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March 10, 2023

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Key Points

Two faunal assemblages suggest that the surface waters of the Gulf of Mexico changed between 320 to 60 ka.

One form of variation of the Loop Current explains most of the sequence, but a second characterizes the warmest interglacial substages.

Over the Late Pleistocene, fluctuations in the Loop Current link to changes in the water masses of the Caribbean Sea.

Abstract

The Loop Current (LC) in the Gulf of Mexico (GoM) is part of the western North Atlantic circulation. Recording its strength and slowdown variations can help us characterize the regional climate over the Late Pleistocene. To reconstruct the sea surface and the LC intensity in the eastern GoM, we study the distribution patterns of planktonic foraminifera in the core EN-032-18PC, spanning the end of Marine Isotope Stage (MIS) 9 to early MIS-4. We reconstructed a sequence of paleoceanographic events based on stable isotopes ($\delta^{18}O$ and $\delta^{13}C$) of the surface dweller *Globigerinoides ruber* and two faunal assemblages. The first assemblage explains most of the glacial and late interglacial periods, suggesting a subtropical environment with a deep thermocline and a reduced LC due to a moderate inflow of warm Caribbean waters. The second assemblage explains the warmest interglacial substages, dominated by tropical species, a shallow thermocline, and an extended LC, driven by summer insolation. Overall, surface ocean conditions led to more ecological successions and instability during the warmest interglacial substages than during glacial periods, as supported by the stable isotope records. Besides the GoM relationship to AMOC, as a regulator of heat transport to higher latitudes, we suggest that fluctuations in the LC rely on the migration of atmospheric circulation patterns and astronomical insolation forcing.

Plain Language Summary

The Loop Current (LP) in the Gulf of Mexico (GoM) is part of the western North Atlantic circulation. The study of LC variations can help us portray the regional climate during older periods, and of interest, as they could be similar to those we currently live in. We reconstructed past surface ocean conditions during two glacial (cold) and two interglacial (warm) episodes in the GoM. To interpret the past ocean conditions, we studied (1) ancient distribution patterns of surface dwellers' marine microfossils (planktonic foraminifera) and (2) the environmental chemical signals (called stable isotopes) left in their tests. The age of our study spans from 320 ka to 70 ka during the Late Pleistocene. Overall, two scenarios were reconstructed. The first represents subtropical surface waters, suggesting a less intense LC, a deep thermocline, and a moderate inflow of Caribbean waters. The second scenario portrays the warmer periods, which we interpret as tropical waters, an extended LC, a shallow thermocline, and a more significant entry of waters from the Caribbean. We suggest that fluctuations in the GoM primary current are based on atmospheric circulation patterns and Earth's insolation intensity changes over thousands of years.

Keywords: biostratigraphy, factor analyses, Late Pleistocene, stable isotopes, glacial-interglacial, planktonic foraminifera.

1. Introduction

The tropical surface ocean is an essential climatic component as it receives most solar energy. The transport of surface water masses controls the variability of the ocean's interior structure and regional climate (Liu and Philander 2001). Tropical Atlantic circulation strongly influences the deep ocean waters (Lazier et a., 2001). The Gulf of Mexico (GoM) surface currents are part of the Atlantic Meridional Overturning Circulation (AMOC), the engine of the thermohaline circulation and a fundamental element of North American and European weather (Bryden and Imawaki 2001; Johns et al. 2002). The GoM mediates the transference of heat and freshwater from the tropics to the North Atlantic. Moreover, it provides essential services for human support and biological communities (Badan et al. 2005; Gordon 1967; Miloslavich et al. 2010).

The Loop Current System (LC) is the primary current in the GoM. It is determined by wind patterns and the seasonal position of the Intertropical Convergence Zone (ITCZ) (Saha 2010). In the past, the currents experienced changes in circulation and water masses. For example, during glacial Marine Isotope Stage (MIS) 2, it is thought that currents were less intense because the ITCZ migrated south. However, during MIS-1, the currents intensified as the ITCZ migrated north (Schmidt et al., 2004; Schmidt et al., 2006). Model studies suggest that the southward migration of the North Atlantic currents made the atmospheric system and the AMOC unstable and more variable during the last glacial (Sévellec and Fedorov 2013). Palaeoceanographic studies in the GoM have reconstructed surface ocean conditions at different time scales, focusing on the last glacial cycle (i.e., MIS-4 to MIS-1). For instance, efforts have been made over the Holocene (Brown et al., 1999; Poore et al., 2011; Poore et al., 2003), the last glacial termination events (Flower et al., 2004; Lynch-Stieglitz et al., 2011; Schmidt and Lynch-Stieglitz, 2011), the Last Glacial Maximum (Lynch-Stieglitz et al., 1999; Arellano-Torres and Machain-Castillo, 2017). Only a few studies have spanned older times (Brunner 1982; Kennett and Huddlestun 1972) or focused on paleohydrology and sea levels for the penultimate termination (TII) and the last interglacial (MIS-5e) (Simms 2021; Suh et al. 2020). In the region, there are studies focused on the transport intensity of the Florida Strait (Lynch-Stieglitz et al., 2011; Thirumalai et al., 2021), transport changes at the vicinities of the Yucatan Strait (Brunner 1984), and the northern influence of the Mississippi River discharge on the LC intensity (Nürnberg et al. 2008; Ziegler et al. 2008). However, few studies have investigated, at glacial-interglacial scales, the interaction between the LC and the water masses from the Caribbean Sea.

Using planktonic foraminifera as proxies for reconstructing surface ocean conditions is ideal because free-living planktonic protists depend on the hydrographic conditions to control their ecology and populations (Arnold and Parker, 1999; Be, 1977; Kucera and Schönfeld, 2007). Their sensitivity to a variety of environmental parameters, like sea surface temperature (SST), salinity, chlorophyll concentrations, or ocean currents, make them a powerful tool for recording changes at different timescales (Morey et al., 2005; Ravelo et al., 1990; Hemleben 1989; Watkins et al., 1996; Watkins and Mix, 1998). Furthermore, we can use either their assemblages' distributions to reconstruct paleoenvironmental changes (Arnold and Parker, 1999; Be, 1977; Kucera and Schönfeld, 2007) or the geochemical signals in their tests, to investigate the chemical and physical properties of ocean waters (Fischer & Wefer, 1999; Henderson 2002).

In this work, we aim to reconstruct mixed layer conditions and circulation changes in the GoM using assemblages of planktonic foraminifera and stable isotopes ($\delta^{18}O$ and $\delta^{13}C$) of the surface dweller *Globigerinoides ruber*. We will investigate environmental changes during the Late Pleistocene (MIS-9 to MIS-4) in the marine sediment core EN032-18PC collected below the influence of the LC. Faunal changes in planktonic foraminifera will help us to investigate the mixed layer, the intensity of the surface and subsurface waters flowing from the Caribbean to the gulf, and the LC extension.

