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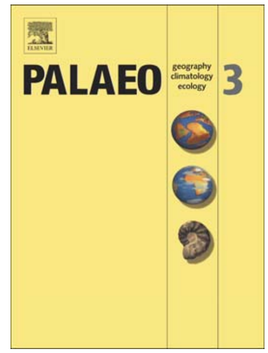

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Insight into the Latest Messinian (5.7-5.2 Ma) paleoclimatic events from two deep-sea Atlantic Ocean ODP Sites.

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Abstract: The results of a multi-proxy study, including quantitative planktonic foraminifera faunal analysis, geochemistry of foraminifera tests, and lithogenic counts (IRD) are presented for two open marine sites. The sites are located in the eastern South Atlantic (ODP Leg 177 Site 1088) and the western tropical North Atlantic (ODP Leg 154 Site 925). Both sedimentary records span the interval 5.7-5.2 Ma (i.e. late Miocene to early Pliocene), which encompasses the time of deposition of the upper evaporites (UE) in the Mediterranean basin. The observations confirm a major oceanographic and climatologic event occurred during the Messinian at the transition between the glacial TG12 and the prominent TG11 warm interglacial at 5.5 Ma. However, some oceanographic changes also occurred at the Miocene-Pliocene (M-P) transition in the northern tropical Atlantic and in the Southern Ocean with the first input of IRD at ODP Site 1088. In contrast to the termination across the lower evaporites (LE) at 5.5 Ma, the M-P transition may not have involved a large change in ice volume. The potential causes behind the data across the major climatic transient are examined in the light of published information, including evidence from polar areas with focus on the climatic impact of fluctuating meridional oceanic circulation (MOC). A thermal seesaw mechanism in pre-Quaternary times is hypothesised as part of the large late Messinian deglaciation across the TG12-TG11 transition. An implication of the major Southern Atlantic warming before

5.5Ma is that an abrupt event freshening the surface of the North Atlantic might be present in the sub-polar Northern Hemisphere, but this has yet to be verified. This deglaciation may have been reinforced by a freshening of the North Atlantic as a result of discontinuous connection of the Mediterranean Sea.

Highlights:

- Major oceanographic/climatologic event at the Late Messinian TG12/TG11 transition.
- Potential for a pre-Quaternary seesaw mechanism.
- Location of the ice-sheet capable of producing a large deglaciation to be identified.

Keywords:

- Late Messinian
- Planktonic Foraminifera
- Climatic Seesaw
- Deglaciation.
- Mio-Pliocene Transition

1. Introduction

1.1. Messinian climatic background

A large late Miocene glaciation (Bart et al., 2005) (Rabassa et al., 2005) and associated glacio-eustatic oscillations have long been suggested as an explanatory factor in the dessication cycles found in rocks of Messinian age from around the Mediterannean Sea (Kouwenhoven et al., 1999). Such climatic trend running into and out of that glaciation is visible in compilations of benthic $\delta^{18}O$ e.g. (Zachos et al., 2001). With regard to deglaciation, in particular, one glacial-interglacial transition at circa 5.55 Ma is often identified as being associated with a major eustatic change of at least 30 metres (Aharon et al., 1993) (Hodell et al., 2001) (Van der Laan et al., 2006). Before the onset of widespread Northern Hemisphere glaciation c. 2.7 Ma ago (Bartoli et al., 2005) and prior to the warm mid-Pliocene interval at c. 3.3 Ma (Dowsett et al., 2005) (Haywood et al., 2005), the Messinian stage (between 7.25 and 5.33 Ma (Lourens et al., 2004)) was a time of large-scale environmental modifications, as recorded in the late Miocene orography (Zhisheng et al., 2001), and of changes to seaway configurations e.g. Arctic (Gladenhov et al., 2002), Panama (Marshall, 1988) and Indonesia (Li et al., 2006)), which altered the overall planetary geography to more closely resemble that of modern times. Unsurprisingly, there are numerous signs of continental climate (Mercer and Sutter, 1982; Fortelius et al., 2006) and oceanic circulation changes (Lear et al., 2003) (Li et al., 2006) (Vidal et al., 2002), and changes to the hydrological cycle (Cane and Molnar, 2001) (Cerling et al., 1997) (Pagani et al., 1999), which collectively suggest a major climatic transition. The climatic reorganisation progressed towards a warming at 5.6 Ma which was superimposed on a longer-term well-defined cooling trend initiated during the middle Miocene (from c. 15 Ma) (Shevenell et al., 2004). The overall cooling is thought to be related to the development of the East Antarctic Ice Sheet and cooling of the global deep ocean (Miller et al., 1987). Evidence for global Miocene cooling is also found in the Northern Hemisphere suchas IRD on the Voring Plateau and around Fram Strait (Thiede et al., 1998). During the interval 15-5 Ma, deep and intermediate circulation changes with increasing ventilation of intermediate water are recorded in the Atlantic and Pacific oceans by changes in the benthic $\delta^{13}C$ (Hodell and Venz-Curtis., 2006). Ocean palaeocirculation patterns were also affected globally by progressive modification of connections between the main oceanic basins, and are recorded by Cd/Ca records (Delaney, 1990). A 0.8 ‰ decrease in $\delta^{18}O$ across the TG12 to TG11 transition was observed at ODP Site 982 (Hodell et al., 2001) and more recently in the Eastern Pacific at ODP Site 1241, (Steph et al., 2006), which is most simply explained as reflecting a global deglaciation event.

