



## Evolution of the Late Miocene Mediterranean–Atlantic gateways and their impact on regional and global environmental change

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### ARTICLE INFO

#### Article history:

Received 27 March 2015

Received in revised form 24 July 2015

Accepted 19 August 2015

Available online 24 August 2015

#### Keywords:

Marine gateways

Mediterranean

Atlantic

Climate

Messinian Salinity Crisis

### ABSTRACT

Marine gateways play a critical role in the exchange of water, heat, salt and nutrients between oceans and seas. As a result, changes in gateway geometry can significantly alter both the pattern of global ocean circulation and associated heat transport and climate, as well as having a profound impact on local environmental conditions. Mediterranean–Atlantic marine corridors that pre-date the modern Gibraltar Strait, closed during the Late Miocene and are now exposed on land in northern Morocco and southern Spain. The restriction and closure of these Miocene connections resulted in extreme salinity fluctuations in the Mediterranean, leading to the precipitation of thick evaporites. This event is known as the Messinian Salinity Crisis (MSC). The evolution and closure of the Mediterranean–Atlantic gateways are a critical control on the MSC, but at present the location, geometry and age of these gateways are still highly controversial, as is the impact of changing Mediterranean outflow on Northern Hemisphere circulation. Here, we present a comprehensive overview of the evolution of the Late Miocene gateways and the nature of Mediterranean–Atlantic exchange as deduced from published studies focussed both on the sediments preserved within the fossil corridors and inferences that can be derived from data in the adjacent basins. We also consider the possible impact of evolving exchange on both the Mediterranean and global climate and highlight the main enduring challenges for reconstructing past Mediterranean–Atlantic exchange.

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### 1. Introduction

During the late Tortonian (~11.6 to 7.2 Ma), several marine gateways through southern Spain, northern Morocco and potentially Gibraltar, connected the Mediterranean Sea with the Atlantic Ocean (Fig. 1).

Plate tectonic convergence between Africa and Iberia, combined with subduction dynamics in the Alborán region, progressively closed these connections during the Messinian (e.g. Duggen et al., 2003; Gutscher et al., 2002). This tectonic forcing combined with eustatic (e.g. Manzi et al., 2013) and climatic (Hilgen et al., 2007) factors resulted in a complex history of varied Mediterranean–Atlantic exchange and high amplitude environmental fluctuations in the Mediterranean including the formation of the world's most recent saline giant (Warren, 2010).

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Like other marginal basins, the Mediterranean's near-landlocked configuration makes it sensitive to subtle changes in climate (e.g. Thunell et al., 1988). Consequently, the first environmental responses to gradual restriction of exchange with the Atlantic recorded in the Mediterranean (e.g. faunal and isotopic changes; Fig. 2), predate any evaporite precipitation there by a million years or more. The most extreme palaeoenvironmental changes took place during the so-called Messinian Salinity Crisis (MSC; 5.97–5.33 Ma; Fig. 2; Table 1) when extensive gypsum deposits precipitated in the Mediterranean's marginal basins and kilometre thick halite units formed in the deep basins (e.g. Hsü et al., 1973; Ryan et al., 1973). This was followed by a period during which the sediments recorded highly fluctuating conditions varying from brackish to hypersaline, before returning, in the Early Pliocene, to open marine conditions (Fig. 2; Hsü et al., 1972). These Late Miocene low salinity intervals, known as the Lago Mare, may be the product of an additional freshwater source supplied to the Mediterranean from Paratethys, the lacustrine precursor to the Black and Caspian seas. Like other major freshwater sources, this is a key component of the Mediterranean's freshwater budget, which combined with the gateway dimensions determine its salinity.

The large volume of salt preserved in the Mediterranean necessitates that one or more marine connections with the open ocean remained, at least until the end of the halite stage (5.55 Ma; Krijgsman and Meijer, 2008). However, the location of the last gateway(s) remains highly ambiguous. Field studies of the sedimentary basins in southern Spain (the Betic Corridor) and northern Morocco (the Rifian Corridor) thought to be part of the corridor network (Fig. 1), typically indicate that these areas were closed to marine exchange well before the MSC (e.g. Betzler et al., 2006; Ivanović et al., 2013a; Krijgsman et al., 1999b; Soria et al., 1999; van Assen et al., 2006), while the Gibraltar Strait is thought to have first opened at the beginning of the Pliocene (5.33 Ma) bringing the MSC to an end (e.g. Blanc, 2002; Garcia-Castellanos et al., 2009; Hsü et al., 1973, 1977). The key problem is that it is extremely difficult to pinpoint the exact location or timing of closure from field data alone, because the sedimentary successions within the corridors have been uplifted and eroded (e.g. Hüsing et al., 2010). Using other datasets to identify the location of each marine corridor, reconstructing its geometry and reducing uncertainty in the age of closure is therefore critical for constraining the process-response chain linking gateway evolution with the development of the Mediterranean's MSC succession. The Atlantic response to a change in gateway configuration is reliant on changes to the density and volume of Mediterranean outflow and consequently also depends on an ability to reconstruct gateway dimensions and the patterns of exchange.

Several indirect approaches to the study of gateway evolution have been employed in the context of the MSC. These include for example, physics-based mathematical models that quantify gateway configuration (e.g. Meijer, 2006; Meijer and Krijgsman, 2005); and the use of isotopic proxies to elucidate changing connectivity (e.g. Flecker and Ellam, 1999; Ivanović et al., 2013a; Topper et al., 2011). Many other techniques have been applied to successions within the main Mediterranean basin rather than the gateway and the palaeoenvironmental information they provide can be compared with similar data on the Atlantic side of the connection to constrain aspects of exchange. In this paper we review, synthesise and integrate both direct and indirect data relating to the Mediterranean–Atlantic gateways during Late Miocene. Our aim is to use this information to reconstruct exchange before, during and after the MSC, consider the regional and global implications and highlight enduring questions that may be amenable to new research methods.

## 2. Background information

### 2.1. Gateway control on Mediterranean water properties

At present, the Mediterranean Sea loses more water to the atmosphere by evaporation than it receives from rainfall and river runoff.

As a result, its surface waters are subject to an increase in salinity and thus in density. This relatively dense water finds its way to the deeper levels of the Mediterranean. In addition, the Mediterranean is also a major heat sink for the Atlantic (the temperature of Atlantic inflow is ~16 °C, while Mediterranean outflow is ~12 °C; Rogerson et al., 2012). Consequently, in the present-day gateway to the Atlantic, the Strait of Gibraltar, dense Mediterranean water is juxtaposed against lower density Atlantic water (Fig. 3). The resulting pressure gradient drives an outflow to the Atlantic Ocean which occupies, roughly speaking, the lower half of the 300 m deep and 13 km wide strait. Above the outflow, an inflow of Atlantic water occurs, pushed eastwards by the sea surface sloping down in that direction (in its turn a direct response of outflow). For a correct understanding of the working of this gateway-landlocked basin system, it is important to realise that the in- and outflow are much larger than the net evaporation through the sea surface. Of the 0.8–1.8 Sv (1 Sv = 10<sup>6</sup> m<sup>3</sup>/s) worth of inflow, only about 0.05 Sv is “needed” to close the water budget of the Mediterranean Sea, while the rest flows out again. This description of what has been termed anti-estuarine exchange, relates to the situation averaged over several years. On a shorter (seasonal to tidal) time scale, significant deviations from this generalisation occur (Naranjo et al., 2014). The recent review by Schroeder et al. (2012) provides more background on the present-day strait-basin system, including specific references. Bryden and Stommel (1984) and Bryden and Kinder (1991) stand out from the vast literature on the two-way exchange at Gibraltar.

Today, Mediterranean salinity is about 38–39 ppt (1 ppt = 1 g of salt per kilogramme of water, roughly equivalent to 1 g per litre or 1 psu, practical salinity unit), which is higher than at least 90% of the world ocean. During the MSC, Mediterranean salinity attained more extreme levels, reaching gypsum saturation (~130 ppt) or even higher. As net evaporation during the Late Miocene was of a similar order as today (Gladstone et al., 2007), it is clear that during the MSC the dimensions of the Atlantic–Mediterranean ocean gateway must have been significantly different from those of the present Gibraltar Strait.

### 2.2. Geodynamic framework of the gateway region

The Mediterranean is the last remnant of a much larger Tethys Ocean that formed in the Jurassic with the breakup of Pangaea. Continental fragments rifted away from North Africa and collided with Eurasia to form the complex of Mesozoic to early Cenozoic basement that dominates the countries bordering the north Mediterranean coast. As late as the Early Miocene, there was still a marine corridor that linked the Mediterranean with the Indian Ocean (Hüsing et al., 2009b; Rogl, 1999). Closure of this eastern gateway fundamentally changed global oceanic circulation by shutting off the circum-equatorial current (Bryden and Kinder, 1991; Reid, 1979), changed salinity and temperature in both Mediterranean and Paratethys (Karami et al., 2011) and may have played a role in global Middle Miocene cooling (e.g. Flower and Kennett, 1993; Woodruff and Savin, 1989). In this broad context, the current corridor region of Gibraltar and adjacent Morocco and Spain has been the location of the western arm of the seaway and its link to the Atlantic since the inception of the Tethys Ocean.

The external thrust belt of the Gibraltar domain gives the Betic–Rif–Tell arc its characteristic horseshoe shape (Fig. 1). It consists of the Pre- and Subbetic of southern Spain, the Prerif in Morocco and the Tell Mountains of northwest Algeria (Fig. 1). These thrust belts are mostly the remnants of the Iberian and Nubian (African) passive margins that were deformed by thin-skinned folding and thrusting during the Miocene (Fig. 1 inset). In addition, the thrust pile involves deep marine sediments, the Flysch Nappes (Fig. 1), which are thought to be derived from a nearby oceanic basin. The external thrust belt was activated in the Flysch Nappes and Subbetics during the Burdigalian and has accumulated up to 8.5 km crustal material and sediments since then (Fullea et al., 2010). The Rifian corridors in northern Morocco are thought to have been over-thrust by the external thrust belt in



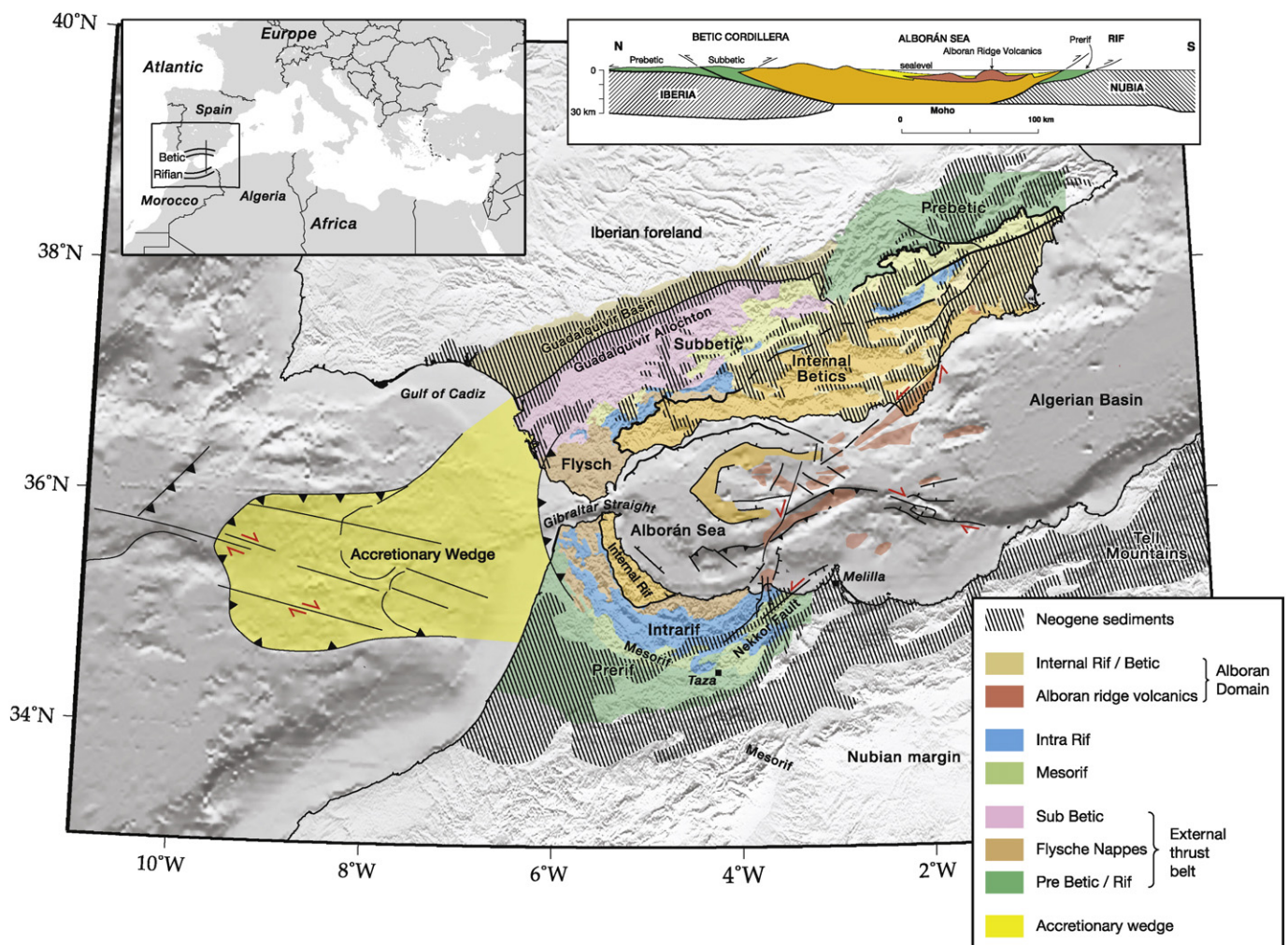
Morocco and the Tell Mountains while the accretionary wedge in the Gulf of Cadiz, which is largely continuous with the external thrust belt (Fig. 1), started building up in the Middle Miocene, initially as a result of westward over-thrusting by the Betic–Rifian belt on to the Atlantic margins (Medialdea et al., 2004).

The internal zone of the arc (Internal Betics of southern Spain and the Internal Rif in northern Morocco) over-thrusts Upper Cretaceous to Lower Miocene Flysch deposits of the Nubian and Iberian palaeomargins. This zone consists mainly of metamorphic rocks that were unroofed during Early Miocene extension that also affected the Alborán Basin (Comas et al., 1999). Remnants of the Guadalhorce gateway (see Section 3.1) are located in the Internal Betic of southern Spain (Martín et al., 2001). The internal zone and the Alborán Basin (Fig. 1) are commonly considered to be genetically linked and are referred to as the Alborán Domain. Here, Miocene extension came to an end in the Tortonian, and strike-slip and thrust faulting were initiated (Lonergan and White, 1997; Platt and Vissers, 1989). Flat-lying Pliocene–Recent sediments that cover the Cadiz accretionary wedge (Iribarren et al., 2007) attest to a significant change in the regional tectonics west of Gibraltar at the end of the Miocene.

One driver of the regional deformation has been the convergence of Nubia and Eurasia. Following closure of the Iberian–Eurasian suture in the Pyrenees around 20 Ma (Vissers and Meijer, 2012), continuing convergence was accommodated in the Betic–Rif–Tell region. However,

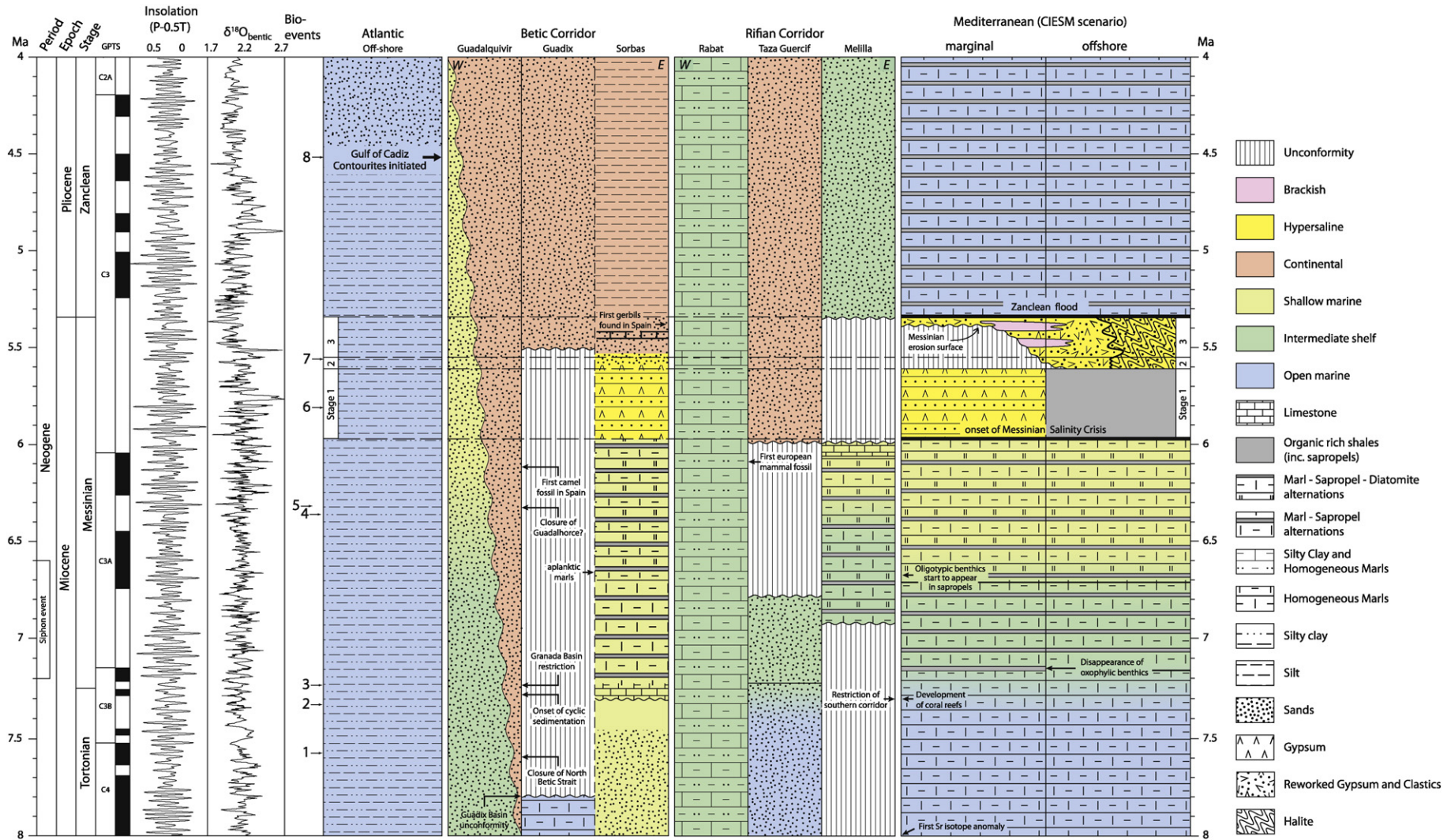
an additional tectonic driver is needed to explain the westward transport and extension of the Alborán Domain since the Miocene in this convergent setting (Dewey et al., 1989; Maldonado et al., 1999). The most likely driver is rollback of the east dipping Gibraltar slab (Gutscher et al., 2002; Lonergan and White, 1997) resulting from subduction that occurred mostly during the Miocene (see Gutscher et al., 2012; Platt et al., 2013 for recent reviews). New seismological observations support the idea of Duggen et al. (2003) that delamination of the lithospheric mantle and upwelling of asthenosphere beneath northwest Africa and southern Spain accompanied the rollback of the slab (Thurner et al., 2014).

