

**SST variability in the  
central-western  
Mediterranean Sea  
during the last 2700  
years**

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**Sea surface temperature variability in the  
central-western Mediterranean Sea during  
the last 2700 years: a multi-proxy and  
multi-record approach**

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Received: 28 September 2015 – Accepted: 6 October 2015 – Published: 18 November 2015

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Published by Copernicus Publications on behalf of the European Geosciences Union.

CPD

11, 5439–5508, 2015

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## Abstract

This study analyses the evolution of sea surface conditions during the last 2700 years in the central-western Mediterranean Sea based on six records as measured on five short sediment cores from two sites north of Minorca (cores MINMC06 and HER-MC-MR3).

5 Sea Surface Temperatures (SSTs) were obtained from alkenones and *Globigerina bulloides*-Mg/Ca ratios combined with  $\delta^{18}\text{O}$  measurements to reconstruct changes in the regional Evaporation–Precipitation (E–P) balance. We reviewed the *G. bulloides* Mg/Ca-SST calibration and re-adjusted it based on a set of core top measurements from the western Mediterranean Sea. According to the regional oceanographic data,  
10 the estimated Mg/Ca-SSTs are interpreted to reflect spring seasonal conditions mainly related to the April–May primary productivity bloom. In contrast, the Alkenone-SSTs signal likely integrates the averaged annual signal.

A combination of chronological tools allowed synchronizing the records in a common age model. Subsequently a single anomaly stack record was constructed for each  
15 proxy, thus easing to identify the most significant and robust patterns. The warmest SSTs occurred during the Roman Period (RP), which was followed by a general cooling trend interrupted by several centennial-scale oscillations. This general cooling trend could be controlled by changes in the annual mean insolation. Whereas some particularly warm SST intervals took place during the Medieval Climate Anomaly (MCA) the  
20 Little Ice Age (LIA) was markedly unstable with some very cold SST events mostly during its second half. The records of the last centuries suggest that relatively low E–P ratios and cold SSTs dominated during negative North Atlantic Oscillation (NAO) phases, although SST records seem to present a close positive connection with the Atlantic Multidecadal Oscillation index (AMO).

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# 1 Introduction

The Mediterranean is regarded as one of the world's highly vulnerable regions **in front of** the current global warming situation (Giorgi, 2006). This high sensitivity to climate variability has been evidenced in several studies focussed in past natural changes (Rohling et al., 1998; Cacho et al., 1999a; Moreno et al., 2002; Martrat et al., 2004; Reguera, 2004; Frigola et al., 2007; Combourieu Nebout et al., 2009). Paleo-studies focussed mostly in the rapid climate variability of the last glacial period have presented solid evidences of a tied connection with changes in North Atlantic oceanography and climate over the Western Mediterranean Region (Cacho et al., 1999b, 2000, 2001; Moreno et al., 2005; Sierro et al., 2005; Frigola et al., 2008; Fletcher and Sanchez-Goñi, 2008). Nevertheless, climate variability during the Holocene and, particularly during the last millennia, is not so well described in this region, although its understanding is crucial to place the nature of the 20th century trends in the recent climate history (Huang, 2004).

Some previous studies have already proposed that Holocene centennial climate variability in the western Mediterranean Sea could be linked to NAO variability (Jalut et al., 1997, 2000; Combourieu Nebout et al., 2002; Goy et al., 2003; Roberts et al., 2012; Fletcher et al., 2012). In particular, nine Holocene episodes of enhanced deep convection in the Gulf of Lion (GoL) and surface cooling conditions were described at the same location than this study (Frigola et al., 2007). These events have also been correlated to intensified upwelling conditions in the Alboran Sea and tentatively described as two-phase scenarios driven by distinctive NAO states (Ausín et al., 2015). A growing number of studies reveal considerable climate fluctuations during the last 2 kyr (Abrantes et al., 2005; Holzhauser et al., 2005; Kaufman et al., 2009; Lebreiro et al., 2006; Martín-Puertas et al., 2008; Kobashi et al., 2011; Nieto-Moreno et al., 2011, 2013; Moreno et al., 2012; PAGES 2K Consortium, 2013; Esper et al., 2014). However, there is not uniformity about the exact time-span of the different defined cli-

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matic periods such for example the Medieval Climatic Anomaly (MCA), term coined originally by Stine (1994).