1.1. Study Area

The GoM is a semi-closed subtropical basin in North America that connects to the Caribbean Sea and the North Atlantic Ocean (Figure 1). The basin is quasi-circular, 1,500 km in diameter, with an average of 1,615m (INEGI 2014). The carbonate platforms in the GoM, ranging from 100 to 300 km wide, are provinces from cyclical sedimentation derived from Pleistocene sea-level changes (Coleman et al. 1991). Dominated by low sedimentation rates, most sediments on the continental shelf are characterized by terrigenous muds and sands with variable amounts of organic remains (Davis 2017; Galloway 2008, Salvador and Salvador 1991). The deep-sea sediments show a combination of terrigenous and biogenic components of planktonic origin, like oozes of foraminifera and coccolithophores (Davis 2017).

The continental climate in the vicinities of the gulf varies from tropical to subtropical, seasonally controlled by the influence of the latitudinal migration of the Intertropical Convergence Zone (ITCZ), the passage of polar fronts, and the formation of tropical storms (Lamb 1974; Saha 2010). During the dry and cold season (December – April), the ITCZ is at its southernmost position, and the cold fronts favor temperature drops. During the rainy and hot season (August – October), the ITCZ is at its northernmost position, and the trade winds favor the northern and western displacement of tropical storms (Herzka-Llona et al. 2020; Lamb 1974). In the ocean, the general surface circulation is part of the AMOC (Schmitz and McCartney 1993). The western Atlantic surface current system moves north when the Guinea Current enters the Caribbean Sea through the Antilles arc (Gordon 1967). The Caribbean Current is the precursor of the Yucatan Current (Badan et al. 2005), which rotates clockwise to become the LC, then the Florida Current (Figure 1a), and finally the Gulf Current in the western North Atlantic (Molinari and Morrison 1988; Candela et al. 2019).

Over most of the year, the GoM develops little differences in surface salinity and temperature between the west and the east, as well as weak vertical convection and strong vertical stratification (Badan et al. 2005). The intense water transport through the Yucatan Channel (Athié et al. 2011) characterizes the LC system into two forms of intrusion: retreated and extended. The first occurs during autumn and winter, reaching up to 24º N. The second represents a more extensive intrusion during spring and summer, reaching up to 28º N (Alvera-Azcarate et al., 2009; Delgado et al., 2019) (Figures 1b and 1c). The extended intrusion into the GoM is influenced by the advection of the Mississippi River plume into the Florida Strait (Alvera-Azcarate et al. 2009; Hu et al. 2005). In the center of the gulf, the LC generates anticyclonic gyres of different sizes and frequencies (on average ~ 3 to 17 months) (Elliott 1982; Sturges and Leben 2000), crucial to mixing the surface (Alvera-Azcarate et al. 2009; Portela et al. 2018).

Nowadays, three surface and subsurface water masses domain the gulf (Portela et al. 2018): the Caribbean Surface Water (CSW; depth interval 50-150 m), the Subtropical Underwater (SUW; depth interval 150-230 m), and the Gulf Common Water (GCW; depth interval 50-100 m). The CSW origins at the surface layer of the Caribbean Sea; it is more oligotrophic and saltier than the rest of the waters in the Caribbean. The CSW enters the gulf during the maximum extension of the LC, being minimum during winter (Delgado et al. 2019; Morrison and Nowlin 1982; Muller-Karger et al. 2015). The SUW lies below the CSW and restricts the LC influence region. It forms at the center of the subtropical gyre in the North Atlantic, then sinks to the upper pycnocline, and in the Yucatan Channel shows a salinity maximum (> 36.92) (Cervantes-Díaz et al. 2022; Qu et al., 2013; Shcherbina et al. 2015). The GCW forms on the gulf western shelf by the collision of anticyclonic gyres (Elliott 1982; Portela et al. 2018). It originates from the saline water exchange in the anticyclonic gyres that derive west during the autumn-winter, linked to the deepening of the mixing layer in the north (Herzka-Llona et al. 2020). Although, the GCW could also be formed from the mixture of CSW with Tropical Atlantic Central Water (TACW; depth interval 300-700 m) in the western gulf (Cervantes-Díaz et al. 2022).

The nutrient distribution in the surface waters shows seasonal variability, mainly controlled by the entrance of CSW oligotrophic waters and the convective mixture of thermocline waters by northern winds. In spring and summer, there is little nutrient availability on the surface. The mixed layer and the thermocline are shallow (30 m) due to higher insolation and weaker SE winds, unable to develop vertical mixing but causing strong stratification (Herzka-Llona et al. 2020; Muller-Karger et al. 2015). During spring-summer, the LC is at its maximum extent. The advection of oligotrophic and warm waters from the CSW reduces the gulf's chlorophyll concentration, biological productivity, and apparent oxygen utilization (AOU) (Biggs and Ressler 2001; Delgado et al. 2019; Muller-Karger et al. 2015). In contrast, nutrient concentrations increase, as well as primary productivity and AOU during the autumn and winter (Pasqueron de Fommervault et al. 2017). The mixed layer and thermocline deepen (85 m) due to convective mixing and wind forcing (Muller-Karger et al. 2015). During autumn-winter, the LC is at its minimum extent because less CSW inputs the gulf, but with more intense northern winds, there is a mixing of surface waters and more GCW (Cervantes-Díaz et al. 2022; Herzka-Llona et al. 2020).

2. Materials and Methods

Marine sediment core

The core EN-032-18PC (from now on EN32-18PC) was collected with a piston corer from the eastern GoM (24º33.5' N, 86º35.4' W; water depth of 2030 m) during an oceanographic campaign onboard the R/V Endeavor (Figure 1a). The core length equals 725 cm with 7 cm in diameter. A total of 137 samples of 5 cm^3 were subsampled every 5 cm by request to the Marine Geological Samples Laboratory at the University of Rhode Island.

We took 2 to 3 g of each sediment sample to separate the sand size fraction (> 62 μm) from the mud (< 62 μm). To isolate the planktonic foraminifera tests from the sediment, we washed the sediments with running water on a MONT INOX brand sieve of 62 μm mesh size. The samples were dried at 25ºC and then sieved through a 150 μm mesh size to collect the adult tests needed to perform micropaleontological and stable isotope analyses ($\delta^{18}O$ and $\delta^{13}C$).

Micropaleontological analyses

To study the planktonic foraminifera distribution along the core, we divided each washed sample with an Otto micro-splitter to obtain between 300 to 500 tests from the size fraction > 150 μm. We picked the specimens under a stereomicroscope Velab VE-S5 using a fine brush (#000). The species were identified using specialized taxonomic references (Brummer and Kucera 2022; Kennett and

Srinivasan 1983). All tests show excellent morphological preservation, with no apparent evidence of dissolution or damage. In each sample, we calculated and plotted the species' absolute abundance per gram (ind/g), the percentage relative abundances (%), and the Shannon-Wiener equitability index to verify the preservation of fragile and resistant species.

Temporal framework

2.3.1 Biostratigraphy and $\delta^{18}O$ isotope curve

The core EN32-18PC chronology was based on defined tie points at selected depths based on the age relationship between the planktonic foraminifera faunal zones and the oxygen isotope curve ($\delta^{18}O$). We used the repeating patterns of absence/presence between the *Globorotalia menardii* group and *Globoconella inflata* to establish that the core lies within the Late Pleistocene (Kennet and Huddlestun, 1972).