Regional GPS velocities (Koulali et al., 2011) and seismicity (Stich et al., 2006; Zitellini et al., 2009) suggest that the active plate boundary aligns with the boundary between the Alborán Domain and exterior thrust belt in southern Spain (Fig. 1). In northern Morocco, the active boundary follows the outer boundary of the Pre-rif in the south (Toto et al., 2012) and the Nekkor fault in the east (Fig. 1). GPS observations also indicate that the west Alborán–Cadiz block currently moves independently to the SW relative to Africa at a rate of a few mm/yr, i.e. velocities deviate significantly from the relative velocity of stable Africa with respect to stable Eurasia (4.3–4.5 mm/yr; Fernandes et al., 2003). This suggests that, in addition to the convergence of Africa and Iberia there is another driver for these motions (Gutscher et al., 2012).



**Fig. 1.** The main structures and tectonic units of the Betic–Rif–Tell arc. Stippling represents the location of Late Miocene marine sediments which approximate the position of the pre-Gibraltar Strait connections between the Mediterranean and Atlantic. These connections are the Betic Corridor in southern Spain and the Rifian Corridor in northern Morocco.





**Fig. 2.** Summary of the Late Miocene–Early Pliocene stratigraphy of the Mediterranean, Atlantic gateway region. From left to right the columns are: age; Geomagnetic Polarity Time Scale (GPTS); P-0.5 T which is normalized precession minus half normalized obliquity. This is almost identical to summer insolation on 25th June at a latitude of 65° North and takes into account the time lag between astronomical forcing and climatic response for the different frequency bands (Lourens et al., 1996); benthic foraminifera  $\delta^{18}O$  curve generated for the Rabat section, Morocco (Hodell et al., 2001) and the Plio–Pleistocene stack (Lisiecki and Raymo, 2005); bioevents where 1 = LCO *G. menardii* 4, 2 = FCO of *G. menardii* 5, 3 = FCO of *G. miotumida* group; 4 = *N. aconstaensis* s/d, 5 = FO *G. margaritae*, 6 = bottom acme *G. margaritae*, 7 = influx *G. menardii* sinistral, 8 = FO *G. puncticulata*; sedimentation, environments and events in the Atlantic adjacent to the gateway area, Betic and Rifian corridors and the Mediterranean (split into shallow and deeper water settings according to the CIESM scenario; Roveri et al., 2008).

**Table 1**

List of acronyms and their definitions used in this paper.

Acronym	Definition
AMOC	Atlantic Meridional Overturning Circulation
AMW	Atlantic Mediterranean Water
CU	Complex Unit
DSDP	Deep Sea Drilling Program
GCM	General Circulation Model
GIN seas	Greenland–Iceland–Norwegian seas
LGM	Last Glacial Maximum
LU	Lower Unit of the seismic trilogy in the western Mediterranean
IODP	Integrated Ocean Discovery Program
MES	Messinian Erosion Surface
MO	Mediterranean Outflow
MSC	Messinian Salinity Crisis
MU	Mobile Unit of the seismic trilogy in the western Mediterranean
NAC	North Atlantic Current
NADW	North Atlantic Deep Water
NEADW	North Eastern Atlantic Deep Water
ODP	Ocean Drilling Program
TES	Top Erosional Surface
UU	Upper Unit of the seismic trilogy in the western Mediterranean

### 2.3. Messinian stratigraphy of the Mediterranean and Atlantic basins

The stratigraphy that characterises the Late Miocene sediments of the Mediterranean–Atlantic gateway region results from the relative influence of both the adjacent basins and the tectonic evolution of the area itself. It is beyond the scope of this paper to describe in detail the extensive stratigraphic work that has been carried out on Late Miocene Mediterranean and Atlantic sediments (Roveri et al., 2014a). However, a brief description of the Messinian sedimentary sequences in both basins is provided here, along with a summary figure illustrating their relationship with the corridor sediments (Fig. 2).

Knowledge of the Late Miocene Mediterranean sedimentary evolution is heavily biased by the onshore exposures. Around the margins of the Mediterranean there are numerous sections of pre-MSC sediments and much of the lower part of the MSC succession is also exposed. A continuous deep basinal sedimentary succession is however still lacking as it has not yet been possible to drill in the deep Mediterranean and recover a complete MSC succession.

#### 2.3.1. Mediterranean onshore successions

The pre-evaporite succession onshore consists of cyclic alternations of homogeneous marls, sapropels and diatomites (Fig. 2) that were deposited in response to astronomically-driven climate oscillations, predominantly precession, and were amplified as the Mediterranean became progressively isolated from the open ocean (Bellanca et al., 2001; Blanc-Valleron et al., 2002; Hilgen and Krijgsman, 1999; Hilgen et al., 1995; Krijgsman et al., 1999a; McKenzie et al., 1980; Sierro et al., 1999, 2001; Suc et al., 1995). This orbital signal combined with biostratigraphic and palaeomagnetic data allows the age of these sediments to be precisely constrained with an error of a single precessional cycle ( $\pm 10$  kyr; Krijgsman et al., 1999a).

The onset of the Messinian Salinity Crisis occurred at 5.97 Ma (Fig. 2; Manzi et al., 2013) with synchronous precipitation of gypsum in marginal basins around the Mediterranean (Krijgsman et al., 1999a, 2002). This first period of the MSC (Stage 1; Fig. 2) is characterised by deposition of 16–17 gypsum horizons with interbedded laminated clastic sediments (Krijgsman et al., 2001; Lugli et al., 2010; Manzi et al., 2013). Little is known about coeval deposition in the deep basinal areas, but evidence from the Apennines and geochemical relationships suggest that organic-rich shales and dolomites rather than gypsum accumulated here (Fig. 2; de Lange and Krijgsman, 2010; Manzi et al., 2007; Roveri and Manzi, 2006).

Unsurprisingly, evaporite-bearing successions lack clear biostratigraphic markers and have a weak palaeomagnetic signal. Consequently,

the independent age time points that are used so successfully to confirm the age of pre- and post-MSC sediments are not available during the MSC itself. However, assuming precession also controlled the periodicity of gypsum cyclicity, the top of Stage 1 Lower Evaporites has been estimated at 5.61 Ma (Fig. 2; CIESM, 2008; Hilgen et al., 2007; Krijgsman et al., 2001; Roveri et al., 2014a).

In some marginal basins, the upper boundary of Stage 1 is marked by a major erosive event known as the Messinian Erosion Surface (MES; Figs. 2 and 4) which can also be traced offshore in seismic data where it splits into several surfaces not all of which are erosional (Fig. 4c; Lofi et al., 2011a; Lofi et al., 2011b). The marginal erosion surface is likely to have been caused by a sea-level drop which resulted in reworking and transport of sediments basinward, including significant gypsum, and their deposition in intermediate to deep basinal settings (Fig. 2; e.g. Clauzon et al., 1996; Lofi et al., 2005; Maillard et al., 2006; Manzi et al., 2005; Roveri et al., 2008). However, whether the associated seismic surfaces offshore were generated in subaerial or subaqueous environments is still highly contentious.

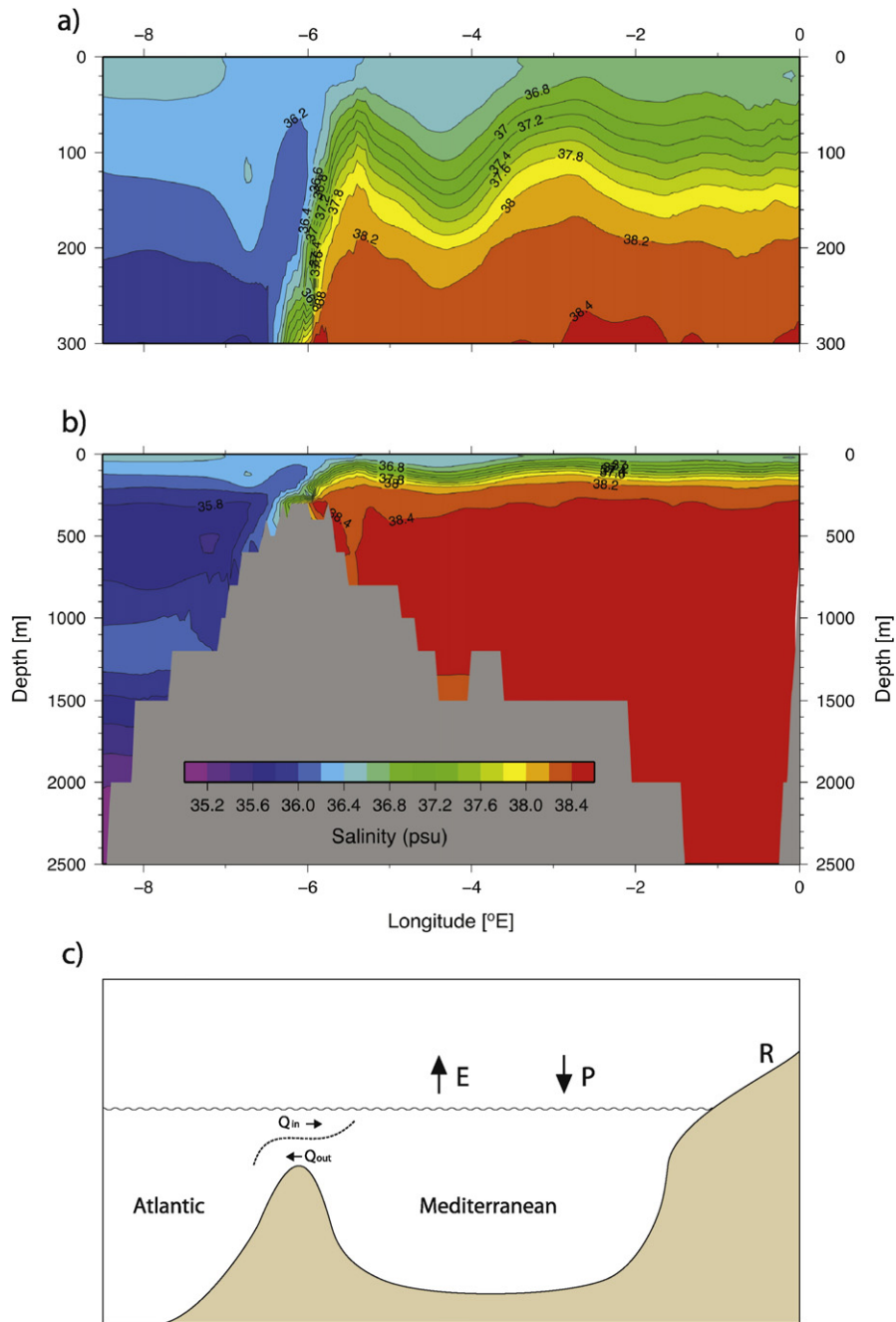
The halite phase of Stage 2 only developed in the deep basins (Figs. 2 and 4). By contrast, Stage 3 sediments which, in the upper part includes intervals known as the Lago Mare, have been recognized in many of the Mediterranean's marginal basins (Fig. 2). These sediments contain fauna thought to live in brackish water conditions (Bassetti et al., 2003, 2006; Orszag-Sperber, 2006); on Sicily, these deposits are interbedded with gypsum (Manzi et al., 2009; Rouchy and Caruso, 2006). In the western Mediterranean and within the Betic corridor, however, the Lago Mare comprises grey to white marls interbedded with deltaic or fluvial conglomerates (Fig. 2; Fortuin and Krijgsman, 2003; Omodeo Salé et al., 2012). Material recovered from ODP and DSDP holes that correlates with the seismic Upper Unit, indicates that Lago Mare sediments overlie the deep sea evaporites (Hsü et al., 1973, 1977; Orszag-Sperber, 2006; Roveri et al., 2014b). Unfortunately, the correlative relationship between these offshore deposits and the marginal sediments remains unclear. During Stage 3 the Mediterranean may have been isolated from the open ocean until the return to normal open marine conditions after the Miocene–Pliocene boundary (Fig. 2) at 5.33 Ma (Lourens et al., 1996).

#### 2.3.2. Mediterranean offshore successions

The interpretation of the Mediterranean's basinal succession is mainly reliant on seismic data. These data show that in the deepest parts of the western basin, a seismic trilogy is developed comprising a bedded Lower Unit (LU) of unknown age and lithology with low frequency reflectors; a transparent Mobile Unit (MU; Fig. 4c) consisting mainly of halite and showing plastic deformation; and a bedded Upper Unit (UU) which occurs just below Lower Pliocene sediments (Fig. 4; Lofi et al., 2011a; Lofi et al., 2011b). Of these, only the Upper Unit has been drilled (e.g. sites 122 and 372; Montadert et al., 1978; Ryan et al., 1973). The absence of well-ties and significant differences between the onshore and offshore MSC successions makes correlation highly controversial.

Seismic mapping of the Mobile Unit indicates that a large volume of salt, more than 2 km thick in some places, accumulated in the deepest parts of the Mediterranean (CIESM, 2008; Hsü et al., 1973; Lofi et al., 2005). It has been argued that this extensive salt deposit is equivalent in time to the major base-level drawdown that eroded the margins (e.g. Stage 2; Fig. 2; CIESM, 2008; Roveri et al., 2014a). This temporal coincidence suggests a major change in the pattern of Mediterranean–Atlantic exchange such that while Mediterranean outflow decreased to very low levels reducing salt export to the Atlantic, ocean inflow continued providing a constant supply of the ions required for gypsum and halite formation (Krijgsman and Meijer, 2008; Lugli et al., 2010; Meijer, 2006).

**2.3.2.1. Mediterranean offshore succession adjacent to the corridors.** The Western Mediterranean is made up of a series of basins that contain



**Fig. 3.** Vertical east–west section along the Strait of Gibraltar showing the present-day salinity distribution a) between 0–300 m water depth and b) down to 2500 m. Depicted is the climatological field, averaged latitudinally across a  $0.5^\circ$  wide slice through the Strait (MEDAR Group, 2002); c) cartoon of present day Mediterranean–Atlantic exchange where E = evaporation, P = Precipitation, R = river inflow and  $Q_{in}$  and  $Q_{out}$  are the exchange fluxes at the gateway.

