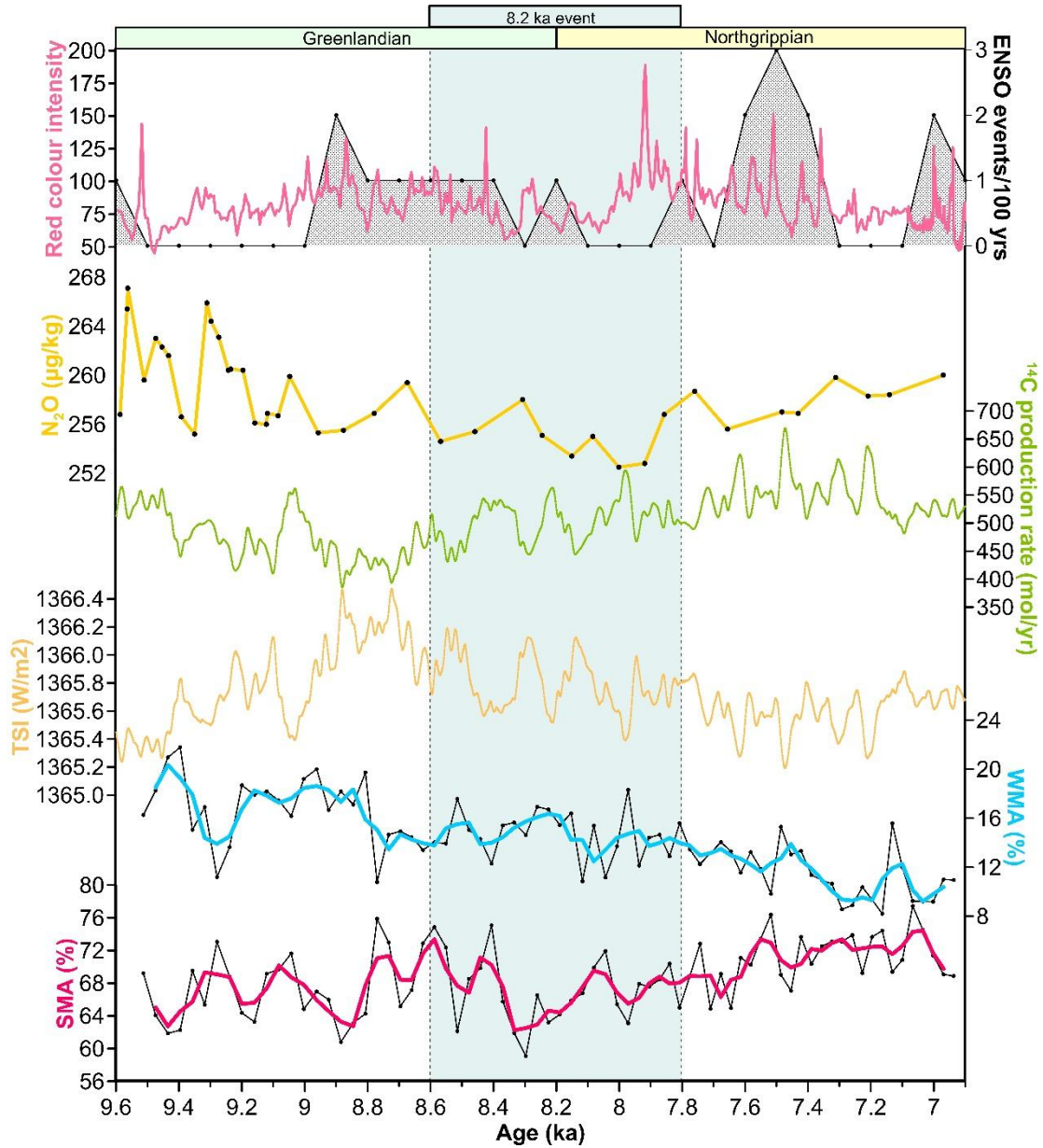


Figure 5.6: The changes in summer and winter monsoon (SMA and WMA), total solar irradiance (TSI, w/m^2), ^{14}C production rate (mol/yr), N_2O ($\mu\text{g/kg}$), and ENSO activity during the 8.2 ka event.

intensity during the 8.2 ka event as also evident from a significant decrease in planktic foraminiferal abundance (**Fig. 5.3; 5.4**).

The southwestern Bay of Bengal has pronounced seasonality characterized by distinct summer and winter phases (Schott and McCreary, 2001). These seasonal phases influence the sea surface temperature, salinity, and nutrient availability, which governs the planktic foraminiferal abundance. The surface distribution and ecological studies

demonstrated that different planktic foraminiferal species thrive in specific monsoon conditions, making them effective indicators of seasonal hydrographic changes (Salman and Saraswat, 2024; Maeda et al., 2021; Stoll et al., 2007; Gupta et al., 1997; Schulz et al., 2002). *Globigerinoides ruber* has a preference for warmer, oligotrophic, and stratified sea surface conditions, while *G. bulloides* prefers cooler, eutrophic, and upwelling/vertical mixing (Salman and Saraswat, 2024; Stoll et al., 2007). Therefore, the *G. bulloides*/*G. ruber* ratio can be used as an SST, productivity, upwelling, and stratification proxy. The low abundance indicates warmer, oligotrophic, and well-stratified conditions and vice versa. The reduction in *G. bulloides*/*G. ruber* suggests warmer, oligotrophic, and stratified conditions during the 8.2 ka event (Fig. 5.4). The decrease in MLES, CaCO₃%, and C_{org}%, and the increase in MLOS also suggests low productivity during the 8.2 ka event. Finally, SMA and WMA suggested a weaker summer monsoon and moderately intensified winter monsoon, and the prominent shift in CaCO₃, C_{org}, C_{org}/N, *G. bulloides*/*G. ruber*, MLES, MLOS confirms the initiation of the 8.2 ka event at 8.6 ka rather than 8.2 ka and its duration of 800 years.

5.4.3 Response of upper water column stratification to the 8.2 ka event

The difference in $\delta^{18}\text{O}$ of *G. ruber* (a mixed layer or surface dweller species) and *N. dutertrei* (a thermocline or subsurface dweller species) ($\Delta\delta^{18}\text{O}$), incorporating different isotopic signatures due to different depth habitats/calcification depths, was used to understand the upper water column stratification. The depleted or low $\Delta\delta^{18}\text{O}$ indicates a strong stratification while an enriched or high $\Delta\delta^{18}\text{O}$ suggests a weak stratification or vertical mixing (Rai et al., 2025; Bhadra et al., 2023; Mulitza et al., 1997). A strong stratification associated with high freshwater influx leads to a warmer mixed layer or sea surface and shallow thermocline where the high-temperature gradient exists in the thermocline zone (Steph et al., 2009). The evaporation-precipitation changes also create a salinity gradient in the upper water column affecting the $\Delta\delta^{18}\text{O}$. Therefore, the $\Delta\delta^{18}\text{O}$ has been used to trace the changes in the mixed layer and thermocline depths.

Previously, a strong stratification and shallow thermocline has been suggested in the northern Bay of Bengal and Andaman Sea due to stronger summer monsoon (Sijinkumar et al., 2021; 2016). However, in the swBoB, strong winter monsoon flux creates a large salinity gradient (Fig. 2.2; 2.3). As the freshwater flux from the north and through Cauvery River increases (decreases) due to the strengthening (weakening) of the summer monsoon, the influence of southeastern Arabian Sea high salinity water also

increases (decreases) maintaining the salinity in the swBoB (Salman and Sarasawat, 2024; Vinayachandran et al., 1999; Fig. 3.1). In addition to that, the fluctuation in winter monsoon flux creates a salinity variation at the core location (Fig. 2.2; 2.3). The summer monsoon was weaker and the winter monsoon was relatively stronger during the 8.2 ka event which might have created a large salinity gradient. The depleted $\delta^{18}\text{O}_{\text{sw-ivc}}$ and the decrease in mixed-layer salinity between 8.6 and 7.8 ka (Fig. 5.5) suggest a large salinity gradient due to a stronger winter monsoon, resulting in a strong stratification and shallow thermocline. However, it was enriched before and after the event. The high abundance of *G. ruber* and *T. sacculifer* suggests oligotrophic conditions while the decrease in *N. dutertrei* suggests a shallower thermocline during the 8.2 ka event (Fig. 5.3). The more depleted $\Delta\delta^{18}\text{O}$ during the 8.2 ka event unlike before and after the event (Fig. 5.5) suggests that the sea was well stratified and the thermocline was shallower during the 8.2 ka event. A well-stratified water column results in weaker cyclonic eddies during the winter season in the basin. These cyclonic eddies upwell cold and nutrient-rich water to the sea surface increasing the primary productivity (Jayalakshmi et al., 2015). The productivity decreased during the 8.2 ka event as suggested by MLES and MLOS assemblages. Therefore, the intensity of cyclonic eddies was weak leading to well-stratified water column during the 8.2 ka event.

The 8.2 ka event was a global cooling event in the Northern Hemisphere (McKay et al., 2024; Dixit et al., 2024). With the decrease in atmospheric temperature, the summer monsoon winds weakened due to less pressure gradient between land and ocean. The weaker winds result in weaker upwelling of the subsurface cold and nutrient-rich water to the surface, leading to warmer and stratified conditions during the 8.2 ka event. Due to weaker winds, the intensity of summer monsoon current also decreased, which carries high salinity water from the Arabian Sea to swBoB. Therefore, both the weaker summer monsoon winds and the decreased advection of high salinity water enhanced the upper water column stratification during the 8.2 ka event. The advection of Arabian Sea high salinity water is one of the major factors responsible for a deeper mixed layer in swBoB (Narvekar and Prasanna Kumar, 2014). Finally, the weaker summer monsoon, less influence of high salinity water, reduced eddy intensity during both the seasons, and relatively stronger winter monsoon led to the well stratified conditions in swBoB during the 8.2 ka event.

5.4.4 Teleconnections between North Atlantic cooling and northern Indian Ocean

The short-lived abrupt climatic events like 8.2 ka event have been linked with the north

Atlantic cooling. The climatic signals propagated via oceanic and atmospheric processes that affected the regional climate systems far away from the region of its origin (Mohtadi et al., 2014; Alley and Agustsdottir, 2005). However, Dixit et al. (2024) emphasize signal propagation via oceanic processes to reach its influence in the Indo-Pacific Warm Pool (IPWP) region. The onset of drying at ~8.15 ka suggested a lag of 100 years between north Atlantic cooling and weakening of the monsoon at a centennial scale. However, a lag of 100 years in climate signal transmission may be attributed to uncertainties in dating methods and age models (Dixit et al., 2024). Interestingly, we report the onset of the event at 8.6 ka based on our decadal to multidecadal scale marine record from the swBoB. The significant lead of the changes in the swBoB over those in the high northern latitude, requires additional teleconnections, if these regions are mutually interconnected climatically.

Before the 8.2 ka event, the North Atlantic experienced multiple episodes of enhanced meltwater discharge from the decaying Laurentide Ice Sheet between 8.9 and 8.2 ka (Rush et al., 2023; Lawrence et al., 2016). The high-resolution relative sea-level (RSL) records from southwest Scotland show that RSL rose abruptly in three steps at 8.760–8.640, 8.595–8.465, and 8.323–8.218 ka, correlating with short-lived events in the North Atlantic proxy climate and oceanographic records. These RSL steps indicate that the meltwater discharge increased in three distinct episodes, with rise in seastand of 0.35 m, 0.7 m, and 0.4 m, respectively (Lawrence et al., 2016). Therefore, the onset of the event at 8.6 ka matches with the first freshwater pulse release between 8.760 and 8.640 ka in the North Atlantic, suggesting a fast propagation of climatic signals reaching far away somewhere in the northern Indian Ocean, most likely transmitted via atmospheric teleconnections. A majority of the records focussed only on the third freshwater pulse between 8.323–8.218 ka, missing the first and second freshwater pulses (Dixit et al., 2024; Cheng et al., 2009).

A widely accepted hypothesis proposed by Broecker (1998), suggests that the freshening of the North Atlantic triggered a temporary cessation of AMOC, resulting in cooling in the northern regions while causing warming in the far southern areas, creating a see-saw effect. A slowdown in the North Atlantic sinking weakens the ocean current that transfers heat from the southern to the Northern Hemisphere. This keeps more heat in the south, leading to warming (Alley and Agustsdottir, 2005). Therefore, it is evident that the disruption in AMOC caused cooling in the North Atlantic region. Now, the question is whether this cooling signal propagated via deep-water circulation or through atmospheric teleconnections to reach far away somewhere in the Indian and Pacific

Ocean. If the cooling of the North Atlantic is transmitted through deep water circulation, we would expect a lag of more than 100 years. This is due to the cycle of sinking that begins with deep water formation in the North Atlantic. The water then upwells in various regions of the Indian and Pacific Oceans before returning to the North Atlantic to sink again. This process, known as the global oceanic conveyor belt, can take ~1000 years to complete a full cycle. Therefore, the lag is expected to be >500 years in the northern Indian Ocean. However, the quick response of North Atlantic cooling at 8.6 ka in the Bay of Bengal suggests its faster propagation via atmospheric teleconnection rather than the oceanic circulation. Most likely, the transmission of cooling effect via North Atlantic Deep Water circulation was mitigated or partially propagated due to warming in the Southern Hemisphere before reaching the northern Indian Ocean. The see-saw mechanism suggested for Heinrich stadials and the Younger Dryas works well as these events lasted for a longer time span than the 8.2 ka event. The propagation of signals between the tropical Indian and high latitude regions was via atmosphere, whereby slowdown of AMOC affected the tropical Indian Ocean by altering the westerly wind patterns (Hadley circulation), and a southward shift in ITCZ (Mohtadi et al., 2014). Based on simulations and proxy records, it is suggested that the dry conditions prevailed in the northern Indian Ocean realm, while the southern region experienced increased precipitation during North Atlantic cold events (Mohtadi et al., 2014). Therefore, we rely more on atmospheric teleconnections, in case of short-term climatic events like 8.2 ka event, rather than the oceanic circulation.

