Variations in Mediterranean–Atlantic exchange across the late Pliocene climate transition

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Abstract. Mediterranean–Atlantic exchange through the Strait of Gibraltar plays a significant role in the global ocean–climate dynamics in two ways. On one side, the injection of the saline and warm Mediterranean Outflow Water (MOW) contributes to North Atlantic deep-water formation. In return, the Atlantic inflow is considered a sink of less saline water for the North Atlantic Ocean. However, while the history of MOW is the focus of numerous studies, the Pliocene Atlantic inflow has received little attention so far. The present study provides an assessment of the Mediterranean–Atlantic exchange with a focus on the Atlantic inflow strength and its response to regional and global climate from 3.33 to 2.60 Ma. This time interval comprises the mid-Pliocene warm period (MPWP; 3.29–2.97 Ma) and the onset of the Northern Hemisphere glaciation (NHG). For this purpose, gradients in surface δ18O records of the planktonic foraminifer Globigerinoides ruber between the Integrated Ocean Drilling Program (IODP) Hole U1389E (Gulf of Cádiz) and Ocean Drilling Program (ODP) Hole 978A (Alboran Sea) have been evaluated. Interglacial stages and warm glacials of the MPWP revealed steep and reversed (relative to the present) W–E δ18O gradients suggesting a weakening of Mediterranean–Atlantic exchange likely caused by high levels of relative humidity in the Mediterranean region. In contrast, periods of stronger inflow are indicated by flat δ18O gradients due to more intense arid conditions during the severe glacial Marine Isotope Stage (MIS) M2 and the initiation of NHG (MIS G22, G14, G6–104). Intensified Mediterranean–Atlantic exchange in cold periods is linked to the occurrence of iceberg-rafted debris (IRD) at low latitudes and a weakening of the Atlantic Meridional Overturning Circulation (AMOC). Our results thus suggest the development of a negative feedback between AMOC and exchange rates at the Strait of Gibraltar in the latest Pliocene as it has been proposed for the late Quaternary.

1 Introduction

Mediterranean–Atlantic water mass exchange through the Strait of Gibraltar is driven by a two-directional current system in which the westward outflow of warm and saline Mediterranean Outflow Water (MOW) at the bottom, driven by excess evaporation in the Mediterranean, is compensated for by the inflow of colder and less saline North Atlantic Central Water (NACW) at the surface (Bormans et al., 1986; Ochoa and Bray, 1991; Vargas-Yáñez et al., 2002). This exchange plays an important role for regional and global climate in two respects. First, the injection of warm and salty MOW into the North Atlantic contributes to deep-water formation at high latitudes and thus to the dynamics of the ocean–climate system (Bryden and Kinder, 1991; Ivanovic et al., 2014; Reid, 1979; Rogerson et al., 2010). Periods of stronger MOW correspond to weak phases of Atlantic Meridional Overturning Circulation (AMOC) resulting in a negative feedback between exchange and AMOC recognized during Heinrich Stadials in the late Quaternary (Rogerson et al., 2010, 2012). Previous research has focused strongly on MOW variability since the Pliocene (e.g., Bryden and Stommel, 1982; Hernández-Molina et al., 2014; Iorga and Lozier, 1999;
Kaboth et al., 2016; Voelker et al., 2006) but has neglected the Atlantic surface water component (Rogerson et al., 2010). Furthermore, there are a number of studies on Mediterranean–Atlantic exchange during the early–mid Pliocene and early Pleistocene (e.g., Bahr et al., 2015; García-Gallardo et al., 2017; Grunert et al., 2017; Hernández-Molina et al., 2014; Kaboth et al., 2017; Khélifi et al., 2009, 2014; Van der Schee et al., 2016; Voelker et al., 2015), leaving a gap in our knowledge about the late Pliocene climate transition comprising the mid-Pliocene warm period (MPWP) and the initiation of the Northern Hemisphere glaciation (NHG).