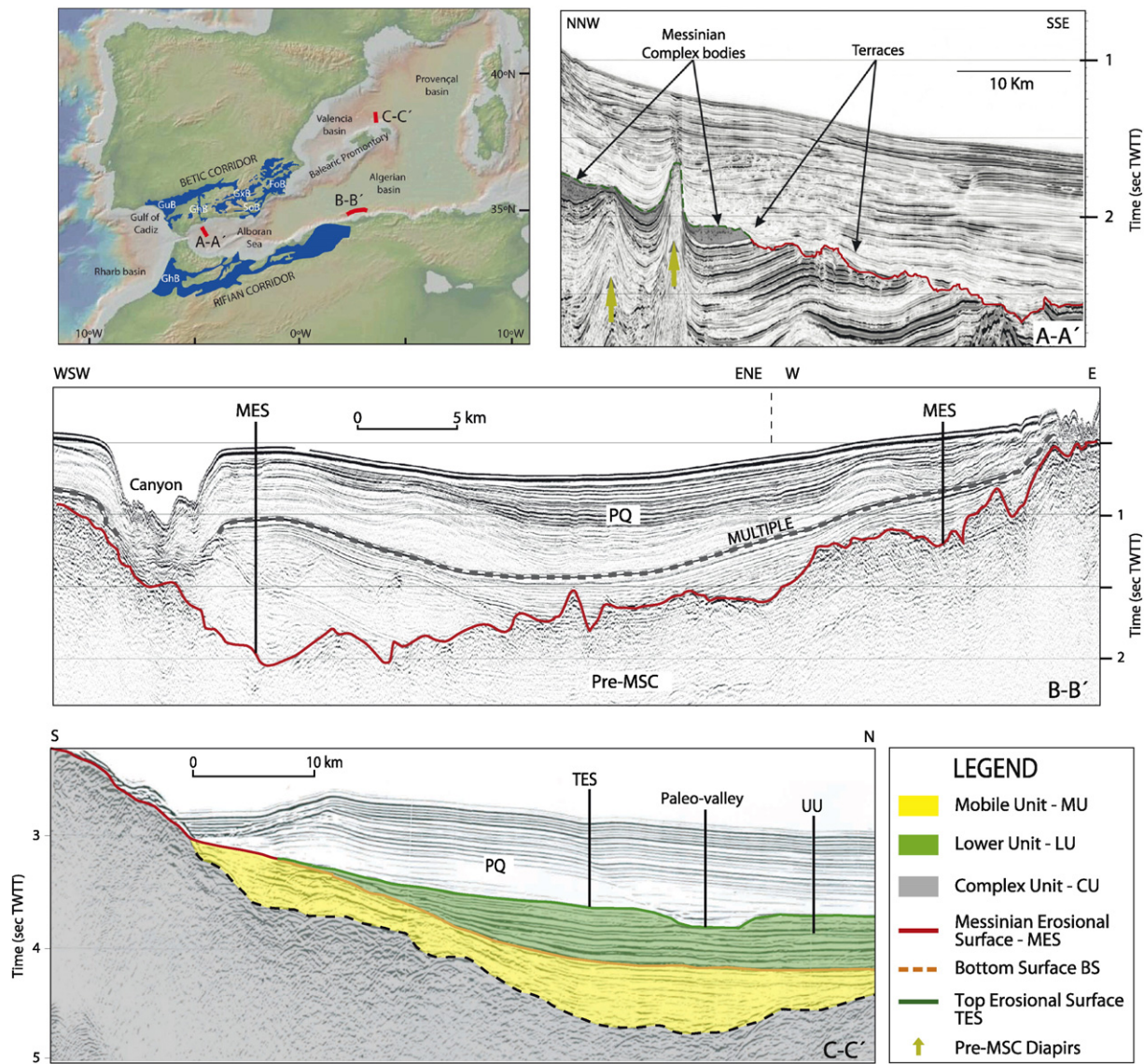
sediments deposited in both marginal and deep water settings during the MSC e.g. Alborán, Algerian, Provençal and Valencia basins (Fig. 4). They provide insight into the relationship between the marine corridors and the deeper Mediterranean basin to which they were connected.

The Algerian Basin, located between the Alborán Basin to the west, Sardinia to the east, and the Balearic Promontory to the north (Fig. 4) has a complex geological history resulting from the Alpine orogeny (Mauffret et al., 2004). The basin is characterised by having young oceanic basement, relatively thin pre-MSC sediments and significant salt-driven deformation. The deep basin MSC trilogy has been identified offshore (Capron et al., 2011; Obone-Zué-Obame et al., 2011). The Messinian Erosion Surface is, as elsewhere, traced on the margin between pre-MSC and prograding Pliocene deposits, where it preserves

several deep canyons (Fig. 4b; Obone-Zué-Obame et al., 2011). The connection between the onshore Algerian corridors and their offshore extension has yet to be fully explored.

The MSC seismic trilogy is absent from the Alborán Basin which is dominated by the presence of a widespread erosion surface (MES, Fig. 4a; eg. Estrada et al., 2011). The MES separates Miocene units from Plio–Quaternary sediments and appears as a prominent, high-amplitude, and laterally continuous reflector (Martínez-García et al., 2013) that occurs in terraces at different depths (Fig. 4a; Estrada et al., 2011). The origin of this erosional surface is thought to be polygenic since it has been shaped by multiple erosional stages related to erosion during the MSC acme and the Zanclean reflooding through the Gibraltar Strait (Estrada et al., 2011; Martínez-García et al., 2013).





**Fig. 4.** a) Location map of the Atlantic and Mediterranean offshore basins adjacent to the gateway region and the location of seismic sections A–A', B–B' and C–C'. GuB Guadaquivir Basin, GhB Guadalupe Basin, SoB Sorbas Basin, GXB Guadix Basin, FoB Fortuna Basin, GhB Gharb Basin. b) Seismic section from the Alborán Sea close to the Guadalquivir corridor, part of the Betic connection. c) Seismic section from the Valencia Basin. d) Seismic section from the Algerian Basin close to onshore Neogene corridor sediments.

Where the basin has been drilled, the Pliocene typically unconformably overlies the pre-MSC series (e.g. Leg 13 site 121; Leg 161 sites 977, 978). Where MSC sediments are recovered (CU unit in Fig. 4a), they consist of pebbles, shallow-water carbonates, sandy turbidites and gypsum or anhydrite deposits (Iaccarino and Bossio, 1999; Jurado and Comas, 1992), which are interpreted as being deposited in small isolated lacustrine basins (Martínez-García et al., 2013) equivalent in age to the Lago Mare (Fig. 2).

Late Miocene marine sediments exposed along the north coast of Morocco adjacent to the Alborán Sea (Fig. 1) indicate the nature of this eastern extension of the Rifian Corridor. The sections near the Melilla volcanic centre (Fig. 6) record alternations of volcanic ashes interbedded with diatomites and marls that grade up and laterally into reef carbonates (Fig. 2; Cornée et al., 2002; Cunningham and Collins, 2002; Cunningham et al., 1994; Cunningham et al., 1997; Münch et al., 2001; Münch et al., 2006; Roger et al., 2000; Saint Martin and Martín, 1996; Saint Martin et al., 1991; Van Assen et al., 2006; Azdimousa et al., 2006; Barhoun and Wernli, 1999). Where the sections have been astronomically tuned and/or Ar/Ar dated, the resulting interpretation suggests that cyclic sedimentation began around 6.84 Ma (Van Assen et al., 2006). Sedimentation was terminated

following exposure at or above sea level at ~6.0 Ma (Fig. 2; Münch et al., 2006; Van Assen et al., 2006). In the near-by Boudinar Basin (Fig. 6) clastic sedimentation began around late Tortonian (Azdimousa et al., 2006) followed by a transgressive–regressive sequence comprising alternations of marls and sands. The sequence is similar to that seen in the Melilla sections, but in the Boudinar Basin the top sandy-marl intervals are Early Pliocene in age (Barhoun and Wernli, 1999).

The southern limit of the Valencia Basin is formed by the Balearic Promontory which is considered to be the offshore continuation of the Betic Range (Fig. 4). The asymmetric basin formed during an Oligo–Miocene extensional phase linked to a concurrent transpressional–compressional system (Alfaro et al., 2002; Doglioni et al., 1997; Fontboté et al., 1990; Gueguen et al., 1998; Maillard and Mauffret, 2011, 2013; Roca and Guimerà, 1992; Vegas, 1992). The intermediate depth of the Valencia Basin and its location adjacent to well exposed onshore successions of Messinian Salinity Crisis sediments (e.g. in the Sorbas Basin) mean that it has great potential as a source of information linking marginal sequences with deep-marine seismic units (Fig. 2). In the Valencia Basin, seismic profiles show only an Upper Unit (generally <200 m thick; Fig. 4c) rather than the complete deep basin MSC succession (Maillard et al., 2006) that is observed further east, in the Liguro–

Provençal basin. This Upper Unit overlies an erosion surface which truncates pre-evaporitic reflectors and is itself cut by an upper erosion surface which has been interpreted as having been generated during the latest Messinian (Maillard et al., 2006). Diachronism, a polygenetic origin, and local expression of the MES hinder assessment of its exact relationship with the MSC event (Fig. 4c; Lofi et al., 2005; Maillard et al., 2006; Lofi et al., 2011b).

### 2.3.3. Atlantic successions

Late Miocene Atlantic sediments recovered from the present day Moroccan and Iberian shelf are dominated by nannofossil ooze, clays, marls and silty marls (Fig. 2, Hayes et al., 1972; Hinz et al., 1984). Where Atlantic margin sediments are exposed today onshore Morocco and Spain in the mouths of the two corridors, they comprise long and continuous successions of cyclic silty marls.

**2.3.3.1. The Guadalquivir Basin.** During the Middle and Late Miocene, all the Betic corridors were connected to the Atlantic Ocean through the Guadalquivir Basin which extends offshore into the Gulf of Cadiz (Fig. 1). The sediments of the Guadalquivir Basin are not well exposed on land and most studies are reliant on seismic data, well logs and boreholes (e.g. Berástegui et al., 1998; Fernández et al., 1998; Larrasoña et al., 2008; Pérez-Asensio et al., 2012b; Ríaza, 1996). These indicate that the succession comprises marine and continental sediments that range in age from late Tortonian to Recent (González-Delgado et al., 2004; Sierro et al., 1996). After closure of the North Betic strait, the Guadalquivir Basin was established as a wide marine embayment open to the Atlantic (Pérez-Asensio et al., 2012b), and marine sedimentation continued throughout the late Tortonian and Messinian into the Pliocene (Fig. 2). Recent studies of the Montemayor borehole have resulted in a magnetobiostratigraphic framework (Larrasoña et al., 2008), a palaeoenvironmental scheme based on benthic foraminifera (Pérez-Asensio et al., 2012a,b, 2013) and the reconstruction of local vegetation and sea level changes (Jiménez-Moreno et al., 2013). These results show a shallowing trend before and during the MSC (Fig. 2), with two distinct cooling periods and associated sea-level falls, one of which is identified as contributing to the onset of the MSC. However, the chronostratigraphic framework of the Montemayor borehole on which these results are based (Pérez-Asensio et al., 2012a) is constructed from magneto- and biostratigraphic tie-points combined with an age model based on extrapolation of stable isotope data. An improved age model by van den Berg et al. (in revision) was based on astronomical tuning of the cycles found in the geochemical composition of the borehole sediments. This shows that the cooling period cannot be associated with the onset of the MSC since the oxygen isotope records follow global ocean trends. These results also imply that there was no direct influence of Mediterranean outflow on this part of the Guadalquivir Basin during the Late Messinian.

**2.3.3.2. The Gharb–Prerif Basin and classic successions near Rabat.** The Gharb–Prerif Basin (Figs. 1 and 6) straddles the Moroccan Atlantic coastline and is a Miocene foreland basin which developed parallel to the southern edge of the Rif (Michard, 1976; Flinch, 1993; Pratsch, 1996). During the Late Miocene and Early Pliocene this area, the western extension of the Rifian corridor, was a large marine gulf (e.g. Esteban et al., 1996; Martín et al., 2009). The Gharb–Prerif Basin is commonly divided into two main areas. The northern part links onshore sub-basins in the Rifian corridor's northern strand or Prerif Zone (e.g. Taounate and Dar Souk) with the Gharb Basin which extends offshore (Fig. 1), while the southern arm comprises the Sais Basin west of Taza-Guercif and the South Rifian Trough, which also extends offshore. The stratigraphy of the Gharb–Prerif Basin is subdivided into Mesozoic pre-foredeep and Upper Miocene (mainly Messinian) to Pliocene foredeep successions in the north (Flinch, 1993) into the lower part of which deformed Triassic–Miocene sediments of the Prerif Nappes were emplaced (Michard, 1976; Flinch, 1993). During the late stages

of the Rif–Betic orogeny, a series of satellite extensional mini-basins were generated within the Gharb Basin that are filled with thick Upper Tortonian to Pleistocene clastics (Flinch and Vail, 1998). These are deformed by a late Neogene phase of flexurally-induced normal faulting and syn- to post-thrusting of the Rif nappes (Zouhri et al., 2002) interpreted as being either coeval (Flinch, 1993), or as part of two episodes of deformation; an extensional Tortonian–Messinian and a compressional Plio–Quaternary episode (Litto et al., 2001).

In the South Rifian Trough (Fig. 1), the pre-foredeep succession is overlain by Middle to Late Miocene clastics (Wernli, 1988; Flinch, 1993) thought to have been deposited in a shallow foreland basin (Michard, 1976; Flinch, 1993). Unlike the Gharb Basin, it is not disrupted by the Nappe system and the Neogene succession is almost structurally undeformed with a north dipping basement leading to a thicker Neogene and Quaternary succession to the north (e.g. Zouhri et al., 2002). In contrast with the relatively coarse clastic-rich, occasionally turbiditic successions that dominate the central areas of the Rifian corridor, the well-studied sections near Rabat on the Atlantic coast comprise regular alternations of indurated marls that are blue when freshly exposed and softer reddish clay-rich marls (Fig. 2) unconformably overlying Devonian limestones (Hilgen et al., 2000). These well exposed successions have been astronomically tuned and therefore have an exceptionally high resolution age model (Hilgen et al., 2000; Krijgsman et al., 2004). In offshore parts of the basin, data is sparse and interpretations are of much lower resolution and often restricted by poor quality seismic data.

Despite being outside the Mediterranean and consequently significantly buffered by the Atlantic Ocean, the tuning of these Late Miocene–Pliocene successions indicates a strong pattern of sedimentary response to precession with secondary components of obliquity and eccentricity (e.g. van der Laan et al., 2005). Fluctuations in the faunal assemblage that mirror this strong precessional signal are similar to, but of lower amplitude than those seen in time equivalent sediments in the Sorbas Basin (Fig. 2; van der Laan et al., 2012). One possible explanation for the insolation-driven sedimentation that persists in the Rabat successions before, during and after the MSC is that the controlling mechanism linking orbital forcing to its sedimentary response is, in this area, independent of the state of connectivity between the Atlantic and Mediterranean and therefore probably lies outside the Mediterranean. A possible explanation for the colour cycles are precession driven climate oscillations that are linked to the Atlantic system (Bosmans et al., 2015; Brayshaw et al., 2011; Tzedakis, 2007) rather than to the African monsoon as generally assumed for Mediterranean sapropels (Kutzbach et al., 2014). The Atlantic system brings rain in the winter-halfyear, and the dominantly precession controlled changes in this system may have been operative at least from 7.8 Ma to the Recent, although climatic interpretations differ (Sierro et al., 2000; Moreno et al., 2001; Bozzano et al., 2002; van der Laan et al., 2012; Hodel et al., 2013).

## 3. Evolution of the Atlantic–Mediterranean gateways: direct approach

The distribution of Late Miocene marine sediments across Northern Morocco and Southern Spain resembles a complex network of channels connecting the Mediterranean and Atlantic (Figs. 1, 5 and 6) and it is this now uplifted area that is considered to be the Late Miocene gateway region (e.g. Santisteban and Taberner, 1983). It is less clear, however, how much of the bifurcating channel pattern reflects the primary configuration of the Late Miocene marine corridors and how much is a function of the preservation of the sediments after uplift and erosion. Much of the sedimentation of this region, particularly at either end of the corridors, is characterised by clay–silt grade material (Fig. 2) with few if any current structures. This is indicative of low energy conditions rather than the coarser material anticipated from high velocity currents such as those seen in the Gibraltar Strait today. Even in central areas of



the corridors, where sands and conglomerates are common (Fig. 2), these may result from uplift-driven higher energy processes relating to slope transport for example, rather than being the direct product of inflow or outflow current transport and deposition. Here we present a summary of the sedimentological data preserved in the exposed gateway region.

### 3.1. The Betic Corridor

The Betic Corridor can be subdivided into four distinct connections that link the Atlantic with the Mediterranean during the Late Miocene: the North-Betic strait; the Granada Basin; the Guadix Basin and the Guadalhorce Basin (Fig. 5). All these basins contain large-scale palaeocurrent structures (Fig. 5) typically in coarse-grained sand or conglomeratic sediments indicating high energy currents (Martín et al., 2014). These coarse clastics which commonly form the last part of the preserved succession are difficult to date. However, some of the Betic successions also contain evaporite and continental sediments that predate the MSC suggesting that these connections were conduits for Mediterranean–Atlantic exchange before the formation of the Mediterranean's saline giant.

#### 3.1.1. North Betic strait

This most northerly corridor connects the Guadalquivir Basin and the Mediterranean through the Fortuna and Lorca basins (Fig. 5; Martín et al., 2009). Müller and Hsü (1987) initially suggested that the North Betic strait was open throughout the Messinian, allowing an ocean water flux and providing the salt required for evaporite deposition to reach the Mediterranean. This flux was thought to have been modulated by glacio-eustatic sea level change (Santisteban and Taberner, 1983). By contrast, Benson et al. (1991) envisaged the North Betic Strait closing just before the onset of the MSC, but serving as a channel for unidirectional Mediterranean outflow during the precursor 'Siphon event'. Subsequently, integrated stratigraphic studies were carried out on the Fortuna Basin at the eastern end of the North Betic Strait (Garcés et al., 1998; Krijgsman et al., 2000b). These indicate that sedimentation changed from marls to diatomites and evaporites at 7.8 Ma (the 'Tortonian salinity crisis' of the eastern Betics; Krijgsman et al., 2000a), before deposition of continental deposits at ~7.6 Ma. The dating of this tectonically driven restriction event (Krijgsman et al., 2000a) suggests that the North Betic Strait cannot have been the route by which ocean water reached the Mediterranean during the MSC.