5.4.5 Drivers of 8.2 ka event

As discussed in the previous section, the shutdown of AMOC during the 8.2 ka event led to cooling of the North Atlantic and accumulation of heat in the southern regions. This phenomenon caused a southward shift of the Atlantic ITCZ, which led to dryer conditions (Frierson et al., 2013; Alley and Agustsdottir, 2005). In the Indian monsoon region also, the weakening of summer monsoon has been linked with the southward shift of summer ITCZ (Dixit et al., 2024, 2014; Mohtadi et al., 2014). However, a speleothem record from the Oman margin suggested that after 7.8 ka, the summer ITCZ continuously migrated southward and summer monsoon decreased gradually in response to seasonal insolation changes (Fleitmann et al., 2007). Interestingly, the total solar irradiance representing the changes in solar radiation by orbital positioning and sunspot activity peaked around 8.6 ka. It decreased abruptly afterwards suggesting that the short-term solar storms also modulated the summer and winter monsoon during the 8.2 ka event (**Fig. 5.6**). The

coronal holes on the sun's surface emit solar winds carrying magnetic fields that deflect galactic cosmic rays based on the intensity of magnetic fields. The strong solar winds result in a more significant deflection of galactic cosmic rays and a reduced production of cosmogenic radionuclides (Roth and Joos, 2013). The production of ^{14}C radionuclide increased after 8.6 ka, confirming a significant decrease in sunspot activity during the 8.2 ka event (Fig. 5.6). Both, SMA and WMA suggest a prominent shift in summer and winter monsoon between 8.6 and 7.8 ka (Fig. 5.6). The summer monsoon peaked at 8.6 ka when TSI was high and decreased afterward. The winter monsoon weakened at 8.6 ka when ^{14}C production rate was low (Fig. 5.6). Therefore, short-term sunspot activity played a crucial role in driving the changes in summer and winter monsoon during 8.2 ka event on a decadal scale resolution.

Ocean denitrification, a typical phenomenon associated with the oxygen minimum zone (OMZ), is an important source of atmospheric nitrous oxide (N_2O). The Holocene variability of atmospheric N_2O in the northern Indian Ocean has been linked with the Indian monsoon (Azharuddin et al., 2024; Rixen et al., 2020; Altabet et al., 2002). North Atlantic cooling and slowdown of AMOC shifted the ITCZ southward, which affected the monsoon and OMZ strength in northern Indian and Pacific oceans (Pichevin et al., 2007; Wang et al., 2005). Therefore, abrupt climatic events of Holocene might have affected the intensity of OMZ and production of N_2O in the Arabian Sea and Bay of Bengal. Azharuddin et al. (2024) have suggested a prolonged decrease in atmospheric N_2O between 8.8 and 6.2 ka correlated with the Asian monsoon and intensification of Southern Hemisphere monsoon. The lowest concentration of atmospheric N_2O between 8.6 and 7.8 ka suggests a significant decrease in OMZ intensity caused by lower productivity during the 8.2 ka event (Fig. 5.6). The decrease in fourth most effective greenhouse gas (N_2O) might have exaggerated the atmospheric cooling during the 8.2 ka event resulting in weaker monsoon as inferred from the changes in SMA and WMA (Fig. 5.6).

The Indian monsoon is significantly influenced by Indian Ocean Dipole (IOD) and ENSO activities, which are primarily driven by SST variability in the tropical Indian and Pacific Oceans (Ratna et al., 2024). However, the response of these processes to the 8.2 ka event remains inadequately documented in the northern Indian Ocean. The intensity and frequency of ENSO events was relatively low prior to 7.0 ka, but increased markedly thereafter (Moy et al., 2002). Notably, ENSO events diminished between 8.4 and 7.9 ka, with a corresponding decrease in intensity during this period (Moy et al., 2002; Fig. 5.6). During El Niño phases, the Indian subcontinent typically experiences drought conditions,

whereas La Niña phases bring substantial rainfall (Ratna et al., 2024). Furthermore, ENSO activity tends to increase with global warming (Latif and Keenlyside, 2009). The 8.2 ka event was characterized by cooling in the Northern Hemisphere (Dixit et al., 2024; He et al., 2022), which may have suppressed ENSO activity during this period. Consequently, the reduced ENSO activity, attributed to the colder conditions in northern regions, likely resulted in a weaker summer monsoon during the 8.2 ka event. The positive IOD phase is initiated by cooling in the south-eastern tropical Indian Ocean (STIO) and warming in the western tropical Indian Ocean (WTIO) during May and June, while the negative IOD phase begins with warm STIO and cool WTIO in December and January (Brown et al., 2009; Saji et al., 1999). A pronounced positive IOD phase can lead to heavy rainfall over East Africa and the Indian subcontinent (Saji et al., 1999). However, concurrently, a strong El Niño phase may induce drought conditions in the Indian subcontinent (Brown et al., 2009). During the early to middle Holocene, weak IOD variability was linked to a strengthened monsoon when ENSO activity was suppressed, whereas strong IOD variability correlated with a weakened monsoon during the late Holocene when ENSO events became more frequent (Brown et al., 2009; Abram et al., 2007). It is likely that the weak ENSO and IOD variability were associated with the 8.2 ka event. Unfortunately, records of IOD variability during this period are lacking. Therefore, high-resolution paleo records from both the eastern and western tropical Indian Ocean are essential to comprehensively understand the seasonal implications of IOD events throughout the Holocene (Brown et al., 2009).

5.5 Conclusions

Based on the decadal resolved marine sediment core, the onset, duration, and impact of the 8.2 ka event in the Bay of Bengal is inferred in detail. A large shift in the monsoon, productivity, and surface stratification is reported during the 8.2 ka event. The onset of the event occurred around 8.6 ka and lasted for 800 years, as confirmed by shift in sediment's geochemistry, planktic foraminifera species, and their isotopic signatures. The key indicators, including planktic foraminifera species and their assemblages suggest a weakening of summer monsoon and a moderate strengthening of the winter monsoon accompanied by oligotrophic conditions during the 8.2 ka event. Furthermore, changes in upper water column stratification, marked by the $\delta^{18}\text{O}$ difference between *G. ruber* and *N. dutertrei*, suggest enhanced stratification and a shallow mixed and thermocline layer during the 8.2 ka event, a likely consequence of reduced upwelling, large salinity gradient due to stronger winter monsoon, weaker cyclonic eddies, and reduced advection of high

salinity waters. The North Atlantic cooling likely influenced the Bay of Bengal through atmospheric teleconnections rather than oceanic circulation and played a prominent role in regional climate response during the 8.2 ka event, highlighting the importance of high-resolution marine records in understanding the decadal to multi-decadal climatic shifts. The solar forcing linked to changes in sunspot activity also modulated the changes in monsoon dynamics during the 8.2 ka event. This was likely exacerbated by weak ENSO and IOD variability during this period. The 8.2 ka event had a profound and prolonged impact on the Indian monsoon system, with significant implications for both regional hydroclimate patterns and global climate dynamics.

Chapter 6

Multiple, High-amplitude, Short-term Climatic Events during the Last Glacial Maximum

6.1 Introduction

A prolonged global cooling centered around 21 ka (19–24 ka) is defined as the Last Glacial Maximum (LGM). This period is marked by the most recent extreme cold phase in Earth's climate history, with the maximum ice sheet coverage in the Northern Hemisphere (Clark et al., 2009; Schneider von Deimling et al., 2006). The cooling during LGM was not uniform all around the globe, with the maximum cooling of 28°C over the Northern Hemisphere ice sheets and a moderate cooling in the tropical regions (Schneider von Deimling et al., 2006). A pronounced cooling (about $8\pm 2^\circ\text{C}$) was also observed in Antarctica (Vimeux et al., 2002; Jouzel et al., 2003). The intensity of cooling over the land was 1.6 times higher than over the oceans (Shi et al., 2023). A coupled climatic model based on proxy data suggested a global cooling of about 6°C, out of which 1.5°C of the temperature anomaly was attributed to the glacial atmospheric dust and vegetation cover (Schneider von Deimling et al., 2006). Based on Mg/Ca ratio of the foraminiferal shells, it has been estimated that the global decrease in the equatorial sea surface temperature during the LGM was $\sim 2.3^\circ\text{C}$ as compared to the Holocene (Saraswat, 2011). The sea surface temperature of the southeastern Arabian Sea was $2.7 \pm 0.5^\circ\text{C}$ cooler during the LGM and began to warm at 18.6 ka at the onset of deglaciation (Saraswat et al., 2013). Another record from the eastern Arabian Sea reveals that the LGM was 4°C cooler than the present (Govil and Naidu, 2010). However, the western Bay of Bengal was $\sim 3^\circ\text{C}$ cooler during the LGM as compared to the early Holocene (Rashid et al., 2011). The widespread but heterogeneous cooling affected multiple climate processes with varying intensity and extent. Therefore, it is imperative to understand the regional response to LGM.

Earth's orbital changes primarily influenced the past glacial-interglacial cycles (Clark et al., 2009). During the LGM, the Earth experienced a period of relatively low axial tilt, which led to cooler temperatures at high latitudes when averaged over the year. Furthermore, the timing of aphelion during the Northern Hemisphere's summer maintained the cooler summer temperatures, supporting the growth of large ice sheets during the LGM (Berger, 1977). Additionally, the LGM cooling was also influenced by

significantly lower levels of greenhouse gases compared to the preindustrial era, as well as the presence of Laurentide and Fennoscandia ice sheets, which are no longer present today. The expanded land-ice cover during this period contributed to a global sea level drop of approximately 130 meters and altered the distribution of land and sea (Peltier and Fairbanks, 2006; Yokoyama et al., 2001). The strength of the LGM Atlantic overturning circulation was weaker, and the depth was shallower than its modern extent. This boundary between North Atlantic Deep Water (NADW) and Antarctic Bottom Water (AABW) was relatively shallower (~2 km) as suggested by low-nutrient water mass of northern origin and high nutrient water below it (Lynch-Stieglitz et al., 2007). Such a widespread reorganization of the ice sheets, sea level and water masses must have strongly affected the climate and oceanographic processes.

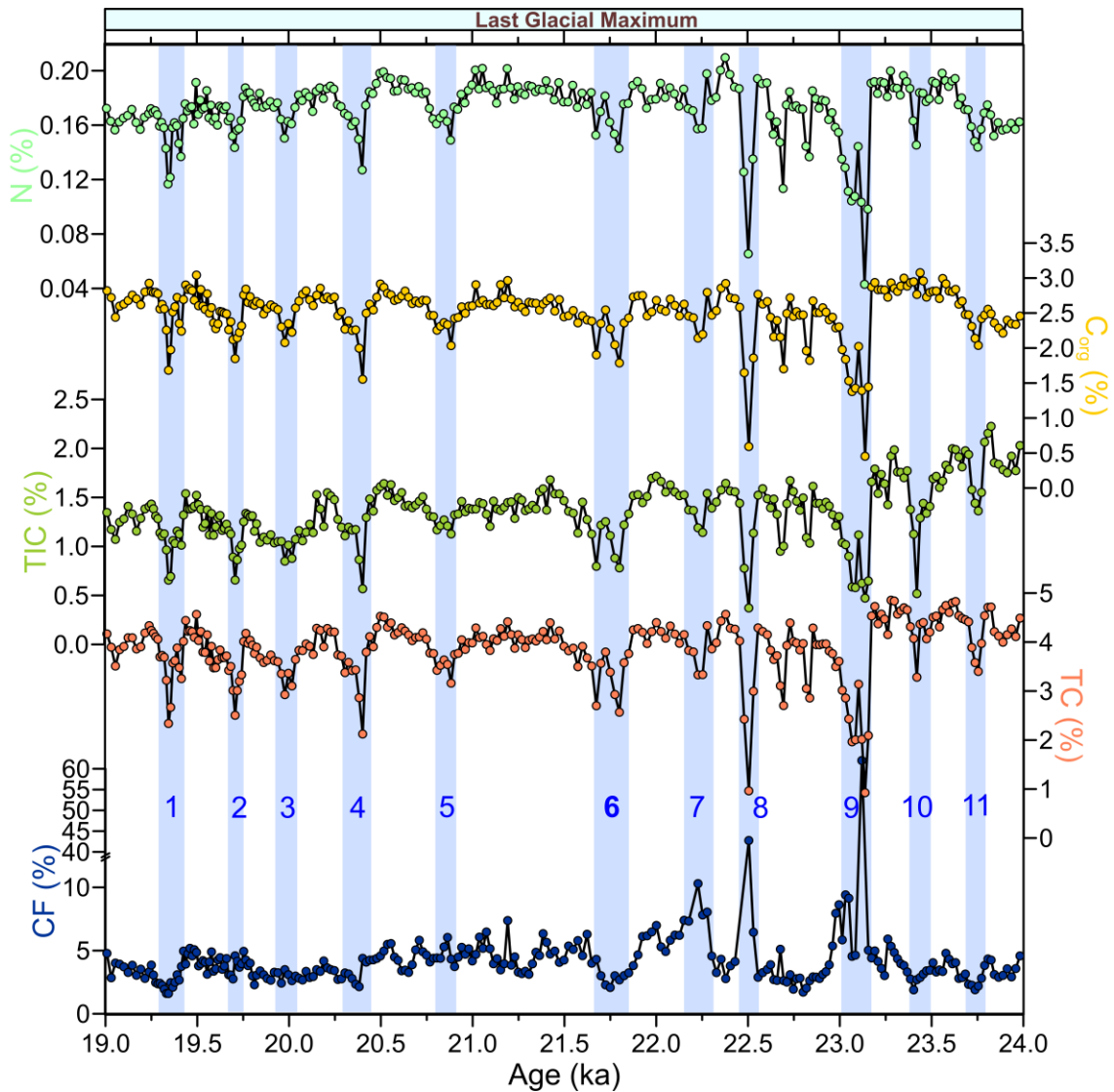


Figure 6.1: Variation in coarse fraction (CF), total carbon (TC), organic carbon (C_{org}), total inorganic carbon (TIC), and nitrogen (N) percentage between 19 and 24 ka in the southwestern Bay of Bengal. All the climatic events from 1 to 11 are marked by the light blue bar.