Biostratigraphy in the GoM is based on defined faunal zones (Ericson and Wollin 1968, Kennett and Srinivasan 1983, Martin et al. 1993), which have been associated with the $\delta^{18}O$ curves in Caribbean and Gulf cores over the last 400 ka (Martin et al. 1993; Martinez, Mora and Barrows 2007; Antonarakou et al., 2019). The gulf's planktonic foraminifera biostratigraphy relies on the relative distribution of the *Globorotalia menardii* complex and *Globoconella inflata* as two of the most notable species. The *G. menardii* complex is formed by *Globorotalia menardii, Globorotalia tumida* and *Globorotalia flexuosa*, as they are considered part of the same ecological group (Kennett and Srinivasan 1983) (from now on, the *G. menardii* group).

To identify the planktonic foraminifera faunal zones in core EN32-18PC, we follow the biostratigraphic descriptions by Ericson and Wollin (1968), Kennett and Srinivasan (1983), Martin et al. (1993), Martinez et al., (2007). Zone V is divided into subzones V2 and V1, matching the interval between the Marine Isotope Stage (MIS) 9 and the first half of MIS-6. Subzones V2 and V1 are characterized by the sudden percentage increase (decrease) of the *G. menardii* group (*G. inflata*). Next, the limit V-W agrees with the abrupt decline of the *G. menardii* group and *Pulleniatina obliquiloculata,* just after a gradual increase of *G. inflata*. Zone W coincides with the second half of MIS-6, distinguished by the general scarcity of warm-water species like *G. menardii group* and *P. obliquiloculata*, cold-water species like *G. inflata*, and the high percentages of *G. truncatulinoides*. In general, *G. inflata* vary inversely with *Neogloboquadrina dutertrei*. The limit W-X agrees to the following faunal variations: the abrupt decrease of *G. inflata*, a gradual decrease of *G. truncatulinoides*, the increase of *N. dutertrei*, *Globorotalia crassaformis* and *Globigerinoides conglobatus*, and the abrupt increase of *G. menardii group* and *P. obliquiloculata*. Zone X coincides with MIS-substages 5e to 5c, characterized by the steep rise and high abundance of the *G. menardii group*; the cold-water species were much less significant, but *G. truncatulinoides, G. dutertrei* and *G. crassaformis* are prominent. The severe reduction of the G. menardii group establishes the X-Y limit. Zone Y includes eight subzones spanning from MIS-5b to MIS-2, determined by maxima abundances of *G. inflata*, along with minima values or the complete absence of the *G. menardii* group and *P. obliquiloculata*. The complete disappearance of *G. inflata,* marking the beginning of Zone Z (MIS-1), was never found in core EN32-18PC.

Stable isotope analysis

In core EN32-18PC, we analyzed 137 samples by duplicate or triplicate to obtain an $\delta^{18}O$ and $\delta^{13}C$ isotope curves. For each sample, we picked twenty specimens of *Globigerinoides ruber* white variety (*G. ruber*$_w$) from the size fraction between 250 and 300 µm. We selected the species *G. ruber*$_w$ because it is the most representative and abundant in all samples (Bé, 1977). We followed the methodology established by Barker et al. (2003) to clean the tests. We analyzed the samples at the Laboratory of Stable Isotopes Analysis (LAIE), *Unidad Académica de Ciencias y Tecnología de Yucatán, UNAM*. The oxygen and carbon isotopes composition in the tests were determined via phosphorolysis using an Isotope Ratio Mass Spectrometer (IRMS) *Thermo Scientific Delta V*. The reproducibility of the NSB-18 and NSB-19 standards measurements was, for $\delta^{18}O$ and $\delta^{13}C$ better than 0.18‰ and 0.15‰, respectively, and relative to the PDB scale. The accuracy of 70 sample replicas was $\delta^{18}O \pm 0.21$‰ and $\delta^{13}C \pm 0.14$‰. Finally, to define the Marine Isotope Stages (MIS) in the $\delta^{18}O$ curve of the core EN32-18PC, we correlated to the $\delta^{18}O$ LR04$_{benthic}$ stack (Lisiecki and Raymo 2005). Also, we applied a visual comparison of glacial to interglacial transitions with existing $\delta^{18}O$ curves from the southwestern Caribbean Sea (ODP 999A; Martinez et al., 2007) and northern GoM (MD02-2575; Nürnberg et al., 2008).

Q-mode factor analysis (QFA)

The multivariate factor analysis was used to evaluate the data matrix in a simplified form by determining similar groups of entities, i.e., groups of planktonic foraminifera according to their ecological preferences (Adam, 1976; Klovan and Imbrie, 1971; Klovan, 1975). The factor analysis assumes that the species' response along the environmental gradients shows a normal distribution, with one peak and two tails (Morey et al., 2005). We chose the appropriate number of factors following (1) the scree plot, i.e., the number of factors vs. eigenvalues, (2) eigenvalues > 1, and (3) the variance explained (Song and Belin, 2008). The factor loading is the correlation between each level (i.e., core depth (cm)) and each factor. A factor loading of more than 0.30 usually indicates a moderate correlation between the item and the factor (Tavakol and Wetzel, 2020). In this study, to avoid values overlapping, when the factor loadings are positive (>0.6) or negative (<0.6), imply that a group of levels has a more significant influence on the factor. The factor scores are the assemblage calculated as the weighted sum of each species. In the core EN32-18PC, to create the data matrix, we used 16 species with relative frequencies (%) > 1% of the total population and 137 depth levels, later transformed into a time scale (ka). We performed a QFA (Klovan and Imbrie, 1971) to determine the planktonic foraminifera assemblages using the software *PAST* (Hammer et al., 2001) (CABFAC algorithm, varimax rotation). Before the QFA, we applied a *row normalization length* transform to the faunal database to prevent the highly abundant species from dominating the analysis. The *Q-mode* was chosen over the *R-mode* as we aim to obtain the relationship between the foraminifera species, not between the levels' properties (Imbrie and Kipp, 1971).

2. Results

Marine sediments

The sediments in core EN32-18PC are primarily biogenic along the 660 cm length, i.e., a calcareous ooze with abundant planktonic foraminifera and coccolithophorids beside the remains of various organisms. After washing the sample, we removed the mud fraction (<62 µm) from the sand size fraction (>62 µm). On average, the mud represents 68% of the bulk sample and the sand size 33%, which contains all the foraminifera tests (Arellano-Torres et al., 2023a).

Planktonic foraminifera faunal distribution

The distribution of individuals per gram (ind/g) coincides with the distribution of the sand-size fraction. We found an average of 1756 ind/g, with a minimum of 179 ind/g at 466 cm depth and a maximum of 5876 ind/g at 115 cm depth. We recognized seven families (Candeinidae, Globigerinidae, Globigerinitidae, Globorotaliidae, Hastigerinidae, Pulleniatinidae and Sphaeroidinidae), including 16 genera and 31 species (Table 1). According to the equitability index Shannon – Wiener, all levels are considered diverse when values are > 0.5 (Figure 2b). There are 16 main species with relative abundances >1%. The species *G. ruber* is the most abundant in the entire sedimentary sequence (27.9% on average) (Figure 2c). The species *N. dutertrei, T. sacculifer* and *O. universa* contribute 30.2% on average (Figure 2d-2f). The 15 minor species with relative abundance <1% only represent 3.3% (Arellano-Torres et al., 2023a).