The MPWP (3.29–2.97 Ma) has been identified by the United States Geological Survey’s PRJSM (Pliocene Research, Interpretation and Synoptic Mapping; Dowsett et al., 2010) group as a potential analogue for the future of global climate due to remarkable similarities with model predictions (Dowsett et al., 2012; Robinson et al., 2008). These include increased levels of greenhouse gases up to 350–450 ppmv, global temperatures increased by 1–5 °C relative to today, and increased annual precipitation and elevated sea level as projected from data and models (Budyko et al., 1985; Faquetter et al., 1998; Haywood and Valdes, 2004; Lunt et al., 2012; Pagani et al., 2010; Raymo et al., 1996; Seki et al., 2010). At ~3.3 Ma, a severe ice sheet expansion initiated glacial Marine Isotopic Stage (MIS) M2, a precursor for the initiation of NHG at 2.95 Ma (Bartoli et al., 2006). The strong glacial at MIS M2 has been considered the first severe cold period before NHG, in which long-term intensification of Mediterranean–Atlantic exchange occurred (Khélifi et al., 2014; Sarnthein et al., 2017). The present study aims to reconstruct variations in Mediterranean–Atlantic exchange with a strong focus on the Atlantic inflow throughout the MPWP and the onset of NHG. The Integrated Ocean Drilling Program (IODP) Hole U1389E has been used as the reference location for the Gulf of Cádiz and Ocean Drilling Program (ODP) Hole 978A for the Alboran Sea (Fig. 1). Planktonic δ18O records obtained from both sites were used for the reconstruction of isotopic gradients as indicators of Atlantic inflow strength across the late Pliocene climate transition.

2 Regional setting

2.1 Alboran Sea – ODP Hole 978A

ODP Hole 978A is located in the Alboran Sea north of the Al-Mansour Seamount (36°13′N, 02°03′W; Fig. 1) at 1930 m water depth (Comas et al., 1996). Circulation in the Alboran Sea is driven by three water mass layers. At the surface (0~200 m), inflowing Atlantic water enters the Mediterranean basin (Millot, 1999). On its way, it mixes with upwelled MOW within the Strait of Gibraltar (Folkard et al., 1997) and with surface waters of the Alboran Sea, creating the Modified Atlantic Water with temperatures of 15–16°C and a salinity of 36.5 (Millot, 1999), which follows two anticyclonic gyres, the Western Alboran Gyre (WAG) and the Eastern Alboran Gyre (EAG; Fig. 1) (Gascard and Richez, 1985; Vargas-Yáñez et al., 2002). The northern Alboran Sea is affected by upwelling along the Spanish coast providing nutrients and enhanced primary productivity (Minas et al., 1991; Peeters et al., 2002; Sarhan et al., 2000). The bottom layer is represented by the Western Mediterranean Deep Water (WMDW; water temperature ~13°C, salinity ~38.5) below 1000 m, formed due to overturning in the Gulf of Lions (Bryden et al., 1994; Hernández-Molina et al., 2006; Millot, 1999; Rhein, 1995). The intermediate layer found between ~200 and 1000 m is composed of the salty (up to 39.1) and warm (14.7 to 17°C) Levantine Intermediate Water (LIW) which originates from overturning in the Eastern Mediterranean (Fig. 1; Millot, 2013; Wüst, 1961). Finally, WMDW, together with LIW, exits the Mediterranean basin through the Strait of Gibraltar and forms MOW in the Gulf of Cádiz (Bryden et al., 1994).

2.2 Gulf of Cádiz – IODP Hole U1389E

IODP Hole U1389E is located in the northern Gulf of Cádiz (36°25.515′N, 07°16.683′W; Fig. 1) at 644 m water depth under direct influence of MOW (Fig. 1). It constitutes a key site for the recovery of an upper Pliocene contourite succession (Stow et al., 2013). Once LIW and WMDW exit the Strait of Gibraltar and form MOW, this water mass splits into two plumes due to the complex morphology of the continental slope in the Gulf of Cádiz. The upper plume flows between 500 and 800 m while the lower plume flows between 800 and 1400 m (Ambar and Howe, 1979; Borenás et al., 2002; García et al., 2009; Llave et al., 2007; Madelain, 1970; Marchès et al., 2007; Serra et al., 2005; Zenk, 1975). Surface circulation in the Gulf of Cádiz is governed by the Gulf of Cádiz slope current (Fig. 1), flowing eastward along the western Iberian margin, and the offshore inflow, which meet at the Strait of Gibraltar and enter the Mediterranean basin (Peliz et al., 2009).