#### 3.1.2. Granada Basin

The Granada Basin connects the Guadalquivir Basin to the Mediterranean via the Zagra strait (Martín et al., 2014). The stratigraphy of this basin was first described by Dabrio et al. (1978), who suggested tectonism caused basin restriction through the uplift of the present day Sierra Nevada, leading to the deposition of evaporites in the Granada Basin (Fig. 5) that pre-date the MSC (Fig. 2). According to Martín et al. (1984) this restriction took place around the Tortonian–Messinian transition. Braga et al. (1990) suggested instead that restriction began on the eastern side of the basin in the late Tortonian, spreading to the southern and western margins which uplifted around the Tortonian–Messinian boundary, gradually isolating and desiccating the basin and filling it with continental deposits. The restriction of the Granada Basin has recently been more precisely dated using biostratigraphy by Corbí et al. (2012) demonstrating that a short phase of evaporite precipitation occurred between 7.37 to 7.24 Ma, followed by a less well constrained phase of continental sedimentation. This confirms the conclusion of earlier studies that the Granada Basin was not a conduit for Mediterranean–Atlantic exchange for any part of the Messinian.

#### 3.1.3. Guadix Basin

This was a relatively open marine passage (around 12–15 km wide), probably permitting two-way flow, with coarse grained sediments

deposited on the edges while marls accumulated in its central part. Later it evolved into a narrow strait with strong bottom currents flowing from the Mediterranean to the Atlantic, based on huge bioclastic sand and conglomerate dunes, displaying internally cross-bedding several meters high (Fig. 5; Betzler et al., 2006). Contemporaneous with these bottom currents there were Atlantic surface current flowing southwards (Puga-Bernabeu et al., 2010). Although there is broad consensus that the Guadix Basin corridor (or Dehesas de Guadix strait; Martín et al., 2014) was open during the late Tortonian (Betzler et al., 2006; Hüsing et al., 2010; Soria et al., 1999), the detailed timing of the closure is disputed. According to Betzler et al. (2006) the strait narrowed to about 2 km and was finally blocked at ~7.8 Ma by a tectonic swell fringed by reefs (7.8–7.4 Ma). This is contradicted by more recent magnetobiostratigraphic results for the same section (La Lancha, Hüsing et al., 2010) which show that there is a major hiatus of at least 2 Myr between open marine sediments of ~7.85 Ma and continental deposits, dated at 5.5 Ma (Fig. 2) in agreement with the presence of the MN13 mammal biostratigraphy (Hüsing et al., 2012; Minwer-Barakat et al., 2012). This unconformity means that closure of the Guadix Basin corridor is not recorded and consequently, the possibility that a shallow marine connection remained during MSC Stages 1 and 2 cannot be excluded.

#### 3.1.4. Guadalhorce Corridor

The history of the Guadalhorce corridor (Fig. 5; Martín et al., 2001) is less well known than the other Betic basins. Its sedimentary record consists predominantly of siliciclastics containing unidirectional cross-beds (Fig. 2) with sets over 100 m in length and ranging from 10 to 20 m in thickness. These structures have been interpreted as indicating that the corridor was at least 60–120 m deep and subject to an extremely fast ( $1.0\text{--}1.5\text{ m s}^{-1}$ ) unidirectional current flowing northwest (Fig. 5; Martín et al., 2001). Foraminifera-bearing marls intercalated with carbonates in one of the outcrops towards the bottom of this unit have an early Messinian age (7.2–6.3; Martín et al., 2001). Consequently, the Guadalhorce corridor was considered to be a conduit for Mediterranean outflow prior to the MSC in accordance with the 'siphon' model (Benson et al., 1991). Pérez-Asensio et al. (2012b) also took the view that Guadalhorce had an important role in Mediterranean to Atlantic outflow. These authors assume that ultimately all other corridors were closed during the MSC and only the Guadalhorce supplied Mediterranean water to the Atlantic via the Guadalquivir Basin during this period. They interpreted a change in benthic  $\delta^{18}\text{O}$  record in the Guadalquivir basin at 6.18 Ma, as indicating closure of the Guadalhorce corridor. However, as there are no MSC-aged sediments preserved evidence of this from within the Guadalhorce corridor itself remains to be found.

In summary, of the four possible Betic corridors that may have supplied Atlantic water to the Mediterranean during the Late Miocene, two are known to have been closed during the MSC (the North Betic Corridor and the Granada corridor) while the successions of the remaining two (Guadix and the Guadalhorce corridors) contain large unconformities and uncertainties that span the critical Late Miocene period. It is therefore not currently possible to rule out definitively an open or intermittent connection within the Betic corridor area during the MSC.

### 3.2. The Rifian Corridor

The distribution of late Tortonian and Messinian sediments through northern Morocco is broadly divided into a northern and a southern strand (Fig. 6); the intramontane basins form the northern part of this gateway; the Nador–Taza–Guercif–Rabat axis forms the better-studied southern arm. These strands merge at their western end which, in the Late Miocene, formed a broad Atlantic-facing embayment (Fig. 6). By contrast, their eastern connections to the Mediterranean are distinct, with the northern strand reaching the Mediterranean between Boudinar and Melilla and the southern strand joining further east in Algeria. Given that the area between the two strands is a thrust nappe

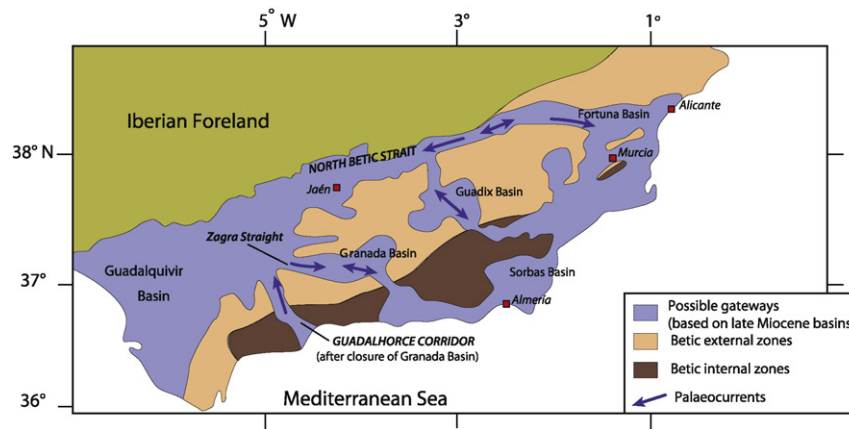


Fig. 5. Detail map of the Betic Corridor showing the main geological units, corridors and palaeocurrent indicators (modified after Santisteban and Taberner, 1983). Palaeocurrents from Santisteban and Taberner (1983), Benson et al. (1991), Martín et al. (2001) and Martín et al. (2014).

pile (Chalouan et al., 2008; Flinch, 1993; Feinberg, 1986) which is locally overlain by Late Miocene marine sediment, one possibility is that the Rifian corridor was in fact a single wide strait which has subsequently been uplifted and eroded leading to the preservation of a more complex and segmented pattern of Late Miocene marine sediments (Fig. 6). Detailed palaeogeographic data are required to distinguish between these two hypotheses.

Understanding of the evolution of the Rifian corridor is strongly biased by the location of well-studied sections, some of which have been astronomically tuned. To be specific:

1. There is almost no published information about the onshore Algerian section of the corridor;
2. For reasons of exposure, nearly all information available regarding the western embayment comes from an area to the south, near Rabat (Fig. 6);
3. The only astronomically tuned section from the central part of the corridor is located near the southern margin of the southern strand, in the Taza-Guercif Basin (Fig. 6).

### 3.2.1. North Rifian connection

The age of the northern strand of the Rifian Corridor currently relies on biostratigraphic analysis of marine sediments carried out during the eighties (Wernli, 1988). These sediments are all assigned to an undifferentiated Tortonian–Messinian marine zone (M6), which spans 11.6 to

5.3 Ma but does not allow more accurate age subdivisions. Nevertheless, the highly diverse faunal assemblage suggests sediments occurring in these apparently distinct subbasins (Taounate, Dhar Souk, Boured; Fig. 6) experienced open marine conditions during the Late Miocene. Sections near the town of Taounate and in the other intramontane basins record the onset of clastic sedimentation with poorly sorted conglomerates and sandy marls, followed by thick alternations of marls and sands which indicate abundant sediment input from the basin margins. The top of the marine sequence is generally truncated by erosion and covered by Quaternary deposits (Wernli, 1988).

### 3.2.2. South Rifian Corridor

Astronomical tuning of the central southern Taza-Guercif Basin section (Fig. 6) indicates that closure of this part of the corridor (e.g. the transition from marine to continental sedimentation) occurred between 6.7 to 6.0 Ma (Fig. 2; Krijgsman and Langereis, 2000; Krijgsman et al., 1999b), but this evidence cannot preclude there being a concurrent Mediterranean–Atlantic connection further north. The Nd isotope data from the Taza-Guercif Basin suggests that restriction of the southern strand of the corridor took place to the east around 7.2 Ma, cutting the connection with the Mediterranean, but allowing the connection between the Taza-Guercif Basin and the Atlantic to persist (Ivanović et al., 2013a; see Section 4.1.2). The timing is coincident with a significant shallowing event seen in the palaeobathymetric reconstructions for the thick clastic successions found in the central

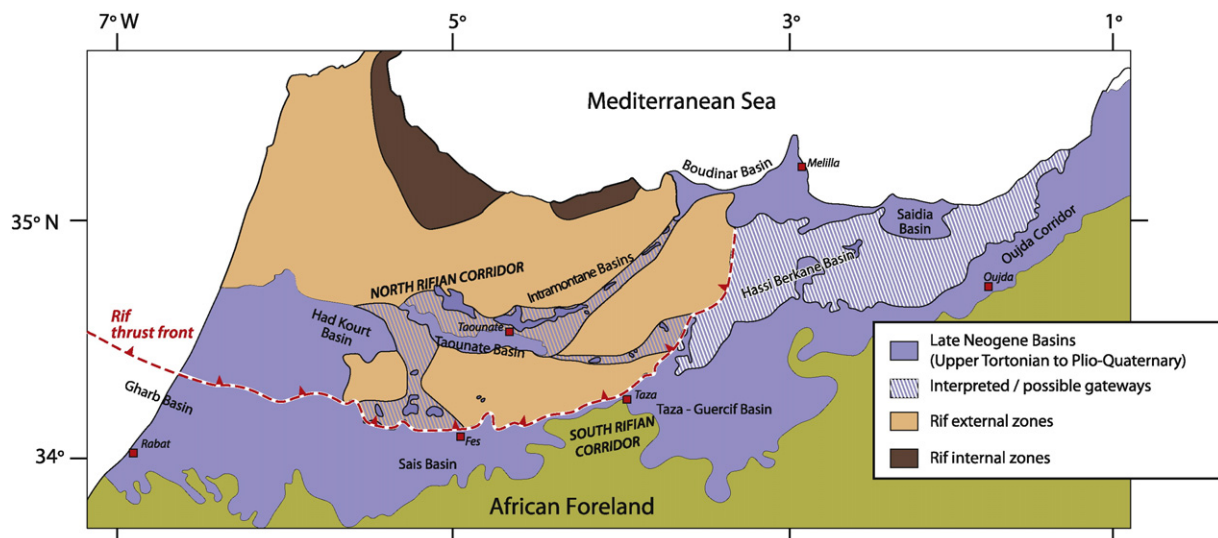


Fig. 6. Detail map of the Rifian Corridor showing the main geological units and corridors.



Taza-Guercif Basin (Krijgsman et al., 1999b). The sedimentation pattern in the Taza-Guercif Basin is consistent with a pre-MSC closure of the Rifian corridor as a whole, for which there is a growing and diverse body of evidence including evidence from mammal fossils which indicates that rodents were able to migrate between Morocco and Spain before 6.1 Ma (Agustí et al., 2006; Benammi et al., 1996; Gibert et al., 2013).

Palaeomagnetic studies in the Rifian Corridor demonstrate that no vertical axis rotations have occurred in the Taza and Gharb foreland basins and in the post-thrusting Melilla Basin (Fig. 1) since the Tortonian–Messinian (Krijgsman and Garcés, 2004). In the Rifian part of the External Thrust Belt (Fig. 1), however, upper Miocene thrust top sequences show anticlockwise rotations (Cifelli et al., 2008). Southward propagation of the thrust load appears to be a significant driver of closure of the Rifian corridors along with regional uplift attributed to slab dynamics (Duggen et al., 2004).

### 3.3. Strait of Gibraltar

The Strait of Gibraltar connects the Mediterranean with the Atlantic today (Fig. 1) and is commonly assumed to have formed at the Miocene/Pliocene boundary (5.33 Ma; Hsü et al., 1973; Hsü et al., 1977) with rapid (5–800 days according to Garcia-Castellanos et al., 2009) refilling of the Mediterranean (Blanc, 2002; Meijer and Krijgsman, 2005; Loget and Van Den Driessche, 2006). The assumption of a major ‘Zanclean flood’ is based on seismic evidence of an incised channel that dips from the Gibraltar Strait across the Alborán Sea and can be followed for at least 300 km towards the east (Estrada et al., 2011). Here, the deep U-shaped incision merges laterally with the Messinian Erosion Surface (MES) and is locally filled with Plio–Quaternary sediments (Campillo et al., 1992; Garcia-Castellanos et al., 2009). The cause of the initial rupture at Gibraltar is still debated. If a eustatic trigger is excluded (van der Laan et al., 2006), it may have been caused by tectonic collapse (Govers, 2009), possibly in combination with headward river erosion (Hsü et al., 1973; Blanc, 2002; Loget and Van Den Driessche, 2006).

Detailed geomorphological studies (Esteras et al., 2000; Blanc, 2002) revealed the asymmetric structure of the strait and identified an eastern and western domain. The eastern domain is narrower and deeper (750 to 960 m) while the western domain has an irregular bottom morphology composed of sharp submarine hills up to 80 m high, large flat banks and ditches up to 650 m deep. To explain such physiography Esteras et al. (2000) suggested that large gravity slides of unknown age may have occurred, derived from both the Iberian and Moroccan margins. A major slide may have been contemporaneous with the Zanclean flood (Blanc, 2002). These gravitational collapses locally filled the two excavated channels with flysch mud-breccia, forming shallow sills.

Since the clastic sediments at the base of the canyon infill are not dated, and the age of the main erosion event that shaped the channels remains uncertain (Esteras et al., 2000), the possibility that (at least temporary) opening of the Strait of Gibraltar preceded the Miocene/Pliocene boundary cannot be ruled out. This missing information is arguably the main reason for the uncertainty around the evolution of Mediterranean–Atlantic exchange before, during and after the MSC.

### 3.4. Biotic evidence of corridor evolution

Faunal groups impacted by the evolution of the Mediterranean–Atlantic gateways include marine species utilising the marine corridors (e.g. foraminifera, fish, corals) and land-based species able to migrate when these corridors were closed (e.g. mammals). Analysis of faunal datasets provides insights into the dimensions of the marine gateways and the timing of closure.

#### 3.4.1. Marine biota

From the perspective of gateway evolution, evidence from marine faunal palaeoecology contributes in two important areas: (1) palaeoecological differences between the Mediterranean and Atlantic sides of the corridors, and (2) palaeo-water depth reconstructions of the corridors themselves. During the Tortonian–earliest Messinian and from the Early Pliocene onwards, open Mediterranean–Atlantic connections resulted in similar planktic and benthic microfossil assemblages on both the Atlantic side of the gateways and in the homogeneous, marly sediments that span the entire Mediterranean basin (most data pertain to foraminifera and calcareous nannoflora: e.g. Baggley, 2000; Barbieri and Ori, 2000; Corbí et al., 2012; Feinberg, 1986; Hilgen et al., 2000; Hodell et al., 1989; Hodell et al., 1994; Hüsing et al., 2009a; Kouwenhoven et al., 2003; Kouwenhoven et al., 2006; Lozar et al., 2010; Pérez-Asensio et al., 2012b; Raffi et al., 2003; Santarelli et al., 1998; Sierro et al., 1993; Van de Poel, 1991; Zhang and Scott, 1996 and references therein). During the intervening Messinian interval, there are differences in the Mediterranean and Atlantic assemblages to a greater or lesser extent. These biological distinctions are sometimes mirrored by sedimentological differences which suggest synchronous chemical differentiation between the two basins. Many of the hypotheses that relate the Mediterranean's distinct Late Miocene sedimentary record to its connectivity with the Atlantic are based on these observations and the inferences that can be drawn about the nature of exchange.

**3.4.1.1. Stepwise restriction of Mediterranean–Atlantic exchange.** Not all faunal groups record environmental changes at the same time. Benthic foraminifera (and stable isotope records; see Section 4.2.2) indicate changing Mediterranean deep water environments as early as 7.17 Ma (Fig. 2), just after the Tortonian–Messinian boundary (7.251 Ma; Hilgen et al., 2000) with the disappearance of open marine, oxyphilic taxa (e.g. Kouwenhoven et al., 2003). This is more or less coincident with shallowing in the central Rifian Corridor (Fig. 2; Krijgsman et al., 1999b), although a causal relationship is not certain and changes in bottom water oxygenation are dependent on freshwater runoff as well as circulation. The first sapropel in the Faneromeni section (Crete, Hilgen et al., 1995) and the onset of cycle LA1 of the lower Abad marls in the Sorbas Basin (Sierro et al., 2001) indicate that restriction of water exchange occurred at this time such that the Mediterranean became sufficiently isolated to record a lithological response to subtle orbital variation. From 7.17 Ma onwards, benthic foraminifera typical of organic-rich suboxic waters and more tolerant to high environmental stress become increasingly abundant in bottom-water environments of the Mediterranean. In this development towards the MSC rather discrete steps are recognized around 6.7, 6.4 and 6.1 Ma (Seidenkrantz et al., 2000; Bellanca et al., 2001; Blanc-Valleron et al., 2002; Kouwenhoven et al., 2003, 2006; Iaccarino et al., 2008; Orszag-Sperber et al., 2009; Di Stefano et al., 2010; Lozar et al., 2010). Eventually oligotrophic, *Bolivina*–*Bulimina* dominated faunas developed (Fig. 2) and deeper basins are barren of foraminifera (Sprovieri et al., 1996; Violanti, 1996; Kouwenhoven et al., 2003) whereas contemporaneous assemblages in the Atlantic are largely unaffected (Kouwenhoven et al., 2003; Pérez-Asensio et al., 2012b).