The existing Mediterranean climatic records for the last 1 or 2 kyr are mostly based on terrestrial source archives such as tree rings (Touchan et al., 2005, 2007; Griggs et al., 2007; Esper et al., 2007; Büntgen et al., 2011; Morellón et al., 2012), speleothem records (Frisia et al., 2003; Mangini et al., 2005; Fleitmann et al., 2009; Martín-Chivelet et al., 2011; Wassenburg et al., 2013), or lake reconstructions (Pla and Catalan, 2005; Martín-Puertas et al., 2008; Corella et al., 2011; Morellón et al., 2012). All of these archives can be good sensors of temperature and humidity changes but often their proxy records mix these two climate variables. Recent efforts have focussed in integrating these 2 kyr records into a regional climatic signals and they reveal a complexity in the regional response but also evidence the scarcity of marine records to have a more complete picture (PAGES, 2009; Lionello, 2012).

In reference to marine records, they are often limited by the lack of adequate time resolution and accurate chronology to produce detailed comparison with terrestrial source records, although and they have the potential to provide a wider range of temperature sensitive proxies. Until the moment, not numerous paleoclimate records derived from marine data are available from the last 2 kyr in the Mediterranean Sea (Schilman et al., 2001; Versteegh et al., 2007; Piva et al., 2008; Taricco et al., 2009, 2015; Incarbona et al., 2010; Fanget et al., 2012; Grauel et al., 2013; Lirer et al., 2013, 2014; Di Bella et al., 2014; Goudeau et al., 2015) and they are even more scarce in the Western Basin. The current disperse data is not enough to admit a potential commune pattern of marine Mediterranean climate variability for these two millennia (Taricco et al., 2009; Nieto-Moreno et al., 2011; Moreno et al., 2012 and the references therein).

The aim of this study is to characterise changes in surface water properties from the Minorca margin in the Catalan-Balearic Sea (central-western Mediterranean), contributing to a better understanding of the climate variations in this region during the last 2.7 kyr. Sea Surface Temperature (SST) has been reconstructed **in base to** two independent proxies, Mg/Ca analyses on the planktonic foraminifera *Globigerina bulloides*

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and alkenone derived SST (Villanueva et al., 1997; Lea et al., 1999; Barker et al., 2005; Conte et al., 2006). The application of *G. bulloides*-Mg/Ca as a paleothermometer in the western Mediterranean Sea is tested through the analysis of a series of core top samples from different locations of the western Mediterranean Sea and the calibration reviewed consistently. Mg/Ca thermometry is applied with  $\delta^{18}\text{O}$  in order to evaluate changes in the Evaporation–Precipitation (E–P) balance of the basin ultimately linked to salinity (Lea et al., 1999; Pierre, 1999; Barker et al., 2005). One of the limitations for the study of climate evolution of the last 2 kyr is that often the intensity of the climate oscillations is at the limit of detection of the selected proxies. In order to identify significant climatic patterns within the proxy records, the analysis have been performed in a collection of multicores from the same region, and their proxy records have been stacked. The studied time periods have been defined as follows (years expressed as BCE = Before Common Era and CE = Common Era): Talaiotic Period (TP; ending at 123 BCE); Roman Period (RP; from 123 BCE to 470 CE); Dark Middle Ages (DMA; from 470 until 900 CE); Medieval Climate Anomaly (MCA; from 900 to 1275 CE); Little Ice Age (LIA; from 1275 to 1850 CE) and Industrial Era (IE) as the most recent period. The limits of these periods are not uniform across the Mediterranean (Lionello, 2012) and here, the selected ages have been chosen according to historical events in Minorca Island and also to the classic climatic ones defined in literature (i.e. Nieto-Moreno et al., 2011, 2013; Moreno et al., 2012; Lirer et al., 2013, 2014).

## 2 Climatic and oceanographic settings

The Mediterranean Sea is a semi-enclosed basin located in a transitional zone between different climate regimes, from the temperate zone at the north, to the subtropical zone at the south. Consequently, the Mediterranean climate is characterized by mild wet winters and warm to hot, dry summers (Lionello et al., 2006). Interannual climate variability is very much controlled by the dipole-like pressure gradient between the Azores (high) and Iceland (low) system known as the North Atlantic Oscillation (NAO) (Hurrell, 1995;

Lionello and Sanna, 2005; Mariotti, 2011; Ausín et al., 2015). But the northern part of the Mediterranean region is also linked to other midlatitude teleconnection patterns (Lionello, 2012).