3 Material and methods

3.1 Sample material and data collection

This study relies on published and newly acquired stable oxygen isotope (δ18O) records of the planktonic foraminifer Globigerinoides ruber from upper Pliocene (3.33–2.60 Ma) sediments at ODP Hole 978A (Alboran Sea), IODP Hole U1389E (Gulf of Cádiz), and the Rossello section in Sicily (Fig. 1). Sediment cores from ODP 978A were recovered during ODP Leg 161 (Comas et al., 1996). Khélifi et al. (2014) have established a δ18O record from Gs. ruber for this site which ranges from 3.60 to 2.70 Ma (cores 26R–20R, 438.38–371.00 m b.s.f.). For the purpose of our study, this record was extended to 2.60 Ma through analyses of
new samples collected every 50 cm from core sections 19R-5 through 16R-4 at the Bremen Core Repository.

Sediment cores from IODP Hole U1389E were obtained during IODP Expedition 339 (Stow et al., 2013). A δ18O record of *G. ruber* ranging from 3.70 to 2.60 Ma (cores 70R–41R, 982.78–703.62 m b.s.f.) has been established by Grunert et al. (2017) and adopted for this study.

The stacked δ18O record of Lourens et al. (1992, 1996) from the Rossello outcrops in Sicily has been adopted for stratigraphic calibration of the new δ18O record at ODP Hole 978A.

### 3.2 Stable isotope analysis

Details on laboratory protocols and isotopic analyses performed by Grunert et al. (2017), Khélifi et al. (2014), and Lourens et al. (1996) can be found in the respective publications. For the continuation of the ODP 978A record, 46 samples from core sections 19R-5 to 17R-1 were analyzed every 50 cm. Sediment samples were dried, weighed, washed through sieves of 250 and 63 µm, and dried. Whenever possible, 10 to 20 well-preserved specimens of *G. ruber* > 250 µm were picked for isotopic analysis. Shells were crushed, cleaned in an H2O dest : methanol (2 : 1) mixture, and bathed ultrasonically for 1 min. Clean shells were reacted with 100% phosphoric acid at 70°C using a Gasbench II connected to a ThermoFisher Delta V Plus mass spectrometer at the GeoZentrum Nordbayern (Erlangen). All values are reported in per mill relative to Vienna Pee Dee Belemnite (VPDB). Reproducibility (±1σ) and accuracy were monitored by replicate analysis of laboratory standards calibrated by assigning δ13C values of +1.95‰ to NBS19 limestone and −46.6‰ to LSVEC lithium carbonate and δ18O values of −2.20‰ to NBS19 and −23.2‰ to NBS18.

### 3.3 Age model

Age constraints for ODP Hole 978A are established from 3.6 to 2.8 Ma by Khélifi et al. (2014), and those for IODP Hole U1389E are reported in Grunert et al. (2017). The latter study establishes the correlation of the δ18O record of IODP U1389E with ODP 978A from Khélifi et al. (2014). However, in the upper part (< 2.75 Ma), IODP U1389E is correlated with the Rossello section because the ODP 978A record ends there. To ensure comparability, the newer δ18O record obtained in this study for the upper ODP 978A has been visually correlated with the Rossello section. Biostratigraphic events have been adopted from Comas et al. (1996) for ODP 978A and from Lourens et al. (1996) for the Rossello section to guarantee a precise correlation (Fig. 2a; Table 1).