1.2. Controls on the Messinian climate

Besides, the dominant control of greenhouse gas concentrations on climate during the Cenozoic (DeConto and Pollard, 2003), oceanic currents strength and seasonality are thought to have influenced Miocene climate (Kurschner et al., 2008). On long geological time scales, plate tectonic tends to favour either (i) circum-equatorial oceanic circulation, thus promoting a warm climate, (ii) equator to pole circulation, which contributes to the development of glacial conditions (Smith and Pickering, 2003). Today, with a meridional Atlantic Ocean and very limited equatorial oceanic connections except for the narrow Indonesian gateway and minor Aguhlas “leakage”, which together enable mixing of warm Indo-Pacific equatorial surface water with that of the South Atlantic. The main factor affecting latitudinal heat transfer is water density enabeling a meridional overturning circulation (MOC). MOC is controlled by salinity variations in polar to sub-polar surface waters, involving winter cooling and/or sea-ice formation promoting sinking. The net effect of the MOC is to slightly cool the tropics. In the reverse scenario (i.e. less MOC), as demonstrated by modelling experiments such as that of the thermodynamic Bipolar Seesaw (Stocker and Johnsen 2003) heat tends to accumulate in the Southern Hemisphere, while the sub-polar Northern Hemisphere cools (Vellinga and Wood, 2002). Also, a reduced MOC modifies the global precipitation pattern, as the Inter-Tropical Convergence Zone (ITCZ) moves to the south. Despite relatively large uncertainty about the causal relations between events, Molnar (2008 and references therein) explored the potential role of closing the Panama gateway which is usually regarded as having played a key role in Northern Hemisphere glaciations (Driscoll and Haug, 1998), on the global climate evolution in the late Neogene. In particular, northward heat transfer via the Gulf Stream current is thought to have been enhanced after the Panama isthmus experienced progressive shoaling (Haug and Tiedemann, 1998) and occasional closures after the Messinian as shown by the great faunal interchange (Marshall, 1988), until it finally closed at around 2.7 Ma. More recently, Lunt et al., (2008) challenged this view of the onset of Pliocene glaciation, and on the basis of modelling outputs attributed it to declining atmospheric CO_2 levels. Because it is largely controlled by polar convection of dense waters from the surface to the abyss, the MOC appears to be easily destabilised by relatively small but rapid inputs of fresh water from collapsing Northern Hemisphere ice sheets during the last glacial interval (Labeyrie et al., 1995) or during older glacials back into Pliocene times (Bartoli et al., 2005). However, Heinrich events, linked causally with collapses of the Laurentide ice sheet, are observed after the middle Pleistocene (Hodell et al., 2008) suggesting considerable variations in size and locus and the likely recent involvement of the Laurentide Ice Sheet in rapid climatic events. Despite some recent advances (Nielsen et al., 2007), detailed accounts of similar massive ice sheet collapses in the Southern Hemisphere remain sparse. Modern global circulation depends on salinity gradients and on the input of pre-formed water masses with distinct salinities into the North Atlantic at sensitive locations. For example, the dense Mediterranean water flowing out through the Strait of Gibraltar contributes to the saltiness of North Atlantic waters and to the composition of North Atlantic Deep Water (NADW). Conversely, the presence of fresher Arctic surface water over the site of NADW convection exerts a negative feedback on NADW formation. However, brine rejection when sea-ice forms can also exerts a positive feedback on deep water formation (Haley et al., 2008) and may even be the major process that ventilates the North Atlantic during parts of glacial periods (Dickson et al., 2008). The entry of these pre-formed water masses occurs either through the return of surface water via shallow seaways connecting oceanic basins on the “shallow and warm route of the conveyor” (Gordon, 1986), or else through the input of dense water at intermediate water depths, such as the Mediterranean Outflow Water (MOW) or Red Sea water in the South Atlantic (de Ruijter et al., 2002). MOW is thought to have invigorated the MOC during the last glacial termination (Rogerson et al., 2006). Because changing land/sea configurations are affected by relative sea level changes, they effectively control water exchanges through shallow gateways (e.g. the Bering, Panama, Indonesian and Gibraltar straits) and modify global salinity gradients. In addition, the transfer of atmospheric moisture across the Panama isthmus, provides a stabilising feedback during episodes of weaker thermohaline circulation (Lohmann, 2003). Palaeoclimatic inferences must be made cautiously particularly prior to circa 5.5 Ma and the establishment of modern geography (Gladenkov et al., 2002) because even limited surface-water freshening in the sub-Arctic Ocean can influence the operation of the MOC (Bryan, 1986). Indeed, as pointed out by Molnar (2008), causal inferences between oceanic gateways and climate are liable to be oversimplified and at circa 5.5 Ma, the situation is particularly complicated (i.e. several oceanic straits were either slowly changing or periodically closed). As a potential analogue, the most recent deglaciation from 20,000 to 13,000 years BP, is well documented (e.g. Stott et al., (2007), and other references therein). An important mechanism involved during a full deglaciation is an oceanic tele-connection called the thermal Bipolar Seesaw that is set up across the Atlantic. Its action implies that the Southern Hemisphere, and the South Atlantic (Barker et al. 2009), becomes warmer when the Northern Hemisphere cools and vice versa (Stocker, 1998). In particular, the influence of fresh water from melting Northern Hemisphere ice sheets is believed to have been a major driver of the seesaw mechanism during the last glacial and subsequent deglaciation (Dokken and Nisancioglu, 2004).

1.3. Messinan in the Mediterranean basin:

Recently, the wealth of observations in the Mediterranean Sea area accumulated over the last 30 years has led to both a better understanding of the Messinian nature and evolution as well as to the developement of a stratigraphic synthesis presented graphically in Manzi et al., (2013) on Figure 1. Evaporite sequences found in and around the Mediterranean basin and representing the end of the Messinian stage (5.9-5.33 Ma) (Krijgsman et al., 1999) can be correlated with North Atlantic records (Hodell et al., 2001). Although the onset of deposition of evaporites at 5.96 Ma (Krijgsman et al., 1999); revised to 5.97 Ma by Manzi et al. (2013) is related to tectonic events affecting the Strait of Gibraltar (Duggen et al., 2003), their cyclicity testifies to isolation events involving sea level changes and therefore glacial cycles. Alternative explanation about such evaporitic cycles involved changing water budgets (Precipitation-Evaporation) around the Mediteranean Sea.

1.4. Messinan ice-sheets evolutions

In the context presented above, the transition from TG12 to TG11 that marks the onset of the UE seems particularly significant (Van der Laan et al., 2006).

Nearer to Antarctica, geological evidence is scarce. However, seismic records of multiple fluctuating ice sheet grounding lines of late Miocene age (7 to 5 Ma) off the Antarctic Peninsula testify to large changes in ice sheet configuration (Bart et al., 2005). Moreover, high-resolution data from sediment cores obtained from multi-sensor studies suggest orbitally controlled size variations of the Antarctic Ice sheet in the late Miocene (Grutzner et al., 2003). Further afield, evidence for several glacial episodes is found in Patagonia between 7 and 5 Ma (Rabassa et al., 2005). These observations are also consistent with major coeval changes in deep oceanic circulation recorded in sediments in the Argentine basin (Ledbetter and Bork, 1993; Hernandez-Molina et al., 2006) and thought to be related to episodic changes in Antarctic Bottom Water (AABW) flow. Close to West Antarctica in the Bellingshausen Sea, opal mass accumulation rate (a productivity-related proxy) denotes a steep change in sea ice coverage from a maximum circa 5.6 Ma to a minimum at 5.3 Ma (Hillenbrand and Ehrmann, 2005). Such a dramatic change could only be related to a large-scale regional warming and sea-ice reduction. Evidence from Antarctica currently suffers from dating uncertainties, but terrestrial outcrops are available from this period. These are of great significance in providing an additional check on global paleoclimate theories based on more distal marine records. For example, several localities, on James Ross Island and the Antarctic Peninsula, contain sedimentary deposits interbedded with volcanic rocks that are well dated by the $^{40}Ar/^{39}Ar$ method (Jonkers et al., 2002), (Pirrie et al., 1997) (Smellie et al., 2006), (Hambrey et al., 2008)). $^{87}Sr/^{86}Sr$ dating of locally abundant pectinid shells within the sediments indicates a variety of late Miocene-Pliocene ages, several of which are essentially coincident with, or are within error of, the 5.5 Ma global late Messinian event (Smellie et al., 2006) (Smellie et al., 2008). In the Northern Hemisphere, in addition to long marine sequences recovered through ODP drilling (Jansen and Sjoholm, 1991) (Larsen et al., 1994),the best exposed outcrop repository of late Cenozoic glaciomarine rocks is the Yakataga Formation (Gulf of Alaska), which dates the earliest marine tidewater glacier incursions to no older than the late Miocene (Lagoe et al., 1993).

2. Oceanographic conditions at the selected sites

2.1. Location

The present study is based on samples from two ODP Sites from the Atlantic Ocean. (1) ODP Leg 154 Site 925 is located in the western equatorial Atlantic, at 4°N 43°W and at 3041 meters water depth on the Ceara Rise (Curry et al., 1995). The Ceara Rise is one of the few places in the Atlantic where the Middle to Late Miocene was drilled continuously in a high sedimentation rate section. (2) ODP Leg 177 Site 1088 is situated at 41.8°S 13.3°E and 2300 metres depth on the Agulhas ridge in the South Eastern Atlantic (Gersonde et al., 2003).