Seven of the ten most abundant species show shifting patterns varying from maximum to minimum abundance peaks (Figures 2d-2i, 2l). In contrast, three species show unpredictable distribution, i.e., *Globigerinella. siphonifera, Hastigerina. pelagica* and *Globigerinita. glutinata* (Figures 2p-2r). Two species show the most considerable amplitude changes: *G. menardii group* and *G. inflata*, representing 8.7% of the total fauna (Figures 2g-2h). The species *G. menardii group* has five relative maximums (50, 150, 280, 410 and 580 cm) with six minimums between them. The opposite pattern is followed by *P. obliquiloculata, G. conglobatus* and *G crassaformis* (Figures 2i-2k). The species *G. inflata* has relative maximums overlapping with the minimums of the *G. menardii group* (100, 220, 340 and 480 cm), except for the last 100 cm. Simultaneously, *G. bulloides* follow the *G. inflata* pattern but with more irregularity (Figure 2n).

Temporal framework

As shown in the Supp. Figure S1, the chronology of the core EN32-18PC is based on faunal zones relative to the percent changes of *G. ruber, G. menardii group, P. obliquiloculata, G. inflata, G. crassaformis* and *G. truncatulinoides*, their biostratigraphic boundaries, and their temporal relationship with the $\delta^{18}O$ curve. We identified the Late Pleistocene faunal zones (V, W, X, and Y) based on Ericson and Wollin (1968), Kennett and Huddlestun (1972), Martin et al. (1993), Martinez, Mora and Barrows (2007). Later, the identification of the Marine Isotope Stages (MIS) in the $\delta^{18}O$ record from core EN32-18PC was made by tuning to scale the global $\delta^{18}O$ $LR04_{benthic}$ stack (Lisiecki and Raymo, 2005), besides comparing to regional $\delta^{18}O$ curves from the Caribbean Sea (Martinez et al. 2007) and the GoM (Nürnberg et al. 2008). We assigned an age (ka) to eight tie points according to the $\delta^{18}O$ curve and the biostratigraphy (Arellano-Torres et al., 2023b). We applied linear interpolation between tie points to complete the core chronology. As a result, the core spans three interglacial (MIS-9, MIS-7, and MIS-5) and two glacial periods (MIS-8 and MIS-6).

Stable isotopes in G. ruber tests

Overall, the $\delta^{18}O_{G.ruber}$ values vary between -1.93‰ and 1.60‰, with an average value of 0.1‰ (Figures 3 and 4). The most negative values are between 310-280 ka (transition MIS-9 to MIS-8), 245-210 ka (MIS-7), and 170-160 ka (mid-MIS-6), with variable values between 130-100 ka (first half of MIS-5). Overall, we found a difference between ~2 to 4‰ during the glacial-interglacial transitions. Values of $\delta^{13}C_{G.ruber}$ range between 0.21‰ and 1.36‰, with average of 0.85‰. The carbon isotope record shows a tendency to increase their values from MIS-9 to MIS-4. However, over the end of MIS-9 to MIS-8 (320-270), we find average values ~0.7‰, but over MIS-6 to MIS-4 (178-70 ka), average values ~1.0‰ (Figure 4) (Arellano-Torres et al., 2023a).

Planktonic foraminiferal assemblages

Based on QFA, we found two main factors in the core EN32-18PC, which significantly explain 86.3% of the accumulated variance, and eigenvalues >10 (Table 3). The factors scores group two faunal assemblages in agreement with their analogous (values > 1) or antagonistic (values < -1) ecological preferences. Factor 1 explains 78.7% of the variance; the species with a positive factor score is *G. ruber,* but the negative is the *G. menardii* group. Factor 2 explains 7.7% of the variance; the species with a positive score are the *G. menardii* group*, T. sacculifer*, *O. universa*, and *N. dutertrei*. Considering only the factor loadings > 0.6, we observe the following (Figure 4). Factor 1 is widely distributed along the core, between the intervals 320 - 310 ka (MIS-9), 280 - 240 ka (MIS-8 to the beginning of MIS-7), 200 - 120 ka (end of MIS-7 to MIS-6) and variably between 100 - 70 ka (end of MIS-5). Factor 2 distributes between 310 - 280 ka (MIS-9 to mid-MIS-8), 240 - 215 ka (first half of MIS-7), 160 - 155 ka (mid-MIS-6), 125 - 100 ka and 85 - 80 ka (MIS-5).

3. **Discussion**

Chronostratigraphy

We selected eight anchor points to construct the core chronology through linear interpolation (Table 2) using the scale of the global δ^{18}O curve LR04$_{benthic}$ stack (Supp. Figure S1) (Arellano-Torres et al., 2023b). The age of the core ranges between 319 ka to 68 ka. Site sedimentation rates are low and vary between 2.02 and 3.37 cm/ka, coinciding with former studies on the Yucatan Channel and Slope showing similar values (Brunner, 1984; Dıaz-Asencio et al., 2020). The low sedimentation rates may be a product not only of the reduced sediment transport from the Yucatan Peninsula but also because the core locates below the LC’s average position (Brooks et al. 2020; Díaz-Asencio et al. 2020), which might prevent high sediment fluxes to the seafloor.

The δ^{18}O curve from core EN32-18PC shows high variability in intervals like MIS-5 if compared to regional δ^{18}O records (Figure 3). In some *G. ruber* tests, photographic evidence with Scanning Electron Microscopy showed pores reduction and clays adhered to the sutures, which suggests incipient diagenetic effects as clays were impossible to remove with the cleaning protocol (Arellano-Torres et al., 2023b). Nevertheless, we tuned five isotope stages using biostratigraphic and isotopic comparison techniques. We consider the chronology congruent, as we did not find evidence of hiatuses, the calculated sedimentation rates are as expected, and the faunal patterns do not evidence sequence alterations.

Isotope records of $\delta^{18}O$ and $\delta^{13}C$

Previous palaeoceanographic studies have used oxygen isotopes ($\delta^{18}O$) in planktonic foraminifera to help monitor variations in the vertical structure of the surface ocean based on its relationship to hydrographic conditions (Whitman & Berger, 1993; Spero et al., 2003; Steph et al., 2009, Waelbroeck et al., 2005). In general, calcite $\delta^{18}O$ composition could reflect the thermocline structure during different stratification regimes depending on their calcification depth and the water $\delta^{18}O$ signal during calcite formation. Various environmental and preservation factors can affect their $\delta^{18}O$ signal (Lisiecki and Raymo, 2005; Spero et al., 2003; Steph et al., 2009), but various studies reinforce it as a tool to reconstruct mixed layer features. For instance, Waelbroeck et al. (2005) report that recent fossils of planktonic foraminifera show values ~0.2–0.8 ‰ higher than living specimens, linking such discrepancy to the stratification of the upper water mass, which increases at low latitudes. Steph et al. (2009) suggest that planktonic foraminifera $\delta^{18}O$ values increase with the depth of calcification. Thus, low isotope values indicate calcite formation in the surface layer, whereas high values help track deeper dwellers.