From ~6.7 Ma planktic assemblages within the Mediterranean mirror restriction of gateway exchange through decreasing abundances and amplification of the orbital-scale changes of the planktic communities (Santarelli et al., 1998; Sierro et al., 1999, 2003; Bellanca et al., 2001; Perez-Folgado et al., 2003; Flores et al., 2005). Warm-water oligotrophic planktic foraminifera (summed as *Globigerinoides* spp.), and between 6.6 and 6.4 Ma *Globigerina bulloides* show 60–100% abundance shifts (Sierro et al., 1999, 2003). On the Atlantic side of the Rifian corridor, in the Ain el Beida section (6.46–5.52 Ma) the amplitude of these precessional changes in warm-water planktic foraminifera abundance is in the order of 20% (van der Laan et al., 2012; Fig. 2). Decreasing

abundances of Mediterranean planktic assemblages and a trend towards low diversity are attributed to increasingly adverse conditions of the surface waters preceding the MSC. Eventually, just below the MSC virtually barren samples are recorded (Sierro et al., 1993, 2003; Blanc-Valleron et al., 2002; Krijgsman et al., 2004; Lozar et al., 2010). The most extreme divergence between the planktic and benthic communities on the Mediterranean and Atlantic sides of the corridors is associated with the onset of evaporite deposition at 5.971 Ma (Manzi et al., 2013), when marine organisms disappear from the Mediterranean. Although planktic and benthic micro- and macro-organisms have been described from Stage 1, their presence is probably mainly the result of reworking and most marly layers intercalated in the gypsum are barren (Rouchy and Caruso, 2006 and references therein).

Brackish-water assemblages are recorded during the Lago-Mare phase in the latest Messinian (Section 2.3.1; e.g. Iaccarino and Bossio, 1999; Aguirre and Sanchez-Almazo, 2004; Pierre et al., 2006; Orszag-Sperber, 2006; Rouchy et al., 2007; Guerra-Merchan et al., 2010). This period is also characterised by intervals containing well preserved fossil fish of open marine origin suggesting at least an episodic biological connection to the Atlantic during the Lago Mare phase (Carnevale et al., 2006).

Only after the Miocene–Pliocene boundary (5.33 Ma; Lourens et al., 1996; Van Couvering et al., 2000) did planktic and benthic communities of the Mediterranean recover (e.g. Wright, 1979; Thunell et al., 1991; Pierre et al., 2006; Rouchy et al., 2007; Sprovieri and Hasegawa, 1990; Sgarrella et al., 1997, 1999). Open ocean assemblages persist throughout the MSC in both the Guadalquivir (Spanish) and South Rifian (Moroccan) corridors (Feinberg, 1986; Kouwenhoven et al., 2003; Pérez-Asensio et al., 2012b).

**3.4.1.2. Biostratigraphic implications.** Increasing restriction of Mediterranean–Atlantic exchange has impacted biostratigraphic datums. Biostratigraphic correlations within the Mediterranean and between the Mediterranean and Atlantic are unequivocal for most of the pre-MSC Messinian, up to the sinistral-to-dextral coiling shift of *Neoglobobulimina acostaensis* (e.g. Sierro et al., 1993, 2001; Hilgen et al., 2000; Krijgsman et al., 2004; Larrasoana et al., 2008). However, the last common occurrence of *Globobulimina miotumida* at 6.28 Ma in the Atlantic (Sierro et al., 1993; Krijgsman et al., 2004) is a diachronous bioevent in the Mediterranean where the last influx of the species has been identified as early as 6.6 Ma, while some rare specimens were found in levels as young as 6.25 Ma (Sierro et al., 2001). This may relate to environmental sensitivity in planktic foraminifera. For the same reason, calcareous nannofossil datums have been found to deviate between Atlantic and Mediterranean strata (e.g. Raffi et al., 2003).

**3.4.1.3. Palaeodepth reconstructions in the corridors.** Reliable records of water-depth change in the Mediterranean–Atlantic corridors are critical to reconstructing gateway exchange. Several methods based on microfossils have been used to reconstruct palaeo-water depths, including (1) the ratio between planktic and benthic foraminifera (P/B ratios), (2) the occurrence of benthic foraminiferal species with restricted depth ranges that are assumed to be constant (e.g. Van Hinsbergen et al., 2005), and (3) transfer functions (e.g. Van der Zwaan et al., 1990; Hohenegger, 2005; Báldi and Hohenegger, 2008). For instance, vertical movements of the Taza-Guercif basin (Rifian Corridor), reconstructed using P/B ratios indicate rapid shallowing from outer shelf-upper slope (~400 m depth) to near-shore depths of ~40 m between 7.2 and 7.1 Ma (Fig. 2; Krijgsman et al., 1999b). P/B ratios were also used to reconstruct palaeodepths in the Murcia-Cartagena and Sorbas basins, to constrain differential vertical movements and tectonic uplift in the Murcia basin (Krijgsman et al., 2006). Problems associated with the exclusive use of P/B ratios are discussed in Van Hinsbergen et al. (2005) and Pérez-Asensio et al. (2012b). Apart from the effect of environmental factors such as oxygen and food on benthic assemblages

(causing, among others, a correlation with astronomical cyclicity; Van Hinsbergen et al., 2005), these problems include reworking and down-slope transport of sediment. In Late Miocene sediments from the Guadix section in the Betic Corridor, significant reworking of Cretaceous foraminifera and sediment transportation precluded the use of P/B ratios for depth reconstruction. Instead, depth-diagnostic species were used to approximate palaeodepths, suggesting a rather rapid shallowing from ~500 to ~200 m (Hüsing et al., 2010). In the Guadalquivir area, palaeodepth reconstructions which also document shallowing as the basin infilled are based on a combination of methods including a transfer function (Pérez-Asensio et al., 2012b) or a transfer function alone (Pérez-Asensio et al., 2013). Despite each method having its limitations, valuable information can be extracted regarding differential vertical movements within the corridor area. Disentangling whether the shallowing seen is caused by tectonic uplift, sedimentary infill and/or Mediterranean sea-level fall remains challenging.

#### 3.4.2. Terrestrial biota and mammal migration

The progressive closure of the Mediterranean–Atlantic marine connections should have created a land-bridge that permitted mammal migration between Africa and Spain. The continental (fossil mammal) biostratigraphic record is less detailed than the marine record, because the richest fossil localities are usually found in small scattered outcrops, and well-dated successions are scarce. Nevertheless, it has long been recognized that typical African species like camels and gerbils (desert rats) appear in the Messinian fossil records of Central Spain (Jaeger et al., 1975; Pickford et al., 1993; van Dam et al., 2006). Similarly, typical European rodent species are observed in the Messinian successions of northern Africa (Coiffait et al., 1985; Garcés et al., 1998; Jaeger, 1977).

One of the most significant Messinian mammalian events is marked by the entry into southern Spain of the murid *Paraethomys miocaenicus* (Agustí et al., 2006; Gibert et al., 2013) and camels of the genus *Paracamelus*. Camels originated in North America and *Paraethomys* in Asia, but the absence of their fossil remains in Western Europe suggest that they reached Iberia via Africa (Pickford et al., 1993; Van der Made et al., 2006). The first African immigrants (*Paraethomys* and *Paracamelus*) in Spain are magnetostratigraphically dated at ~6.2 Ma (Fig. 2; Garcés et al., 1998, 2001; Gibert et al., 2013). Another Messinian mammalian event is characterised by the dispersal of gerbils into Southern Spain. Gerbils are subdesertic rodents that today inhabit the dry landscapes of northern Africa and southwestern Asia. Their first record in Spain is found just after the basal Pliocene transgression (Garcés et al., 2001) and their presence in Europe is probably directly related to the onset of the MSC and the spread of subdesertic conditions in the Western Mediterranean Basin. The identification of gerbils in the latest Messinian reddish continental beds of the Zorreras Formation, in the Sorbas Basin, is in agreement with this hypothesis (Fig. 2; Martín-Suárez et al., 2000).

The first age constraint on Africa–Iberia mammal exchange came from magnetostratigraphic dating of European mammalian fossils in the Ait Kandoula Basin of Morocco, which are correlated with chron C3An.1n at an age of ~6.2 Ma (Benammi et al., 1996). These data indicate that mammal exchange in the Gibraltar area took place in both directions more than 200 kyr before the onset of the MSC (Fig. 2) and this points to an ephemeral Messinian land bridge between Morocco and Spain, indicative of a pre-MSC (partial) closure of the Mediterranean–Atlantic gateways (Agustí et al., 2006; Gibert et al., 2013).

#### 4. Evolution of the Mediterranean–Atlantic gateways: indirect approach

The geological records of the Late Miocene marine connections that are now exposed on land are incomplete as a consequence of the unconformities associated with corridor closure and uplift (Fig. 2) and the uncertainty as to corridor location. As a result, important information about the nature and timing of Mediterranean–Atlantic exchange can only be derived from records outside the corridors that respond to



some aspect of the exchange or the lack of it. In the Atlantic a variety of water mass tracing methods (contourites and Nd and Pb isotopes) capture aspects of Mediterranean outflow and consequently provide information about the timing of closure and the vigour of exchange. In the Mediterranean, Sr isotopes serve as an indicator of isolation from the global ocean. The successions preserved within the Mediterranean and elsewhere during the Late Miocene provide geological constraints for numerical modelling experiments, both those investigating the processes occurring during the MSC and those considering its consequences for global ocean circulation and climate. This section summarises the information about Mediterranean–Atlantic exchange that can be deduced from these diverse indirect approaches.

#### 4.1. Evidence of Mediterranean–Atlantic connectivity from outside the Mediterranean

##### 4.1.1. Mediterranean outflow

As a mid-latitude semi-enclosed marginal basin, the Mediterranean Sea plays a fundamental role in supplying dense waters to the global ocean (Price and Baringer, 1994; Price et al., 1993) impacting the thermohaline structure of the North Atlantic (Artale et al., 2002; Hecht et al., 1997; Mauritzen et al., 2001) and ultimately global climate (Li, 2006). Today, two-layer flow exists in the Straits of Gibraltar, and colder, more saline Mediterranean water (Mediterranean Outflow/Overflow, MO) flows down the continental slope. En route it entrains significant quantities of Atlantic water (Baringer and Price, 1999) which decreases the density and velocity of the resulting water mass and causes it settle out into the Atlantic at intermediate depths (~500–1400 m; Fig. 3; Ambar and Howe, 1979). This distinctive water mass which is the combination of MO and ambient Atlantic water, we will refer to as Atlantic Mediterranean Water (AMW) as defined by Rogerson et al. (2012). Although MO undergoes rapid dilution due to mixing and entrainment processes (Dietrich et al., 2008), the resulting AMW remains a well-defined water mass in the Gulf of Cadiz (Figs. 1 & 7). Subsequently AMW divides into two distinct pathways and can be traced both westward to the Bermuda Rise and northward, over most of the central North Atlantic basin (Armi and Bray, 1982; Curry et al., 2003; Iorga and Lozier, 1999; Lozier and Stewart, 2008). AMW influences the heat and salt balance of the North Atlantic (Dietrich et al., 2008) and contributes to deep-water formation by keeping relatively high salinities at the surface (Price and Baringer, 1994; Reid, 1979).

The interaction of AMW with the Iberian margin's slope system results in an extensive contourite depositional system in the Gulf of Cadiz (Fig. 1), visible in both seismic profiles and bathymetry (García et al., 2009). Thick sedimentary deposits generated by these currents and by bottom currents on the eastern (Alborán Sea) side of the Gibraltar Straits, provide extensive records of past Mediterranean–Atlantic dynamics (Rogerson et al., 2010; Stow et al., 2013). These records, along with observational data suggest that Mediterranean–Atlantic exchange exhibits significant variability over seasonal (García Lafuente et al., 2007), interannual (Lozier and Sindlinger, 2009), and glacial–interglacial (Rogerson et al., 2005; Voelker et al., 2006) time scales.

Mediterranean–Atlantic exchange through the Straits of Gibraltar, is assumed to have been established immediately after the Zanclean flood (5.33 Ma; e.g. Iaccarino et al., 1999). However, to date, no direct evidence exists to enable the characterization of MO just after the opening of Gibraltar. IODP drilling in the Gulf of Cadiz recovered turbidites and debrites deposited between ~4.5–4.2 Ma. These indicate the presence of relatively high flow strength in the Early Pliocene (Hernández-Molina et al., 2013, 2014). From 3.8 Ma onwards these deposits developed into an extensive contourite depositional system. AMW circulation strengthened from 3.2–2.1 Ma, where two major sedimentary hiatuses from 3.2–3.0 Ma and 2.4–2.1 Ma indicate strong bottom water currents (Hernández-Molina et al., 2014). The first hiatus has been linked with geochemical evidence of a rise in MO density and it

has been suggested that this intensified Upper North Atlantic Deep Water (NADW) formation (Khélifi et al., 2009, 2014). Rogerson et al. (2012) concluded that a AMW pathway comparable to that of today (Fig. 7) could have been established around 1.8 Ma. (e.g. Llave et al., 2001, 2007; Llave, 2003; Brackenridge et al., 2013). Finally, changes in the distribution and splitting of the upper and lower AMW is likely to have been caused by diapiric reactivation that can be correlated to tectonic events and sea level changes (García et al., 2009; Llave et al., 2007; Rodero et al., 1999).

Consequently, exchange through the Late Miocene Mediterranean–Atlantic gateways may have been quite different to that seen today in the Gibraltar Strait, and the evolution of the Atlantic's sedimentary and geochemical response to MO may well reflect the evolution of the gateway itself. Three hypotheses can be formulated to explain the absence of evidence for MO before the earliest turbidites and debrites. Firstly, the outflow was insufficiently powerful to generate such current flow related deposits (Hernández-Molina et al., 2013). This is consistent with the interpretation of the onset of Gulf of Cadiz contourites and the geochemical signal of outflow as a strengthening or intensification of MO rather than its initiation (Hernández-Molina et al., 2013; Khélifi et al., 2009). Secondly, Mediterranean and Atlantic waters may have had physical properties too similar to leave traceable geochemical evidence of MO in the Atlantic (Rogerson et al., 2010). During the Early Pliocene global climate was warmer and the gateway may have been deeper (Raymo et al., 2006; Esteras et al., 2000). It is suggested that such conditions would not be met at the time of the gateway opening, but there is no proof of this (Rogerson et al., 2012). A third possibility for the absence of Early Pliocene contourites from Leg 339 Sites is that Early Pliocene MO did exist, but AMW did not follow the present-day route along the Iberian margin and therefore it was not recovered during IODP Expedition 339. However, Rogerson et al. (2012b) demonstrated that the relationship between the salinity of Mediterranean Outflow and its flow pathway while not intuitive, is predictable. Increased salinity of MO could result in increased flow velocity and not in a variation of the plume's depth.

##### 4.1.2. Geochemical tracers of Mediterranean outflow – Nd and Pb isotopes

Attempts have been made to deduce the presence of MO in the Atlantic during the Miocene using geochemical tracers such as neodymium (Nd) and lead (Pb) isotopes. While Nd has a residence time in seawater on the order of 200–1000 years (Tachikawa et al., 1999), on a global average Pb is removed from the water column within 10–100 years (Henderson and Maier-Reimer, 2002). These relatively short residence times enable the Pb and Nd isotope systems to vary regionally in seawater. Nd isotopic compositions (expressed as  $\epsilon_{\text{Nd}}$ , the ratio of  $^{143}\text{Nd}/^{144}\text{Nd}$  in a sample normalized to the bulk earth value in parts per  $10^4$ ; Jacobsen and Wasserburg, 1980) are used as water mass tracers for open ocean palaeocirculation reconstructions (e.g. Frank et al., 2002; Robinson et al., 2010; Thomas et al., 2003). Within the Mediterranean–Atlantic gateway region, clarifying  $\epsilon_{\text{Nd}}$  signal provenance is complicated by riverine and aeolian input (e.g. Henry et al., 1994; Sholkovitz and Szymczak, 2000) and boundary exchange at the sediment–bottom water interface (see Lacan and Jeandel, 2005). Nevertheless, since MO and Atlantic Inflow Water (AIW) have measurably different  $\epsilon_{\text{Nd}}$  (–9.4 and –11.8 respectively, Fig. 8; Piepgras and Wasserburg, 1983; Spivack and Wasserburg, 1988; Tachikawa et al., 2004), this isotope system theoretically has the potential to monitor past exchange. Two marine Nd archives have been exploited to investigate Late Miocene Mediterranean–Atlantic exchange: hydrogenous ferromanganese (FeMn) crusts derived from Atlantic seamounts and marine microfossils from Atlantic and palaeo-corridor locations.