The last occurrence (LO) of the calcareous nannofossil *Discoaster tamalis* (2.78 Ma) delimits ODP 978 at the bottom of the studied interval (Fig. 2), while the top is demarcated by the LO of *D. pentaradiatus* (2.52 Ma; Comas et al., 1996). In the Rossello section, the bottom of the studied interval includes the first occurrence (FO) of the planktonic foraminiferal *Neogloboquadrina atlantica* (sin.) (2.72 Ma), while the top is delimited by the LO of the calcareous nannofossils *D. pentaradiatus* and *D. surculus* (2.51 and 2.55 Ma, respectively; Lourens et al., 1996). Marine Isotopic Stages (MISs) have

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**Figure 1.** Location of IODP Site U1389, ODP Site 978, and the Rossello section. AI: Atlantic Inflow; EAG: Eastern Atlantic Gyre; GoCSC: Gulf of Cádiz Slope Current; MAW: Modified Atlantic Water; MOW: Mediterranean Outflow Water; WAG: Western Atlantic Gyre. The map was generated with GeoMapApp (http://www.geomapapp.org), using the default basemap, Global Multi-Resolution Topography (GMRT) Synthesis (Ryan et al., 2009).
Figure 2. (a) Stratigraphic framework established for the ODP Hole 978A record based on indicated biostratigraphic tie points (FO: first occurrence; LO: last occurrence) and visual correlation of the δ¹⁸O record with the Rossello section (Lourens et al., 1996). (b) Calculated sedimentation rates.

Table 1. Biostratigraphic events of ODP Hole 978A (Comas et al., 1996) and the Rossello section (Lourens et al., 1996) used for the establishment of age constraints of the upper ODP 978A (Fig. 2a). FO: first occurrence; LO: last occurrence.

<table>
<thead>
<tr>
<th>Biostratigraphic event</th>
<th>Position sedimentary cycles (m)</th>
<th>Age (Ma)</th>
</tr>
</thead>
<tbody>
<tr>
<td>ODP Hole 978A</td>
<td></td>
<td></td>
</tr>
<tr>
<td>LO D. pentaradiatus</td>
<td>323.74</td>
<td>2.52</td>
</tr>
<tr>
<td>LO D. tamalis</td>
<td>372.70</td>
<td>2.78</td>
</tr>
<tr>
<td>Rossello section</td>
<td></td>
<td></td>
</tr>
<tr>
<td>LO D. pentaradiatus</td>
<td>138.05</td>
<td>2.51</td>
</tr>
<tr>
<td>LO D. surculus</td>
<td>135.05</td>
<td>2.55</td>
</tr>
<tr>
<td>FO N. atlantica (sin.)</td>
<td>83.53</td>
<td>2.72</td>
</tr>
</tbody>
</table>

been identified by visual correlation with the Rossello section (Lourens et al., 1996; Figs. 3, 4a), which was in turn correlated with the LR04 benthic stack (Lisiecki and Raymo, 2005).

3.4 Glacial–interglacial δ¹⁸O gradients across the Strait of Gibraltar

A compilation of the late Pliocene (3.33–2.60 Ma) δ¹⁸O records from the two studied sites is shown in Fig. 3 (Supplement, Table S1). Based on the average δ¹⁸O values of glacial–interglacial MISs (see Supplement, Table S2), gradients between the corresponding stages in the Gulf of Cádiz
Figure 3. $\delta^{18}O$ records of IODP Hole U1389E and ODP Hole 978A from 3.33 to 2.60 Ma. Intervals I, II, and III with continuous $\delta^{18}O$ records from both cores are indicated. Marine Isotopic Stages (MISs) have been identified from Lourens et al. (1996) and Khélifi et al. (2014).