In core EN32-18PC, we analyzed the $\delta^{18}O_{G.ruber}$ isotope signal, whose calcification depth is in the surface layer (0-40 m water depth) (Mulitza et al., 2004). As expected, we found higher $\delta^{18}O_{G.ruber}$ values during glacial periods (MIS-8 and MIS-6) and lower during interglacial periods (MIS-9, MIS-7, and MIS-5) due to changes in global sea level and global temperature (Lisiecki and Raymo, 2005). However, when comparing our core to neighboring $\delta^{18}O_{G.ruber}$ records (Figure 3), average shifts can be observed. Overall, an ~0.4‰ shift toward higher values is observed when core EN32-18PC ($\delta^{18}O_{average}$=0.1‰, Figure 3a) is compared to the northern GoM ($\delta^{18}O_{average}$=0.45‰, Figure 3b), and a ~1‰ shift higher relative to the Caribbean Sea ($\delta^{18}O_{average}$=0.85‰, Figure 3c). Such differences might relate to isotope fractionation along the currents' system or regional changes in the hydrological gradients (Mulitza et al., 2003; Waelbroeck et al., 2005). In this sense, during interglacial periods, the heat transport and the contributions of fluvial or less saline waters increase in tropical regions (Spero and Williams, 1990); hence, the surface and subsurface stratification strengthen (i.e., vertical density gradient). During glacial periods, the ocean heat transport and fluvial contribution reduce, and the stratification weakens favoring more homogeneous waters and depth penetration of the wind-driven currents. In core EN32-18PC, during interglacial periods, we found minor differences between the $\delta^{18}O_{G.ruber}$ records from the Caribbean and the GoM (i.e., $\Delta\delta^{18}O_{Car-GoM}$ = 0.5-0.75‰), but more prominent during glacial periods (i.e., $\Delta\delta^{18}O_{Car-GoM}$ = of 1.0-1.5‰). However, the limited amount of $\delta^{18}O$ records spanning MIS-9 to MIS-4 in the study area, prevents from resolving the processes behind the local to regional hydrological variations controlling the oxygen isotope curves (e.g., reduced tropical seasonality or increased fluvial or precipitation inputs) (Waelbroeck et al., 2005).

Using stable carbon isotopes in paleoceanography ($\delta^{13}C$) offers a broad perspective for reconstructions (Mackensen & Schmiedl, 2019; Waelbroeck et al., 2005). In foraminifera tests, the $\delta^{13}C$ signal is controlled by the fractionation of inorganic carbon from seawater ($\delta^{13}C_{DIC}$), and to vital effects (Spero et al. 2003). Planktonic foraminifera use $CaCO_3$ to form their shells and incorporate a proportion of heavy to light carbon isotopes controlled mainly through: (1) CO_2 gas fractionation in surface waters, (2) the concentration and fixation of carbonate ions during calcite precipitation, and (3) the metabolic reactions of degradation and remineralization in deeper waters (Mackensen & Schmiedl, 2019; Zhang et al., 1995). The $\delta^{13}C_{foraminifera}$ can be used as a tracer of water masses and potentially indicate changes in ocean circulation or ventilation patterns (Broecker & Peng, 1993; Mix et al., 1991). Therefore, surface and newly formed waters show higher $\delta^{13}C$ than deep and old water masses (Mix et al., 1991).

In the core EN32-18PC, we observe a trend toward higher $\delta^{13}C_{G.ruber}$ values from MIS-9 to MIS-4 (i.e., from 0.5‰ to 1.2‰) (Figure 4). Also, we observe a 0.5‰ change during the transitions between MIS-8/MIS-7 and MIS-7/MIS-6 until reaching a maximum value of 1.2‰ during MIS 6-MIS 5. According to previous studies, lower $\delta^{13}C_{planktonic}$ values in the surface ocean suggest a supply of lighter isotopes (^{12}C) brought by the convection of deeper waters (Ninnemann & Charles, 2002). In the GoM, during MIS-9 to MIS-7, an overall transport of subsurface waters to shallow depths might help explain the low $\delta^{13}C_{planktonic}$. However, during MIS-6 to MIS-4, higher $\delta^{13}C_{planktonic}$ values suggest the opposite.

In the Atlantic, previous studies indicate that a long-term increase in $\delta^{13}C$ values might partially reflect the activation of the North Atlantic Thermohaline Circulation, with the subsequent increase in the contribution of North Atlantic Deep Water (NADW) (Spero and Lea, 2002; Spero et al., 2003). The NADW is known as a low-nutrient, high $\delta^{13}C_{DIC}$ water mass, opposite to the high-nutrient, low $\delta^{13}C_{DIC}$ waters from the Southern Ocean (Ninnemann & Charles, 2002; Spero & Lea, 2002). Over the late Pleistocene, a trend to higher $\delta^{13}C_{planktonic}$ values has been identified in records from the tropical Atlantic, the Southern Ocean, the tropical and North Pacific (Banakaar, 2005; Curry & Crowley, 1987; Hall et al., 2001; Shackleton & Pisias, 1985; Yamane, 2003). Therefore, in the core EN32-18PC, the $\delta^{13}C_{G.ruber}$ record coincides with a gradual northward transport of low-nutrient waters, depleted in ^{12}C. At the same time, Mulitza et al. (1999) suggest that in the eastern Caribbean Sea, the changes in the $\delta^{13}C$ signal are transferred from a southern source of intermediate waters. However, if the transferring of light carbon from high to low latitudes is limited, or the source carbon at intermediate waters varies with time, such a $\delta^{13}C$ signal can be recorded in the GoM. Nevertheless, local effects must be additionally explored. For instance, in a modern study from the eastern Caribbean Sea, Jentzen et al. (2019) suggest that during times of higher turbidity driven by rainfall, a reduction in the light attenuation could reduce photosynthesis by symbionts and low measurements on $\delta^{13}C_{G.ruber}$ tests. The western tropical Atlantic is mostly oligotrophic, so its $\delta^{13}C$ signal inherently shows lower values than other regions, i.e., the South Atlantic. Overall, it seems plausible that during the interval between MIS-9 to MIS-7, the low $\delta^{13}C_{G.ruber}$ signal could be explained by changes in the surface layer due to a limited nutrient supply or changes in water thermodynamics (Mulitza et al., 1999). Unfortunately, the unavailability of $\delta^{13}C_{planktonic}$ records during this time frame prevents comparison to neighboring records, thus knowing its $\delta^{13}C$ distribution pattern.

Paleoecological trends in surface waters

Based on the planktonic foraminifera assemblages and their ecological preferences, we reconstruct two paleoceanographic scenarios for the GoM during MIS-9 to MIS-5 (Figures 4 and 5). The first is linked to subtropical waters with a deeper thermocline and a retracted Loop Current (LC), whereas the second is linked to tropical waters with a shallow thermocline and an extended LC.

4.2.1. Factor 1. The subtropical, deep thermocline, and retracted-LC assemblage

The highest factor loadings (>0.6) distribute over most of the sequence (Figure 4), covering the glacial and late interglacial periods (i.e., MIS-9, MIS-8 to the beginning of MIS-7, end of MIS-7 to MIS-6, and the future of MIS-5). The faunal assemblage indicates positive values for *G. ruber*, although negative for the *G. menardii* group.