The Mediterranean Sea is a concentration basin (Béthoux, 1980; Lacombe et al., 1981) and the excess of evaporation **respect** freshwater input is balanced by water exchange at the Strait of Gibraltar (i.e. Pinardi and Masetti, 2000; Malanotte-Rizzoli et al., 2014). The basinwide circulation pattern is prevalently cyclonic (Millot, 1999). Three thermohaline cells promote the Mediterranean deep and intermediate circulation: a basinwide open cell and two separated closed cells, one for the Western Basin and one for the Eastern part. The first one connects the two basins of the Mediterranean Sea through the Sicilia Strait, where water masses interchange occurs at intermediate depths. This cell is associated with the inflow of Atlantic Water (AW) at the Strait of Gibraltar and the outflow of the Levantine Intermediate Water (LIW) that flows below the first (Lionello et al., 2006).

In the north-western Mediterranean Sea, the Northern Current (NC) represents the main feature of the surface circulation transporting waters alongshore from the Ligurian Sea to the Alboran Sea. North-east of the Balearic Promontory is located a surface oceanographic front that separates properly Mediterranean waters transported by the NC from the Atlantic waters that have recently entered the Mediterranean (Millot, 1999; Pinot et al., 2002; André et al., 2005).

Deep convection occurs offshore the GoL due to the action of very intense cold and dry winter winds such as the Tramontana and the Mistral. These winds cause strong evaporation and cooling of surface water thus increasing their density until sinking to greater depths leading to Western Mediterranean Deep Water (WMDW) (MEDOC, 1970; Lacombe et al., 1985; Millot, 1999). Dense shelf water cascading (DSWC) in the GoL also contributes to the sink of large volumes of water and sediments into the deep basin (Canals et al., 2006).

The north-western Mediterranean is subject to an intense bloom in spring when the surface layer stabilizes, and sometimes to a less intense bloom in autumn, when the

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thermocline is progressively eroded (Bosc et al., 2004; D'Ortenzio and Ribera, 2009; Siokou-Frangou et al., 2010). The most important annual bloom in the Mediterranean Sea usually takes place as early in February (Estrada et al., 1985), concomitantly with minima SST, which subsequently increases until maxima summer values during August. Afterwards, a SST drop can be observed on October although with some inter-annual variability (Pastor, 2012).

### 3 Material and methods

#### 3.1 Sediment cores description

The studied sediment cores were recovered from a sediment drift built by the action of the southward branch of the WMDW north of Minorca (Fig. 1). Previous studies carried out at this site already described high sedimentation rates ( $> 20 \text{ cm kyr}^{-1}$ ) (Frigola et al., 2007, 2008; Moreno et al., 2012), which initially suggested a suitable location to carry on a detailed study of the last millennia. The cores were recovered from two different stations at about 50 km north of Minorca Island with a multicore system. Cores MINMC06-1 and MINMC06-2 (henceforth MIN1 and MIN2) ( $40^{\circ}29' \text{ N}$ ,  $04^{\circ}01' \text{ E}$ ; 2391 m water depth; 31 and 32.5 cm core length, respectively) were retrieved in 2006 during HERMES 3 cruise onboard the R/V *Thethys II*. In reference to the recovery of cores HER-MC-MR3.1, HER-MC-MR3.2 and HER-MC-MR3.3 (henceforth MR3.1, MR3.2 and MR3.3) ( $40^{\circ}29' \text{ N}$ ,  $3^{\circ}37' \text{ E}$ ; 2117 m water depth; 27, 18 and 27 cm core length, respectively) took place in 2009 during HERMESIONE expedition onboard the R/V *Hespérides*. The distance between the MIN and the MR3 cores is  $\sim 30 \text{ km}$  and both stations are located in an intermediate position within the sediment drift, which extends in a water depth range between  $\sim 2000\text{--}2700 \text{ m}$  (Frigola, 2012; Velasco et al., 1996; Mauffret et al., 1979), being MIN cores deeper than the MR3 ones by about  $\sim 300 \text{ m}$ .