FeMn crusts faithfully record both the Nd and Pb isotopic composition of overlying bottom seawater (Frank, 2002). Pb is commonly analysed alongside Nd as it contributes complimentary information such as insight into local changes related to continental weathering and other climate-induced signals (Christensen et al., 1997; Gutjahr

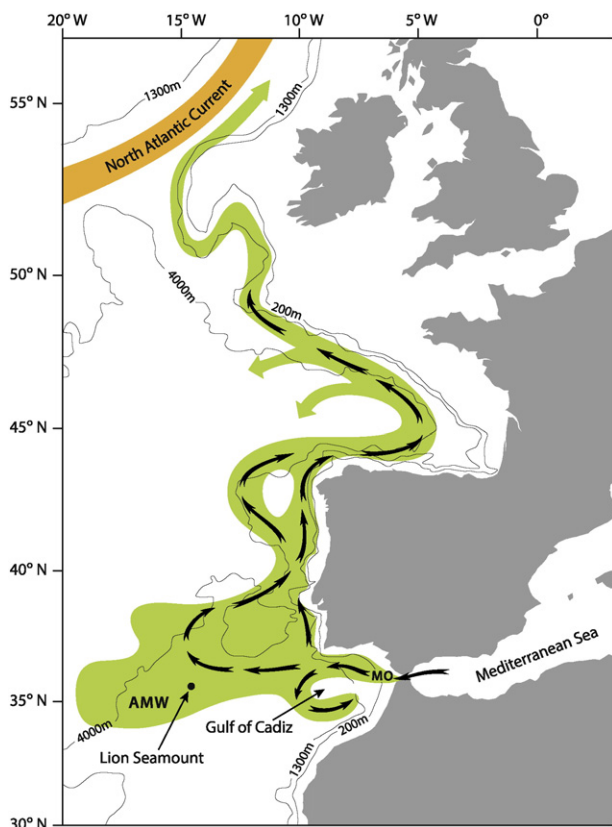


Fig. 7. Overview of the main circulation patterns of modern Mediterranean Outflow (MO) and its pathway in the North Atlantic (Iorga and Lozier, 1999) as Atlantic Mediterranean Water (AMW). In green is the saline tongue centred around 1000 m depth (Reid, 1979) and in orange is the surface North Atlantic Current (NAC). Arrows represent flow direction and the relevant bathymetric features are contoured in black.

et al., 2009; Harlavan and Erel, 2002). Pb and Nd isotope records from the Lion Seamount west of Gibraltar (Fig. 7) which is bathed in AMW today provide no evidence for the cessation of MO during the Messinian (Abouchami et al., 1999; Muiños et al., 2008). Unfortunately, the temporal resolution of both studies is too coarse to clearly rule out changes in Atlantic–Mediterranean exchange during the different stages of the MSC (Fig. 7).

Marine fossils such as fish remains, teeth or bone fragments and foraminifera are an alternative target for Nd isotopic measurements with the potential for much higher resolution records. Ivanović et al. (2013a) studied this archive in sediment samples from the onshore Rifian Corridor. These authors independently estimated palaeo-MO and found palaeo-Atlantic  $\epsilon_{\text{Nd}}$  values which were not inconsistent with Fe–Mn crust studies (Fig. 8; Abouchami et al., 1999; Muiños et al., 2008) and showed that:

- Mediterranean water is likely to have reached the Atlantic through the Rifian Corridor until it closed between 6.64 and 6.44 Ma (Fig. 8),

well before the onset of the MSC. This is consistent with other ages for the timing of closure, but provides tighter constraints;

- The prevailing water in the corridor was likely to have been a mixture of both Atlantic and Mediterranean water between 7.2 and 6.58 Ma (Fig. 8), directly contradicting the Siphon Event hypothesis which advocates that all water passing through the Rifian Corridor during this period was Atlantic in origin (Benson et al., 1991).

Further research is necessary to develop a more robust Nd and Pb isotope signature framework from this time period for the relevant water masses.

#### 4.2. Reconstructing gateway history by comparing Mediterranean and Atlantic records

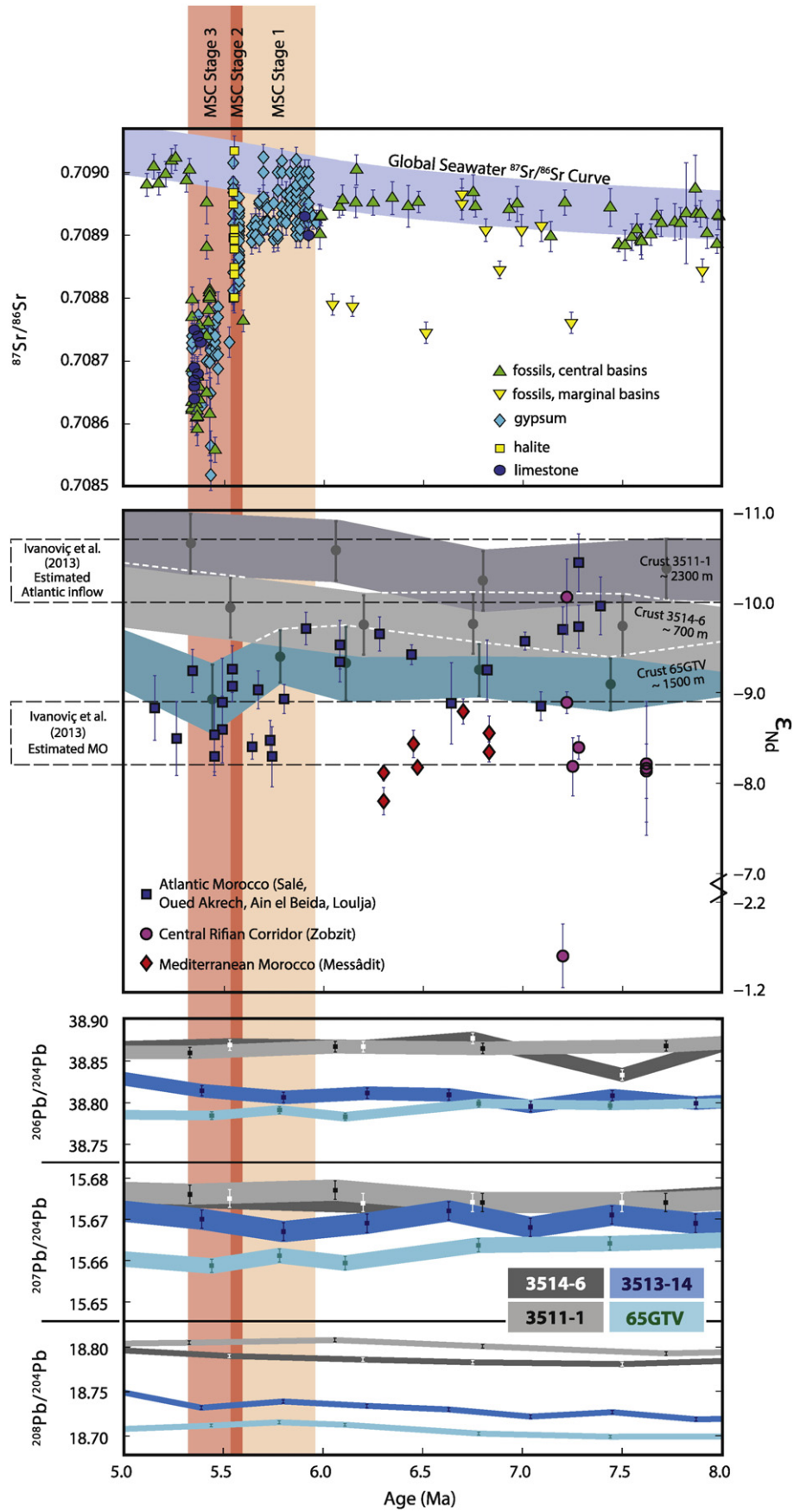
##### 4.2.1. Sr isotopes

Sr isotope stratigraphy is an established marine carbonate dating tool. However, where semi-enclosed basins have limited exchange with the global oceans, they evolve an  $^{87}\text{Sr}/^{86}\text{Sr}$  ratio that deviates from the global ocean water Sr isotope curve (McArthur et al., 2012) towards the  $^{87}\text{Sr}/^{86}\text{Sr}$  of the basin's fluvial input. Such deviating Sr isotope records have been used to infer the connectivity history of various marginal marine systems (e.g. Baltic Sea and San Francisco Bay, Andersson et al., 1994; Ingram and Sloan, 1992). For the Mediterranean, the  $^{87}\text{Sr}/^{86}\text{Sr}$  will deviate measurably from global ocean values when the river discharge constitutes >25% of the total water flux (Topper et al., 2014). It has also been suggested that the changing concentration of Sr in groundwater could affect the  $^{87}\text{Sr}/^{86}\text{Sr}$  values of marginal basins. During periods of aridity (occurring during low insolation), a build-up of Sr in groundwater may occur. On transition to wetter periods that occur during insolation maxima, this excess is flushed into marginal basins, causing  $^{87}\text{Sr}/^{86}\text{Sr}$  to deviate away from global ocean values towards ratios that reflect the local geology (Schildgen et al., 2014). As a large proportion of geology around marginal Mediterranean sub-basins is composed of Mesozoic age carbonate or igneous rock both of which have low  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios (e.g. Albaredo and Michard, 1987; Flecker and Ellam, 1999; Schildgen et al., 2014), the resulting deviation would be towards values lower than those expected from the seawater curve for the age of deposition (McArthur et al., 2001).

Around 8 Ma, Sr isotope values from northern marginal basins indicate a divergence from global values e.g. in southern Turkey (Flecker and Ellam, 1999), the Adriatic (Montanari et al., 1997), and Tyrrhenian Seas (Müller et al., 1990). Coeval records from central Mediterranean areas such as Sicily and Crete (Flecker et al., 2002; Flecker and Ellam, 2006; Sprovieri et al., 2003), maintain global values (Fig. 8). This has been interpreted as indicating significant restriction of these northern marginal basins from the main body of the Mediterranean prior to the MSC (Flecker and Ellam, 1999). At the onset of the MSC, the anomalously low Sr isotope values in the northern marginal basins return back to values within error of the ocean water curve (Flecker et al., 2002). Model analysis by Topper et al. (2011) indicates that although a marine transgression would generate these oceanic Sr isotope values (Flecker et al., 2002), a transgression alone would not account for gypsum precipitation which marks the onset of the MSC. Consequently gateway

Fig. 8. Radiogenic isotope records for the Mediterranean Sea and Mediterranean Outflow (MO) from 8 to 5 Ma, with Messinian Salinity Crisis stages. Top)  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios of Mediterranean water inferred from fossils, gypsum, halite, and limestone, plotted with the global  $^{87}\text{Sr}/^{86}\text{Sr}$  seawater curve. Fossils are divided according to marginal or more central Mediterranean locations. Compilation based on Roveri et al. (2014b) and Topper et al. (2011) and references therein. Middle)  $\epsilon_{\text{Nd(t)}}$ . Point data represent inferred bottom water values at various locations from the Rifian Corridor (present day Northern Morocco), estimated from foraminiferal and fish tooth Nd (Ivanović et al., 2013a). These data are compared to estimates for MO and North Eastern Atlantic Deep Water (NEADW) inferred from microfossils (Ivanović et al., 2013a) and ferromanganese crusts (hereafter 'crusts'; Abouchami et al., 1999; Muiños et al., 2008). 65GTV and 3514-6 are thought to represent MO [4,5]; 3511-1 is thought to represent NEADW (Muiños et al., 2008). 65GTV values recalculated for post-deposition ingrowth of  $^{143}\text{Nd}$  using  $^{147}\text{Sm}/^{144}\text{Nd} = 0.115$  (assuming the value from Muiños et al., 2008) and  $^{147}\text{Sm} \text{ } t_{1/2} = 1.06 \times 10^{11} \text{ y}$ ; all crust value ages recalculated using the latest  $^{10}\text{Be}$  half-life of 1.387 Ma (Chmeleff et al., 2010). Bottom) Relatively flat Pb isotope ratio time series reported for major water masses outside the Gulf of Cadiz as inferred from crusts 65GTV and 3514-6 (potential AMW archives), 3511-1 (potential NEADW archive) and 3513-14 (potential Antarctic Bottom Water archive, Muiños et al., 2008). All errors for Nd and Pb isotope ratios are shown as 2 standard deviations of the external reproducibility, unless the internal reproducibility was larger.





restriction must have played a dominant role in this transition (Topper et al., 2011).

During the MSC, the Mediterranean basin responded to reduced exchange resulting in a deviation away from ocean  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios to lower, fluvial-like values, reflecting a decline in the proportion of ocean water reaching the basin (Fig. 8). The lowest values are recorded during Stage 3 throughout precipitation of the Upper Evaporites and Lago Mare. Recently, Roveri et al. (2014b) assessed the available  $^{87}\text{Sr}/^{86}\text{Sr}$  data for the evaporites and determined correlations between marginal and deep basin evaporite units which, although potentially controversial at present, may lead to an enhanced understanding of evaporite deposition during the MSC. Sr isotope ratios returned abruptly to global ocean values at the Mio–Pliocene boundary (Fig. 8) indicating re-establishment of exchange with the Atlantic.

#### 4.2.2. Stable isotopes and glacioeustatic sea level change

Open ocean oxygen isotopes provide a record of global glacioeustatic sea level change (Fig. 2). High resolution astronomically tuned age-models are available both for sections within the Mediterranean and on the adjacent Atlantic margin. This allows stable isotope records close to the Mediterranean–Atlantic gateway and further afield (e.g. Hodell et al., 2001; Shackleton et al., 1995a,b) to be compared with the timing of key MSC events within the Mediterranean. This comparison provides a means to disentangle glacioeustatic and tectonic controls (e.g. Govers et al., 2009; Krijgsman and Meijer, 2008; Manzi et al., 2013; van der Laan et al., 2006). The idea that expansion of the Antarctic ice-sheet and associated glacio-eustatic sea-level lowering played a critical role in the restriction and subsequent isolation of the Mediterranean during the latest Miocene goes back several decades (e.g. Adams et al., 1977). This early notion is consistent with what now may be called the Messinian glacial interval which is characterized by dominantly obliquity controlled glacial cycles, and started at 6.29 Ma with glacial stage C3An.180.16 and ended with a major deglaciation following glacial TG12 at 5.54 Ma (Hodell et al., 2001; van der Laan et al., 2005, 2006). Although it has become clear that tectonics are likely to have played a more important role than eustasy in the overall isolation of the Mediterranean, individual peak interglacials with sea-level changes in the order of tens of metres can still determine critical steps in the evolution of the MSC.

The oldest conspicuous Messinian glacial C3An.180.16 coincides with a marked change observed in the pre-evaporite succession towards higher surface water salinity and stressful conditions for the marine microfauna (Blanc-Valleron et al., 2002). The onset of evaporite (gypsum) formation at 5.96 Ma did not seem to be related to glacio-eustatic sea-level fall (Krijgsman et al., 2004), although such a claim has recently been made following the inclusion of a discontinuous older gypsum cycle in the Sorbas basin in Spain (Manzi et al., 2013; Pérez-Asensio et al., 2013).

According to current high-resolution age models, the climax phase of the MSC during Stage 2 halite precipitation (Fig. 2) coincides with the twin peak glacials TG12 and 14. It has been suggested that these may correspond to unconformities in the succession when the connection with the Atlantic was fully blocked (Hilgen et al., 2007; Roveri et al., 2008). The onset of the Upper Evaporites may be coincident with the stepwise deglaciation following TG12, potentially explaining the marine faunas observed in this unit (van der Laan et al., 2006; Hilgen et al., 2007). The Zanclean reflooding of the Mediterranean following the MSC has also been related to deglaciation and glacio-eustatic sea-level rise, in this case of interglacial TG5 (Shackleton et al., 1995b; Suc et al., 1997). However, subsequent studies led to a revision of the tuning that did not confirm this causal connection (van der Laan et al., 2006).

#### 4.3. Using box models to understand the Messinian gateway problem

Budget or box models based on the laws of physics and chemistry have provided valuable insight into some of the longstanding, first

order questions about the Mediterranean's Late Miocene water and evaporite budgets and its gateways. Most of these studies are based on the notion of water conservation, which must be maintained for Mediterranean sea level to remain constant:

$$Q_{\text{in}} + \text{Precipitation} + \text{River input} = Q_{\text{out}} + \text{Evaporation} \quad (1)$$

where  $Q_{\text{in}}$  and  $Q_{\text{out}}$  are the fluxes from the Atlantic to the Mediterranean and from the Mediterranean to the Atlantic, respectively (Fig. 3).

Components like evaporite minerals or Sr isotopic ratios which are controlled by the water budget and for which there are palaeo-records, are added to this equation to constrain modelled scenarios. This allows quantitative testing of many of the hypotheses that have been constructed to explain different aspects of the MSC.

Sonnenfeld and Finetti (1985) were the first to apply this approach to the MSC. These authors calculated the volume of fresh water that must be lost from seawater by evaporation in order for gypsum (81%) and halite (90%) to precipitate. This has been built on subsequently by a variety of authors interested in calculating the time taken for different saturation states to be reached (Debenedetti, 1982; Meijer, 2006), or for desiccation and refilling of the basin (Blanc, 2000, 2002; Meijer and Krijgsman, 2005). As to the magnitude of the exchange fluxes, Meijer (2006) shows that in order to reach MSC saturations levels the exchange fluxes have to reduce to a few percent relative to modern values. Also, Krijgsman and Meijer (2008) and Topper et al. (2011) show that during the precipitation of Stage 1 Primary Lower Gypsum (Fig. 2) the Mediterranean outflow must have continued, but may well have been cut off during Stage 2 halite formation.

Including hydraulic-control theory in box models allows the translation of gateway depth and the depth of the interface between the two water mass layers, to their corresponding exchange fluxes. These exchange fluxes can then be further linked to basin salinity or isotope ratios. Rohling et al. (2008) used this approach to calculate basin salinity as a function of strait depth and showed that the gypsum-clastic alternations that dominate Stage 1 of the MSC (Fig. 2) and had been widely considered to be driven by precession (e.g. Vai, 1997; Krijgsman et al., 1999a and b; Hilgen et al., 2007) could result from sea level variation (Rohling et al., 2008). The Mediterranean's cyclic evaporites were also the target record of a study that explored the role of erosion and tectonic uplift on an open gateway to the Atlantic (Garcia-Castellanos and Villaseñor, 2011). Mediterranean sea-level variation related to eustatic changes combined with uplift in the gateway region have been modelled and compared to seismic data by Gargani and Rigollet (2007).