The species *G. ruber* is a spinose, symbiont-bearing foraminifera, dweller of shallow depths (0 to 50 m) in subtropical waters, with a preference for oligotrophic regimes (Bé 1982; Bé and Hamlin 1967; Hemleben et al., 1989; Morey et al., 2005; Tolderlund and Bé 1971; Jentzen et al., 2019). The species has been observed in SST between 16.4 to 29.6ºC (Žarić et al., 2006), with a range of salinities from 35.9 to 36.7 ‰ (Jentzen et al., 2018, 2019; Schmuker and Schiebel, 2002; Tolderlund and Bé 1971). Its optimum SST is around 26.5ºC in waters from the Caribbean to the Florida Strait (Jentzen et al., 2018; Jones 1971; 1968; Schmuker and Schiebel, 2002), where low-nutrients and chlorophyll concentrations make food scarce (Bé 1982, Jentzen et al., 2018). In subtropical environments, it is a primary component of foraminiferal assemblages (Siccha and Kucera, 2017; Morey et al., 2005). In the GoM and the Caribbean Sea, the ranges of temperature, salinity, and nutrient availability coincide with its optimal environmental preferences, being the most biologically prosperous and abundant planktonic foraminifera in sediments and sediment traps (Arellano-Torres and Machain-Castillo, 2017; Brunner, 1979, Jentzen et al., 2018; Poore et al., 2003; Poore et al., 2013). In the core EN32-18PC, the distribution frequency of *G. ruber* suggests that over the MIS-9 to MIS-4, the changes in the environmental parameters have not severely affected the *G. ruber* distribution (Figure 4), remaining optimal, at least during most glacial and late interglacial periods. On the other hand, the assemblage's opposite species, the *G. menardii* group, dwells in tropical and subtropical environments, tolerant to high vertical temperature gradients with a preference for stratification. It is a pycnocline dweller at depths between 25-50 m and up to 100 m (Jentzen et al., 2018; Jones, 1971; 1968; Hilbrecht, 1996; Tolderlund and Bé, 1971; Schmuker and Schiebel, 2002).

To favor the presence of *G. ruber* and reduce *G. menardii*, the environmental conditions must experience an increase in the convective or wind mixing, deepening the thermocline and reducing stratification, unfavorable for the *G. menardii* group. Such conditions allow *G. ruber* to be ecologically dominant over less competitive species, although the community diversity was not necessarily affected (i.e., equitability; Figure 2). Factor 1 represents a subtropical scenario with a deep thermocline and weak surface layer stratification. Today in the GoM, the former conditions occur when the LC retracts (Delgado et al. 2019; Morrison and Nowlin 1982; Muller-Karger et al. 2015). When the waters from the Caribbean Sea rescind to the Yucatan Strait, the LC directly flows to the Florida Strait; thus, cooler SST can be observed across the GoM (Alvera-Azcarate et al., 2009, Portela et al., 2018).

4.2.2. Factor 2. The tropical, shallow thermocline and extended LC assemblage

Its main distribution occurred during the transition MIS-9/MIS-8 (310-295 ka and 290-280 ka), a brief episode during the MIS-6 (163 to 159 ka), early MIS-7 (236 to 220 ka) and MIS-5 (127 to 103 and 92 to 86 ka) (Figure 4). The main species of the assemblage are the *G. menardii* group, *T. sacculifer, O. universa* and *N. dutertrei*.

The *G. menardii* group is formed of non-spinose species that prefer transitional environments between tropical and subtropical water masses (Hilbrecht 1996, Tolderlund and Bé 1971). It prefers depths associated with the thermocline and pycnocline, although, in its late ontogeny, it migrates to greater depths (Bé and Hamlin 1967; Tolderlund and Bé 1971). In a transect from the Caribbean to the Florida Strait, its depth distribution occurs shallower than 40 m in the eastern Caribbean but below the SUW at the Florida Strait (Jentzen et al., 2018). The species are abundant in the Gulf Stream and the western North Atlantic (Morey et al., 2005; Siccha and Kucera, 2017). The *G. menardii* group occurs in temperatures between 13.3° to 30.5°C and salinities from 36.1‰ to

36.5‰ (Žarić et al. 2006). In the Caribbean, GoM and the Florida Strait, the optimum average temperatures are ~24.4°C with salinities of ~36.1‰ (Jones 1971, 1968, Schmuker and Schiebel, 2002, Jentzen et al. 2018).

The species *T. sacculifer* is dominant in tropical to subtropical regions, typical of oligotrophic to mesotrophic environments, as it can host algal symbionts (Žarić et al. 2006). It prefers the upper 50 m of the euphotic zone and generally outcompetes *G. ruber* if nutrient concentrations are high (Bijma et al.,1990; Bijma and Hemleben, 1994). Its temperature and salinity intervals range between 9.7-31ºC and 35.9-36.4‰, but in tropical waters, *T. sacculifer* reaches its optimum abundance at temperatures and salinities ≥22.1ºC and ≥36.43‰, respectively (Bé 1982; Tolderlund and Bé 1971). The species T. *sacculifer* prefers water masses with low seasonality and vertical temperature gradients (i.e., moderate summers or winters) (Bé 1977; Bé 1982; Hilbrecht 1996; Tolderlund and Bé 1971). In sediments beneath the LC, *T. sacculifer* is abundant (Brunner 1979; Poore et al. 2003; Schiebel et al. 2018; 2019) because it prefers the low-salinity tropical waters flowing from the Caribbean Sea. The former conditions have made *T. sacculifer* a thriving species in the western North Atlantic and the GoM, linked to the entry of Caribbean surface water, and an indicator of the LC (Brunner, 1979; Jentzen et al., 2018; Poore et al., 2003; Schiebel et al., 2018; Jentzen et al., 2019).

The species *O. universa* is the most ubiquitous planktonic foraminifera documented, more adapted to subtropical waters (Be, Harrison and Lott 1973; Hemleben et al. 1989). It is a spinous species with a carnivorous diet that inhabits subsurface depths within the euphotic zone above the thermocline (65 – 85 m) (Bé 1982, Jentzen et al., 2018). Its optimal abundance range temperatures between18.2ºC to 29.5ºC and salinities between 35.75 to 36.63‰ (Jentzen et al., 2018; Tolderlund and Bé, 1971). The species is annually present with an abundance of <10% (Poore et al., 2013; Siccha and Kucera, 2017).

Neogloboquadrina dutertrei is a non-spinous herbivorous species of tropical to subtropical waters restricted to warm waters above 200 m depth (Bé 1960; Bé 1982; Hilbrecht 1996). It commonly dwells in the deep chlorophyll maximum (30-60 m depth) of productive tropical environments that provide rich food sources, including ocean margins. From the Caribbean to the Florida Strait, its preferred water temperatures are between 18º and 27ºC and salinities between 35.75 to 36.63 ‰ (Bé 1982; Jentzen et al., 2018; Jones, 1971; 1968; Tolderlund and Bé 1971).