4 Results

4.1 Age model for the new $\delta^{18}O$ record of ODP 978A (369–337 m b.s.f.)

The new $\delta^{18}O$ record from the upper part of ODP 978A is visually correlated with the Rossello section relying on established biostratigraphic tie points (Fig. 2a; Table 1; Gradstein et al., 2012). As the upper part of both records is clearly delimited by the LO of $D$. pentaradiatus (2.52 Ma) and the lower part by the LO of $D$. tamalis (2.78 Ma) at ODP Site 978 and the FO of the foraminifer $N$. atlantica (2.72 Ma) in the Rossello section, the first 0.06 Myr in ODP 978A stays out of visual correlation with the Rossello section. ODP 978A ranges from 2.78 to 2.52 Ma, which corresponds to 2.6 Myr in around 50 m of sediment core, therefore equivalent to 0.05 Myr/10 m. Accordingly, the visual correlation should start around 10 m upward of the LO of $D$. tamalis (2.72 Ma) in the Rossello section, the first 0.06 Myr in ODP 978A and the FO of the foraminifer $N$. atlantica and the lower part by the LO of $D$. tamalis (2.78 Ma) at ODP Site 978 (Fig. 2a). From that point (∼362 m b.s.f.), a similar number of $\delta^{18}O$ cycles can be observed in both records, resulting in the best solution, observed in Fig. 2a. Our age model for ODP 978A indicates that the Pliocene–Pleistocene boundary at ∼2.58 Ma is located at ∼340 m b.s.f. (Fig. 2a). Sedimentation rates calculated for this interval vary from 0.06 to 0.57 m kyr$^{-1}$ (mean: 0.26 m kyr$^{-1}$; Fig. 2b).

4.2 Glacial/interglacial $\delta^{18}O$ gradients

$\delta^{18}O$ values for intervals I to III are shown vs. longitude in Fig. 3. IODP Hole U1389E suffers from notable gaps throughout the studied interval due to poor core recovery (Fig. 3; Stow et al., 2013; Grunert et al., 2017). For this reason, the section is subdivided into three well-recovered intervals with continuous $\delta^{18}O$ records, which will be the focus of our study (Interval I: 3.33–3.27 Ma; Interval II: 3.02–2.88 Ma; Interval III: 2.73–2.60 Ma; Figs. 3, 4a).

In Interval I (3.33–3.27 Ma), mean $\delta^{18}O$ values of IODP U1389E (min.: −1.22‰; max.: +0.46‰; mean: −0.34‰) and ODP 978A (min.: −1.34‰; max.: +0.49‰; mean: −0.39‰) are close to each other (Fig. 4a1; Table 2). $\delta^{18}O$ gradients between ODP 978A and IODP U1389E change direction from interglacial MIS MG1 to glacial MIS M2 (Figs. 4a1, b1). $\delta^{18}O$ data from interglacial MIS MG1 reveals higher values in the Gulf of Cádiz compared to the Alboran Sea resulting in an isotopic gradient of −0.08‰ degree$^{-1}$. In contrast, glacial period MIS M2 shows an opposite isotopic gradient ranging from +0.05 to +0.09‰ degree$^{-1}$ for the two intermittent $\delta^{18}O$ maxima M2.1 and M2.2, respectively (Fig. 4b1).

In Interval II (3.02–2.88 Ma), mean $\delta^{18}O$ values of IODP U1389E (min.: −0.99‰; max.: +0.98‰; mean: −0.16‰) are considerably heavier compared to ODP 978A (min.: −1.64‰; max.: +0.50‰; mean: −0.50‰) (Fig. 4a2; Table 2). The records are particularly well separated during $\delta^{18}O$ minima, whereas they converge during maxima (Fig. 4a2). Interglacial periods MIS G21, G19, and G15 and glacial periods MIS G22, G20, and G16 show isotopic gradients ranging from −0.06 to −0.13‰ degree$^{-1}$ and −0.04 to −0.12‰ degree$^{-1}$, respectively (Fig. 4b2). The only exception occurs during glacial period MIS G14, which shows a positive and relatively flat gradient of +0.03‰ degree$^{-1}$.

In Interval III (2.73–2.60 Ma), mean $\delta^{18}O$ values are again closer to each other (IODP U1389E – min.: −0.85‰; max.: +0.70‰; mean: −0.05‰; ODP 978A – min.: −1.11‰; max.: +0.54‰; mean: −0.20‰) (Fig. 4a3, Table 2). A wide range of $\delta^{18}O$ gradients has been calculated for Interval III (Fig. 4b3). Interglacial periods MIS G7, G5, and G3 as well as glacial period MIS G4 show negative $\delta^{18}O$ gradients between −0.03 and −0.10‰ degree$^{-1}$). In contrast, glacial periods MIS G6, G2, and 104 reveal positive $\delta^{18}O$ gradients ranging from +0.02 to +0.06‰ degree$^{-1}$ (Fig. 4b3).
Figure 4. (a) Details of δ¹⁸O records of IODP Hole U1389E and ODP Hole 978A for intervals I–III. Periods of lowered NADW and increased influx of IRD from DSDP Site 607 (Mid-Atlantic Ridge) reported in Kleiven et al. (2002) are indicated. (b) Calculated δ¹⁸O gradients between the Gulf of Cádiz and the Alboran Sea for glacial and interglacial stages in intervals I–III (see Supplement, Table S2).