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MIN cores were homogeneously sampled at 0.5 cm resolution in the laboratory while for MR3 cores a different strategy was followed. MR3.1 and MR3.2 were initially sub-sampled with a PVC tube and splitted in two halves for XRF analyses in the laboratory. Both halves of core MR3.1, MR3.1A and MR3.1B, were used for the present work as replicates of the same core. All MR3 cores were sampled at 0.5 cm resolution for the upper 15 cm and at 1 cm for the rest of the core, with the exception of half MR3.1B that was sampled at 0.25 cm resolution. MR3 cores were formed by brown-orange nanofossil and foraminifera silty clay, lightly bioturbated, with the presence of enriched layers in pteropods and gastropods fragments and some dark layers.

Additionally, core top samples from seven multicores collected at different locations in the western Mediterranean have also been used for the correction of the Mg/Ca-SST calibration from *G. bulloides* (Table 1; Fig. 1).

### 3.2 Radiocarbon analyses

Twelve  $^{14}\text{C}$  AMS dates were performed on cores MIN1, MIN2 and MR3.3 (Table 2) over 4–22 mg samples of planktonic foraminifer *Globigerina inflata* handpicked from the > 355  $\mu\text{m}$  fraction. Ages were calibrated with the standard marine correction of 408 years and the regional average marine reservoir correction ( $\Delta R$ ) for the central-western Mediterranean Sea using Calib 7.0 software (Stuiver and Reimer, 1993) and the MARINE13 calibration curve (Reimer et al., 2013).

### 3.3 Radionuclides $^{210}\text{Pb}$ and $^{137}\text{Cs}$

The concentrations of the naturally occurring radionuclide  $^{210}\text{Pb}$  were determined in cores MIN1, MIN2, MR3.1A and MR3.2 by alpha-spectroscopy following Sanchez-Cabeza et al. (1998). Concentrations of the anthropogenic radionuclide  $^{137}\text{Cs}$  in core MIN1 were measured by gamma spectrometry using a high purity intrinsic germanium detector. Gamma measurements were also used to determine the  $^{226}\text{Ra}$  concentrations via the gamma emissions of  $^{214}\text{Pb}$ , used to calculate the excess  $^{210}\text{Pb}$  concen-

trations. Sediment accumulation rates for the last century were calculated using the CIC (constant initial concentration) and the CF:CS (constant flux: constant sedimentation) models (Appleby and Oldfield, 1992; Krishnaswami et al., 1971), constrained by the  $^{137}\text{Cs}$  concentration profile for core MIN1 (Masqué et al., 2003).

### 3.4 Bulk geochemical analyses

The elemental composition of cores MR3.1B and MR3.2 was obtained with a XRF Core-Scanner Avaatech System (CORELAB, University of Barcelona), which is equipped with an optical variable system that allows determining in an independent way the length (10–0.1 mm) and the extent (15–2 mm) of the bundle of beams-X. This allows obtaining qualitative information of the elementary composition of the materials. The core surfaces were scraped cleaned and covered with a 4 mm thin SPEXCertiPrep Ultralene foil to prevent contamination and minimize desiccation (Richter and van der Gaast, 2006). Sampling was performed every 1 cm and scanning took place directly at the split core surface. Among the several measured elements this study has mainly use the Mn profile in the construction of the age models.

### 3.5 Planktonic foraminifera analyses

Samples for stable isotopes and trace elements analyses were washed over a 63  $\mu\text{m}$  sieve to remove the clay and silt fractions and then, dry-sieved between 355–250  $\mu\text{m}$  for picking the planktonic foraminifera *Globigerina bulloides*. In core MR3.1B, picking was often performed in the < 355  $\mu\text{m}$  fraction due to the small amount of material (sampling every 0.25 cm). Additionally, quantitative analysis of planktonic foraminifera assemblages was carried out in core MR3.3 and on the top of core MR3.1A by using the fraction size above 125  $\mu\text{m}$ . The 42 studied samples presented abundant and well-preserved planktonic foraminifera.