The assemblage species *G. menardii*, *T. sacculifer, O. universa* and *N. dutertrei* suggest an environment of warm tropical waters with intense currents (Tolderlund and Bé 1971; Siccha and Kucera, 2017). Along the western Atlantic margin, *Globorotalia menardii* prefers warm and stratified environments, *O. universa* and *N. dutertrei* prefer strong currents, and *T. sacculifer*, the tropical low-salinity waters from the Caribbean (Brunner, 1979; Jentzen et al., 2018; Poore et al., 2003; Poore et al., 2004; Poore et al., 2013,). The assemblage represents interglacial conditions where the warmest and a more stratified surface prevailed. Today in the GoM, the former conditions are present when large fluxes of Caribbean waters intensify the LC (Delgado et al. 2019; Morrison and Nowlin 1982; Muller-Karger et al. 2015), and the nutrient content increase as mesoscale gyres develop to form the LC (Pasqueron de Fommervault et al. 2017). Therefore, the assemblage suggests stratified tropical waters with shallow thermocline, an intense LC derived from the persistent transport of Caribbean waters (Nürnberg et al., 2008; Schmidt et al., 2006; Ziegler et al., 2008).

4.2.3. Long term trends in G. truncatulinoides, G. crassaformis and T. sacculifer

In addition to the faunal assemblages, three species show trends that complement our understanding of the surface ocean and ecological successions in core EN32-18PC (Figures 5e-5g). The general trend of *G. truncatulinoides D* decreased from late MIS-9 to MIS-4 (Figure 5e). Today, the species is a non-symbiont deep dweller controlled by stratification and the thermocline depth structure, as it dwells between 600-800 m in the western North Atlantic (Lohmann and Schweitzer 1990). In the Caribbean, it shows greater abundance between 100-300 m and could be used as a tracer of the regional and temporal distribution of the SUW (Schmuker and Schiebel 2002). In the GoM, the decreasing abundance of *G. truncatulinoides* from MIS-9 to MIS-4 could suggest a reduction of advected subsurface waters from the Caribbean Sea, fed by warm waters from the tropical western Atlantic (Martinez et al. 2007). The second species with an increasing trend from MIS-9 to MIS-4 is *G. crassaformis* (Figure 5f), preferably present during interglacial and scarcer during glacial periods. Not much is known about the ecological preferences of *G. crassaformis* besides its calcification depth between 450 and 700 m (Cléroux et al. 2013, Steph et al. 2009). In the GoM, the TACW is the water mass between 300-700 m, likely to originate in the Angola Dome, South Atlantic (Portela et al. 2018). Its low oxygen concentration has partially defined the TACW <3 ml/L (Morrison and Nowlin 1982). Although the causes of today's oxygen depletion remain obscure, it could result from diverse coastal processes during its path through the Caribbean Sea (Morrison and Nowlin 1982, Portela et al. 2018). Therefore, *G. crassaformis* could benefit from the long-term changes in the physicochemical properties of a fluctuating TACW over time. For instance, along the Atlantic, Cléroux et al. (2013) document a positive match between the *G. crassaformis* calcification depth and the dissolved oxygen levels (relative oxygen minimum layer of 3.2 ml/l). The third species with an overall decreasing abundance is *T. sacculifer* (Figure 5g). As previously mentioned, it is recognized as a sedimentary indicator of the LC variations (Brunner, 1979; Poore et al., 2003; Poore et al., 2013; Richey et al., 2007), and the average position of the ITCZ (Richey et al., 2007). At least in the eastern gulf, at the glacial-interglacial scales, we confirm that trends of *G. truncatulinoides*, *G. crassaformis* and *T. sacculifer* could be used as tracers of changes in the Caribbean surface and subsurface water masses. Over MIS-9 to MIS-4, these tracers suggest a retracted LC form and an overall transition to more variable oceanographic conditions.

Paleoceanography of the GoM from MIS-9 to MIS-4

In a sediment core collected from the eastern GoM, we investigated isotope records of $\delta^{18}O$ and $\delta^{13}C$ and two planktonic foraminifera assemblages (Figure 5i-5h). Over the late MIS-9 to early MIS-4, we link their changes to variations in the mixed layer, water mass transport from the Caribbean Sea to the GoM and LC reorganizations (i.e., retracted or extended mode) (Figure 5).

In agreement with the planktonic foraminifera assemblages, (1) during the glacial and late interglacial periods, the assemblage indicates the predominance of a mixed surface layer with a deep thermocline. Additionally, the reduced sea level and relatively colder SST in the region (Figure 5b to 5d) could impose reduced inputs of Caribbean surface and subsurface water masses to the gulf (Figure 6a). A reduction in the gulf's heat transport might lead to more limited biological activity due to the reduced formation of mesoscale gyres. The overall flux of water masses should be slower and colder by constantly presenting a retracted LC, just as Nürnberg et al. (2008) previously suggested.

The second assemblage, (2) during the transition MIS-9/MIS-8 and the warmest interglacial substages (Figure 5i), suggests that the surface ocean was significantly more dynamic with a community experiencing frequent ecological successions. Possibly oligotrophic and stratified waters, with a prominent input of Caribbean surface waters, lead to a most extended LC (Figure 6b). Furthermore, when comparing the highest F2 loadings (<0.6), they coincide with the highest global sea levels (Waelbroeck et al. 2002), highest SST reconstructions from the Caribbean and the northern GoM (Nürnberg et al., 2008; Schmidt et al., 2006), and maxima summer insolation values (30ºN) (Laskar et al. 2004) (Figures 5b, 5c, 5d, 5k).

The interval between MIS-9/MIS-8 could imply significant variability of surface water masses between the coldest and warmest stages, greater nutrient availability in the euphotic zone, and sufficient Caribbean waters to feed the LC. According to previous reconstructions, the end of interglacial MIS-9 was warm but variable, and the glacial MIS-8 has been defined as a weak glacial (Hughes, Gibbard and Ehlers 2020); thus, paleoceanographic conditions had to be less extreme than subsequent MIS. On average, the glacial MIS-8 experienced higher sea levels and surface temperatures than the subsequent glacial MIS-6 and MIS-2 (Hughes et al., 2020). In the northern GoM, Nürnberg et al. (2008) reported cooler and less saline conditions in the mixing layer during glacial times, mainly influenced by the intense discharge of the Mississippi River and a decrease in warm water transport through the LC. Lastly, during substage MIS-5e (about 123 ka) (Müller 2008), warmer and wetter conditions prevailed over the rest of MIS-5. In the northern hemisphere, relative to today, MIS-5e was ~4°C warmer in high latitudes, ~ 1°C warmer in low latitudes, and the global sea level was 11 m higher (Shackleton et al. 2003, Waelbroeck et al. 2002).

The role of the GoM in the North Atlantic

During the Late Pleistocene and Holocene, modifications in the LC intensity and SST reconstructions could be a consequence of the overall migration in the ITCZ (Arellano-Torres and Machain-Castillo 2017; Martinez et al. 2007; Nürnberg et al. 2008; Schmidt et al. 2006). The ITCZ position is controlled by changes in the atmospheric pressure gradients and insolation (Saha 2010). At centennial, millennial and orbital scales, several studies have linked a cooling in the North Atlantic to the southward migration of the ITCZ (Black et al., 1999; Broccoli et al., 2006; Schmidt et al., 2004; Peterson and Haug, 2006).