Table 2. Minimum, maximum, mean, and standard deviation of δ¹⁸O values of IODP Hole U1389E and ODP Hole 978A in the long term and within intervals I, II, and III.

<table>
<thead>
<tr>
<th>Period</th>
<th>Long term (3.33–2.60 Ma)</th>
<th>Interval I (2.73–2.60 Ma)</th>
<th>Interval II (3.02–2.88 Ma)</th>
<th>Interval III (3.33–3.27 Ma)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Site ID</td>
<td>IODP Hole U1389E</td>
<td>ODP Hole 978A</td>
<td>IODP Hole U1389E</td>
<td>ODP Hole 978A</td>
</tr>
<tr>
<td>Min.</td>
<td>−1.22</td>
<td>−1.81</td>
<td>−1.22</td>
<td>−1.34</td>
</tr>
<tr>
<td>Max.</td>
<td>0.98</td>
<td>0.68</td>
<td>0.46</td>
<td>0.49</td>
</tr>
<tr>
<td>Average</td>
<td>−0.14</td>
<td>−0.41</td>
<td>−0.34</td>
<td>−0.39</td>
</tr>
<tr>
<td>SD</td>
<td>0.37</td>
<td>0.51</td>
<td>0.38</td>
<td>0.52</td>
</tr>
</tbody>
</table>

5 Discussion

5.1 Direction of δ¹⁸O gradients

Our late Pliocene data set is based on the planktonic Gs. ruber, yet recent δ¹⁸O gradients between the Gulf of Cádiz and the Alboran Sea have been established from the planktonic Globigerina bulloides (e.g., Cacho et al., 2001; Rogerson et al., 2010). A seasonal δ¹⁸O offset between Gs. ruber (blooming in spring and summer) and G. bulloides (fall–winter) is evident from core top and parallel down-core records of δ¹⁸O in the late Quaternary (Fig. 5a, b; Voelker et al., 2009). However, despite this offset, the δ¹⁸O gradients obtained from both species (G. bulloides: Rogerson et al., 2010; Gs. ruber: Rohling, 1999, and Salgueiro et al., 2008) show the same gra-
dient direction with lighter $\delta^{18}$O values in the Gulf of Cádiz and heavier values in the Alboran Sea (Fig. 5b). Previous studies further show that the direction of this W–E gradient has not changed over the last ∼25 000 years (Cacho et al., 2001; Rogerson et al., 2010). For the late Pliocene, however, our data suggest that the $\delta^{18}$O gradient was considerably more variable, particularly during glacial stages (Fig. 4a, b). While all studied interglacial stages and glacial stages G22–G16 (except G14) of Interval II show a reversed gradient with respect to the present, the strong glacialsd M2, G14, G6, G2, and 104 show a gradient in line with present-day observations (Figs. 4b, 5b).

Seasonal variations in Atlantic inflow reported in previous studies (e.g., Bormans et al., 1986; Ovchinnikov, 1974; Parada and Cantón, 1998; Vargas-Yáñez et al., 2002) can be the cause of sea surface salinity (SSS) and sea surface temperature (SST) variability. While seasonal changes in SSS are <0.5, SST variability is more prominent and may result in brief temporary reversals of SST gradients (MEDATLAS, 2002; Rogerson et al., 2010). The Alboran Sea shows colder temperatures than the Gulf of Cádiz during all seasons due to upwelling, with occasional exceptions in summer under the influence of easterly winds (Bakun and Agostini, 2001; Folkard et al., 1997; Peeters et al., 2002; Rogerson et al., 2010; Sarhan et al., 2000; Shaltout and Omstedt, 2014). The $\delta^{18}$O composition of present-day seawater is thus considered to be largely determined by changes in SST (Rogerson et al., 2010).