Further investigation of hydraulic-control theory by Meijer (2012) shows: (1) the depth of the marine corridor connecting to the Atlantic must be reduced to a few tens of metres to result in gypsum saturation in the Mediterranean; (2) blocked Mediterranean outflow results if the depth is reduced to a few metres and (3) the relationship between salinity and corridor depth is highly non-linear. This suggests that even a gradual shallowing of the gateway sill can result in an abrupt salinity rise in the enclosed basin leading to an apparent evaporite precipitation event.

Typically, Late Miocene budget calculations consider the Atlantic gateway to resemble Gibraltar (e.g. Meijer, 2012 and references therein). This is clearly a simplification given the evidence of more than one concurrent corridor and the possibility that these may have had multiple strands (Figs. 1, 5 & 6). In order to investigate the gateway dimension problem, Topper and Meijer (2015) used a regional ocean model to examine how a restricted Mediterranean–Atlantic connection would influence both Mediterranean thermohaline circulation and water properties. This study indicates that Mediterranean–Atlantic exchange is proportional to sill depth and the results of these simulations can therefore be used to interpret the Late Miocene sedimentary record both preceding and during the MSC Topper and Meijer (2015). Sill depths below 10 m result in blocked Mediterranean Outflow, which could represent a plausible scenario for the second stage of the

MSC (Fig. 2), characterised by the rapid accumulation of halite in the Mediterranean (Topper and Meijer, 2015).

## 5. Causes and consequences of Late Miocene Mediterranean–Atlantic exchange

### 5.1. Tectonic drivers of gateway evolution

There are a variety of different tectonic processes that contributed to the evolution of the Mediterranean–Atlantic connections during the Late Miocene–Pliocene. The main tectonic drivers of the location of the connections are likely to have been the combined influence of African–Iberian convergence with slab rollback and westward motion of the Alborán Domain. Regional uplift will also have played a significant role in the closure of these marine connections and this is likely to have resulted from lithospheric delamination and asthenospheric upwelling in the region (Duggen et al., 2003, 2004). In Morocco for example, southward propagation of the thrust load appears to be a significant driver of both formation and closure of the Rifian corridors (Fig. 1) along with regional uplift attributed to slab dynamics (Duggen et al., 2004).

During the MSC itself, loading and unloading of the lithosphere provides another tectonic mechanism for vertical movement in the gateway area. For example, loading of the lithosphere with thick evaporites results in flexural subsidence of the Mediterranean's basin and uplift of its margins contributing to reduced connectivity while sea level lowering has the opposite effect, but not necessarily by the same magnitude (Govers et al., 2009). The relative importance of these two processes depends on the thickness and location of evaporites precipitated, the amount of Mediterranean sea level fall and the relative timing of the two. All three of these are controversial with the largest degree of uncertainty associated with the relative timing of events during stage 2 (Fig. 2). This is perhaps best illustrated by the persistence of three contrasting scenarios for halite emplacement and sea level fall each of which would result in a different subsidence/uplift history for the marginal gateway region:

1. deep water emplacement where halite is precipitated during a moderate base level fall (Roveri et al., 2008);
2. halite precipitation during a large relative sea level fall (Lofi et al., 2011b; Ryan, 2009); and
3. halite precipitation after desiccation of the Mediterranean basin (Bache et al., 2012; Clauzon et al., 1996).

Despite this ambiguity, it is clear that vertical lithospheric adjustment significantly affected river canyons, topographic slopes and erosion rates in and around the basin and was an important contributory driver of the Mediterranean's connectivity history through its impact on the gateway region. One extension of this was the work by Garcia-Castellanos and Villaseñor (2011) who explored the interplay between uplift of the gateway region and erosional deepening of the corridor as a result of Atlantic inflow. The numerical model they produced illustrates that this interplay provides a mechanism for maintaining at least Atlantic inflow over relatively long timescales despite progressive tectonic uplift. The harmonic behaviour of this interplay also led them to suggest that the cyclicity in the evaporites might have a tectonic-erosion origin rather than the widely held climatic one (Garcia-Castellanos and Villaseñor, 2011).

Another example of the possible role of tectonics and erosion in the evolution of the Mediterranean–Atlantic connectivity concerns the end of the MSC. At this point, as a result of evaporite loading, the Alborán Basin was bordered by a peripheral dam along the Spanish, Gibraltar and northwest Africa margins (Govers et al., 2009). This impeded the reconnection of the Mediterranean to the global oceans. Breaching this barrier with the opening of the Gibraltar Strait has been attributed to regional subsidence as a consequence of the evolution of the Gibraltar slab (Govers et al., 2009).

While no-one disputes the role of tectonics in the evolution of the Mediterranean–Atlantic gateway, disentangling the relative importance of the different processes for individual events is challenging. What the sedimentary record of the Mediterranean does allow us to do is identify key moments at which restriction or opening of the gateway must have occurred. It is then possible to evaluate the probable roles of tectonics, eustatic sea level change and erosion in effecting this change. These are summarised in Fig. 9.

### 5.2. Climatic drivers of the MSC and their impact on Mediterranean–Atlantic exchange

It is clear that climate played an important role in the evolution of the MSC. Numerical modelling has been used to explore aspects of this relationship, particularly in relation to the Mediterranean's hydrologic budget at times of restricted Mediterranean–Atlantic connection and may help to explain the salinity fluctuations that took place during the MSC (see also Krijgsman and Meijer, 2008). As a consequence of salinity changes in the Mediterranean, the density contrast between the Mediterranean and Atlantic will have varied thus impacting the vigour of exchange. Using a global atmosphere-only GCM, Gladstone et al. (2007) demonstrated that the Mediterranean freshwater budget in the Late Miocene may have been closer to a neutral position than it is today making it easier for climatic change to switch the sign of the hydrologic budget from negative, which would result in higher salinities, to positive, which reduces Mediterranean salinity below ocean water values. Gladstone et al. (2007) also estimated river runoff around three times greater than today mostly as a consequence of input from North African rivers feeding the Eastern part of the basin. Many of these rivers, which are dry today, are thought to have transported water from the south (Griffin, 1999) as a result of a stronger African summer monsoon (Marzocchi et al., 2015; Gladstone et al., 2007). However, on precessional time scales, wetter periods in the Mediterranean region may also have resulted from enhanced wintertime storm track activity in the Atlantic and associated increased precipitation (Kutzbach et al., 2014). It is these two independent precession-driven processes which may be responsible for the formation of similarly cyclic Late Miocene sedimentation both within the Mediterranean, responding mainly to the North African monsoonal signal and outside it along the Atlantic coast where the driver is rainfall and runoff from Atlantic storms. Evaluation of the role of the Mediterranean's freshwater fluxes in controlling both its environmental evolution and exchange through its gateways is in its early stages, hampered by inadequate rainfall data as well as model-data mismatch on temporal as well as spatial scales.

Another modelling approach used is to combine GCMs with regional ocean-only models. This can be achieved by forcing ocean-only models with output from fully-coupled global simulations, and then running them at sufficiently high resolution to resolve more realistically the different gateway scenarios. For instance, Meijer and Tuenner (2007) combined an intermediate complexity, global-scale atmosphere-ocean model with a more detailed regional model for the circulation of the Mediterranean Sea to investigate the consequences of precession-induced changes in the Mediterranean freshwater budget, linking it to the Late Miocene sedimentary cyclicity.

### 5.3. The impact of the MSC on global climate

Today MO entrained in the North Atlantic current is thought to contribute to North Atlantic circulation by supplying warm saline waters to sites of North Atlantic Deep Water formation in the Greenland–Iceland–Norwegian (GIN; Fig. 7) seas and northernmost Atlantic (McCartney and Mauritzen, 2001; Reid, 1978, 1979). This hypothesis has been investigated through the removal of MO water from several North Atlantic ocean circulation modelling experiments, both for the present day (Ivanović et al., 2014a; Wu et al., 2007; Kahana, 2005;



Chan and Motoi, 2003; Artale et al., 2002; Rahmstorf, 1998) and the Quaternary (Rogerson et al., 2010; Bigg and Wadley, 2001). According to these studies, the presence of AMW in the North Atlantic appears to have a negligible effect on modern climate. An enduring question is therefore whether the MSC had any significant impact on global climate.

In order to answer this question it is first necessary to establish the differences between the Miocene and present day climatic system. Major changes include:

- A Central American Seaway that linked the Atlantic and Pacific oceans during the Late Miocene to Early Pliocene (Duquecaro, 1990; Keigwin, 1982; Osborne et al., 2014; Sepulchre et al., 2014). This implies a significant difference in ocean circulation patterns and may have resulted in considerably weaker Messinian North Atlantic Deep Water formation than today (Boehme et al., 2008; Herold et al., 2012; Lunt et al., 2008; Molnar, 2008; Murdock et al., 1997; Prange and Schulz, 2004; Schneider and Schmittner, 2006; Zhang et al., 2012).
- Major post-Miocene uplift of the Himalayas, Andes, Rockies, Alps and East African Plateau (see Bradshaw et al., 2012 and references therein) that will have impacted significantly on the patterns of atmospheric circulation.
- An extensive Late Miocene Antarctic ice-sheet (e.g. Lewis et al., 2008; Shackleton and Kennett, 1975), but more limited Northern Hemisphere glaciation than today (Kamikuri et al., 2007; Moran et al., 2006).
- In addition, despite the overall cooling trend during the Cenozoic (Zachos et al., 2001), the global climate proxy record for the Late Miocene suggests that it was generally hotter and/or wetter than today (Bradshaw et al., 2012; Bruch et al., 2007; Eronen et al., 2010; Pound et al., 2011, 2012; Utescher et al., 2011).

These significant differences in the climatic configuration mean that evaluation of the impact of the MSC on global climate requires Miocene specific model experiments.

Possible mechanisms for an MSC influence on climate include: extreme changes in MO density and volume reaching the Atlantic; a substantial reduction in sea level and evaporative flux during draw-down; associated changes in circum-Mediterranean vegetation; and a significant reduction in global ocean salinity as a result of salt sequestration in the Mediterranean. General Circulation Models (GCMs) are a powerful tool for simulating the interactions between the main components of the global climate system. They have therefore been used to assess some aspects of the possible impact of the MSC on global climate.

A significant reduction in Mediterranean outflow is an essential aspect of raising Mediterranean salinity and although removal of AMW in the North Atlantic has a negligible impact on modern climate (Ivanović et al., 2014a,b; Wu et al., 2007; Kahana, 2005; Chan and Motoi, 2003; Artale et al., 2002; Rahmstorf, 1998), during stages of weaker Atlantic Meridional Overturning Circulation (AMOC) such as the Younger Dryas, it does impact North Atlantic Ocean circulation. This matches the Younger Dryas sea surface salinity and temperature records from the Iberian margin and Alborán Sea (Penaud et al., 2011; Voelker et al., 2006), and implies that, since AMOC was weaker in the Messinian, it should have been more sensitive to Atlantic salinity and temperature variations driven by AMW (Ivanović et al., 2014a,b).

The impact of extreme changes in MO properties on ocean circulation and climate has also been investigated. Coupled ocean–atmosphere GCM simulations suggest that elevating Mediterranean salinity enhances salt export from the Mediterranean, which in turn modifies the rate of deep water formation and circulation in the Atlantic Ocean (Ivanović et al., 2014a,b). In the model, this triggered cooling (9 °C in the Boreal winter–spring) in Northern mid-high latitude surface air temperatures, but did not result in significant changes in atmospheric circulation or precipitation patterns (Ivanović et al., 2014b). However, episodes of extremely elevated or negative Mediterranean salt-export are most likely to have occurred only intermittently (Thierstein and

Berger, 1978) and for short periods of time (Meijer and Krijgsman, 2005). This scenario is not well captured by published model experiments which maintain MO salinity levels for hundreds of years until climate equilibrium is achieved in the model. The modelling results of Ivanović et al. (2014b) exhibit an initial decadal-scale overshoot in ocean circulation, suggesting that the shorter-term (transient) ocean circulation impact of the MSC may therefore have been far more extreme than the multi-decadal averages calculated in these published studies.

The dimensions of the Straits of Gibraltar (nearly 60 km long, 12 km at its narrowest point, and with a maximum depth of 300 m; Candela et al., 1990) make realistically simulating the thin, dense current spilling over the sill (the MO) in numerical models very challenging (Dietrich et al., 2008). In models with a relatively coarse resolution where the model grid cannot resolve the features of the shallow, narrow strait, the exchange is simulated either using a less realistic wider and deeper open seaway (Ivanović et al., 2013b; Rogerson et al., 2010; Bigg and Wadley, 2001), or using an empirical parameterisation of the exchange (Ivanović et al., 2013b; Wu et al., 2007; Chan and Motoi, 2003; Rahmstorf, 1998). Clearly, the same resolution problem also applies to both the Late Miocene Mediterranean–Atlantic seaways.

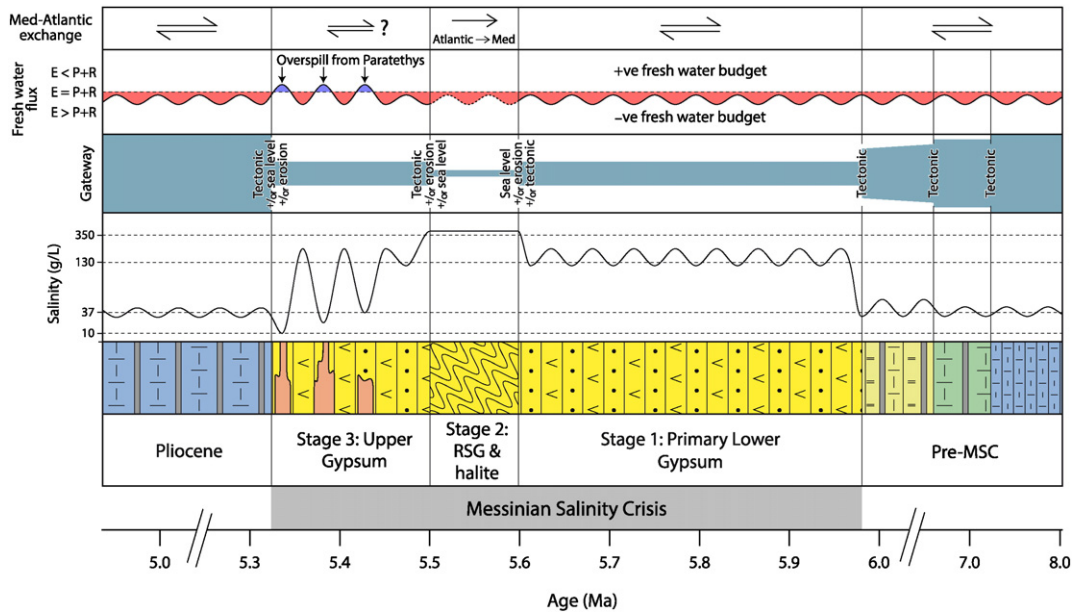
Climate-modifying aspects of the MSC other than MO have also been investigated through modelling. The impact of MSC-driven Mediterranean sea level and vegetation change have been explored using an atmosphere-only GCM and an intermediate complexity Earth system model, respectively. These experiments suggest that lowering Mediterranean sea level results in strong cooling over the North Pacific, which may have enhanced high latitude glaciation in the Late Miocene (Murphy et al., 2009). The vegetation experiments did not exhibit dramatic climate changes close to the margins of the Mediterranean basin during the MSC, but instead indicated cooling in Central, Northern and Eastern Europe (Schneck et al., 2010).

In summary, the impact of the MSC on global climate remains poorly established. This is due partly to shortcomings in palaeoclimate model configurations (Ivanović et al., 2014b), partly to difficulties modelling a realistic MSC scenario with an appropriately small gateway, and more generally to sparse and irregular distribution of Late Miocene climate proxy data (Bradshaw et al., 2012) generating difficulties in evaluating the plausibility of model simulations.

#### 5.4. The challenges of deducing palaeosalinity and its impact on reconstructing exchange

The main control on Mediterranean–Atlantic exchange is the salinity contrast between the two water masses and this is driven by the efficiency or size of the gateway and net evaporation (Section 2.1). Quantifying the evolution of Mediterranean salinity is therefore a key step in reconstructing exchange. The current constraints on extreme palaeosalinity, however, only provide a few threshold values with which to reconstruct the Mediterranean's complex salinity history (e.g. 350 g/L for halite; 130–180 g/L for gypsum; ~50 g/L the upper tolerance limit of foraminifera; 5–20 g/L for brackish water fauna). Consequently, it is not currently possible to quantify the pre-MSC salinity contrast between the Mediterranean and Atlantic. For example, the absence of planktic foraminifera in some intervals of the pre-evaporitic succession of the Sorbas Basin (Sierro et al., 2001), merely indicates that surface water salinity has exceeded their tolerance (e.g. ~49 psu; Fenton et al., 2000 and references therein). Although these salinity thresholds are used to constrain specific exchange scenarios suitable for particular salinity conditions (e.g. gypsum or halite precipitation; Flecker et al., 2002; Topper et al., 2011; Meijer, 2006), another complication is that brine concentration is not always the only possible control on the mineralogical or biological changes observed.