During the LGM, a significant reduction in annual mean global monsoon precipitation was observed, and the decrease in summer monsoon was more prominent than that in winter monsoon (Yan et al., 2016). The extended ice sheets and the decrease in greenhouse gases might have reduced the air temperature and water vapour, causing a significant weakening of the global monsoon. The less seasonal variability in insolation could have also contributed to the weakened global monsoon during the LGM (Yan et al., 2016). A lot of efforts have been made to reconstruct the Indian monsoon variability from the northern Indian Ocean. All of them have reported a weaker summer monsoon and relatively stronger winter monsoon during the LGM (Suokhrie and Saraswat, 2024; Fathima and Saraswat, 2023; Gautam et al., 2021; Verma et al., 2022; Naidu et al., 2019; Sijinkumar et al., 2016; Govil and Naidu, 2011; Rashid et al., 2011, Rashid et al., 2007). All these records suggest a cold and very stable climate during the LGM. However, LGM spanned from 19 to 24 ka. A majority of the records have a very coarse resolution with only a few data points representing the entire LGM. The high-resolution sub-centennial to multi-decadal records of the monsoon seasonality, carbon burial and other associated climate variability during the LGM are scarce from the northern Indian Ocean. Initially, the Holocene was also considered a very stable period, but high-resolution records revealed many short-term climatic anomalies within it. Similarly, the short-term climatic events can exist within the LGM. Here, a first-ever decadal to multi-decadal marine sediment record of the multiple abrupt climate excursions during the LGM is presented from the northeastern Indian Ocean.

6.2 Objective

Based on the above mentioned observations, the given research objective was finalized.

- To reconstruct the monsoon seasonality and carbon burial response to multi-decadal climatic events during the last glacial maximum.

6.3 Results

A total of eleven abrupt climatic disruptions (represented by numbers 1 to 11), were observed within 19–24 ka (**Fig. 6.1**). During the events 7, 8, and 9, the coarse fraction ($>63\ \mu\text{m}$) increased abruptly and reached the maximum up to 60%, highlighting the presence of sandy fraction. However, during the events 1, 2, 3, 4, 5, 6, 10, and 11, the CF decreased (**Fig. 6.1**). The total carbon (TC), total inorganic carbon (TIC), organic carbon (C_{org}), and nitrogen (N) content of the sediments decreased abruptly during all the events (**Fig. 6.1**). The calcium carbonate content ($\text{CaCO}_3\%$) also decreased suddenly during all the 11

climatic excursions (**Fig. 6.2**). Interestingly, the planktic foraminifera abundance significantly decreased during all the 11 climatic events and it varied between 36 and 380 specimen/g sediment (**Fig. 6.2**). The relative abundance of summer monsoon assemblage (SMA) was little high between 24 to 23.2 ka. However, it was low from 23.2 to 22.25 ka while the winter monsoon assemblage (WMA) was relatively high at the same time (**Fig. 6.2**). Similarly, a significant shift in the relative abundance of *G. ruber* and *G. bulloides* was observed during all the events but their response was not uniform (**Fig. 6.2**). A high abundance of *G. bulloides* and relatively low abundance of *G. ruber* were observed from 24 to 23 ka (**Fig. 6.2**). The C_{org}/N ratio decreased during all the events except 10. It abruptly increased during event 10 (**Fig. 6.3**).

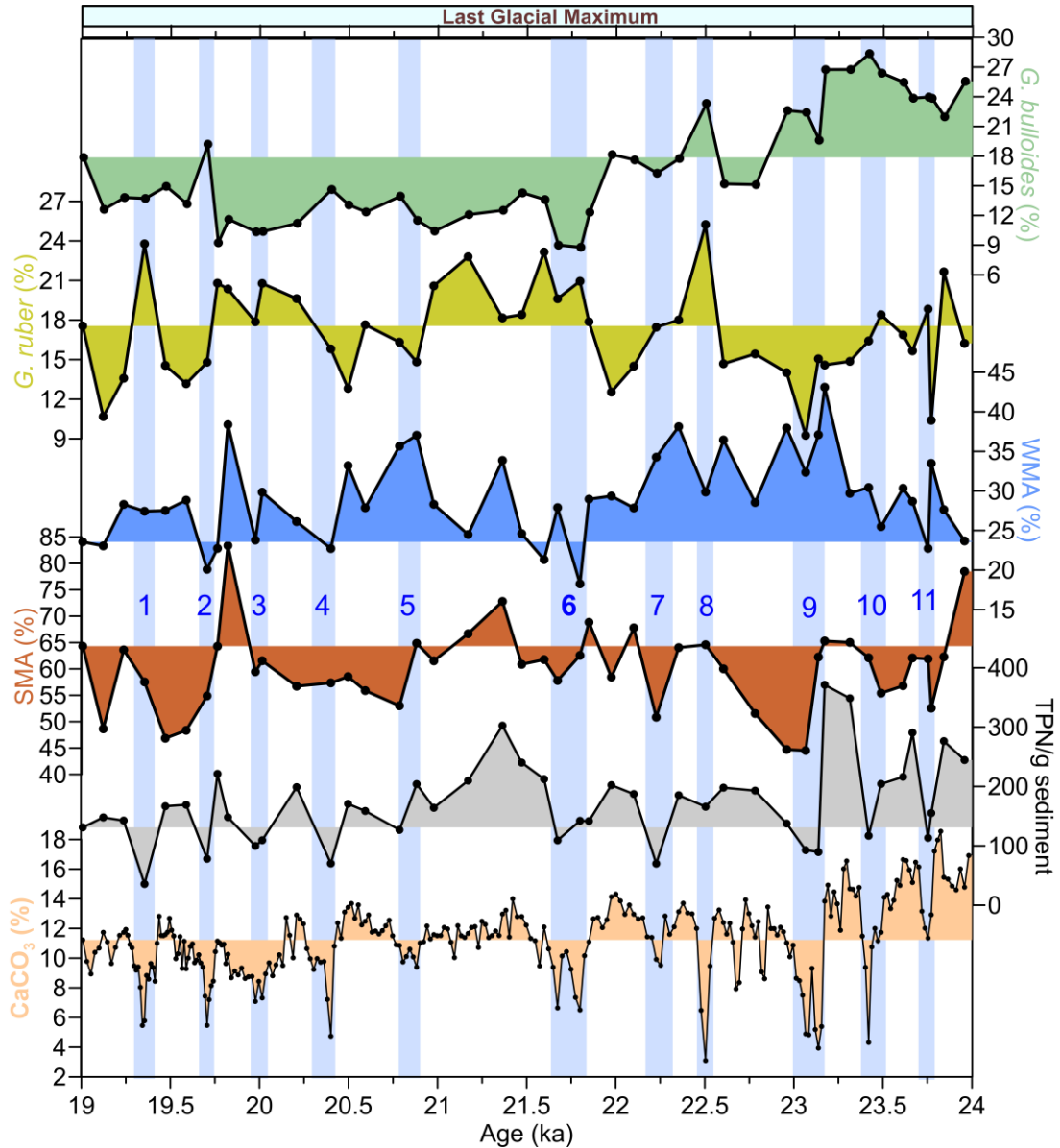


Figure 6.2: Variation in calcium carbonate ($CaCO_3$), absolute abundance of planktic foraminifera (TPN/g sediment), summer and winter monsoon assemblages (SMA and WMA) and relative abundance of *Globigerinoides ruber* and *Globigerina bulloides* between 19 and 24 ka. All the climatic events from 1 to 11 are marked by the light blue bar.

Both spectral and wavelet analyses of $CaCO_3$ and C_{org} data were performed to understand the cyclicity and frequency of these climatic events. The calcium carbonate

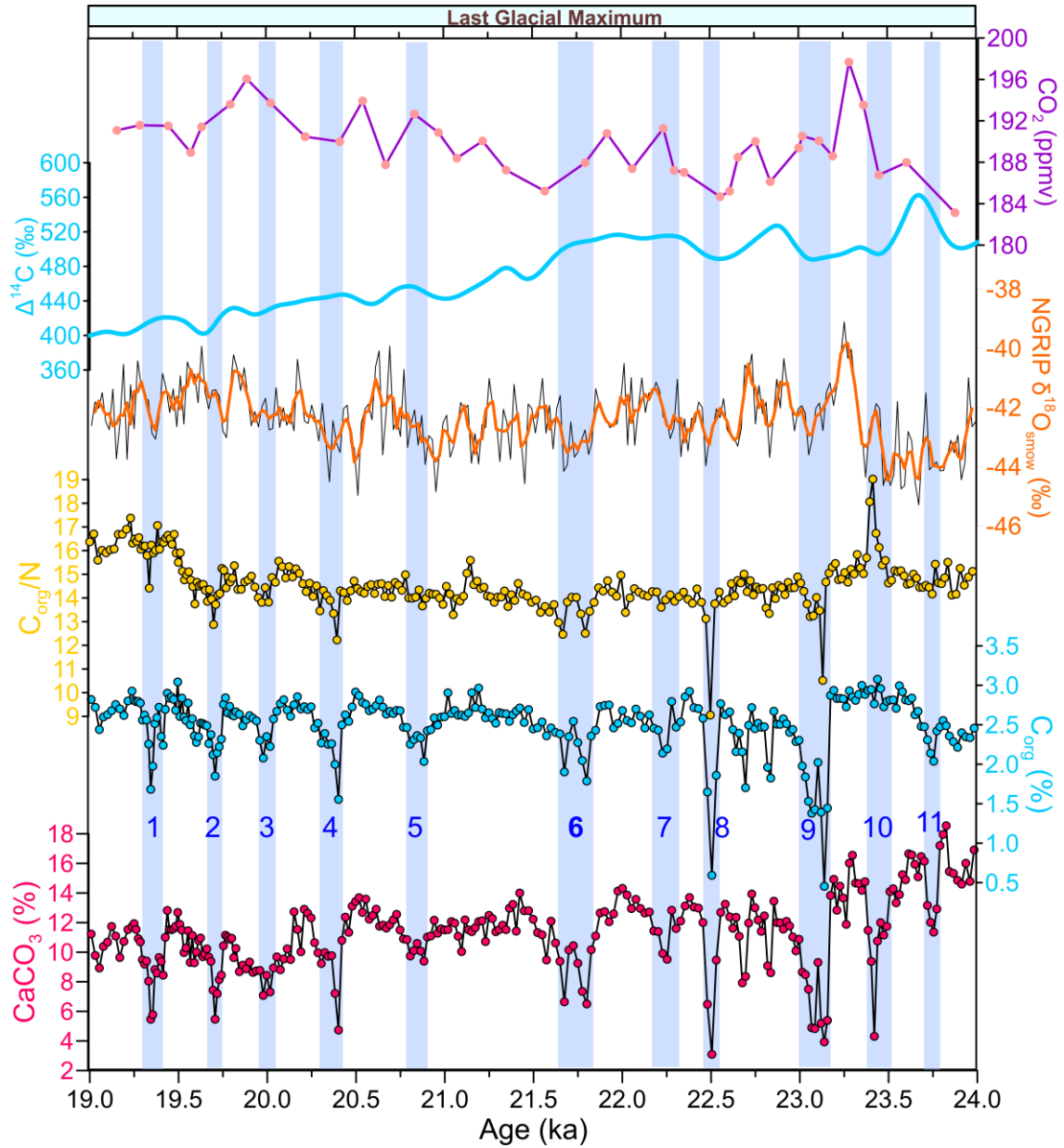


Figure 6.3: Variation in calcium carbonate (CaCO_3) and organic carbon (C_{org}) percentage, organic carbon to nitrogen ratio ($\text{C}_{\text{org}}/\text{N}$), Greenland atmospheric temperature ($^{\circ}\text{C}$) from the Northern Greenland Ice Core Project (NGRIP, Davtian and Bard, 2023), 20 year means of $\delta^{18}\text{O}$ from ice core NGRIP (Andersen et al., 2007), atmospheric $\Delta^{14}\text{C}$ (Heaton et al., 2020) and atmospheric CO_2 concentration (Bereiter et al., 2015) between 19 and 24 ka. All the climatic events from 1 to 11 are marked by the light blue bar.

data had significant periodicity of 312 yr, 340 yr, 1538 yr, and 2778 yr (**Fig. 6.4**). Similarly, the C_{org} also had significant periodicity of 312 yr, 342 yr, 705 yr, and 1087 yr. The periodicity of 312 yr and 340 yr were common to both CaCO_3 and C_{org} (**Fig. 6.4**). In wavelet plots the red and blue colour areas indicate the specific time when certain frequencies were particularly strong or weak (**Fig. 6.4**). CaCO_3 and C_{org} show the high frequency of climatic events all throughout the LGM with longer time cycles at low

frequencies or at the bottom of the plot and shorter time cycles at high frequencies (**Fig. 6.4**).

6.4 Discussion

6.4.1 Transient climate events in the northern Indian Ocean during the Last Glacial Maximum

The decadal to multi-decadal record of sedimentological (CF) and geochemical (CaCO_3 , C_{org} , and N) proxies helped to discover eleven short-lived climatic events, centered at 19.35 ka, 19.72 ka, 20.0 ka, 20.35 ka, 20.85 ka, 21.75 ka, 22.25 ka, 22.5 ka, 23.10 ka, 23.40 ka, and 23.75 ka (**Fig. 6.1**). The organic and inorganic carbon and nitrogen deposited in marine sediments includes both autochthonous and allochthonous particles. The origin and accumulation of these particles is controlled by the monsoon and ambient environmental conditions (Meyers, 1997). Therefore, the abundance of CaCO_3 , C_{org} and $\text{C}_{\text{org}}/\text{N}$ ratio is used to reconstruct the marine and terrestrial paleoenvironments (Saraswat et al., 2023; Meyers, 1997). This is the first such record of 11 distinct high-amplitude climate events during the LGM from the northern Indian Ocean. A millennial record from the North Atlantic also revealed a series of abrupt climatic events during the Holocene and last glacial period. However, only four episodes of abrupt climate shift at 19 ka, 21 ka, 22 ka, and 24 ka, were found during LGM (Bond et al., 1997; Bond et al., 2001).