Reversed gradients in the late Pliocene could imply a different SST gradient due to a different current regime and/or lack of upwelling in the Alboran Sea in the Pliocene. Unfortunately, there is little data on SST available from the late Pliocene which would allow further evaluation. Khélifi et al. (2014) provide an alkenone-based SST record from ODP 978A from ∼26 to 27°C (Fig. 5b). The comparison thus suggests that the δ18O gradient between both basins. For comparison, the gradient obtained from the δ18O composition of G. bulloides by Rogerson et al. (2010) has been added.

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Figure 5. (a) Comparison of Pleistocene δ18O values of Globigerina bulloides and Globigerinoides ruber at sites MD99-2336 and -2339 in the Gulf of Cádiz (Voelker et al., 2009). (b) Recent δ18O gradients of G. ruber between core top samples from the Gulf of Cádiz (M39022-1; Salgueiro et al., 2008) and the Alboran Sea (stations KS82-30 and KS-82-31; Rohling, 1999). Red circles represent mean values of the respective data sets; the red line denotes the gradient between both basins. For comparison, the gradient obtained from the δ18O composition of G. bulloides by Rogerson et al. (2010) has been added.

A warmer and more humid climate implies higher runoff and freshening of the Mediterranean surface waters leading to depleted δ18O records as reported from Pliocene to Holocene data and models in previous studies (e.g., Gudjonsson and van der Zwaan, 1985; Kaboth et al.,
We thus consider the $\delta^{18}O$ depletion of Mediterranean waters during a warm and humid paleoclimate as the most likely explanation for the reversed gradients observed for interglacial stages from all three studied intervals as well as for the comparably warm glacial periods MIS G22, G20, G16, and G4 (Fig. 4b). Conversely, arid conditions are indicated by pollen data only for the strong glacial stages M2, G6, G2, and 104, which herald increasing Northern Hemisphere glaciation and for which our data suggest $\delta^{18}O$ gradients similar to the present (Fauquette et al., 1998; Figs. 4b, 5b).

5.2 Steepness of $\delta^{18}O$ gradients

While humidity likely explains variations in normal and reversed $\delta^{18}O$ gradients in the late Pliocene relative to the present-day trend, slope steepness is considered sensitive to the strength of surface water exchange across the Strait of Gibraltar (Rohling, 1999; Rogerson et al., 2010). On the basis of the previous statement, flat gradients are indicative of well-connected basins and enhanced exchange, whereas steeper gradients suggest more restricted conditions and reduced exchange (Rogerson et al., 2010).

The steepness of present-day gradients varies from 0.05 to 0.13‰ degree$^{-1}$ depending on the time of calcification of *Gs. ruber* and *G. bulloides* (Fig. 5b; this study; Rogerson et al., 2010; see chap. 5.1). In addition, Rogerson et al. (2010) reported a gradient of 0.26‰ degree$^{-1}$ during the Last Glacial Maximum and of 0.01 and 0.14‰ degree$^{-1}$ during Heinrich Stadials 1 and 2, respectively. Regardless of the direction, the steepness of late Pliocene slopes ranges from 0.05 to 0.13‰ degree$^{-1}$ in all interglacial stages (except MIS G5) and glacial stages M2, G20, G16, and 104 (Fig. 4b), thus indicating exchange rates which are similar to or reduced compared to present-day values. In contrast, fluctuating gradients during glacial stages MIS G22, G14, G6, and G2 (0.02 to 0.04‰ degree$^{-1}$; Fig. 4b, b3) point to enhanced exchange during these cold periods.