An excellent example of this problem is the salinity reconstruction of the gypsum–clastic cycles that occur around the margins of the Mediterranean during the precipitation of Stage 1 Primary Lower Gypsum



**Fig. 9.** Summary figure illustrating the main features of the Mediterranean's exchange history in the Late Miocene–Pliocene including lithology, Mediterranean salinity, a qualitative representation of gateway size and the probable drivers (tectonics, erosion, sea level) of changing dimensions, the Mediterranean's fresh water flux where  $E$  and  $P$  = evaporation and precipitation over the Mediterranean respectively and  $R$  is the river discharge into the Mediterranean Sea, and arrows representing one-way or two-way exchange between the Mediterranean and Atlantic.

and Stage 3 Upper Gypsum (Fig. 2). The clastic intervals in these alternations may be organic rich. They are always either abiotic, or if they do contain foraminifera, these show signs of being reworked (e.g. Rouchy and Caruso, 2006). Three different salinity records are possible all of which have different implications for exchange at the time:

1. The clastic sediments between the gypsums may have been deposited when the brine was diluted by additional fresh or marine water generating salinity conditions that were too high to support marine fauna, but lower than the  $\sim 130$  g/L salinity and  $5.25$  g/L  $\text{CaSO}_4$  concentration (Topper and Meijer, 2013) required for gypsum precipitation. The density contrast with the Atlantic was therefore at its highest during gypsum precipitation implying the smallest, least efficient gateway with an increase in exchange during clastic deposition;
2. The clastic sediments between the gypsums may have been deposited during higher salinity conditions when brine concentration exceeded the maximum associated with gypsum precipitation ( $\sim 180$  g/L NaCl). Were this to have happened, exchange would have been most limited during deposition of the clastics;
3. Finally it is possible that the lithological variation does not reflect a change in salinity but rather the availability of sulphate (Natalicchio et al., 2014). Gypsum precipitates when sulphate is available and the salinity is between  $130$ – $180$  g/L; once the sulphate is used up, it is not possible to precipitate gypsum and clastic sediments accumulate instead. This is the same process that leads to the lateral and depth related shift from shallow water gypsum to deeper water organic-rich shales during Stage 1 Primary Lower Gypsum phase (de Lange and Krijgsman, 2010). The consequences for exchange are that there is very little variation throughout this period.

## 6. Evolution of Mediterranean–Atlantic exchange during the Late Miocene

The Mediterranean's hydrologic budget is controlled by both the efficiency of the gateway(s) and net evaporation over the Mediterranean (precipitation + runoff–evaporation,  $P + R - E$ ; Section 2.1). The combination of these two drivers along with a Mediterranean

circulation system where surface water flows east becoming more saline, sinks in the Eastern Mediterranean and flows west at depth, typically results in a density contrast between Mediterranean and Atlantic water at the gateway where Mediterranean water is both more saline and colder than surface Atlantic water (Rogerson et al., 2012). It is this density contrast at the gateway that drives the pattern and vigour of Mediterranean–Atlantic exchange. While tectonic forcing is the dominant driver of the width, depth, length, location and age of the different gateways, with sea level a contributory factor, for any specific gateway configuration, variability in exchange is modulated by climate which drives the salinity ( $P + R - E$ ) and temperature of the Mediterranean and hence its density contrast with the Atlantic (Section 2.1). There is evidence from the Quaternary that illustrates the independent behaviour and impact of gateway efficiency and net evaporation. The Last Glacial Maximum (LGM) and its associated sea-level fall, provides an example of gateway efficiency dominating exchange and associated salinity rise. Modelling studies suggest that Mediterranean salinity rose to around  $44$  psu as a result of sea-level driven reduction in inflow and outflow during the LGM (e.g. Bethoux, 1984; Rohling, 1999). There is no direct evidence of this salinity rise in the Mediterranean since it is not high enough to exceed the salinity tolerance of planktic foraminifera ( $\sim 49$  psu; Fenton et al., 2000) and  $\delta^{18}\text{O}$  residuals which can be used for reconstructing palaeo-salinity elsewhere appear to be non-proportional to salinity and more influenced by run-off in the Mediterranean (Rohling, 1999; Rohling et al., 2015). However, the same LGM sea-level fall across the silled connection between the Red Sea and the Indian Ocean does reduce exchange sufficiently to produce high salinities that contribute to the development of aplanktic horizons in the Red Sea (Fenton et al., 2000). By contrast, Heinrich Events provide an example of climate-driven variation in exchange. These ice-rafting events generated fresher surface water in the North Atlantic. Coeval with Heinrich Events are episodes of coarser-grained contourites in the Gulf of Cadiz (Voelker et al., 2006). Since contourite grain size is indicative of higher energy MO this succession is interpreted to result from a larger, climate driven density contrast between the Mediterranean and Atlantic as a result of reduced Atlantic salinity and consequently more vigorous exchange (Rogerson et al., 2010; Voelker et al., 2006) during a period when the gateway configuration remained effectively constant.

In the Late Miocene, before the onset of Northern Hemisphere glaciation, a mechanism for modifying Atlantic salinity does not appear readily available. Mediterranean–Atlantic exchange is likely to have varied in concert with Mediterranean salinity on a variety of timescales. Controlling factors include:

1. Tectonic driven changes to the gateways controlling Mediterranean salinity via exchange efficiency (Meijer, 2012);
2. Glacio-eustatic sea-level oscillations with an increasing impact on the efficiency of exchange as the gateways shallow;
3. Precessional fluctuations in the freshwater-flux relating to the supply of North African monsoonal rainfall (Marzocchi et al., 2015) to the Eastern Mediterranean via the Eshabi and Nile rivers (Gladstone et al., 2007; Griffin, 1999, 2002);
4. Longer-term climate change resulting in changing net evaporative loss over the Mediterranean basin.

During the Late Miocene there is little evidence of any significant, long-term change in climate that might account for triggering or terminating the MSC (Bertini, 2006; Bertini et al., 1998; Roveri et al., 2014a; Suc and Bessais, 1990; Suc et al., 1995). Consequently, using what is known about Late Miocene Mediterranean salinity, it is possible to reconstruct the history of Mediterranean–Atlantic exchange and consider the relative impact of tectonics and orbital variability in determining the evolution of the MSC (Fig. 9).

Plio–Quaternary Mediterranean successions are dominated by strong precessional cyclicity that is visually enhanced by the barcode-like dark stripes of the sapropels (e.g. Rohling et al., 2015). Very similar sediments were also deposited in the Mediterranean prior to the MSC (Fig. 2). This similarity combined with the observation that the sedimentary response is coupled to orbital variation, suggests that before and after the MSC the basin responded to orbital-induced climate change in essentially the same way despite the different gateway configurations (e.g. de la Vara et al., 2015) and the warmer and wetter Late Miocene conditions (Bradshaw et al., 2012). Consequently, Late Miocene exchange is assumed to resemble that occurring through the Gibraltar Strait today, with a similar degree of restriction (Fig. 9) leading to an overall slightly enhanced salinity in the Mediterranean (38 g/L) relative to the Atlantic (35 g/L). Just as is seen in the Plio–Quaternary, this tectonically controlled exchange is modulated by precessional changes to the Mediterranean's freshwater flux ( $P + R - E$ ) which remains negative throughout (Fig. 9).

Sediments deposited before the first evaporite precipitated contain evidence indicative of step-wise restriction of exchange. The progressive loss of oxic benthic faunal species (Kouwenhoven et al., 2003; Section 3.1; Fig. 2) and divergence of marginal basin Sr isotope values from coeval ocean values (Flecker et al., 2002; Section 4.2.1; Figs. 2 and 8) suggest enhanced water column stratification and reduction in Atlantic influence. The development of aplanktic marls similar to those seen in the Red Sea during the LGM (Fenton et al., 2000) incorporated into the precessional cyclicity of Sorbas Basin sediments (Sierro et al., 2001; Section 3.4.1; Fig. 2) shows that, here at least, the amplitude of salinity variation increased to levels above planktic foraminiferal tolerance (49 psu; Fenton et al., 2000; Fig. 9). This period between 8–5.97 Ma is the interval during which geological evidence demonstrates the closure of most strands of the Betic and Rifian corridors (Fig. 2). This suggests that exchange was progressively reduced with respect to earlier periods, probably by a series of tectonic events with superimposed ~100 and 405-kyr orbital cyclicity (e.g. Hüsing et al., 2009a; Krijgsman et al., 1999a) and the eustatic sea-level fall associated with glacial C3An.180.16 giving the step-wise pattern of restriction (Fig. 9). Precessional modulation of the freshwater flux continued ( $P - E + R$ ) and remained negative (Fig. 9). However, this resulted in an increasingly high-amplitude salinity response in the Mediterranean (Fig. 9) as a consequence of reduced gateway efficiency.

Primary gypsum dominates both the Primary Lower Gypsum in Stage 1 (Fig. 2) and the Upper Gypsum in Stage 3. Gypsum precipitation

requires salinities of 130–160 g/L. Modelling results suggest that maintaining this salinity can only be achieved if two-way exchange persists (Fig. 9), but is reduced significantly by decreasing the efficiency of the gateway, probably by shallowing it to a few tens of metres (Debenedetti, 1976; Meijer, 2012). Topper et al. (2011) demonstrated that the transition from pre-MSC sediments to gypsum at 5.97 Ma requires a reduction in gateway efficiency and consequently, we infer a further tectonic and/or sea level-driven (Manzi et al., 2013) reduction in the gateway as the trigger for the onset of the MSC (Fig. 9) although a significant increase in Mediterranean stratification may also play a role. Both Stage 1 and Stage 3 comprise regular gypsum–clastic alternations (Figs. 2 and 9) which are also thought to be precessionally driven (Hilgen et al., 2007; Krijgsman et al., 2001; Roveri et al., 2014a) and this suggests on-going precessional modulation of the freshwater flux (Fig. 9). However, because of the problems reconstructing palaeosalinity for these gypsum–clastic cycles (Section 5.3), neither the amplitude nor the phasing of the salinity response is clear (see Topper and Meijer, 2013 for model-based insight).

Stage 2 gypsum is the reworked erosional product of Stage 1 primary gypsum (CIESM, 2008). This, combined with its association with the Messinian Erosion Surface (Figs. 2 and 4) indicates a Mediterranean base-level fall and the timing of Stage 2 is coincident with glacials TG12 and TG14. The implications for Mediterranean–Atlantic exchange are clear, but the scale of this sea-level drop is still contentious (see Roveri et al., 2014a for review). In the CIESM (2008) model, thick (~1.5 km) halite precipitated in the deep Mediterranean basin at the end of Stage 2. This required salinities of >350 g/L and a supply of seawater, most easily sourced from the Atlantic. To achieve these high salinities, outflow must have been negligible so that one-way flow from the Atlantic to the Mediterranean is envisaged (Fig. 9). Because the deep basinal halite has not been drilled, the only direct access to part of this succession is through a mine on Sicily. Here, annual bands in the salt have been identified, but while the peak glacials may correspond to unconformities in the succession, a precessional signal has not so far been demonstrated (Lugli et al., 1999). Given that the precessional signal is clearly visible in Stage 3 and Plio–Quaternary sediments (Fig. 9), it is unlikely that precessional modulation to the freshwater flux was switched off during Stage 2. Alternative explanations include the possibility that this area around Sicily was in some way protected from the impacts of the freshwater flux driven by North African runoff; or that any salinity variation generated by the precessional freshwater flux was small by comparison with the extreme salinity of the Mediterranean as a whole at this time and consequently never moved the basin out of the halite window so that no sedimentary response to the freshwater flux is recorded (Fig. 9).

In terms of Mediterranean–Atlantic gateway exchange, the most enigmatic phase of the MSC is the Stage 3 Lago Mare association of local evaporites and sediments some of which contain fresh to brackish water fauna and flora. These low salinity assemblages which increase in abundance and diversity with time (Roveri et al., 2008) and spread progressively westward, resemble those found in the brackish-water Paratethyan lake system (Orszag-Sperber, 2006; Rouchy and Caruso, 2006; Roveri et al., 2008). This suggests prolonged and increasing connectivity between the Mediterranean and Paratethys, although the location of the Mediterranean's Paratethyan gateway at this time is just as enigmatic as its Atlantic counterpart. The traditional interpretation of the Lago Mare phase is that it occurred during a period when there was negligible connectivity with the Atlantic (e.g. Hsü et al., 1977; Orszag-Sperber, 2006). However, Stage 3 successions also include a variety of features that suggest that the Mediterranean did receive at least periodic incursions of Atlantic water. These include the presence of Atlantic open marine fish (Carnevale et al., 2006, 2008), and dwarf foraminifera (Iaccarino et al., 2008). In addition, the same arguments that are used to infer two-way Mediterranean–Atlantic exchange for Stage 1 Primary Lower Gypsum (e.g. Meijer, 2012; Fig. 9) can also be invoked to explain the similar Stage 3 gypsum–clastic cycles. These indicators



of at least episodic Atlantic connectivity, along with the reduction in Mediterranean salinity from its peak halite concentration, suggests that the transition from Stage 2 to 3 is triggered by more efficient Atlantic gateway exchange resulting either from a tectonic driver and/or erosional opening of the gateway and/or stepwise deglaciation following TG12 (Hilgen et al., 2007; Roveri and Manzi, 2006).

The paradox is that associated with this more efficient gateway are Sr isotope ratios indicative of an environment that is more dominated by continental run-off than at any other time during the MSC (Fig. 8). Consequently, although Stage 3 gypsum–clastic cycles resemble those of Stage 1, the geochemistry indicates distinctly different water sources. The obvious contender as an additional water source is Paratethys, with its low Sr isotope ratio (Flecker and Ellam, 2006; Major et al., 2006) and brackish water salinity and fauna. The Mediterranean during Stage 3 appears to be equivalent to the Black Sea today, with a Bosphorus-like connection with the Atlantic.

The mechanisms for achieving very low salinity conditions in the Mediterranean are either substantial dilution by fresh water from Paratethys (Orszag-Sperber, 2006; Rouchy and Caruso, 2006; Roveri et al., 2008) and/or a change in climate leading to a switching of the Mediterranean's hydrologic budget from negative to positive (Gladstone et al., 2007). Since the Mediterranean successions that contain these Lago Mare sediments commonly show the same strong cyclicity as the gypsum–clastic alternations with which they are interbedded (the Eraclea Minoa section on Sicily is a good example), it is likely that precessional modulation of the freshwater flux is still the driver of changes to both hyper and hypo-saline conditions in the Mediterranean (Fig. 9). The mechanism for this is not yet clear, but the cyclicity may suggest that precession-driven Mediterranean–Paratethys connectivity caused episodic fluctuations between negative and positive P-E + R (Fig. 9). Superimposed on this is the concept of on-going base-level rise in the Mediterranean which may or may not be related to latest Messinian deglaciation (see Roveri et al., 2014a). One possibility supported by the increasingly widespread evidence of low salinity in the Mediterranean is therefore that exchange with Atlantic involved only periodic inflow to a partially filled Mediterranean during Stage 3.

Once tectonic and/or erosional opening of the Atlantic gateway further increased gateway efficiency at the Mio–Pliocene boundary (Fig. 9), the Paratethyan component of the freshwater flux is no longer visible in the Mediterranean sedimentary record. This may be because Paratethys was no longer connected, but more likely, as it is today, the more efficient Mediterranean–Atlantic gateway diminished the amplitude of the Mediterranean's response to subtle hydrologic change via its freshwater flux.

## 7. Conclusions

Marine gateways are an important control on both local environmental change and global climate. The Late Miocene Mediterranean gateway system that linked to the Atlantic is a good example of this and much can be learnt about the processes and impacts of gateway closure from the study of the sediments preserved within the ancient marine corridors in southern Spain and northern Morocco. Uplift and erosion resulting from the same tectonic drivers of gateway closure, has led to the preservation of incomplete sedimentary successions punctuated by unconformities. Despite this, it appears that the main channels, as deduced from the current distribution of Late Miocene sediment in the region, were closed before the precipitation in the Mediterranean of large volumes of halite during the Messinian Salinity Crisis. The whereabouts and dimensions of the connection that supplied Atlantic water to the Mediterranean during this period therefore remain currently unclear.

Additional constraints on Mediterranean–Atlantic exchange have been deduced from studying successions outside the immediate corridor region. Contourites in the Gulf of Cadiz are a direct consequence of Mediterranean Outflow and changes in their properties and location

reflect fluctuations in the vigour of Mediterranean Outflow. Novel isotopic proxies that monitor connectivity have also been used to explore the presence of Mediterranean water in the Atlantic and the amount of Atlantic water reaching the Mediterranean. However, these records are currently too low resolution to capture the sub-precessional scale variability which is such a dominant feature of the hydrologic system active across the region. This in turn limits the constraints the data can provide for modelling experiments that explore both the gateway processes themselves and the impact of variable exchange on regional and global climate. In addition, the lack of a robust salinity proxy able to function across the wide range of salinities that were produced during the Messinian Salinity Crisis is currently a major problem that has consequences for reconstructing gateway exchange because this is driven by the density contrast between the Mediterranean and Atlantic.

Despite these challenges, an integrated review of the wide variety of information pertaining to Mediterranean–Atlantic exchange before, during and after the Messinian Salinity Crisis allows the first order reconstruction of gateway evolution (Fig. 9) and the role of exchange in driving the extreme environmental changes to be deduced.

## Acknowledgements

The research leading to these results has received funding from the People Programme (Marie Curie Actions) of the European Union's Seventh Framework Programme FP7/2007–2013/under REA Grant Agreement No. 290201 MEDGATE. The authors would like to thank Javier Hernández-Molina and Mike Rogerson for their helpful reviews and Dr Carla Sands, MEDGATE's superb Project Manager without whom much of the research would not have happened. CC Martins also thanks CAPES by scholarship support (BEX 5366/12-7).

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