Among all trace elements analysed we have mainly used Mg/Ca to obtain SST and Mn, which has been utilized in age models. Trace elements analyses were performed

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on ~ 45 selected individuals of *G. bulloides*, after crushed them under glass slides to open the chambers and carefully cleaned applying a sequence of clay removal, oxidative and weak acid cleaning steps (Pena et al., 2005). Only samples from core MR3.1A were cleaned also including also the “reductive step”, resulting in Mg/Ca ratios around 23% lower than those measured without the reductive step, as it was expected from the preferential dissolution of the Mg-enriched calcite during the reductive step (Barker et al., 2003; Pena et al., 2005; Yu et al., 2007). Instrumental analyses were performed in an inductively coupled plasma mass spectrometer (ICP-MS) Perkin Elmer in the Scientific and Technological Centers of the University of Barcelona (CCiT-UB). A standard solution with a ratio close to the foraminifera values ( $3.2 \text{ mmol mol}^{-1}$ ) was run every four samples in order to correct any drift over the measurement runs for MR3.1 halves. Standard solution used on the rest of analyses was low ( $1.6 \text{ mmol mol}^{-1}$ ). Procedure blanks were also routinely measured in order to detect any potential contamination problem during the cleaning and dissolution procedure. The average reproducibility of Mg/Ca ratios was 97 and 89% for MIN1 and MIN2 cores, and 99 and 97% for MR3.1A, MR3.1B and MR3.3 cores, respectively. Mn/Ca and Al/Ca ratios were always measured in order to detect any potential contamination problem associated with the presence of Mn oxides and aluminosilicates (Barker et al., 2003; Lea et al., 2005; Pena et al., 2005). Correlation coefficients between Mg/Ca and these two ratios were very low, ( $< 0.4$  in both cases), supporting the no significant influence of any mineralogical phase other than the foraminiferal shells in the measured Mg/Ca ratios.

Stable isotopes measurements were performed on 10 specimens of *G. bulloides* after sonically cleaned in methanol to remove fine-grained particles. Analyses were performed in a Finnigan-MAT 252 mass spectrometer fitted with a carbonate micro-sampler Kiel-I in the CCiT-UB. Analytical precision of laboratory standards for  $\delta^{18}\text{O}$  is better than 0.08‰. Calibration to Vienna Pee Dee Belemnite or V-PDB was carried out by means NBS-19 standards (Coplen, 1996). Furthermore, seawater  $\delta^{18}\text{O}$  ( $\delta^{18}\text{O}_{\text{SW}}$ ) was obtained after removing the temperature effect on the *G. bulloides*  $\delta^{18}\text{O}$  record by applying the Mg/Ca-SST records in the Shackleton Paleotemperature Equation (Shack-

leton, 1974). The results are expressed in the water standard SMOW ( $\delta^{18}\text{O}_{\text{SW}}$ ) after the correction of Craig (1965).

### 3.6 Alkenones

Measurements of the relative proportion of unsaturated C<sub>37</sub> alkenones, namely  $U_{37}^{k'}$  index, were carried out in order to obtain SST records on the studied cores. Detailed information about the methodology and equipment used in C<sub>37</sub> alkenone determination can be found in Villanueva et al. (1997). The precision of this paleothermometry tool has been determined as close as  $\pm 0.5^\circ\text{C}$  (Eglinton et al., 2001). Furthermore, taking into account duplicate alkenone analysis carried out in core MR3.3, the precision achieved results better than  $\pm 0.8^\circ\text{C}$ . Reconstruction of SST records was based on the global calibration of Conte et al. (2006).

## 4 Age model development

Obtaining accurate chronologies for each of the studied sediment cores is particularly critical to allow their direct comparison and for achieving the final objective of producing a stack record that represents the regional climatic signal. With this objective, a wide set of parameters have been combined in order to obtain chronological markers in all the studied sedimentary records, including absolute dates and stratigraphical markers based on both geochemical and micro-paleontological data (Table 3).

### 4.1 $^{14}\text{C}$ , $^{210}\text{Pb}$ and $^{137}\text{Cs}$ dates

Absolute dating with radiocarbon dates was focused on cores MIN1, MIN2 and MR3.3 (Table 2). According to those dates and assuming the sampling year as the core top age (2006 and 2009, respectively), the sedimentation rates of these three cores result in  $13 \pm 1$ ,  $20 \pm 3$  and  $13 \pm 5 \text{ cm ky}^{-1}$ , respectively.