2.2. ODP Site 925

ODP Site 925 is located within the western Atlantic under the area of atmospheric convergence where the two prevailing flows from the Southern and Northern Hadley Cells meet and interact at the Inter-Tropical Convergence Zone (ITCZ). Most of the changes in the area are seasonal and driven by wind patterns strongly controlled by movements of the ITCZ.

During the boreal summer, the site is under the influence of the North Equatorial Counter Current (NECC) whilst in the winter it is affected by the South Equatorial Current (SEC). However, the surface temperature is relatively constant and currently fluctuates only between 26 and 28°C annually (Figure. 1). Although the core site is located on a rise outside the main influence of terrigenous deposits from the Amazon River, the surface salinity varies seasonally depending on migration of the ITCZ across the Amazon basin, with values ranging between 35 and >37‰ (in July and December, respectively). Prior to the final closure of the Panama Isthmus the surface circulation would have been somewhat different within the Caribbean Sea and the nearby Western Equatorial Ocean, in particular with the entry of NECC and the cold circulation systems. However, sea floor sills in the region during the late Messinian probably did not exceed a few hundred metres in depth (Collins et al., 1996), and this would have significantly impeded the exchange of water at surface and upper intermediate levels. ODP Site 925, on the Ceara Rise, intercepts the two main deep-water masses on its flanks, i.e. the saltier warmer NADW and the fresher colder AABW. It is thus an ideal location at which to detect changes in deep-circulation, although not the main focus of this paper. Other water masses on site are the tropical surface water originating from the northern gyre water and the Antarctic intermediate water (AAIW).

2.3. ODP Site 1088

ODP Site 1088 in the south-eastern South Atlantic, was drilled at a water depth within the uppermost abyssal water on the broad north-eastern limit of the Agulhas Ridge. Today, ODP Site 1088 is situated just to the south of the subtropical front and it is weakly under the influence of warm Agulhas circulation, which transfers surface and sub-surface water from the Indian Ocean into the subtropical South Atlantic as part of the surface return route of the MOC (Gordon, 1986). Salinity is fairly stable around 34‰ throughout the year while SSTs fluctuate between 10 and 15°C (Figure. 1). Interesting insights into the operation of this seaway during the late Quaternary can be found in Peeters et al., (2004), and in O'Neil et al., (2007) for the middle Pleistocene.

3. Analytical methods

3.1. Selected samples

The samples used in this study were obtained from 3 sources. All samples from ODP Site 925 used the existing coarse fractions studied by Pfuhl and Shackleton (2004). A few of the samples from ODP Site 1088 were studied by Diester-Haass et al., (2005) but additional intermediate samples prepared using standard procedures described by Vautravers et al., (2004) were added (this study) to obtain a higher resolution. The average spacing between samples at either Site was 10cm, resulting in an average time resolution of one sample every 5 and 10 kyr (ODP Site 925 and 1088, respectively).

3.2. Micropaleontology and coarse fraction study

For each sample, 150 micron size fractions were splitted until each sub-sample contained about 300 whole tests allowing full planktonic foraminiferal associations to be determined. Planktonic foraminifera species were counted following the taxonomy of Kennett and Srinivasan (1983). In the case of the mixed layers species (Figure 2I) individual species were grouped following known ecological preferences and according to Chaisson and Ravelo (1997). For ODP Site 925, the preservation of calcium carbonate ($CaCO_3$) was also assessed using the percentages of planktonic foraminifera fragments from Pfuhl and Shackleton (2004) as well as the individual shell weights obtained prior to the trace metal analysis. In addition, for ODP Site 1088 located south of the subtropical front the content of ice rafted detritus (IRD) >350 microns was also quantified in order to trace icebergs rafting during colder episodes associated with potential northward frontal movements.

3.3. Geochemical study

3.3.1. Stable isotopes

At ODP Site 925, the initial isotopic work was carried out on the benthic species *Nutallides umbonatus* and several species of planktonic foraminifera. In particular, samples made of 30 specimens of *Globigerinoides trilobus*, a species that closely resembles a modern *Globigerinoides sacculifer* without a sac chamber, were selected within the 300-355 micron size fraction (Pfuhl and Shackleton, 2004). For ODP Site 1088 the analyses were performed on *Globigerina bulloides* selected within the 300-355 micron size fraction (this study). All analyses were made in the Godwin Laboratory facility (University of Cambridge, UK). The analytical error is 0.08 ‰ for oxygen and 0.06 ‰ for carbon. Small samples were run on a PRISM mass spectrometer whereas larger samples, typically those of planktonic foraminifera, were run on a VG Isotech SIRA.

3.3.2. Trace metals

Mg/Ca analyses were obtained on *G. trilobus* and *G. bulloides* from the same size fraction on additional specimens picked from the same samples. The Mg/Ca measurements were typically obtained from samples containing 30 specimens (*G. trilobus* for ODP Site 925 and *G. bulloides* for ODP Site 1088). Cleaning procedures were as described by Barker and Elderfield (2002). Measurements were carried out on an ICP-AES, with analytical errors of 1 to 2% equivalent to 0.3 to 0.6°C to which the calibration error (Anand et al., 2003) must be added to account for a total maximal error of 1.1°C. Palaeotemperatures were calculated using the general temperature calibration equation from Anand et al., (2003). δ^{18}Ow was calculated using the temperature equation from Bemis et al., (1998).

3.4. Stratigraphy

The stratigraphic framework for each section examined was obtained through orbital tuning for ODP Site 925 following Shackleton and Crowhurst (1997). Stratigraphy for ODP Site 1088 follows Diester-Haass et al., (2005) using only a few additional tie points. The chronological uncertainty for ODP Site 925 does not exceed 5 kyrs but is larger for ODP Site 1088 (c.10 kyrs).

4. Results

4.1 Stable Isotopes

4.1.1. ODP Site 925

The deep-sea record of benthic foraminiferal isotopes ($\delta^{18}O$) from Pfuhl and Shackleton (2004) is replotted on Figure 2B and provides information about changes in ice volume during the interval of interest. Between 5.6 and 5.3 Ma the $\delta^{18}O$ values fluctuate between 3.5 and 2.5‰ showing a trend towards lighter values across the whole interval (Figure 2). The amplitude of the variability is typically about 0.5‰ from one glacial to the next interglacial. However, the amplitude appears to decrease through the interval. In contrast, at the time of the main deglacial event starting at circa 5.55 Ma, a maximal amplitude of 0.8‰ is recorded. Interestingly, the transition displays 2 steps, which are also visible in the SSTs, in a similar way to observations during the last deglaciation between 18 and 12,000 years BP. The benthic $\delta^{13}C$ (Figure 2F) exhibits large fluctuations, with values varying between 0.5‰ and 1.1‰. However, the heaviest value is rare and represented only by 3 data points c. 5.47 Ma. Similar high values have been reported during Pliocene times (Ravelo et al., 2007). Four intervals have low $\delta^{13}C$ and, apart from one (at the M-P transition), are associated with lower shell weights (Figure 2E). In addition, the low-$\delta^{13}C$ intervals correspond to periods of increased shells fragmentation (Figure 2G) (Pfuhl and Shackleton, 2004). The benthic $\delta^{13}C$ is significantly lower (signalling a glacial) just prior to the M-P transition. This is also shown by lower SSTs but is not confirmed by the benthic $\delta^{18}O$ values.