The coarse fraction ($>63\mu\text{m}$) includes both biogenic shells and sand. Small fluctuation in CF may be because of the biogenic shell flux. The abrupt increase (decrease) in CF indicates the presence (absence) of terrestrial sand fraction. The rapid change in CF is attributed to a sudden release of a huge amount of riverine input during the events 7, 8, and 9. Such abrupt influx of terrestrial sand may be caused by floods or storms/cyclones (**Fig. 6.5**). During LGM, the sea level was also ~ 130 m below the present sea level (Peltier and Fairbanks, 2006). The lower sea level would have exposed the entire continental shelf, due to the gentle slope. The exposed shelf means the rivers meeting the sea on the upper slope, very close to the core location. Therefore, the chances of higher terrestrial sediment load being debouched close to the core location increased, whenever there was an intense monsoon or enhanced stormy condition. CF decreased during the rest of the events, which clearly indicates the absence of terrestrial sand influx. However, the terrestrial clay-silt content may have increased depending upon the energy conditions. Interestingly, a sharp decrease in total planktic foraminifera abundance also suggests the dilution due to sudden release of riverine input (**Fig. 6.2**). Earlier, the decreased abundance of planktic

foraminifera on the shelf along the eastern margin of India was attributed to dilution effect in the shallow marine environment due to riverine influx (Salman and Saraswat, 2024; Bhadra and Saraswat, 2021). The southeastern margin of India receives a huge amount of precipitation during the winter season and is very prone to cyclonic activity (Fig. 1.1, 2.2). It has been reported that the northeast winds and winter monsoon were stronger throughout the LGM (Gautam et al., 2021; Verma et al., 2022; Naidu et al., 2019; Govil and Naidu, 2011). Therefore, the episodes of enhanced cooling may have triggered the cyclonic activity, resulting in transient climatic anomalies during the LGM. A prominent decrease in SMA indicates a weaker summer monsoon during all the transient climatic events. However, WMA doesn't show a uniform behaviour for all the events which may be due coarser resolution because the planktic foraminifera species and their

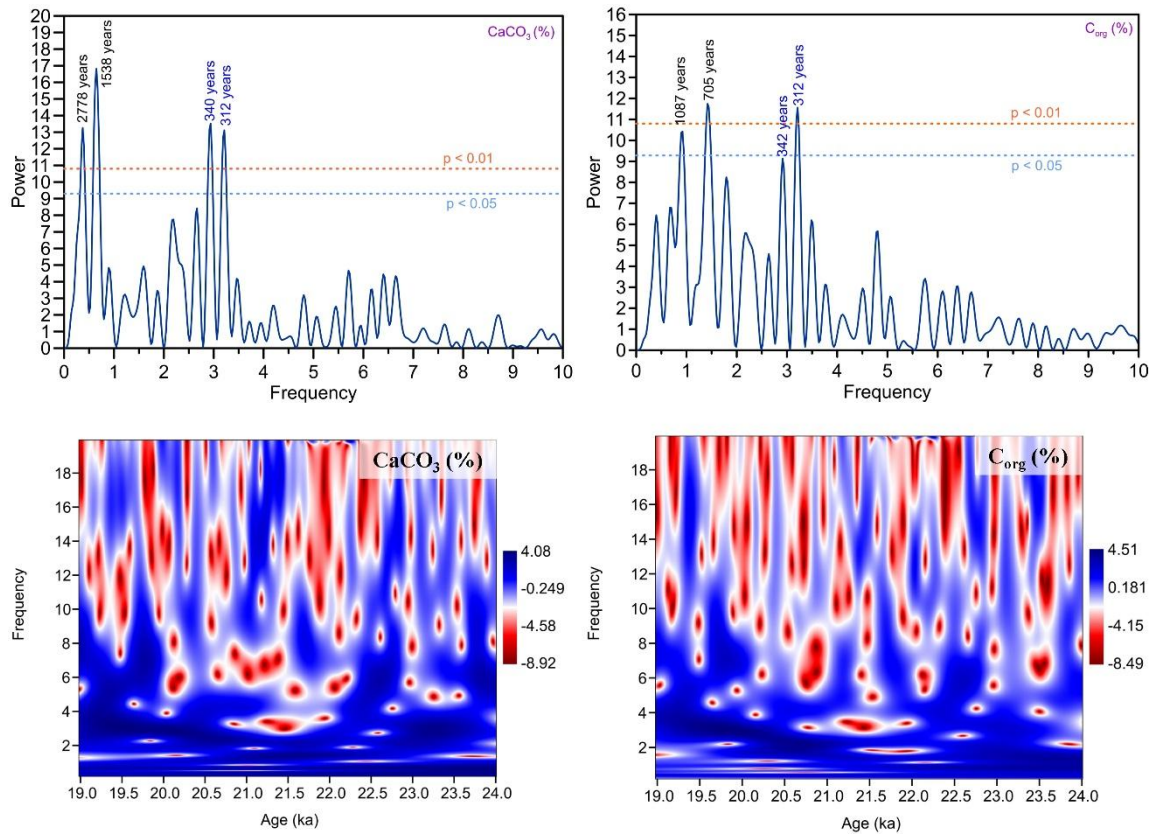


Figure 6.4: The spectral and wavelet analysis of CaCO_3 (%) and C_{org} (%) representing the cyclicity and frequencies of abrupt climatic events during 19–24 ka. In spectral plot, the significant cyclicities are mentioned with the significance level of >99 and 95% represented by orange and blue lines. For the wavelet plot, the red-colored spots indicate long-term cycles at low frequencies and short-term cycles at high frequencies.

assemblages were analysed at 10 cm intervals while geochemical analysis was done at 1 cm. The increase (decrease) in SMA (WMA) suggested a stronger summer monsoon and relatively weaker winter monsoon between 24 and 23.2 ka. In contrast, the winter monsoon was stronger from 23.2 to 22.5 ka (Fig. 6.2). A higher abundance of *G. bulloides* also

suggested a stronger summer monsoon from 24 to 23.2 ka and a weaker summer monsoon for the rest of the section (**Fig. 6.2**). However, *G. ruber* shows a very dynamic behaviour during the studied section. For a better response of planktic foraminifera species to these transient climatic events, the resolution has to be increased, as these events persisted for a very short duration.

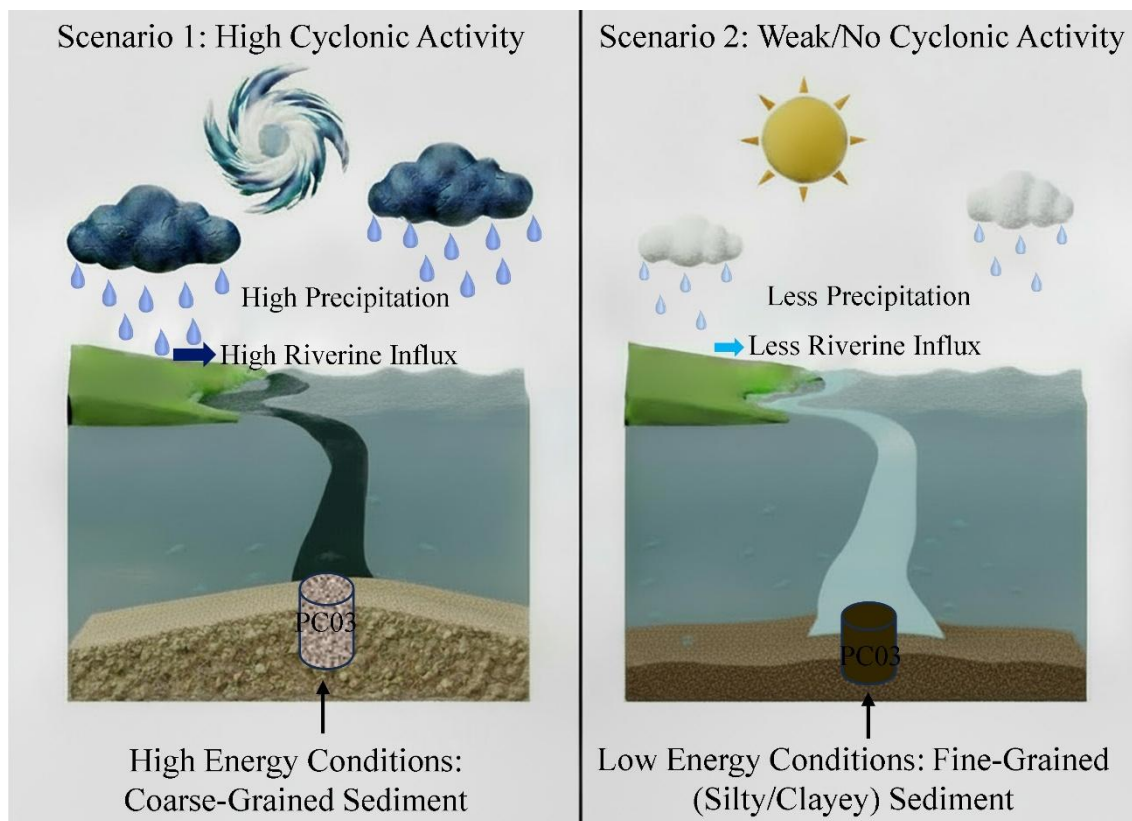


Figure 6.5. Schematic diagram showing the relation between cyclonic activity and sediments grain size in the case of both strong and weak cyclonic activity along the southeast coast of India.

The abrupt decrease in TC, TIC, C_{org} , and N suggests a similar driving mechanism for all eleven climatic events (**Fig. 6.1**). The sudden decrease in TIC, C_{org} , and N can be either due to decreased productivity or increased dilution by terrestrial material. The cyclonic activity along the southeast coast of India enhances the monsoonal precipitation. Additionally, the associated cold-core eddies bring the subsurface nutrient-rich water to the surface, increasing the marine productivity during the winter season (Jayalakshmi et al., 2015). Therefore, source of organic matter can help to assess the relative contribution of marine and terrestrial influx in driving the observed changes. The organic carbon to nitrogen (C_{org}/N) ratio can be used to differentiate the source of organic matter. The higher ratio indicates the higher terrestrial input and a lower ratio indicates higher marine-produced organic matter (Meyers, 1997). The synchronous abrupt decrease in C_{org}/N suggests a higher marine productivity during all the events except the 10th event.

Therefore, the decrease in TIC, C_{org} , and N is attributed to the dilution due to higher terrestrial input indicated by the presence of sandy fraction.

The abrupt changes in terrestrial influx are mainly driven by monsoon precipitation. The Asian monsoon strongly responds to high latitudinal processes. In order to understand the effect of high latitudinal processes in driving the short-term fluctuations in tropical monsoon during the LGM, the sediment record from the northern Indian Ocean were compared with the high latitudinal climate signatures. The Greenland Ice Core $\delta^{18}O$ records show an abrupt decrease in temperature during all the events (**Fig. 6.3**). Therefore, it is evident that the high latitudinal cold events intensified the monsoon and thus the terrestrial influx during the short-term abrupt events. A comparison of the records with the $\Delta^{14}C$ shows a gradual decrease in solar irradiance or sunspot numbers throughout the LGM. But, there was no gradual shift in $CaCO_3$, C_{org} , and C_{org}/N between 19 to 24 ka (**Fig. 6.3**). However, the episodic increase in $\Delta^{14}C$ (‰) matches with the short-lived climate events observed in the northern Indian Ocean, suggesting that an abrupt increase in solar irradiance may have triggered these transient events. The changes in CO_2 concentration don't seem to be the driving force for these abrupt climatic anomalies (**Fig. 6.3**).

The El Niño Southern Oscillation (ENSO), having warm (El Niño) and cool (La Niña) phases, also affects the Indian monsoon (Ratna et al., 2024). The seawater temperature reconstruction based on foraminiferal record from the equatorial Pacific indicates a reduced ENSO variability during LGM compared to the middle to late Holocene (Ford et al., 2015). In contrast, the $\delta^{18}O$ of individual foraminifera from the deep-sea sediments suggests stronger ENSO activity during the LGM (Koutavas and Joanides, 2012). The previous models also suggested that the ENSO was more intense during the LGM than the present (Otto-Bliesner et al., 2003; An et al., 2004; Peltier and Solheim, 2004). ENSO intensity remained elevated even if the influence of greenhouse gas concentration and orbital configurations was removed. This suggests that greenhouse gases and orbital parameters played a limited role in controlling ENSO variability during the LGM (Shi et al., 2023). Interestingly, Koutavas and Joanides (2012) reported a stronger ENSO activity and attributed it to the enhanced inter-annual SST variability. Therefore, ENSO may have further contributed to the cyclonic conditions in the northern Indian Ocean, leading to intensified winter precipitation, and thus the high-frequency flooding events during the LGM.

6.4.2 Cyclic changes during the Last Glacial Maximum

The CaCO_3 content in the core shows the cyclicity of 312 yr, 340 yr, 1538 yr, and 2778 yr (**Fig. 6.4**). Similarly, the C_{org} also shows the cyclicity of 312 yr, 342 yr, 705 yr, and 1087 yr, out of which 312 yr and 340 yr cycles are common for both CaCO_3 and C_{org} (**Fig. 6.4**). Basically, the long-term cycles (2778 yr and 1538 yr) in CaCO_3 content are the multiple of 1087 yr and 705 yr cycles observed in C_{org} . The long-term cycle matched well with the Hallstatt's solar cycle (2100–2300 yr), suggesting its strong influence in driving the long-term climate anomalies. The abrupt climatic events during the Holocene and last glaciation in the north Atlantic had a cyclicity close to 1470 ± 500 years (Bond et al., 1997), which is very close to the 1538 yr cycle observed in CaCO_3 record in the northern Indian Ocean (**Fig. 6.4**).