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In order to evaluate the preservation of the core tops,  $^{210}\text{Pb}$  activity profiles were obtained from cores MIN1, MIN2, MR3.1A and MR3.2 (Fig. 2).  $^{210}\text{Pb}$  concentrations generally decrease with depth in all four cores, down to 3.5 cm in core MIN2 and 3 cm for cores MIN1, MR3.1A and MR3.2. Excess  $^{210}\text{Pb}$  concentrations at the surface and inventories in the MIN cores are in agreement with those published for the Algero-Balear Basin (Garcia-Orellana et al., 2009). However, they were lower in MR3 cores, particularly for core MR3.1A, which we attribute to the loss of the most surficial part of these cores during recovery, corresponding to about 50 yr by comparison to the other cores. The variability in the  $^{210}\text{Pb}$  data denotes the high heterogeneity of this sedimentary system in reference to deep-sea hemipelagic sediments, highlighting the relevance of its study on the basis of a multicore approach (e.g. Maldonado et al., 1985; Martin et al., 1989; Calafat et al., 1996; Velasco et al., 1996; Canals et al., 2006; Frigola et al., 2007).

The concentration profile and inventory of  $^{137}\text{Cs}$  in core MIN1 is also in good agreement with the results reported for the Western Mediterranean Basin (Garcia-Orellana et al., 2009). Its detection down to 3 cm combined with the excess  $^{210}\text{Pb}$  concentration profile suggests the presence of sediment mixing to be accounted for in the calculation of the sediment accumulation rates, which are to be taken as maxima estimates. In doing so, the maxima sedimentation rates for the last 100–150 years are:  $27 \pm 2 \text{ cm ky}^{-1}$  (core MIN1),  $28 \pm 2 \text{ cm ky}^{-1}$  (MIN2),  $28 \pm 4 \text{ cm ky}^{-1}$  (MR3.1A), and  $35 \pm 3 \text{ cm ky}^{-1}$  (MR3.2). These sedimentation rates are in agreement with those previously described in a long sediment record recovered within the contouritic system (Frigola et al., 2007 and 2008), but much higher than those found in the literature from deeper sites of the Balearic Sea, with predominant hemipelagic sedimentation (e.g. Weldeab et al., 2003; Zúñiga et al., 2007; Garcia-Orellana et al., 2009).

## 4.2 Biostratigraphical data based on planktonic foraminifera

Core MR3.3, the best  $^{14}\text{C}$ -dates time-constrained, was chosen in order to perform a taxonomic analysis of planktonic foraminifera. The identified species were: (1) *Globigerina bulloides* including *G. falconensis*, (2) *Globigerinoides ruber* pink and white variety, (3) *Orbulina* spp. including both *O. universa* and *O. suturalis*, (4) *Globigerinoides quadrilobatus* and *G. sacculifer*, (5) *Globigerinatella siphoniphera* including *G. calida*, (6) *Globorotalia inflata*, (7) *Turborotalita quinqueloba*, (8) *Globigerinita glutinata*, (9) *Neogloboquadrina pachyderma* right coiled, (10) *Neogloboquadrina dutertrei*, (11) *Globorotalia truncatulinoides* left coiled and (12) *Clavatorella* spp. The abundance of *G. truncatulinoides* left coiled was also analysed in the top of the core MR3.1A.

In order to improve the time constrain of our cores, percentages records of *G. quadrilobatus* and *G. truncatulinoides* left coiled from core MR3.3 have been correlated with those from a southern Tyrrhenian Sea composite core (Fig. 3), with a very robust age-model (Lirer et al., 2013) based on the combination of different dating methods (radionuclides- $^{14}\text{C}$  AMS dates and tephra-chronology). The pronounced decrease in *G. quadrilobatus* percentages at the base of core MR3.3 (Fig. 3a) can be correlated with the end of the *G. quadrilobatus* acme interval observed in the Tyrrhenian Sea record from 1750 to 750 yr BCE (Sprovieri et al., 2003; Lirer et al., 2013, 2014). This correlation provide to us a control age point in core MR3.3 of  $750 \pm 48$  yr BCE at about 27 cm, consistent with the  $^{14}\text{C}$  dating of  $301 \pm 87$  yr BCE at 24 cm. In the upper part of the MR3.3 record, another control age point can be obtained from the correlation of the pronounced peak of *G. truncatulinoides* left coiled ( $\sim 20\%$  in abundance, Fig. 3b) with a similar peak previously reported in the Tyrrhenian Sea record during the LIA at  $1718 \pm 10$  yr CE (Lirer et al., 2013), and coincident with the Maunder event (Vallefuoco et al., 2012; Lirer et al., 2014). This age point is also consistent with the obtained  $^{14}\text{C}$  date of core MR3.3 at 3.5 cm of  $1434 \pm 51$  yr CE, further supporting the absence of the last two centuries in the core MR3.3. The absence of these centuries is also suggested by the *G. truncatulinoides* left coiled abundance patterns data from the top (1.5–3.5 cm)