At ODP Site 925, planktonic $\delta^{18}O$ values range between -0.8 and -1.6‰, a range comparable in amplitude to the benthic one. The largest fluctuation occurs at the TG12-TG11 boundary

(Figure 2A) and is also recorded by the SSTs. In addition, two other warm periods, one before 5.5 and one circa 5.4 Ma, are shown by both planktonic and benthic isotopic records. This suggests a planktonic and benthic coherence for the major events. However, towards the end of the time corresponding to the UE the planktonic record is more variable than the benthic one. The planktonic $\delta^{18}O$ also contains several discrete events marked, for example, by heavy $\delta^{18}O$ values during TG12 (a glacial). In contrast, the lightest values occur within TG11 (an interglacial). The estimates for the $\delta^{18}O$ of the water (Figure 2C) reconstructed from the planktonic record becomes generally heavier through the interval studied, indicating a trend towards locally saltier surface water through time. In addition, the local surface salinity suggests a relative freshening during four distinct episodes, at 5.5 Ma (the first glacial during the UE) and then for 3 further consecutive glacial episodes just prior to the onset of the Pliocene. It is interesting to note that two of these episodes also correspond to changing deep-water $\delta^{13}C$, indicating an increased influence of AABW at the depth of ODP Site 925.

4.1.2. ODP Site 1088

ODP Site 1088, was studied in much less detail but nonetheless provides some important information about the Southern Ocean (Figure 3). The $\delta^{18}O$ values for *G. bulloides* (Figure 3A) show a big change at circa 5.55 Ma, with an amplitude of 0.8‰. The change is similar to that observed at ODP Site 925 and both are attributed to the TG12-TG11 transition.

4.2. Sea Surface Temperatures

4.2.2. ODP Site 925

The minimal SST recorded for ODP Site 925 is 24°C (Figure 2D). It occurs during TG12, the last glacial in the late lower evaporites, whereas the maximum SST is 27.5°C, during the second Pliocene interglacial. Values for interglacials and glacials vary between 26-27°C and 24-25°C, respectively. In addition, SST derived from Mg/Ca ratios are cyclical and have amplitudes of 1.5°C to 2°C. They reach a maximum value near the middle of the period examined corresponding to deposition of the UE. In addition, the average SSTs calculated for each interval (late Lower Evaporites (LE), UE, and Early Pliocene) increase from LE to UE and up into the early Pliocene. However, interglacial temperatures during the UE are similar to those during the early Pliocene while the glacials one during the UE are only marginally cooler than during the LE. The average values in each of these specific intervals LE, UE, and early Pliocene are 24.8, 25.6 and 26.7°C respectively (solid line on Figure 2D), showing a progressive increase that suggests a warming trend.

4.2.3 ODP Site 1088

The reconstructed temperatures for ODP Site 1088 vary between 15.5°C and 22°C, which is significantly different to the temperatures range observed at the core site today, 12-14°C (Levitus Atlas, 1994), (http://ingrid.ldeo.columbia.edu/SOURCES/.LEVITUS94). However, a major feature of the results for ODP Site 1088 is a prominent 2°C warming that happened during the major deglaciation at c. 5.5 Ma (Figure 3A). During the late Messinian and in particular during the time equivalent to the LE and the LE-UE deglaciation, the temperatures reach a maximum of 22°C and then fall sharply after 5.5 Ma. They then remain relatively low (between 17 and 18°C) and do not exhibit significant variability until after the M-P boundary. Then the SSTs (Figure 3B) show a slight increase at the onset of the Pliocene but they do not regain the preceeding values seen in TG11. More remarkably, the Pliocene is marked by a sudden influx of IRD (Figure 3D). This happens even though the temperatures are not at their lowest but are, in fact, showing a slight increase.

4.3. Planktonic foraminifera faunal assemblages

4.3.1. Equatorial Atlantic ODP Site 925

Faunal assemblages of Miocene age cannot be used for quantitative SST reconstructions (i.e. using transfert function) because of the lack of good modern analogues. However, counts of planktonic associations give qualitative insights into changing surface and subsurface hydrology. For example, the results show that the interglacial preceding the glacial immediately prior to the M-P transition is the first one to contain the species *Globigerinoides hexagonus* (Figure 2H) after a long interruption initiated at the end of the interglacial TG11. Other faunal results, expressed as a percentage of the total for selected indicative species or groups of species such as mixed layer species Chaisson and Ravelo (1997), % "Menardellidae" and % *Neogloboquadrina pachyderma* (left and right coiling) (Figure 2I, J, K respectively).

The mixed layer group is the most abundant of all with a maximum relative abundance close to 75% at 5.48 Ma. This maximum occurs during the glacial following TG11. In addition to being quite variable (range from 45 to75%), the group displays a decreasing general trend (falling to 45% after 5.3 Ma), although with two prominent earlier minima at 5.5 and 5.38 Ma. Before 5.55 Ma, the relative abundance of the mixed layer group is generally high (typically exceeding 60%). Similar to the previous interval (i.e. LE), relatively high values for this group also correspond with lighter benthic $\delta^{13}C$, consistent with glacial intervals. In contrast, the next group of species reported here as "all Menardellidae" (Figure 2J), does not show any trend but the abundance values display a clear cyclicity varying between 5% and 40%. These species are thermocline dwellers and present higher percentages during interglacials. So, they are anti-correlated with the previous group.

The last group discussed here includes polar-subpolar forms ("% cold species" in Figure 2K) % *Neogloboquadrina pachyderma* (left and right coiling). Although never in excess of 10%, the representation of these species increases steadily, starting just after the end of TG11 and continuing into the Early Pliocene. The timing of the onset of this increase in *N. pachyderma* (i.e.cold species) seems unrelated to the major transition at TG12-TG11. Indeed, it clearly postdates it.

4.3.2. South Atlantic ODP Site 1088

At ODP Site 1088, the polar *N. pachyderma* (left) morphotype dominates the *N. pachyderma* species as well as the whole faunal assemblage (up to 40%). *N. pachyderma* (left) represents over 20% during most of the studied interval. It is also only marginally more abundant during the middle interval (i.e. the UE), which therefore appears to be the coldest at that site (also confirmed by the SST from Mg/Ca; Figure 3B). The percentage of "warm " tropical species at ODP Site 1088 suggests two episodes of warming: one at the LE-UE transition, and one at the M-P boundary, with a clear decrease in relative abundance between those two periods. The sum of % *G. bulloides* and %*Globigerinita glutinata*, which together represent the last group examined, fluctuates between 10 and 25% and displays a weak increasing trend through time. Both species indicate high-productivity such as is typically found in upwelling areas (Vautravers and Shackleton, 2006).