Interestingly, the short-term cyclicity of 312 yr and 340 yr are more prominent during the LGM as evident by number of events observed in the spectral and wavelet analysis of CaCO_3 and C_{org} , (**Fig. 6.3 & 6.4**). The wavelet plots of CaCO_3 and C_{org} show the high frequency of climatic events all throughout the LGM with longer cycles at low frequencies and the shorter time cycles at high frequencies (**Fig. 6.4**). A Grand Solar Cycle typically lasts around 330–380 years and is characterized by a modulation of the standard 11-year solar cycle (Zharkova et al., 2023). The 312–340 year cycles match well with the Grand Solar Cycles. This cycle has both grand minima when the solar activity is low and grand maxima when the solar activity is high. Therefore, the short-term periodic changes within the LGM were closely associated with lower solar activity. Hence, the changes in solar activity due to sunspot numbers played a crucial role in driving the high-frequency climatic events during the LGM.

6.5 Conclusions

During the last glacial maximum, the climate remained stable over an extended period, although it was occasionally interrupted by brief climatic events. This study provides a high-resolution reconstruction of LGM transient climatic anomalies from the Bay of Bengal. Eleven short-lived climatic events were identified between 19 and 24 ka. The duration of these climatic events varied from a few decades to a few centuries. The abrupt increase in CF suggests intensified riverine influx likely driven by enhanced winter monsoon and cyclonic activity along the southeastern margin of India. This is further supported by a simultaneous decrease in total planktic foraminifera, TIC, C_{org} , and N, reflecting rapid terrestrial input and a possible dilution effect on marine productivity. The

summer and winter monsoon assemblages suggested a weaker summer monsoon and a stronger winter monsoon during a few of these transient climatic events. From the relative abundance of *Globigerina bulloides*, a stronger summer monsoon is inferred from 24 to 23.2 ka and a weaker summer monsoon during the rest of the section. The spectral and wavelet analysis revealed distinct cyclicity in CaCO_3 and C_{org} content, with notable periodicity of 312 yr, 340 yr, 1538 yr, and 2778 yr. The alignment of long-term cycles with Hallstatt solar and Bond cycles, and ~312–340 yr cycles with Grand Solar Cycles highlight the potential influence of solar activity in driving the short-lived climatic events during the LGM. Furthermore, evidence for an active ENSO during LGM suggests that both inter-annual SST variability and solar-induced monsoonal fluctuations collectively drove the observed climatic anomalies.

Chapter 7

Bottom water oxygenation changes in the western Bay of Bengal since the Last Glacial Maximum

7.1 Introduction

The dissolved oxygen in the oceans strongly modulates marine life and the environment, and also controls the nitrogen and carbon cycles (Rixen et al., 2020). The oxygen concentration in the ocean has been declining since at least the industrial revolution. This global warming driven reduction in the dissolved oxygen concentration has been reported from estuaries and coastal areas, as well as on the continental shelves and in the open ocean (Rixen et al., 2020; Schmitko et al., 2017). The oxygen deficient zones (ODZ) covering around 8% of the total ocean area are a major concern in the context of the warming world (Paulmier and Ruiz-Pino, 2009). The frequent denitrification in ODZ also strongly affects the global nitrogen cycle (Rixen et al., 2020, Naqvi et al., 2010; Paulmier and Ruiz-Pino, 2009). The largest three denitrification centres are located in the Eastern Tropical North Pacific, Eastern Tropical South Pacific, and the Arabian Sea (DeVries et al., 2012). The Bay of Bengal ODZ is the fourth most significant denitrification centre, but remains incomprehensible. Although, the biogeochemical estimates don't show evidence of very strong denitrification in the Bay of Bengal, its ODZ is on the verge of functional anoxia (Azhar et al., 2017; Bristow et al., 2016; Naqvi et al., 1996).

The anthropogenic nutrient input (sewage and fertilizers) resulting in eutrophication in marginal marine regions, is suggested as the primary cause of increasing deoxygenation on short time scale (Rabalais et al., 2010). The productivity changes modulate decadal to centennial and longer term changes in oxygenation even in the open ocean (Hoogakker et al., 2024; Rixen et al., 2020). The freshwater-induced stratification also contributes to the short-term oxygenation changes (Rixen et al., 2020). However, the ventilation changes associated with changes in upwelling intensity, deep water source, ocean gateways, and the overall structure of the ocean basins affect the dissolved oxygen concentration on centennial, millennial, and longer time scales (Hoogakker et al., 2024; Rixen et al., 2020). Therefore, it is important to understand the oxygen levels change on a longer timescale, beyond the anthropogenic influence, in response to different climatic boundaries of the late Quaternary, which will help to understand the future response of

ODZ in a global warming scenario.

The oxygen deficiency strongly affects the benthic community and biogeochemical processes across the Arabian Sea and Bay of Bengal (Saraswat et al., 2022; Rixen et al., 2020; Nigam et al., 2007). Benthic foraminifera, single-celled microorganisms with hard shells, can incorporate the signatures of ambient conditions into their morphology and chemical composition, which helps in long-term preservation of the imprints of these ecological characteristics (Saraswat, 2017). Therefore, benthic foraminifera and their morphological features are extensively utilized to reconstruct the past ODZ. The organic matter availability in marine sediments and its link with ODZ strongly modulates benthic foraminiferal abundance (Saalim et al., 2021; Singh et al., 2021; Suokhrie et al., 2020). Benthic foraminiferal assemblages has the unique advantage of being able to delineate oxic (>1.5 mL/L O_2), suboxic (0.3 – 1.5 mL/L O_2), dyoxic (0.1 – 0.3 mL/L O_2), and anoxic (<0.1 mL/L O_2) facies (Kranner et al., 2022; Saalim et al., 2021; Kaithwar et al., 2020). Additionally, the benthic foraminifera have been divided into a variety of morphogroups typically influenced by their ambient environmental conditions (Singh et al., 2022; Verma et al., 2021; Nigam et al., 2007; Corliss and Fois, 1990). The distinct morphogroups representing a high and low oxygen environment and excellent preservation of their hard shells make them unique to reconstruct the past ODZ variability (Singh et al., 2022; Saraswat and Nigam, 2013; Nigam et al., 2007).

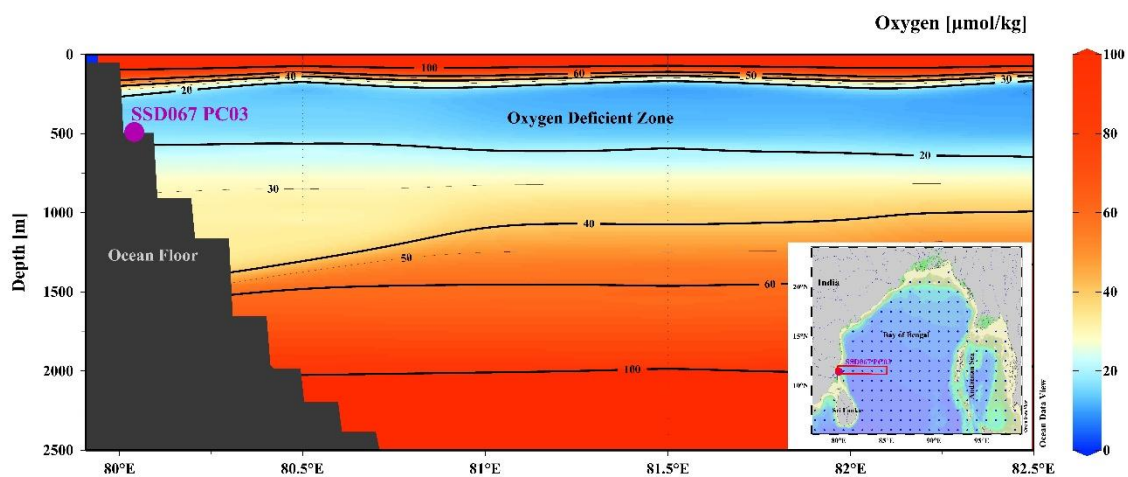


Figure 7.1 Location of the sediment core (SSD067 PC03) in the southwestern Bay of Bengal with a section plot of dissolved oxygen ($\mu\text{mol/kg}$) along the 11.5°N latitude plotted using Ocean Data View software. The dataset of World Ocean Atlas 2023 was taken for the section plot (Garcia et al., 2019). The blue colour band is the modern-day ODZ extent showing the dissolved oxygen level below $20 \mu\text{mol/kg}$ in the southwestern Bay of Bengal.

Based on foraminiferal and geochemical proxies, a lot of efforts have been made to reconstruct the past OMZ variability from the Arabian Sea (Majumder et al., 2022; Gaye et al., 2018; Singh et al., 2015; Naik et al., 2014; Naik et al., 2017; Nigam et al., 2009;

Pichevin et al., 2007; Ivanochko et al., 2005; Altabet et al., 2002). However, the records from the Bay of Bengal are limited. There is no record of ODZ variability during the last glacial and interglacial period (Verma et al., 2021; Pattan et al., 2013; Ma et al., 2019). Therefore, it is important to understand the intensity and extent of ODZ during different climatic boundaries since the last glacial period. Here, a centennial to multi-centennial resolution record of benthic foraminifera morphogroups and sediment geochemistry is used to understand the influence of monsoon-induced productivity and riverine influx on ODZ in the southwestern Bay of Bengal.

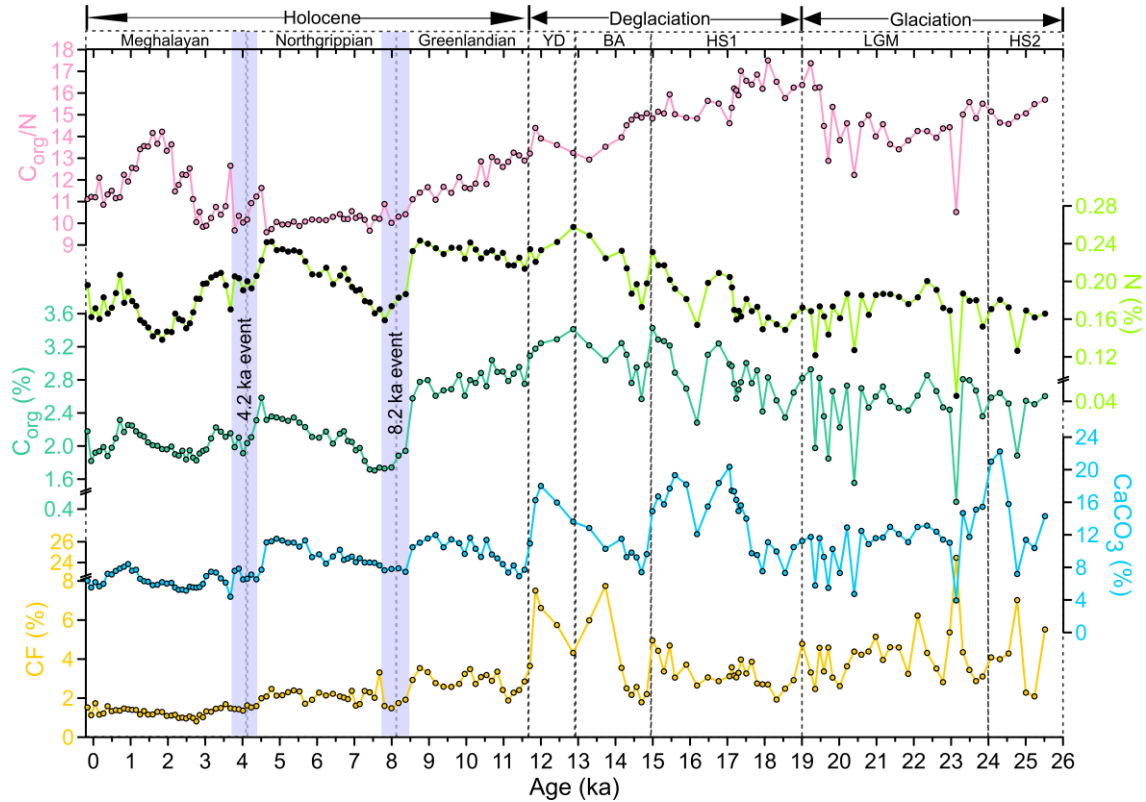


Figure 7.2 Variation in coarse fraction (CF), calcium carbonate (CaCO_3), organic carbon (C_{org}), and nitrogen (N) percentage, and organic carbon to nitrogen ratio ($\text{C}_{\text{org}}/\text{N}$) during the last 26 kyr in the southwestern Bay of Bengal. All the climatic events are marked with the dotted lines. The 8.2 ka and 4.2 ka events are highlighted by the light blue bar.

7.2 Results

7.2.1 Dissolved oxygen concentration at the core location

The annual average dissolved oxygen (DO) concentration data was taken from the World Ocean Atlas 2023. The upper boundary of the ODZ was at 150 m and the lower boundary was at 600 m. The DO concentration was $<20 \mu\text{mol/kg}$ in this depth interval (**Fig. 7.1**). Below 600 m, DO concentration increased. From 600 to 900 m, DO increased from 20 to $30 \mu\text{mol/kg}$ (**Fig. 7.1**). The location of SSD067 PC03 lies well within the ODZ. Therefore,

the changes in bottom water DO will be preserved in benthic foraminiferal morphogroups and geochemical characteristics of the sediments.