Support for our interpretation comes from available studies on MOW development during the late Pliocene, to which the strength of AMOC is directly linked. The onset of the long-term intensification of MOW has been linked to arid conditions during glacial stage M2, a second pulse of MOW occurs at ~2.8–2.7 (Grunert et al., 2017; Hernández-Molina et al., 2014; Khéïfi et al., 2009, 2014; Sarthhein et al., 2017). Both pulses occur during these glacial periods in which we observe gradients similar to today, and their succession is reflected in a flattening of gradients with the onset of NHG (~2.9 Ma) due to arid and cold conditions. Fluctuations of MOW intensity in the late Pliocene are in turn linked to global climate processes. Kleiven et al. (2002) demonstrated that lowered values of benthic $\delta^{13}C$ in the North Atlantic during glacial stages (MIS M2, G22, G16, and G6–104 of our study) were related to a decreased production of North Atlantic Deep Water (NADW; Fig. 4a) at high latitudes. Furthermore, periods of reduced Atlantic Meridional Overturning Circulation are linked to a freshening of surface waters in the North Atlantic due to the increased iceberg melting and influx of ice-rafted debris (IRD) (Kleiven et al., 2002). Despite a strong influx of IRD being recorded at northern latitudes during MIS M2, the first occurrence of IRD at low midlatitudes is recorded during G14 in samples from Deep Sea Drilling Program (DSDP) Site 607, located on the western flank of the Mid-Atlantic Ridge, with following peaks during glacial G6–104 (Bailey et al., 2010; Kleiven et al., 2002; Fig. 4a). Thus, enhanced Mediterranean–Atlantic exchange suggested for MIS M2 parallels a drop in NADW production and weakened AMOC (DeSchepper et al., 2009; Khéïfi et al., 2014; Sarthhein et al., 2017). Continuous occurrences of IRD at low midlatitudes and reduced AMOC during glacial periods G14 to 104 fall together with intensified Mediterranean–Atlantic exchange during glacial periods at the onset of NHG. In contrast, the lack of IRD and increased $\delta^{13}C$ observed during interglacials (Kleiven et al., 2002) is in accordance with more restricted exchange during MIS MG1, G21, G19, G15, G5, and G3 (Fig. 4a). The observed negative feedback between AMOC and Mediterranean–Atlantic exchange at the onset of NHG seems to work similarly as during Heinrich Stadials in the Holocene as reported by Rogerson et al. (2010). However, a higher resolution of our Pliocene records would be necessary to establish an accurate assessment of the timing of these feedback mechanisms within the glacial periods.

6 Conclusions

Late Pliocene $\delta^{18}O$ gradients of the planktonic foraminifer *Globigerinoides ruber* from IODP Hole U1389E (Gulf of Cádiz) and ODP Hole 978A (Alboran Sea) allowed the reconstruction of Mediterranean–Atlantic exchange variations between 3.33 and 2.60 Ma, spanning the transition from the Mid-Pliocene Warm Period (MPWP) into Northern Hemisphere glaciation (NHG). The $\delta^{18}O$ gradients across the Strait of Gibraltar have been analyzed for individual glacial and interglacial stages in terms of direction and steepness.

In contrast to positive gradients in the present-day, elevated levels of humidity during the MPWP caused reversed and steep $\delta^{18}O$ gradients in the late Pliocene, especially during interglacial stages. Increased aridity caused a shift to positive gradients during strong glacial periods at Marine Isotope Stage (MIS) M2 and the onset of NHG (MIS G22, G14, G2–104). Flat slopes indicate enhanced inflow during those cold and arid periods.

Intensified Mediterranean Outflow Water (MOW) has been reported during M2 and from 2.8 Ma coinciding with intense glacial stages. Strengthened Mediterranean–Atlantic exchange occurs at times of reduced Atlantic Meridional Overturning Circulation (AMOC) and North Atlantic Deep
Water Formation (NADW) formation, when a higher influx of IRD arrived at lower latitudes causing the freshening of Atlantic surface waters. Our results thus suggest a negative feedback between AMOC and exchange rates at the Strait of Gibraltar in the late Pliocene, as has been proposed for the late Quaternary.

**Data availability.** Data are available at https://doi.pangaea.de/10.1594/PANGAEA.887137 (García-Gallardo, 2018).

The Supplement related to this article is available online at https://doi.org/10.5194/cp-14-339-2018-supplement.

**Competing interests.** The authors declare that they have no conflict of interest.

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