5. Discussion

5.1. Oxygen and Carbon isotopes

A decrease of 0.8‰ across the TG12/TG11 boundary at ODP Site 925 is recorded by the benthic $\delta^{18}O$ record and strongly suggests a large decrease in global ice volume. Beside this trend the record also shows additional remarkable features relating to subsequent glacial and interglacial intervals. The decreasing amplitude of $\delta^{18}O$ variability is mainly due to the glacials becoming less pronounced while the interglacials remain at similar (low) $\delta^{18}O$ values. The pattern of $\delta^{18}O$ variations seems to indicate that, after an initial large deglaciation at the LE-UE transition, the interval of time corresponding to the deposition of the UE in the Mediterranean basin is characterised by overall diminishing global ice volume. The three heaviest benthic $\delta^{13}C$ values (1.2‰) also occur immediately after TG11 and are attributed to enhanced NADW production as they are close to Holocene values at the site. However, it is worth mentioning that the extreme values are found during an interval that seems to correspond to a glacial as recorded by the benthic $\delta^{18}O$ but at a time when the $\delta^{18}O$ from benthic and planktonic foraminifera diverge. The explanation for this situation is unclear. However, heavier benthic $\delta^{13}C$ values during a glacial might also be related to the formation of glacial North Atlantic Intermediate water (GNAIW), with its high $\delta^{13}C$ value reflecting its origin within subtropical waters. It might be possible to link these features to the opening of the Bering Strait, which is dated to between 5.5 and 5.4 Ma (Gladenkov et al., 2002). Indeed, this last event may have been responsible for a large-scale surface freshening in the North Atlantic, when the relatively fresher Pacific Surface Water became able to cross the Arctic Ocean via the newly-opened Bering Strait. Such an inter-oceanic link and influx of less saline water would have perturbed the NADW, displacing its zone of formation further South. Overall, during the entire UE interval the benthic $\delta^{13}C$, proxy for deepwater circulation frequently exhibits heavy values, close to 0.8-0.9‰, that are close to the present values for NADW. If the opening of the Bering Strait was responsible for these effects, they should be detectable in records similarly resolved at other sites in the North Atlantic. In addition, low $\delta^{13}C$ values (circa 0.5‰) are found in three younger periods (centred at around 5.45, 5.4 and 5.34 Ma). These indicate a recurrent enhanced influence of AABW at ODP Site 925. It is significant that episodes with lighter $\delta^{13}C$ do not always correspond to glacial intervals, as identified by their heavier benthic $\delta^{18}O$. However, the correspondence between shell weight, shell fragmentation and benthic $\delta^{13}C$ (Figure 2E, F, G) suggests that at ODP Site 925 southerly-derived cold deep water is a major control on carbonate dissolution during these three events.

Because of the tropical location of the core, it is likely that the planktonic $\delta^{18}O$ record is affected by periodic changes in local surface salinity related to changing rainfall over eastern Brazil. Indeed, at face value, the trend of increasing planktonic $\delta^{18}O$ values throughout the UE interval suggests a cooling, in contradiction to the Mg/Ca SST record obtained from the same samples, that shows a coincident warming (Figure 2D), and at a time when benthic $\delta^{13}C$ does not show any obvious evidence for a changing influence of NADW versus AABW (apart from some discrete events, as described above). The combination of evidence thus supports a possible link to changing Brazilian rainfall patterns (i.e. enhanced influx of rainwater during TG11). An apparent contradiction to this result comes from the % shell fragments reported here (Figure 2G; after Pfuhl and Shackelton, (2004)), which shows a decreasing trend that could be consistent with a decreasing influence of southerly-derived deep-water. However, this occurs in the absence of changes in deep-water circulation (i.e. benthic $\delta^{13}C$) that could drive a shoaling of the lysocline. The only possible explanation for the trend in preservation is that specimens of *G. trilobus* are more calcified, as traced by the shell weight, increasing toward the end of the interval. The explanation for this pattern is either global and linked to decreasing greenhouse gas concentration (although not captured by the existing Miocene low resolution pCO_2 record (Pagani et al., 1999)) or that it is under the control of the local tropical surface hydrology that may affect the biomineralisation of *G. trilobus*.

The $\delta^{18}O$ isotopic data indicate that the most important climatic event was the TG12-TG11 deglaciation at c. 5.5 Ma. It was followed by a sustained warming trend reflected in decreasing glacial severity that continued through the UE and into the early Pliocene. A possible, although contentious, driver of this warming might be a renewed and progressive influx of MOW into the North Atlantic circulation. The data also suggest that the largest deglaciation took place at the LE-UE boundary and not at the Miocene-Pliocene transition (Van der Laan et al., 2006), and that, at ODP Site 925, it may have been related to a change in deep-water circulation. However, a remarkable feature of the M-P transition, especially when contrasted with the transition between the LE and UE, is that there are significant changes in $\delta^{13}C$ whereas the benthic $\delta^{18}O$ is unaffected. This may indicate that the M-P transition was affected by some large oceanographic changes, perhaps originating near Antarctica and affecting the AABW production, but that it did not involved a significant change in global ice volume. By contrast, the changes observed at the TG12-TG11 transition, would have required substantial melting of an as yet unidentified ice sheet.

5.2. Sea Surface Temperatures (SSTs)

The SSTs during Messinian interglacials were close to that observed today at ODP Site 925, whereas those during Messinian glacials were about 2°C cooler. During glacial episodes, any southward displacement of the ITZC would have caused sustained surface cooling. For example, during UE glacials, the lack of salty Mediterranean Overflow Water (MOW) flowing into the North Atlantic would have reinforced the tendency for a slower MOC, causing the ITZC to be pushed further south and cooling the SSTs more intensely at ODP Site 925. This suggests that the periodic disconnection of the Mediterranean Sea had a far-reaching effect on tropical SSTs. If true, this suggests that MOW might be a leading influence on MOC and consequently on tropical climate during the late Messinian. In agreement with model predictions by Vellinga and Wood (2002) of a 2°C cooling in this area is to be expected in case of a MOC collapsed due to a freshwater influx to the North Atlantic (applied to Holocene boundary conditions). In addition, in the case of the Messinian, fresh water derived from melting of a northern hemisphere ice sheet is a possibility. However, well-dated sites in the North Atlantic have not yet identified evidence, such as an IRD peak, that might verify a melting event (Hodell et al., 2001). Therefore, the plausible alternative mechanism for a slowing of MOC during the latest Miocene might be the lack of salty MOW normaly feeding into the North Atlantic during glacials. TG12 represents only a single glacial period and the record at ODP Site 925 is probably too short for a definitive conclusion, but the existing data suggest that prior to TG11, both glacial and interglacial periods were cooler than during TG11 or younger periods. For example, although only a single pre-TG11 Miocene interglacial is documented, it is the coldest observed in the whole sequence with maximum SSTs of c. 26°C, about 2°C lower than modern. Therefore, it seems that in a context of increased isolation of the Mediterranean Sea, surface conditions were definitively cooler in the tropical Atlantic. This observation suggests that later the waters in the mixed layer became progressively warmer due to renewed inputs of MOW to the general oceanic circulation during successive interglacials, resulting in a northward migration of the ITCZ as MOC was re-invigorated.

The model of MOC collapse described by Vellinga and Wood (2002) alters precipitation patterns and, specifically, increases precipitation in northern Europe, Alaska and eastern Brazil. A reduction in the MOC particularly influences the wet season in Brazil (it tends to become much wetter than normal), whereas the dry season is relatively unchanged. Therefore, the warming that is observed in successive Messinian interglacials, and the trend of increasing $\delta^{18}O$ in the planktonic *G. trilobus*, suggests that local SSTs and salinity probably evolved from an anomalous glacial LE state to a more normal state by the early Pliocene. Indeed, the trend in surface water $\delta^{18}O$ (Figure 2A) suggests that precipitation in eastern Brazil decreased and that a more normal global circulation pattern was gradually re-established throughout the last interval (UE) of the late Messinian.

5.3. Microfauna

During the Late Quaternary, *G. hexagonus*, which today is endemic to the Indo-Pacific Ocean and absent from the Atlantic Ocean (Kennett and Srinivasan, 1983), reappears in the Atlantic during interglacials. Peeters et al., (2004) found it living at intermediate depths (at ~600 meters water depth) and associated with a ring of Indian Ocean water that had been transferred into the South Atlantic via the "Agulhas leakage". At the end of the Messinian, the Isthmus of Panama had shoaled to less than 500 m so that the periodic occurrence in the Atlantic Ocean of this rare species could only be related to its re-introduction via the Agulhas area (Collins et al., 1996). Major changes in the relative abundance of such a species may be linked to intermediate-depth circulation. Thus, the peak in abundance observed at the end of the Miocene at ODP Site 925 (Figure 2H) may indicate an enhanced influx of Antarctic Intermediate Water. This is also potentially reflected in the prominent changes in the benthic $\delta^{13}C$, also linked to AABW production. These together support the idea that a major (i.e. global) circulation change occurred at the end of the Miocene and that it could have been initiated in the Southern Hemisphere.