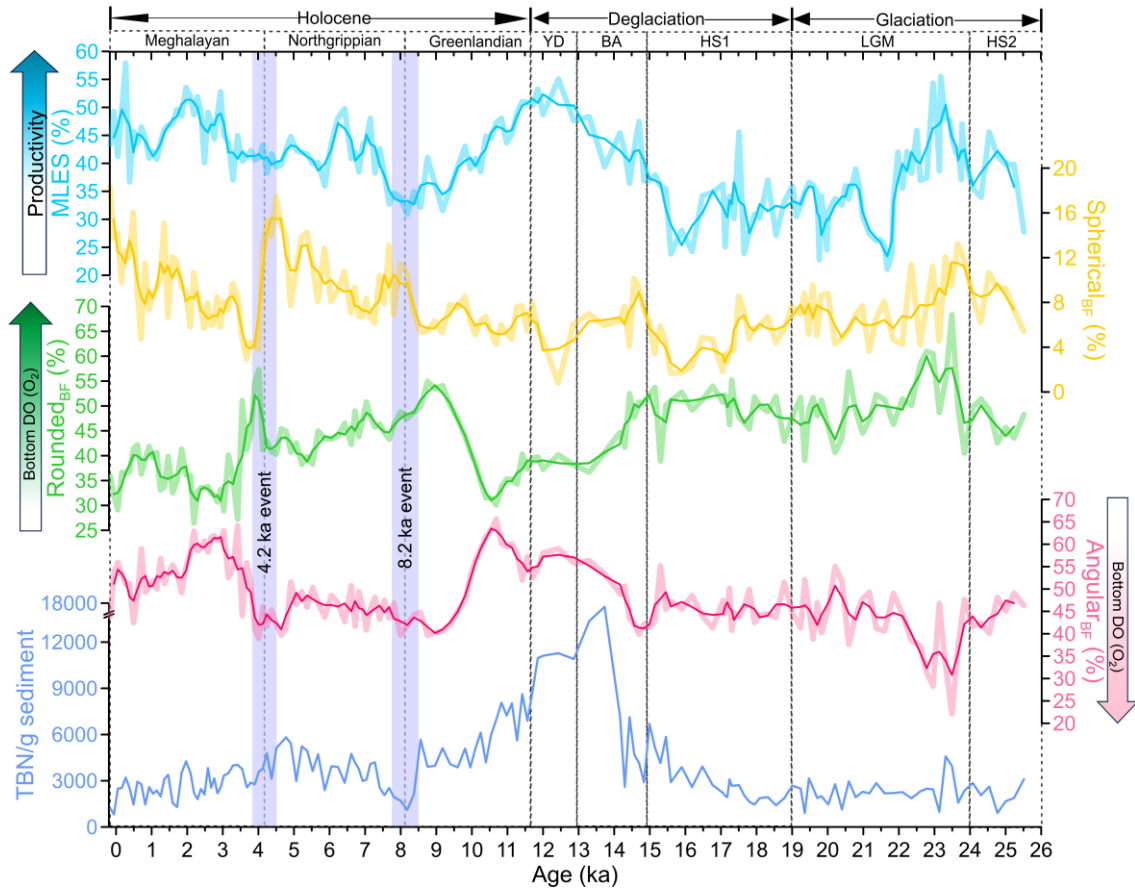


Figure 7.3 Variation in the relative abundance of angular asymmetrical benthic foraminifera ($Angular_{BF}$), rounded symmetrical benthic foraminifera ($Rounded_{BF}$), spherical benthic foraminifera ($Spherical_{BF}$), and an assemblage of mixed layer eutrophic planktic foraminifera species (MLES) during the last 26 kyr in the southwestern Bay of Bengal. All the climatic events are marked with the dotted lines. The 8.2 ka and 4.2 ka events are highlighted by the light blue bar.

7.2.2 Coarse fraction (CF, $>63 \mu m$)

The abundance of coarse fraction ($>63 \mu m$), comprising of both the biogenic shells and the terrigenous sand sized particles, varied from 0.81 to 24.46% (**Fig. 7.2**). The highest CF was observed during the Last Glacial Maximum (LGM) and Younger Dryas (YD). The CF declined at the onset of the Holocene, but it was still high during the Greenlandian Age as compared to the Northgrippian and Meghalayan ages (**Fig. 7.2**). The lowest CF percentage was observed during the Meghalayan Age (**Fig. 7.2**).

7.2.3 Variation in $CaCO_3$, C_{org} , N percentage, and C_{org}/N ratio

The $CaCO_3$ content in the bulk sediments varied from 4.39 to 22.25% with the highest values recorded during the Heinrich Stadial 1 (HS1) and YD periods (**Fig. 7.2**). In contrast,

the CaCO_3 content was notably lower during the LGM and Bølling-Allerød (BA) (**Fig. 7.2**). A significant decrease in CaCO_3 content was observed at 11.7 ka, 8.2 ka, and 4.2 ka. The lowest CaCO_3 was observed during the Meghalayan Age (**Fig. 7.2**). Both, C_{org} and N content were low during the Heinrich Stadial 2 (HS2) and LGM, but began to increase at 19 ka and continued until 13 ka (**Fig. 7.2**). Both, C_{org} and N abruptly decreased at 11.7 ka, 8.2 ka, and 4.2 ka. The highest N was recorded during the YD, Greenlandian, and towards the end of Northgrippian Age (**Fig. 7.2**). The C_{org}/N ratio peaked during the HS1 and reached its lowest point during the Northgrippian Age (**Fig. 7.2**). During the LGM, C_{org}/N was relatively low compared to HS2 and HS1. Prior to the Holocene, from 26 ka to 11.7 ka, C_{org}/N was higher, whereas it was comparatively low during the Holocene (**Fig. 7.2**). Interestingly, C_{org}/N was relatively high during the Greenlandian and Meghalayan Ages as compared to the Northgrippian Age (**Fig. 7.2**).

7.2.4 Benthic foraminiferal morphogroups

The total benthic foraminiferal number (TBN/g sediment) was low during HS2 and LGM but began to increase at 19 ka and continued until ~13.5 ka (**Fig. 7.3**). Interestingly, the abundance of total benthic foraminifera was significantly higher during BA, YD, and Greenlandian Age, coinciding with higher concentration of C_{org} and N (**Fig. 7.2, 7.3**). There was a marked decrease in benthic foraminiferal abundance during the 8.2 ka and 4.2 ka events, with lower values observed during the Meghalayan Age as compared to the Greenlandian and Northgrippian ages (**Fig. 7.3**).

The relative abundance of low oxygen and high food availability preferring angular asymmetrical benthic foraminiferal morphogroup ($\text{Angular}_{\text{BF}}$) (Nigam et al., 2007), was high during the HS2, BA, YD, Greenlandian and Meghalayan ages, suggesting an intensification of ODZ (**Fig. 7.3**). In contrast, $\text{Angular}_{\text{BF}}$ were less abundant during the LGM and Northgrippian Age. The rounded benthic foraminiferal morphogroup ($\text{Rounded}_{\text{BF}}$) which prefers oxygenated bottom water conditions (Nigam et al., 2007), was less abundant during the HS2, BA, YD, Greenlandian and Meghalayan ages, while showing high abundance during LGM, HS1, and Northgrippian Age, suggesting weaker ODZ and more oxygenated bottom water conditions (**Fig. 7.3**). The spherical benthic foraminiferal morphogroup ($\text{Spherical}_{\text{BF}}$), thriving in moderate to high oxygen environments (Singh et al., 2022), also indicate well-oxygenated conditions during the LGM and Northgrippian Age (**Fig. 7.3**).

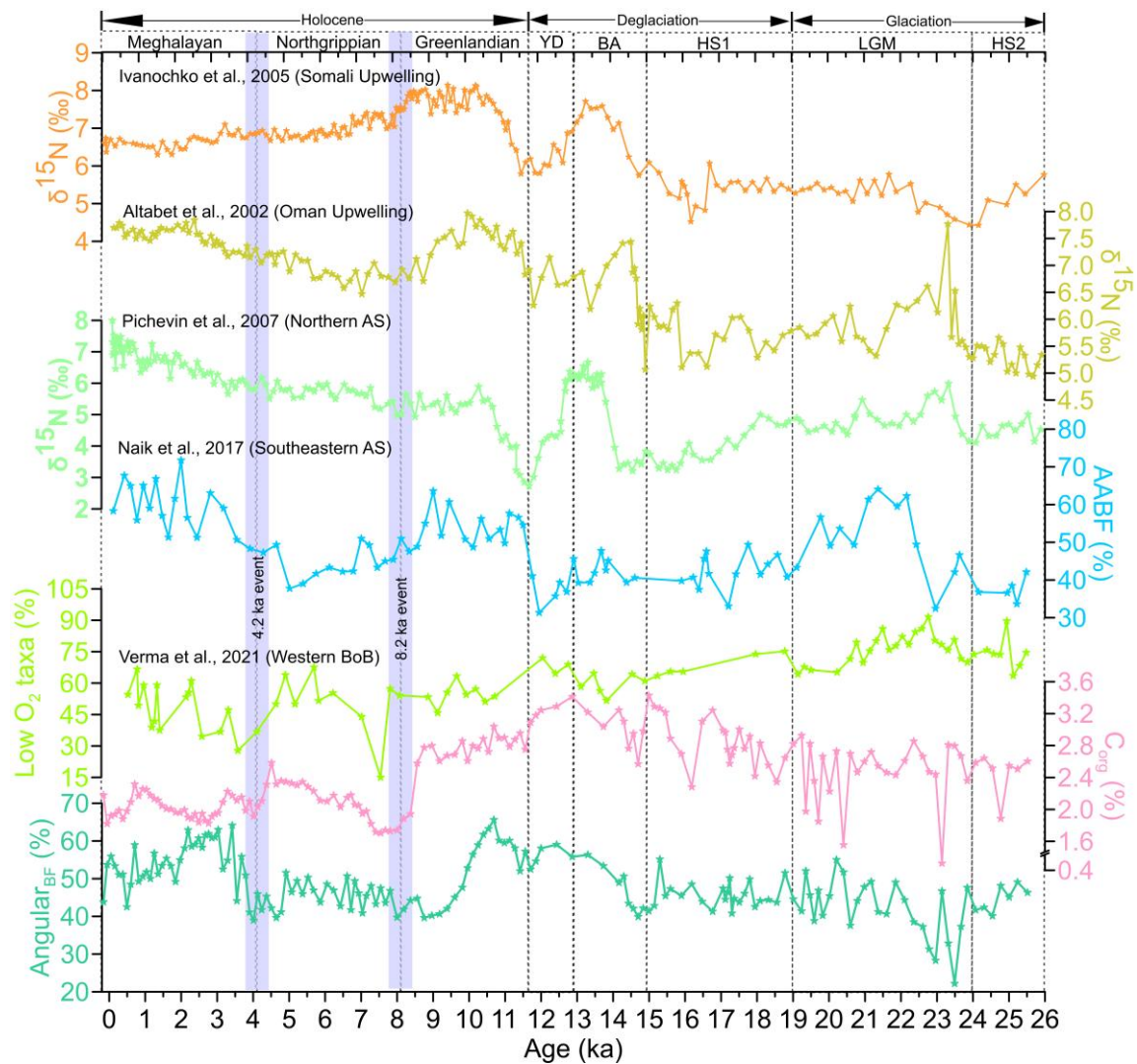


Figure 7.4 A comparison of data of this core (SSD067 PC03) with other past oxygen deficient zone records from the Bay of Bengal and Arabian Sea [Verma et al., 2021 (Western Bay of Bengal); Naik et al., 2017 (Southeastern Arabian Sea); Pichevin et al., 2007 (Northern Arabian Sea); Altabet et al., 2002 (Oman upwelling); Ivanochko et al., 2005 (Somali upwelling)]. All the climatic events are marked with the dotted lines. The 8.2 ka and 4.2 ka events are highlighted by the light blue bar.

7.2.5 Planktic foraminiferal assemblage

The assemblage of mixed-layer planktic foraminifera species including *Globigerinita glutinata*, *Globigerina bulloides*, and *Globigerina falconensis* was used as surface productivity indicator. These three species of planktic foraminifera are classified as mixed-layer eutrophic species (MLES) because they flourish in cooler, nutrient-rich eutrophic environments (Salman and Saraswat, 2024; Verma et al., 2022; Gupta et al., 1997). The higher relative abundance of mixed-layer eutrophic species (MLES) suggests high surface productivity during HS2, BA, YD, and Greenlandian Age while it was low during the LGM and HS1 (**Fig. 7.3**). There was a significant decrease in MLES during the 8.2 ka and

4.2 ka events. The high abundance of MLES suggests a higher productivity during the Meghalayan Age (**Fig. 7.3**).

7.3 Discussion

7.3.1 Glacial-interglacial changes in bottom water oxygenation

The highest and lowest relative abundance of both Angular_{BF} and Rounded_{BF} were similar. Although, most of the picked specimens were clubbed in either of these morphogroups, the spherical specimens were grouped separately. For the large part of the last 26 kyr, the relative abundance of Angular_{BF} and Rounded_{BF} morphogroups varied ~50%. The comparable relative abundance of these two morphogroups suggests a moderately oxygenated environment, occasionally tipping towards either more or less oxic conditions. The food availability due to organic carbon flux to the seafloor, supported by primary productivity, and bottom water oxygen largely controls the benthic foraminiferal abundance and diversity in a deep-sea environment (Singh et al., 2022, 2021; Saalim et al., 2021; Suokhrie et al., 2020; Kaithwar et al., 2020). Based on modern ecology and distribution of benthic foraminifera in surface sediments along the southern and western continental margin of India, several authors have suggested that the Angular_{BF} or infaunal benthic foraminifera flourish in low oxygen and high food availability environments, while Rounded_{BF} or epifaunal benthic foraminifera prefer oxygenated and oligotrophic environments (Singh et al., 2022; Nigam et al., 2007). The Spherical_{BF} morphogroup, however, includes benthic foraminifera inhabiting shallow infaunal microhabitats and preferring sufficient food availability and excess oxygen (Singh et al., 2022). Corliss and Fois (1990) have suggested their preference for low levels of marine organic matter. Interestingly, the high abundance of similar-looking spherical species has been reported from high riverine flux areas in front of the Ganga-Brahmaputra rivers (Manasa et al., 2016). The relative abundance of Spherical_{BF} morphogroup positively correlated with C_{org}/N ratio, suggesting its preference for more refractory type organic matter (Singh et al., 2022; Verma et al., 2018).