The tropical waters are undoubtedly linked to the climate in the North Atlantic. For instance, Schmidt et al. (2004) analyzed two sediment cores from the Caribbean Sea through Mg/Ca-paleotemperatures and $\delta^{18}O_{planktonic}$ records to reconstruct salinity in the tropical Atlantic. They found higher salinity prevailed during cold MIS-6, 4 and 2 but lower salinities during warm MIS-5 and 3, limiting North Atlantic deep water formation. Similarly, at the beginning of the Bølling/Allerød warm interval (14.5 ka ago), the surface Caribbean salinity decreased sharply, suggesting that the advection of salty tropical waters into the North Atlantic amplified the thermohaline circulation and contributed to high latitudes warming (Schmidt et al., 2004). The former coincides with $\delta^{13}C_{G.ruber}$ record interpretation, evidencing that transfer of low nutrient but high in $\delta^{13}C$ waters from the GoM migrated north to the NADW area of formation.

Along with the Loop and Florida currents, the Gulf Stream certainly influences the ocean circulation and climate of the North Atlantic at various scales. For instance, during glacial periods for the last 150 ka, the study by Crowley (1981) in the Azores region found higher than expected SST linked to the intensified Gulf Stream based on planktonic foraminifera assemblages. Studies at geologic and millennial scales coincide with the former. Kaneps (1979) found that during the Pliocene, the intervals when Gulf Stream intensified were consistent with periods of glaciation.

Lynch-Stieglitz et al. (1999) suggest that during the Last Glacial Maximum (LGM, 26.5-20 ka ago), the AMOC weakened. Their records of $\delta^{18}O_{benthic}$ indicate a reduction of the density gradients as much as two-thirds of the current value, indicating that the Gulf Stream was significantly weaker than today. On shorter scales, Lynch-Stieglitz et al. (2014) found analog results during the Younger Dryas cold event (YD, 12.9-11.7 ka ago). Also, in the Florida Straits, they found a consistent relationship between lower temperatures, density reduction and lower geostrophic transport by the Gulf Stream. In turn, a study based on nutrient content indicators such as $\delta^{13}C$ and cadmium (Boyle and Keigwin,1987) support that during the YD, the North Atlantic deep waters were enriched in nutrients, then slower ventilation and the reduced transport of waters through the Gulf Stream, might cause a reduced AMOC.

In this way, the reconstructed environmental variability over MIS-9 to MIS-4 highlights the importance of studying the GoM at different timescales, helping us understand its influence in the North Atlantic and the AMOC.

4. Conclusions

In the marine sediment core EN32-18PC, we studied changes in the surface waters and ocean circulation from late MIS-9 to early MIS-4 based on planktonic foraminifera assemblages and stable isotopes. At the glacial to interglacial scale, the primary shifts occurred in the mixed layer, the transport of surface and subsurface water masses from the Caribbean, and two forms of variation of the LC.

During glacial and late interglacial periods, the evidence denotes a subtropical environment with reduced stratification, vertical gradients, and low of surface and subsurface Caribbean waters inputs, leading to a retracted LC. During the warmer interglacial substages, a tropical environment prevailed with deep and stratified thermocline, prominent inputs of Caribbean surface waters to feed and extended LC, leading to more variable oceanographic conditions.

In agreement with previous regional studies, variability over MIS-9 to MIS-4 connect changes in the GoM to the AMOC. Studying older glacial to interglacial stages during the late Pleistocene helps us understand how atmospheric pressure gradients, the position of the ITCZ, and changes in solar insolation drive different forms of ocean variability.

Acknowledgments

We are grateful to Rebecca Robinson, curator at the Marine Geological Samples Laboratory, University of Rhode Island (URI), for providing us with the marine sediment samples of the core

EN-032-18PC. This study was financed through the project grant UNAM-PAPIIT no. IN118220 to EAT. We want to thank Korynthia Lopez-Aguiar, for her laboratory and technical assistance. We appreciate the comments by two anonymous reviewers and the editor, to improve our manuscript.

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Data Availability Statement

The absolute and relative abundances of the planktonic foraminifera used for Q-mode factor analysis, and the stable isotope data (δ^{18}O-PDB and δ^{13}C-PDB) (‰) of *Globigerinoides ruber* (white), used to reconstruct the water masses and surface circulation in this study, are available at Zenodo (Arellano-Torres et al., 2023a) with license, Creative Commons Attribution 4.0 International (CC BY 4.0).

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Figure Captions

Figure 1. Map of the study area in the Gulf of Mexico. a. Bathymetry (depth), geographical location of the studied core EN-032-18PC (white circle) in the eastern gulf, and prominent patterns of seawater surface circulation (red arrows), indicating the gulf's intrusion of the LP: expanded (solid red line) and contracted (broken red line). b. World Ocean Atlas (WOA18) satellite data of Sea Surface Temperature (SST) during summer. c. WOA18 satellite data of SST during winter. Notice that each SST image shows a different scale (ºC). Black diamond shows the location of other cores in the gulf: ODP-625B (Martin et al., 1993; Whitacker, 2008), MD02-2575 (Ziegler et al., 2008; Nürnberg et al., 2008).

Figure 2. Taxa abundance of planktonic foraminifera along the core depth (cm) EN32-18PC. a. Relative (%) proportion of the sediment fractions > 62um (sand size) and <62um (mud size). b. Equitability index Shannon-Wienner of diversity. Percentage (%) distribution of the most abundant planktonic foraminifera: c. *G. ruber*, d. *N. dutertrei*, e. *T. sacculifer*; f. *O. universa*, g. *G. menardii* group, h. *G. inflata*, i. *P. obliquiloculata*, j. *T. quadrilobatus*, k. *G. conglobatus*; l. *G. crassaformis,* m. *G. truncatulinoides* dextral and n. sinistral, o. *G. bulloides,* p. *G. siphonifera;* q. *H. pelagica;* r. *G. glutinata*. The blue dotted lines show the mean % value.

Figure 3. Identified Marine Isotope Stages (MIS) in the $\delta^{18}O_{G.\ ruber}$ record from a. Core EN32-18PC (this study), where the solid black line is the smooth loess regression between data points (small grey dots); b. Core MD02-2575 (Ziegler et al., 2008); c. Core ODP 999A (Martinez et al., 2007); and $LR04_{benthic}$ stack (Lisiecki and Raymo, 2005). The blue dotted line shows the std. dev. (σ) and the short dash line, the mean value.

Figure 4. Stable isotope records ($\delta^{18}O$ and $\delta^{13}C$) and percentage (%) distribution of the five main planktonic foraminifera (see Table 3. Factor scores) and Factor Loadings (F1, F2, F3) along the core. Grey bands correspond to the Marine Isotope Stages based on the limits of the LR04 stack (Lisiecky and Raymo, 2005).

Figure 5. Distribution of the factor loadings in the core EN32-18PC and comparison with other records of global or regional relevance. a. $LR04_{benthic}$ stack (Lisiecki and Raymo, 2005). b. Relative Sea Level (RSL) (Waelbroeck, et al., 2002). c, d. SST reconstruction in the core MD02-2575 (Nürnberg et al., 2008) and ODP-999A (Schmidt et al., 2006); e. to g. Planktonic foraminifera with a linear trend along the core (m); h, i. F1 and F2 factor loadings >0.6. j. $\delta^{13}C_{G.ruber}$ (‰, this study). k. summer insolation 30°N and l. Obliquity calculations (Lanskar, 2004).

Figure 6. Idealized reconstructed paleoceanographic scenarios based on the two planktonic foraminifera assemblages. The solid black line represents the general circulation of the Loop Current, and the orange arrow represents the Caribbean surface water mass flux.