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of the core MR3.1A (Fig. 3b). MR3.1A data is in agreement with the drop of the peak in core MR3.3 and  $^{210}\text{Pb}$  measurements (Fig. 2) have corroborated the presence of the most recent sediment in core MR3.1A.

### 4.3 Bayesian accumulation models

A preliminary age model for cores MIN1, MIN2 and MR3.3 was initially generated by means of available  $^{14}\text{C}$  ages, the two biostratigraphical dates from core MR3.3 and maximum sedimentation rates derived from  $^{210}\text{Pb}$  concentration profiles from cores MIN1 and MIN2. This preliminary age model was built using the Bayesian statistics software Bacon with the statistical package R (Blaauw and Christen, 2011).

Considering that the two independent sedimentation rates estimations based on  $^{14}\text{C}$  and  $^{210}\text{Pb}$  have significant uncertainties inherent to the methods and considering the different sampling resolution, averaged sedimentation rates obtained from the two methods have been taken into account in the Bayesian accumulation models. Regarding the core top ages, it was considered to be the recovering year ( $2006 \pm 10$  yr CE) in MIN cores and  $1718 \pm 10$  yr CE for core MR3.3, coinciding with the peak in the *G. truncatulinoides* record. The program settings for thickness of the sections and memory were chosen to fulfil the criterions of the best mean 95% confidence range and to maintain good correlation between prior and posterior accumulation rates. In addition, it was decided to keep the *memory strength* values rather high since the sedimentary context, a contouritic drift, is expected to record highly variable accumulation rates, and due to the smother changes induced by lowering the *memory strength* would no reflect realistic changes in this context.

The best Bayesian models achieved with a confidence mean of 95% provide accumulation rates for cores MIN1, MIN2 and MR3.3 of  $14 \pm 2$ ,  $22 \pm 1$  and  $12 \pm 1$   $\text{cm ky}^{-1}$ , respectively, which correspond to mean time resolutions of 292, 161, and 200 yr, respectively. It should be noted that the largest errors are obtained for core MIN1 because of the only two  $^{14}\text{C}$  dates. These age models reconstruct a rather smooth accumulation history, although significant fluctuations in accumulation rate at centennial or even

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decadal scale can be expected in this sedimentological context. The posterior outputs for accumulation rate (see Fig. 4) and its variability are quite comparable to their prior ones, but in the case of core MR3.3 the posterior output indicates larger memory (more variability) than that assumed a priori. This is due to the strong change in sedimentation rates at about 12 cm (998 yr CE) that the prior output tends to attenuate, and which could be associated with abrupt changes in sedimentation rates at that time (Fig. 4c).

These age models have been then further re-evaluated using other geochemical proxies as stratigraphical tools in order to ensure a common chronological framework for the obtained climate records (Sect. 4.4). Nevertheless, any readjustment has always been kept within the confident rage of the Bayesian models.

### 4.4 Multi-proxy chronostratigraphy

The chronologies of cores MIN1, MIN2 and MR3.3 were finally evaluated and re-adjusted in base to their Mg/Ca records and taking into account the 95 % probability intervals obtained in the Bayesian models.