High percentages of mixed layer species (at ODP Site 925) point to a deep mixed layer. Indeed, increased south-easterly trade winds result in piling up of warm surface waters in the western tropical Atlantic (Chaisson and Ravelo, 1997). Therefore, the presence of high percentages of mixed layer species (in particular before 5.6 Ma and during the glacial following TG11) supports the hypothesis that the late Messinian glaciations were accompanied by enhanced SE trade winds. However, the decreasing trend shown by this group suggests that the mixed layer was becoming shallower through time.

The "Menardellidae" group has a pattern of fluctuation reminiscent of the situation in the tropical Atlantic Ocean during the late Quaternary, where *Globorotalia (menardella) menardii* is known to disappear during glacials but reappear during interglacials. This implies that, during the Messinian, related species were being reintroduced from the Indian Ocean into the Atlantic through the Agulhas leakage on the shallow warm (i.e. surface route) of the MOC (Gordon, 1986). Therefore, their fluctuation in abundance may be an additional qualitative tracer for MOC intensity. However, in support of this hypothesis, and in contrast to the situation for *G. hexagonus*, it is difficult to rule out regular re-introduction of Menardellidae from the Pacific through shallow sills before the final closure of the Panama gateway. However, with the exception of the interval with the highest $\delta^{13}C$ value at 5.48 Ma, a comparison between the abundances of this group and the $\delta^{13}C$ variations shows clearly that a higher proportion of Menardellidae tends to correspond to higher $\delta^{13}C$, pointing to a link between the return of warm surface water to the South Atlantic and the intensity of MOC. The observation that the polar fauna (*N. pachyderma*) increases at the tropical ODP Site 925 starting after 5.5 Ma is puzzling because the SSTs from Mg/Ca measured on *G. trilobus* show an increasing trend during the same interval. However, even such apparently contradictory evidence can be reconciled. The increased representation of cold species could denote some circulation change in northern latitudes being transferred into the north equatorial environment via the North Equatorial Current, possibly under increasing influence of NE trade winds. Furthermore, the onset of the increase in *N. pachyderma* seems to be essentially time coincident with the Bering Strait opening. It is not suggested here that the polar planktonic fauna crossed the Bering Sea, since a genetic study has shown that Pacific and Atlantic populations are distinct (Darling and Wade, 2008). However, the timing of the opening of the Bering Strait and that of the cooling in the North Atlantic Ocean is puzzling. The *N. pachyderma* increase may reflect enhanced upwelling linked to increasing influence of the NE trade winds (after the opening of the Bering Strait) versus a decreasing influence of the SE trades winds that were dominant before 5.5 Ma. Therefore, the faunal changes traced by the polar species into the tropical water clearly suggest a North Atlantic cooling (despite local warming of a thinning of the mixed layer), which would have been transmitted to the tropics by enhanced NE trade winds. This cooling might correspond to the onset of a renewed phase of upper Neogene glaciation, since its timing is close to the onset of localised glaciations in Iceland (Geirsdottir et al., 2007).

5.4. IRD at ODP Site 1088

At ODP Site 1088, the onset of IRD (Figure 3D) very shortly after the Miocene-Pliocene boundary may be a consequence of large icebergs calving from a disintegrating ice shelf in Antarctica. This may indicate significant Antarctic warming and be an evidence for melting of part of the Antarctic Ice Sheet. Although there are signs of warming affecting the extent of sea ice during the early Pliocene on the Pacific margin near West Antarctica and the Antarctic Peninsula Ice Sheet (Hillenbrand and Ehrmann, 2005), and although the proportion of tropical fauna increased slightly at ODP Site 1088, the absence of a well-marked decrease of the $\delta^{18}O$ at the M-P boundary at the site appears to preclude a major deglaciation (i.e. involving a large volume of continental ice carrying light $\delta^{18}O$ melting on site). However, ice rafted detritus in the South Atlantic might also have been transported by drifting sea ice and need not indicate major iceberg activity (Nielsen et al., 2007). Its presence might be related simply to changes in surface oceanography, in particular to the position of the polar front and that of the seasonal sea ice than to increasing ice sheet melting.

Although a direct link between the reduced and periodic influx of MOW and a reduction of MOC during the Messinian remains hypothetical, the SST pattern at ODP Site 1088 also suggests another possibility. Could the thermal Bipolar Seesaw mechanism (Stocker and Johnsen 2003) be in action in pre-Quaternary times between the North and the South Atlantic and explain the observed data? Indeed, by analogy with the situation found for one core in the Cape Basin during the last glacial termination (Barker et al., 2009) an early warming during the initiation of the large TG11 deglaciation, as seen at ODP Site 1088, is consistent with a slowdown of the MOC triggered by the collapse of a large Northern Hemisphere ice sheet and surface water freshening (similar to the possible effects of a Heinrich event during the last glacial termination). However, so far there is no evidence for an IRD pulse at the termination of TG12 within the sub-polar North Atlantic. Because ODP Site 925 is located at the northern subtropical front, the data for the site cannot exclude a role for a melting Antarctic Ice Sheet since any icebergs are likely to have melted before delivering IRD to such a northerly location. Future examination of core sites situated much closer to Antarctica will be crucial to unravelling the contribution of Antarctica to this deglaciation. In addition, for the warming in TG11, the micropalaeontological study reveals that the fauna related to Agulhas leakage (i.e. sensus Peeters et al., 2004) at ODP Site 1088 do not increase in relative abundance and therefore the warming cannot be attributed to increased circulation of warm Agulhas water at the site. This observation is similar to those made in a micropalaeontological study of site TNO57-21 over the last deglaciation (Barker et al., 2009).

6. Discussion

As a consequence of the previous observation, the warming at ODP Site 1088 could be explained by a warming of the Antarctic Circumpolar Current as part of a thermal Seesaw mechanism or by a warming in the surface water travelling from the Indian Ocean into the Atlantic Ocean resulting of changes in the Indonesian Sea throughflow. The “Seesaw” implies that the South Atlantic accumulates heat when the MOC is reduced due to melting of a Northern Hemisphere ice sheet. A decrease in sea ice has been inferred from increasing opal accumulation rates in the Bellingshausen Sea between 5.6 and 5 Ma (Hillenbrand and Ehrmann, 2005), whereas IRD data and associated SSTs (showing signs of warming) suggest that, seasonality increased in the early Pliocene. The data at ODP Site 1088 are fragmentary and low resolution but support the existence of a major deglaciation coincident with the LE-UE transition, thus supporting the idea that refilling of the Mediterranean Sea was partly eustatically driven. However, the data presented here confirm that refilling the Mediterranean at the M-P boundary was not eustatic, but that the Southern Ocean was nonetheless experiencing major changes at the time. Those changes may have included reduced sea ice on the Pacific margin of Antarctica, enhanced production of AABW (Ledbetter and Bork, 1993) and AAIW, and a cooling of the South Atlantic with IRD carried by ice reaching further north after the onset of the Pliocene. The reasons for these changes cannot be inferred solely from the data at ODP Site 1088 but the likelihood of an intense large-scale warming near Antarctica at the onset of the Pliocene is high, and such an event would help to explain the occurrence of calcareous interglacial deposits on James Ross Island at broadly the same time (e.g. Jonkers et al., (2002)). Why a warming and the observed hydrographic changes should occur in the Southern Ocean at that time is uncertain. However, the timing, after 5.5-5.4 Ma, seems to link the changes with the opening of the Bering Strait, perhaps through some atmospheric tele-connections across the Pacific Ocean. The Bering Strait is the only gateway definitely known to have opened at that time. Other gateways (e.g. Indonesian and Panama) were changing comparatively slowly so any claims that their progressive closure may be important for late Messinian climatic evolution should probably be discounted pending further data.