The relative abundance of Angular_{BF} morphogroup indicated that the bottom water was well oxygenated during the early part of the LGM, and less oxygenated during HS2 (**Fig. 7.3**). Interestingly, the high abundance of Rounded_{BF} morphogroup also suggested well oxygenated bottom waters during the LGM (**Fig. 7.3**). Earlier, based on a record of benthic foraminifera assemblage from the deeper depth (>3300 m water depth) in the western BoB, oxygen-poor bottom water condition was inferred during the LGM (**Fig. 7.4**,

Verma et al., 2021). However, other records based on benthic foraminiferal assemblage and $\delta^{13}\text{C}$ suggested bottom water deoxygenation during LGM, and linked it with reduction in North Atlantic Deep Water (NADW) and higher contribution of southern-sourced water into the southern Bay of Bengal (Ma et al., 2019; Raza et al., 2014). A majority of these records were from depths, well below the regions where ODZ impinges on to the

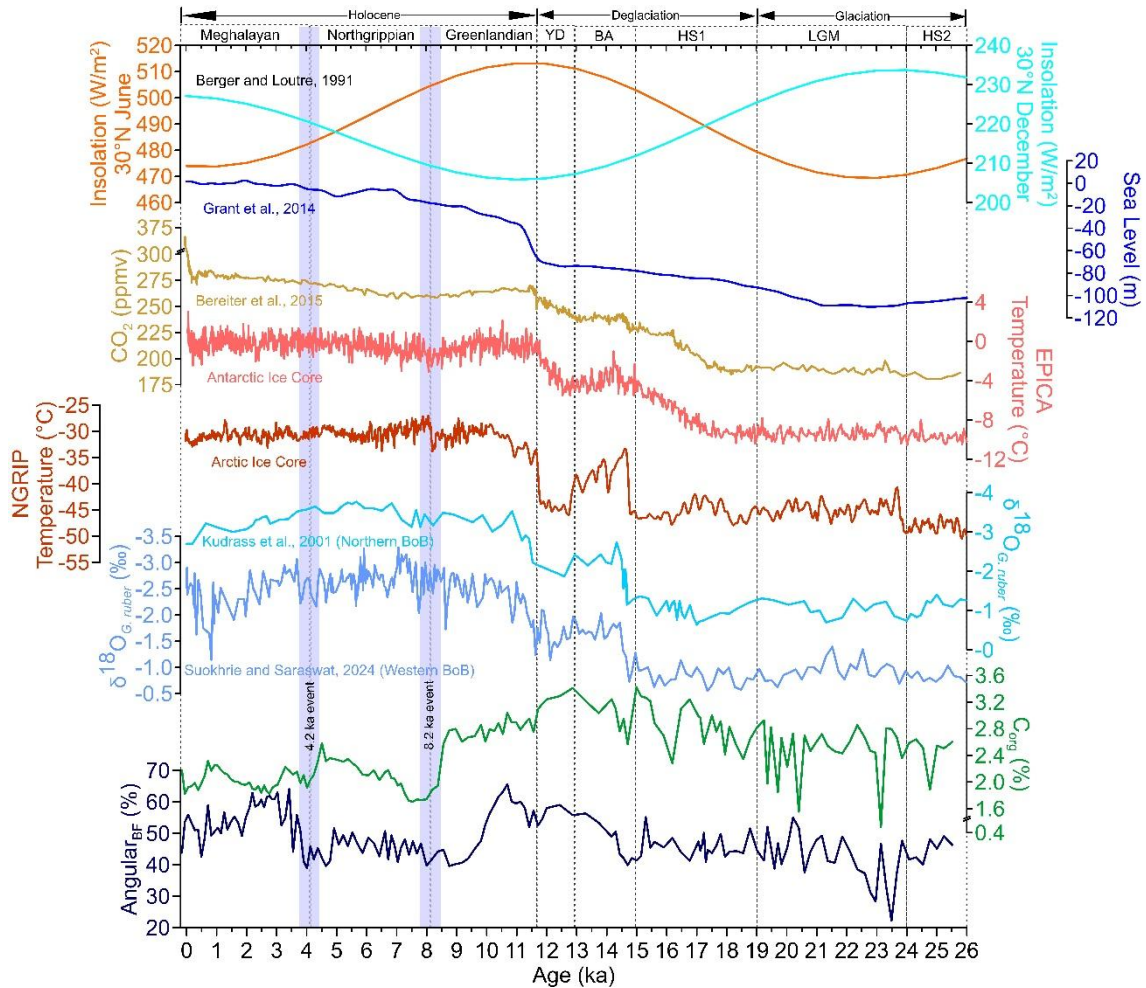


Figure 7.5 Variation in $\delta^{18}\text{O}$ (‰) of *G. ruber* from the western and northern Bay of Bengal (Suokhrrie and Saraswat, 2024; Kudrass et al., 2001), Greenland atmospheric temperature (°C) from the Northern Greenland Ice Core Project (NGRIP, Davtian and Bard, 2023), Antarctica atmospheric temperature (°C) from the European Project for Ice Coring in Antarctica (EPICA, Jouzel and Masson-Delmotte, 2007), atmospheric CO_2 concentration (Bereiter et al., 2015), sea level changes (Grant et al., 2014) and seasonal insolation changes at 30°N during June and December (Berger and Loutre, 1991) during the last 26 kyr. All the climatic events are marked with the dotted lines. The 8.2 ka and 4.2 ka events are highlighted by the light blue bar.

sediments. The present record is from a shallow water depth ($>509\text{ m}$) and lies well within the ODZ (**Fig. 7.1**). Interestingly, benthic foraminiferal shells in the samples with the lowest relative abundance of Angular_{BF} morphogroup during the early part of LGM were dull in lustre, partially abraded, and also included a few relict specimens. The coarse fraction was comparatively higher in these samples and included a large fraction of

terrestrial material. Therefore, the higher relative abundance of Rounded_{BF} morphogroup in this section is attributed to the transport of material from the shallow depth.

The relative abundance of Rounded_{BF} was only marginally higher than that of Angular_{BF} during the latter part of LGM and HS1, suggesting a moderately well oxygenated environment. A drop in oxygen concentration is inferred during the BA, YD and early Greenlandian Age, as the relative abundance of Angular_{BF} was the highest during this interval. The similar low oxygen levels are also inferred during the early Meghalayan Age, as the relative abundance of Angular_{BF} was higher during this interval as well. The Northgrippian is characterized as moderately well oxygenated interval, as the relative abundance of Spherical_{BF} was the highest during this interval.

7.3.2 Effect of water masses in modulating ODZ

Three distinct high salinity water masses, namely Arabian Sea High-Salinity Water (ASHSW), Persian Gulf Water (PGW), and Red Sea Water (RSW), highly influence the oxygen availability in the southwestern Bay of Bengal through currents during the southwest monsoon (Jain et al., 2016). The depth of RSW ranges from 500–1000 m and the presence of PGW exists at 200–450 m in the southwestern Bay of Bengal (Sheehan et al., 2020; Jain et al., 2016). The mixing of oxygenated PGW into the ODZ of Bay of Bengal plays a crucial role in maintaining oxygen levels above the threshold at which denitrification takes place (Sheehan et al., 2020). The outflow from the Persian Gulf was completely cut off during the LGM due to a lower sea level (~130 m) than the present-day water depth (~80 m) of the Strait of Hormuz (Siddall et al., 2003). The outflow from the Red Sea decreased by 85% during the LGM (Lachkar et al., 2019). Still, a moderately-oxygenated condition prevailed at the core site during the LGM. However, Angular_{BF} morphogroup from the southeastern Arabian Sea indicated oxygen-depleted conditions during the LGM (Naik et al., 2017, Fig. 7.4). Similarly, the $\delta^{15}\text{N}$ (‰) record from the northern Arabian Sea and Oman upwelling also suggested deoxygenation during the LGM (Pichevin et al., 2007; Altabet et al., 2002, Fig. 7.4). However, relatively lower $\delta^{15}\text{N}$ (‰) during the LGM suggested oxygenated conditions along the Somali upwelling (Fig. 7.4), which is not affected by RSW and PGW due to its geography location in the south. As the sea level started increasing at the onset of deglaciation, the outflow from the Persian Gulf must have increased. Still, the oxygen-poor conditions prevailed in the southwestern Bay of Bengal during BA, YD, and Greenlandian age (Fig. 7.3). That's a clear indication that, although, the decreased PGW and RSW outflow led to the oxygen depleted conditions in

northwestern to southeastern Arabian Sea, but its impact was not visible in the southwestern Bay of Bengal (**Fig. 7.3**).

7.3.3 Influence of monsoon and its associated processes on ODZ

A large variation was observed in CaCO_3 , C_{org} , N, and $\text{C}_{\text{org}}/\text{N}$ during the last 26 kyr. The dissolved oxygen concentration is closely linked with the primary productivity and subsequent organic matter flux. The low pH, often associated with ODZ, results in dissolution of CaCO_3 . The major part of calcium carbonate (CaCO_3) comes from the shells of microorganisms, like foraminifera and coccolithophores (Schiebel, 2002). The abundance of calcareous shell producing organisms depends on either primary productivity or nutrient availability, the factors strongly modulated by the monsoon in the entire northern Indian Ocean. After death, the hard shells of marine planktons sink to the ocean floor while the shells of benthos are buried in situ within the marine sediments (Saraswat, 2015). However, the organic carbon (C_{org}) and nitrogen (N) content in bulk sediments have both marine-produced and terrestrial-derived sources (Meyers et al., 1994). The oxidation of organic matter is controlled by the availability of oxygen and sediment texture, affecting the C_{org} and N content of the marine sediments. The terrestrial runoff and marine primary productivity in the entire northern Indian Ocean, strongly depends on monsoon. Therefore, monsoon and associated processes primarily modulate the dissolved oxygen variability in this region.

During the LGM, there was a significant reduction in summer monsoon intensity, accompanied by a stronger winter monsoon (Suokhrie and Saraswat, 2024; Verma et al., 2022; Gautam et al., 2021). The stronger winter monsoon must have resulted in a relatively higher winter productivity. However, on an annual average basis, the productivity remained low during the LGM, as indicated by CaCO_3 , C_{org} , N and MLES (**Fig. 7.2, 7.3**). Except the early LGM, MLES assemblage indicates a lower productivity during both late LGM and HS1 (**Fig. 7.3**). The low marine productivity prior to Holocene is also confirmed by the overall higher $\text{C}_{\text{org}}/\text{N}$, as compared to that during the Holocene. The $\text{C}_{\text{org}}/\text{N}$ ratio was ~14 during the LGM and peaked (~17) at the onset of HS1, subsequently declining until the Northgrippian. Therefore, the weaker monsoon and associated low primary productivity led to the moderately oxygenated bottom water during the LGM. *Globigerinoides ruber* $\delta^{18}\text{O}$ records from the western and northern Bay of Bengal also suggested a weaker summer monsoon during the LGM (**Fig. 7.5**). A weaker summer monsoon likely led to a weak stratification during the summer season, which may have

enhanced the supply of oxygenated water during the LGM. The $\delta^{18}\text{O}$ records from the Bay of Bengal, along with temperature records from the Arctic (NGRIP) and Antarctica (EPICA) ice cores, indicate a prominent global cooling during the LGM (**Fig. 7.5**). At the same time, there was a collapse in coastal upwelling in the southwestern Bay of Bengal due to weaker summer winds, as evidenced by a significant decrease in *Globigerina bulloides* abundance (Suokhrie and Saraswat, 2024). As summer winds weakened, it is likely that the intensity of summer monsoon currents also decreased, reducing the transport of high salinity water from the southeastern Arabian Sea to the southwestern Bay of Bengal (Salman and Saraswat, 2024). A significant reduction in oxygen-poor Arabian Sea water during summer, and the influence of oxygenated water from the north as well as through the Cauvery River, may have led to oxygenated bottom water conditions and a weaker ODZ during the LGM. However, it should be noted here that a part of the variability in CaCO_3 , C_{org} , and N percentages can also be due to the changes in sea level. The low sea-stand exposed the entire continental shelf and reduced the distance between the river mouths and core location. The proximity to the river mouth increased the amount of terrestrial influx at the core site, as observed in the early LGM. The increased terrestrial sediment flux reduced the CaCO_3 content due to dilution.

The higher CaCO_3 , C_{org} , and N content suggests increased productivity during HS1. However, the bottom waters were as oxygenated as during the late LGM, as evident from the similar $\text{Angular}_{\text{BF}}$ relative abundance. The relative abundance of MLES was also low during HS1, as it was during the late LGM. Therefore, the marine productivity was definitely low during the HS1. The higher CaCO_3 , C_{org} , and N content despite of the low marine productivity is attributed to the increased terrestrial flux and better preservation. The increased terrestrial contribution to the C_{org} is confirmed by the higher C_{org}/N ratio, peaking during HS1 (**Fig. 7.2**).

The dissolved oxygen concentration decreased during late BA, YD, and Greenlandian Age, as evident from the higher relative abundance of $\text{Angular}_{\text{BF}}$, lower relative abundance of $\text{Rounded}_{\text{BF}}$ and $\text{Spherical}_{\text{BF}}$ morphogroups. This interval is marked by a higher primary productivity as indicated by a prominent increase in MLES and TBN/g sediment (**Fig. 7.3**). The high C_{org} and N content suggests increased organic matter flux to the sea floor that supported the intensification of ODZ during the BA, YD and Greenlandian Age (**Fig. 7.3**). The higher contribution of marine productivity to the C_{org} is confirmed by a prominent decrease in C_{org}/N ratio.

The atmospheric greenhouse gas concentration and seasonal insolation changes were the key drivers of the summer and winter monsoon changes over longer time scale, including glaciation, deglaciation and the Holocene (**Fig. 7.5**). As the intensification of monsoon began with the onset of deglaciation, the ODZ strengthened, leading to oxygen-depleted conditions in the southwestern Bay of Bengal (**Fig. 7.5**). A stronger summer monsoon and the increased influence of oxygen-poor Arabian Sea water, intensified ODZ during the deglaciation and Greenlandian Age.

The bottom waters were well oxygenated during the 8.2 ka and 4.2 ka events as well as during Northgrippian Age (**Fig. 7.3**) when both the summer and winter monsoons were weaker as evident from low MLES and the lowest C_{org}/N ratio (**Fig. 7.2**) (Dixit et al., 2014; 2024). $CaCO_3$, C_{org} , N and MLES suggest an abrupt decline in marine productivity at 8.2 ka and 4.2 ka events (**Fig. 7.2 & 7.3**), which resulted in oxygenated conditions in the southwestern Bay of Bengal. Interestingly, the relative abundance of Rounded_{BF} and Angular_{BF} morphogroups suggests better oxygenated conditions during the Northgrippian than the Meghalayan Age (**Fig. 7.3**). The lowest C_{org}/N ratio also indicates weak summer and winter monsoons. A weak summer monsoon also means a sluggish summer monsoon current, leading to an oxygenated condition during the Northgrippian Age. The high (low) relative abundance of Angular_{BF} (Rounded_{BF}) morphogroups suggest a stronger (weaker) ODZ during Meghalayan Age compared to Northgrippian, HS1, and LGM (**Fig. 7.3**). The MLES assemblage indicates a higher surface productivity but the low C_{org} and N content suggests, degradation of organic matter leading to a high C_{org}/N ratio during the Meghalayan Age (**Fig. 7.2**).