7. Summary and conclusion

At ODP Site 925 a warming is evident in the mixed layer throughout the late Miocene and the earliest Pliocene. However, the evidence also suggests that the mixed layer becomes progressively thinner and saltier through time probably reflecting decreasing south-easterly trade winds and decreasing precipitation over eastern Brazil. It would appear that during the UE depositional period, the absence of MOW was influencing the MOC and cooled the tropical surface climatic conditions. When a normal oceanic connection between the Mediterranean and Atlantic was subsequently re-established at the M-P boundary, the tropics became warmer than during the preceding Messinian salinity crisis. These observations suggest that in the absence of MOW the MOC was reduced and the precipitation patterns were increased in the tropics (data from ODP Site 925) in agreement with Velinga and Wood (2002). Those authors also predicted increased precipitation over Alaska which would have helped grow a large ice sheet there. The main result from ODP Site 1088 is an early warming at the LE-UE transition that is consistent with the action of the bipolar thermal Seesaw mechanism in the Atlantic prior to the Quaternary. It is possible that a major deglacial event in the North Atlantic, similar in magnitude to a Heinrich-type event and signalling an ice sheet collapse in the Northern Hemisphere might have occurred but it has yet to be found at the TG12-TG11 transition. The presence of IRD arriving at ODP Site 1088 during the early Pliocene confirms that oceanographic changes in the South Atlantic are likely linked to large changes around Antarctica. However, the absence of a change in global ice volume at that time (in contrast to that found at the onset of UE deposition) probably rules out a large deglaciation. However, because the site is fairly distant from Antarctica, the lack of IRD at the LE-UE transition is not strong enough evidence in itself to exclude the possibility that the Antarctic ice sheet experienced large-scale melting, sufficient to raise sea levels significantly. Further studies should look closer to Antarctica for potential traces of substantial deglaciation at the TG12-TG11 transition (i.e. ~5.5 Ma ago). Such a search may be in vain, since the new SST data at ODP Site 1088 suggest that in the case of an operant bipolar seesaw in the Messinian much of the melting may have been located in the Northern Hemisphere, where the presence of IRD at the TG12-TG11 transition will also need to be confirmed. If the suggestion that a large scale deglaciation occurring in the Northern hemisphere is confirmed, it would support the proposal, based on evidence on the Antarctic Peninsula, that although parts of the glacial cover in Antarctica have fluctuated in thickness, the ice sheet became progressively thicker toward the present, with no clear evidence for complete melting during the last 7.5 Ma (Smellie et al., 2008) (Naish et al., (2008). However, this result cannot exclude the possibility of smaller scale melting events after the late Miocene. In the North

Atlantic region, onshore studies indicate that Iceland was not significantly glaciated before 4 Ma (Geirsdottir et al., 2007), whereas in Greenland there is evidence for multiple glaciations since the late Miocene. In this last case glaciations may have started in the south of Greenland rather than in the north, pointing to higher precipitation as the key control for the onset of Greenland glaciation (Larsen et al., 1994). Further afield, and at lower latitudes compared to Greenland, another candidate area that may have experienced enhanced precipitation, and thus could have sustained a large-scale ice sheet during the LE period, is the Pacific margin of Alaska. According to Velinga and Wood (2002), a reduction in MOC would have increased precipitation over southern Alaska, which would lead to growth of a Cordilleran Ice Sheet. A large Cordilleran Ice Sheet would then have been prone to significant melting at the end of the Messinian when the MOC returned to a more normal situation. The presence of glaciomarine sediments in the Yakataga Formation confirms that glaciers in the region had attained tidewater status by the late Miocene (Lagoe et al., 2003) and may therefore indicate the presence of a sizeable ice cover at that time capable of affecting sea level. Although purely hypothetical at present, if such a suggestion is verified it would not only confirm the importance of opening the Bering Strait as a major event in Messinian climate evolution, but also raise to centre stage the importance of the glacial history of the North Pacific region for determining global climate evolution in the upper Neogene.

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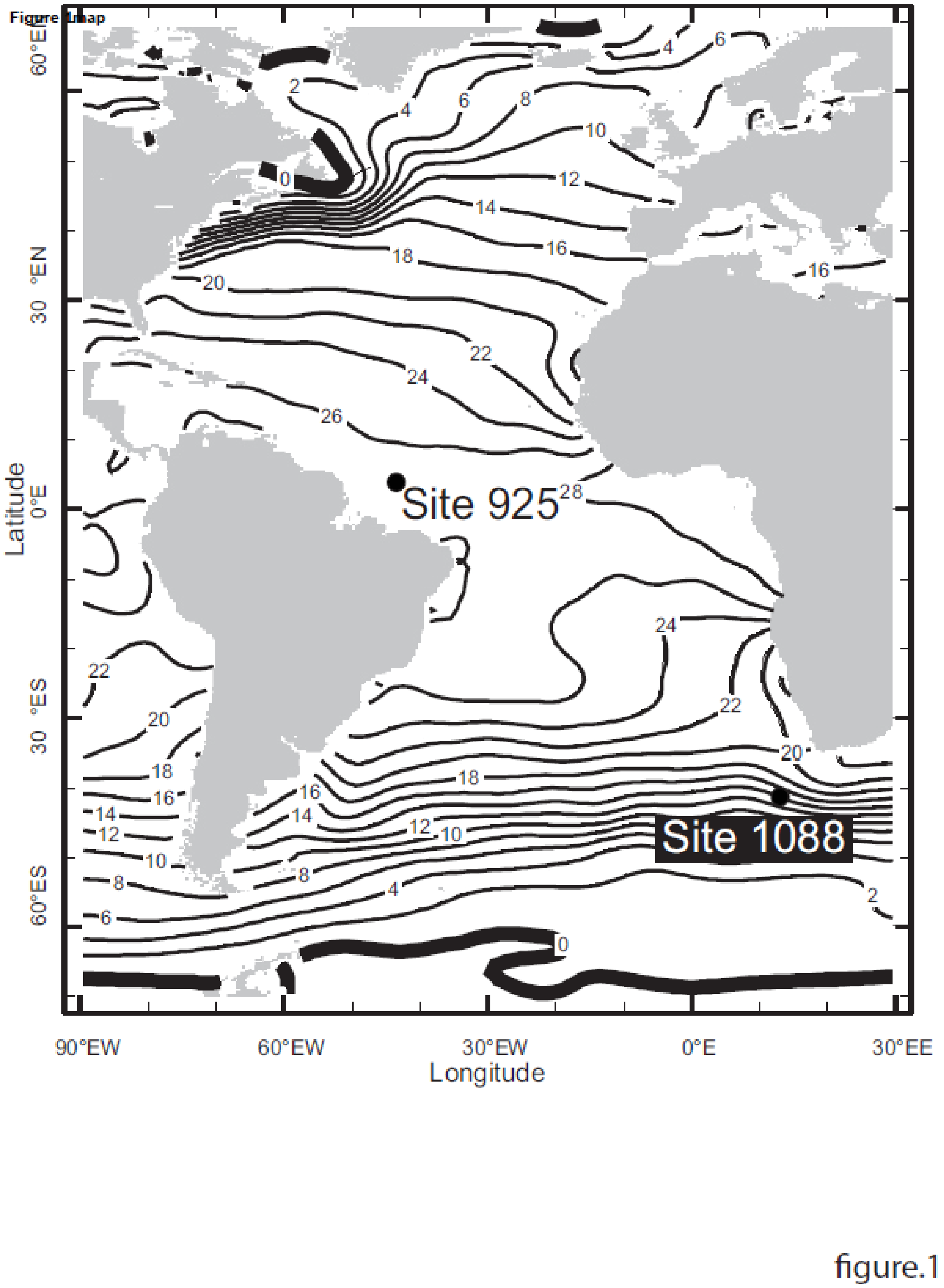
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Longitude
figure.1

Figure1

Figure 2 R2

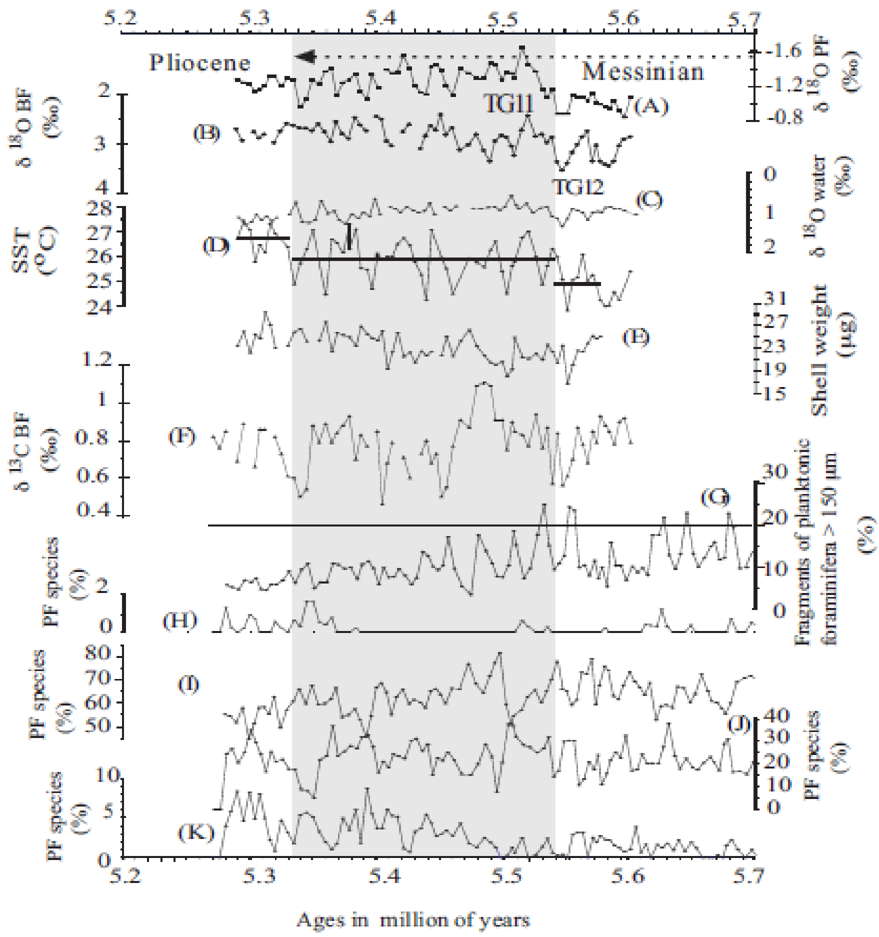

Figure. 2

figure2

Figure 3 R2

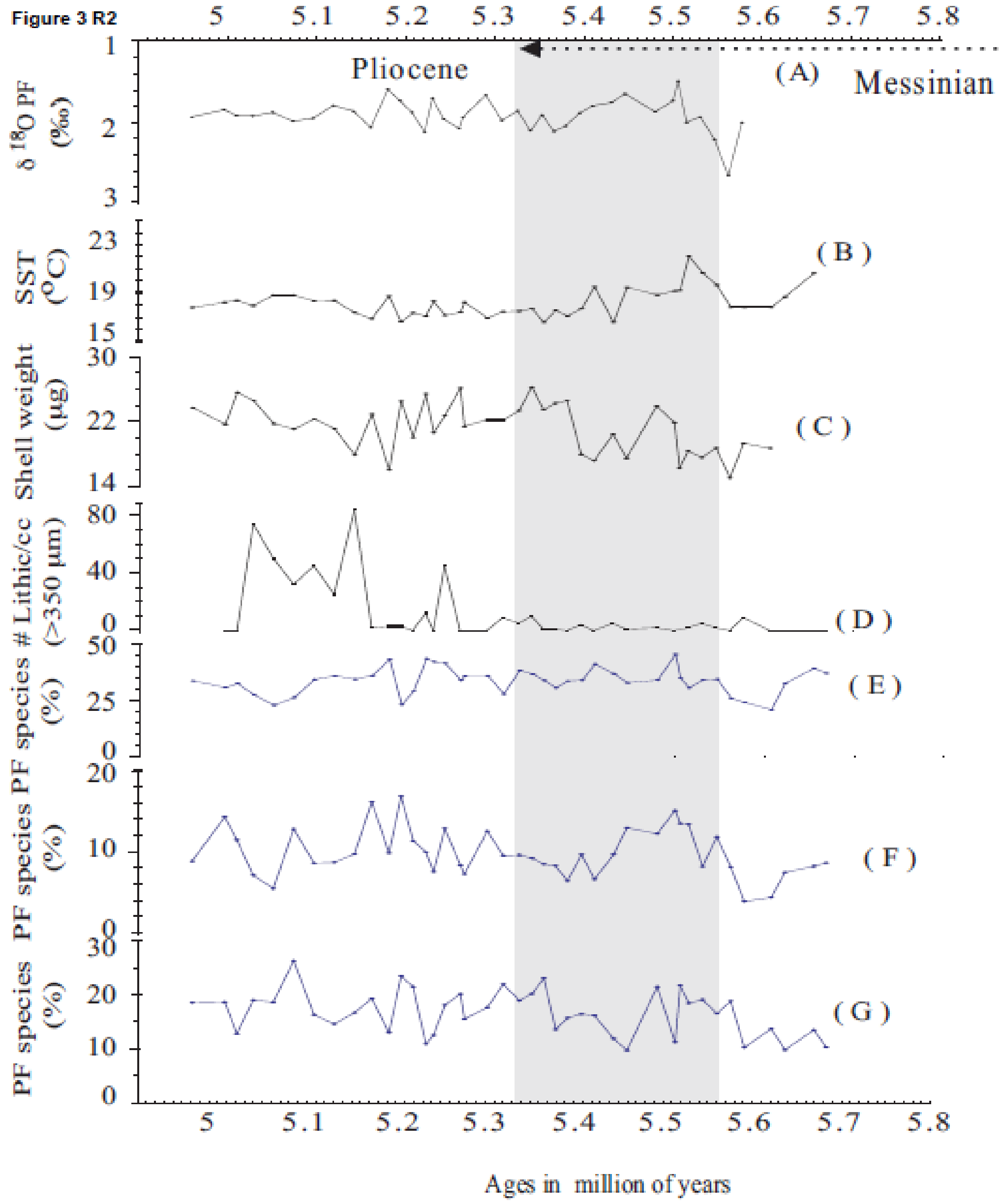

Figure. 3

figure3