7.4 Conclusions

The reconstruction of bottom water deoxygenation and ODZ dynamics in the southwestern Bay of Bengal reveals a strong linkage with monsoon variability and associated oceanographic processes. During the LGM and HS1, reduced summer monsoon intensity and enhanced winter monsoon favoured moderately oxygenated bottom water conditions, as indicated by benthic foraminifera morphogroups and productivity markers. Despite of the reduced inflow of oxygen-rich PGW and RSW during LGM, weaker summer stratification, reduced productivity, influx of oxygenated water due to stronger winter monsoon, and reduced influx of oxygen-poor Arabian Sea water contributed to moderately oxygenated conditions. With the onset of deglaciation, the intensifying monsoon and increased productivity led to enhanced organic carbon flux to the seafloor, triggering

bottom water deoxygenation and strengthening of the ODZ, particularly during the BA, YD and Greenlandian. The Meghalyan Age also experienced an intense ODZ due to high surface productivity. In contrast, periods like 8.2 ka and 4.2 ka events and the Northgrippian Age were marked by weak monsoons, reduced productivity, and thus well-oxygenated bottom waters. Overall, this work emphasizes that monsoon-driven surface productivity, stratification, and water mass exchange are critical in modulating the strength and extent of ODZ in the Bay of Bengal.

Chapter 8

Summary and Future Scope

8.1 Summary

The present work includes the ecology and distribution of planktic and benthic foraminifera in 67 core-top surface samples and a paleomonsoon reconstruction based on a 15.72 m long sediment core (SSD067 PC03) from the southwestern Bay of Bengal. A total of 123 surface samples (67 from 0–1 cm and 56 from 1–2 cm) were used for the modern ecological study, and 1572 subsurface samples were analysed to understand the monsoon and its associated productivity and oxygen level changes during the multi-decadal climatic events since the last glacial maximum. The key findings are summarized below.

Planktic foraminiferal ecology and carbon burial

- A very high planktic foraminiferal number in the swBoB suggests the biannual influence of the Palk Bay and Gulf of Mannar waters during the summer monsoon season as well as the higher productivity during the winter season.
- *Globigerinoides ruber* showed a preference for high salinity and warm waters.
- The relative abundance of *G. bulloides* responds to the upwelling during summer whereas that of *G. falconensis* depends on the conditions during the winter monsoon.
- The chaotic distribution pattern of *G. glutinata* may be the result of cyclonic eddies during the winter season.
- *Neogloboquadrina dutertrei*, *G. menardii*, *G. siphonifera* and *G. conglomerata* were negatively correlated with mixed layer temperature and mixed layer salinity, as well as thermocline temperature and positively correlated with thermocline salinity.
- The high abundance of *T. sacculifer* at deeper depths suggests that *T. sacculifer* was less prone to dissolution as compared to *T. trilobus*.
- The positive correlation of *G. falconensis*, *G. incompta*, *G. bulloides*, *H. pelagica* and *T. quinqueloba* with Chlorophyll-a suggests that the food availability also strongly controlled the planktic foraminiferal abundance.
- The highest carbon burial was observed in the southernmost part of the western BoB. The CaCO₃ burial was much higher in the southern part of swBoB where the planktic

foraminiferal flux was also very high. Therefore, the planktic foraminifera contribute significantly to the total carbon burial in the surface sediments of the swBoB.

Benthic foraminifera ecology and lysocline

- From the surface distribution of living benthic foraminifera, it is inferred that low dissolved oxygen levels, particularly at intermediate depth, correlate with high living benthic foraminifera abundance due to increased food availability and reduced predatory pressure.
- However, as dissolved oxygen concentration increases, a decline in benthic foraminifera abundance is observed, particularly on the shelf and deeper regions where dissolved oxygen is high and C_{org} content is low.
- The food availability is the primary factor controlling the total living benthic foraminiferal abundance in the southwestern Bay of Bengal.
- Furthermore, the research underscores the role of turbulence and energy conditions in shaping the abundance of angular and rounded morphogroups of benthic foraminifera, with dissolved oxygen concentration emerging as the primary factor influencing their distribution.
- The study also highlights the importance of considering the top 2 cm of sediments to assess the living benthic foraminiferal population, as significant fractions may be overlooked if only the top 1 cm is analyzed.
- The bathymetric shift in calcareous and agglutinated foraminifera suggested that the foraminiferal lysocline depth lies at 1250 m in the southwestern Bay of Bengal.
- Overall, both the organic matter content and the dissolved oxygen concentration were the dominant factors affecting living benthic foraminifera abundance in the southwestern Bay of Bengal.

4.2 ka event

- The 4.2 ka event began at 4.65 ka and persisted until 4.1 ka, with a duration of 550 years.
- The event exhibited diachroneity in its onset across different records but resulted in a significant decrease in carbon burial. The weakening of summer monsoon and the strengthening of winter monsoon were key drivers reducing the carbon burial during the 4.2 ka event.

- The synchronous decrease in coarse fraction, CaCO₃ and nitrogen indicated a collapse in productivity. However, the winter monsoon assemblage of planktic foraminifera species and high C_{org}/N ratio suggested a stronger winter monsoon.
- The decrease in species associated with summer monsoon season indicated a weakening of summer monsoon. The behaviour of winter monsoon was opposite to that of the summer monsoon during the 4.2 ka event.

8.2 ka event

- A strong influence of 8.2 ka event on the monsoon in the Indian subcontinent is reported, based on a decadal resolution marine sedimentary record recovered from the southwest Bay of Bengal.
- The abrupt decrease in coarse fraction (>63 µm), CaCO₃, C_{org}, and N at 8.5 ka, points towards a collapse in productivity.
- The decrease in the C_{org}/N ratio at 8.5 ka, also suggests an abrupt weakening of the monsoon that continued till 8.0 ka. The total planktic foraminifera number abruptly decreased between 8.5 to 8.0 ka responding well to the event.
- This event affected the climate for 800 years in the Indian subcontinent and nearby oceans.
- The difference between $\delta^{18}\text{O}_{G. ruber}$ and $\delta^{18}\text{O}_{N. dutertrei}$ was high during the 8.2 ka event, unlike before and after the event, suggesting that the sea was well stratified during the 8.2 ka event.

Multiple high-amplitude short-term climatic events during the LGM

- This study provides a high-resolution reconstruction of LGM transient climatic anomalies from the Bay of Bengal. Eleven short-lived climatic events were identified between 19 and 24 ka.
- The duration of these climatic events varied from a few decades to a few centuries.
- The abrupt increase in coarse fraction suggests an intensified riverine influx likely driven by enhanced winter monsoon and cyclonic activity along the southeast coast of India. This is further supported by a simultaneous decrease in total planktic foraminifera, TIC, C_{org}, and N, reflecting rapid terrestrial input and a possible dilution effect on marine productivity.
- The summer and winter monsoon assemblages suggested a weaker summer monsoon and a stronger winter monsoon during a few transient climatic events. *Globigerina*

bulloides suggested a stronger summer monsoon from 24 to 23.2 ka and a weaker summer monsoon for rest of the section used for this study.

- The spectral and wavelet analysis revealed distinct cyclicity in CaCO_3 and C_{org} content, with notable periodicities of 312 yr, 340 yr, 1538 yr, and 2778 yr. The alignment of long-term cycles with Hallstatt solar and Bond cycles, and ~312–340 yr cycles with Grand Solar Cycles highlight the potential influence of solar activity in driving the short-lived climatic events during the LGM.
- Furthermore, evidence for an active ENSO during LGM suggests that both interannual SST variability and solar-induced monsoonal fluctuations collectively drove the observed climatic anomalies.

Bottom water oxygenation changes since the Last Glacial Maximum

- The reconstruction of bottom water deoxygenation and ODZ dynamics in the southwestern Bay of Bengal reveals a strong linkage with monsoon variability and associated oceanographic processes.
- During the LGM and HS1, reduced summer monsoon intensity and enhanced winter monsoon favoured moderately oxygenated bottom water conditions, as indicated by benthic foraminifera morphogroups and productivity markers.
- Despite of the reduced inflow of oxygen-rich PGW and RSW during LGM, weaker summer stratification, reduced productivity, influx of oxygenated water due to stronger winter monsoon, and reduced influx of oxygen-poor Arabian Sea water contributed to moderately oxygenated conditions.
- With the onset of deglaciation, the intensifying monsoon and increased productivity led to enhanced organic carbon flux to the seafloor, triggering bottom water deoxygenation and strengthening of the ODZ, particularly during the BA, YD and Greenlandian Age.
- The Meghalayan Age also experienced an intense ODZ due to high surface productivity. In contrast, periods like 8.2 ka and 4.2 ka events and the Northgrippian Age were marked by weak monsoons, reduced productivity, and thus well-oxygenated bottom waters.
- Overall, this work emphasizes that monsoon-driven surface productivity, stratification, and water mass exchange are critical in modulating the strength and extent of ODZ in the Bay of Bengal.

8.2 Future scope

This thesis provides deep insights into the current ecology of foraminifera and robust qualitative reconstruction of the monsoon, productivity, dissolved oxygen and associated changes from the western Bay of Bengal. However, the area is still lacking the quantitative estimation of monsoon as well as dissolved oxygen levels during the past. The foraminiferal morphology and chemical composition of shells can be used to quantify the key climatic parameters during the past. In addition to that, the study has also focused on the response of monsoon seasonality and productivity to the short-term climatic anomalies like 8.2 and 4.2 ka events. The sediment core used for this study had a decadal to multi-decadal resolution and covered the last 26 ka. Therefore, a similar multi-proxy approach can be applied for other short-term climatic events. Based on the background provide in this thesis, the following research work can be taken up as a future project.

- Quantification of dissolved oxygen concentration from the western Bay of Bengal by using the benthic foraminiferal morphology and their geochemical composition.
- Reconstruction of upper water column stratification using $\delta^{18}\text{O}$ and elemental ratio based seawater temperature of the mixed layer and thermocline species during the short-term climatic events of the last 26 ka.

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List of Publications

- Salman, M.**, & Saraswat, R. (2024). Intrusion of Arabian Sea high salinity water and monsoon-associated processes modulate planktic foraminifera abundance and carbon burial in the southwestern Bay of Bengal. *Environment Science and Pollution Research*, 31, 24961-24985. <https://doi.org/10.1007/s11356-024-32685-4>.
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- Saraswat, R., Fathima, R., **Salman, M.**, Suokhrie, T., & Saalim, S.M. (2024). Decoupling of carbon burial from productivity in the north-east Indian Ocean. *Science of the Total Environment*, 947, 174587. <https://doi.org/10.1016/j.scitotenv.2024.174587>
- Saraswat, R., Suokhrie, T., Naik, D. K., Singh, D. P., Saalim, S. M., **Salman, M.**, Kumar, G., Bhadra, S. R., Mohtadi, M., Kurtarkar, S. R., & Maurya, A. S. (2023). Large freshwater-influx-induced salinity gradient and diagenetic changes in the northern Indian Ocean dominate the stable oxygen isotopic variation in *Globigerinoides ruber*. *Earth System Science Data*, 15, 171-187. <https://doi.org/10.5194/essd-15-171-2023>.
- Salman, M.**, & Saraswat, R. Precise timing, duration and impact of 4.2 ka event on monsoon seasonality and productivity in the southwestern Bay of Bengal. Under preparation.
- Salman, M.**, Saraswat, R., Singh, D. P., & Maurya, A. S. A pronounced impact of 8.2 ka event on monsoon seasonality, productivity, and upper water column stratification in the southwestern Bay of Bengal. Under preparation.
- Salman, M.**, & Saraswat, R. Multiple, High-amplitude, Short-term Climatic Events during the Last Glacial Maximum. Under preparation.
- Salman, M.**, & Saraswat, R. Bottom water oxygenation changes in the western Bay of Bengal since the Last Glacial Maximum. Under preparation.

List of Conferences (Oral/poster presentations)

- “Glacial-interglacial variability in the bottom water oxygenation in the Bay of Bengal: Insights from benthic foraminifera and sediment geochemistry”, presented in PAGES 2nd PO₂ workshop held from 10-12 December 2025 at Raleigh, North Carolina, USA (Online).
- “Multiple short-lived disruptions in carbon burial during the last glacial maximum”, presented in 15th International Conference on Paleoceanography (ICP15) held from 31-5 September 2025 at the Indian Institute of Science, Bengaluru, India.
- “Precise timing, duration and impact of 4.2 ka event on monsoon seasonality and productivity in the southwestern Bay of Bengal”, presented in PAGES-Young Scientist Meeting and Open Science Meeting 2025 held from 19-24 May 2025 at Tongji University, Shanghai, China.
- “A pronounced impact of 8.2 ka event in the Bay of Bengal”, presented in 29th Indian Colloquium on Micropaleontology and Stratigraphy (ICMS)-2024 held from 17-9 October 2024 at the Department of Geology, Delhi University, Delhi.
- “Anatomy of carbon burial during the 4.2 ka event in the southwestern Bay of Bengal” presented in Samudra Manthan 2023 held from 28-30 September 2023 at CSIR-National Institute of Oceanography Goa.
- “Organic carbon and dissolved oxygen modulate the living benthic foraminifera abundance in the southwestern Bay of Bengal” presented in XXVIII Indian Colloquium on Micropaleontology and Stratigraphy (ICMS)-2022 held from 4-6 May 2022 at Pune University, Pune.
- “Intrusion of Arabian Sea high salinity water and monsoon associated processes modulate planktic foraminifera abundance in the southwestern Bay of Bengal” presented in International Indian Ocean Science Conference (IIOSC)-2022 held from 14-18 March 2022 at CSIR-National Institute of Oceanography, Goa (Online).