Mg/Ca measured in *G. bulloides* is a well-established proxy of Sea Surface Temperatures (Barker et al., 2005). The two sampling stations are only separated by 30 km and thus it is a reasonable assumption to expect comparable and synchronous SST changes in all the studied cores. Visual comparison of the MIN1, MIN2 and MR3.3 records of Mg/Ca show several resemblances in some of the main patterns and structures, which are precisely synchronous with the Bayesian age models (Fig. 5). Consequently, the three records have been tuned in base to the main structures and taking into account the 95 % confidence of the statistical produced models (Fig. 5). The final age-models of cores MIN1, MIN2 and MR3.3 have an average age difference that is bellow 24 years in reference to the Bayesian models and the 75–63 % of the records are into the confidence intervals obtained in the Bayesian models.

The chronology from core MR3.3 has been the base to construct the age model for the other MR3 cores (MR3.1A, MR3.1B and MR3.2) for which no  $^{14}\text{C}$  dates were available (Table 3). The chronostratigraphical tools for core MR3.1 have been again



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the Mg/Ca records (Fig. 5). Additionally, manganese records in all MR3 cores have also been used as an additional chronostratigraphical tool. Mn presence in deep-sea sediments is related to redox processes (Calvert and Pedersen, 1996). Considering that all MR3 cores correspond to the same multicore these Mn rich layers have been used as isochrones. The available Mn records have been measured by two different methods: Mn measured in the bulk sediment by means of XRF Core-Scanner (MR3.1B and MR3.2) and Mn present in the foraminifera samples and measured by ICP-MS (MR3.3, MR3.1A and MR3.1B). Absolute values were very different between those samples measured with ICP-MS after cleaning the foraminifera with the reductive step (MR3.1A) and those without this cleaning step (MR3.3 and MR3.1B) but the same main features can be correlated between the three cores (Fig. 6). In the case of core MR3.1B (Fig. 6b), analysed at ultra-high resolution (0.25 cm slides), the Mn record shows the highest values with peaks over 80 ppb whose Mg/Ca values have been excluded of derived SST records since Mn enrichments can bias Mg/Ca ratios toward higher values and lead to significant overestimation of past seawater temperatures (Boyle, 1983; Pena et al., 2005, 2008). The top 5 cm of cores MR3.1A and MR3.2 have been dated according to the maxima sedimentation rates using the  $^{210}\text{Pb}$  flux.

### 4.5 Final age models and associated sedimentation rates

According to the obtained chronologies, the period covered by the studied sedimentary sequences is from 759 yr BCE to 1988 yr CE, being core MR3.1B the one spanning a longer period (Table 4). Total average of mean accumulation rates is  $17\text{ cm ky}^{-1}$ , with a total mean resolution of 84 years.

The final mean sedimentation rates obtained in MIN cores, 14 and  $25\text{ cm ky}^{-1}$ , are very similar with those derivated from Bayesian model simulations,  $14 \pm 2$  and  $22 \pm 1\text{ cm ky}^{-1}$ , and those previously published by Moreno et al. (2012), 19 and  $23\text{ cm ky}^{-1}$ . The differences in sedimentation rates between all cores except MIN2 are lower than  $3\text{ cm ky}^{-1}$ , variability that is reasonable due the diverse sediment processes that affect the contouritic system.

## 5 Sea surface temperatures and $\delta^{18}\text{O}$ data

### 5.1 Mg/Ca-SST calibration

The Mg/Ca ratio measured in *G. bulloides* is a widely used proxy to reconstruct SST (Barker et al., 2005) although available calibrations can provide very different results (Lea et al., 1999; Mashiotta et al., 1999; Elderfield and Ganssen, 2000; Anand et al., 2003; McConnell and Thunell, 2005; Cléroux et al., 2008; Thornalley et al., 2009; Patton et al., 2011). Apparently, the regional Mg/Ca-temperature response varies due to parameters that have not yet been identified (Patton et al., 2011). A further difficulty arises from the questioned Mg/Ca-thermal signal in high salinity regions such as the Mediterranean Sea where anomalous high Mg/Ca values have been observed (Ferguson et al., 2008). This apparent high salinity sensitivity in foraminifera-Mg/Ca ratios is under discussion and it has not been supported by recent culture experiments (Hönisch et al., 2013), which in addition, could be attributed to diagenetic overprints (Hoogakker et al., 2009; van Raden et al., 2011). In order to test the value of the Mg/Ca ratios in *G. bulloides* from the western Mediterranean Sea and also review its significance in terms of seasonality and depth habitat, a set of core top samples from different locations of the western Mediterranean Sea have been analysed. The studied cores are included in the 35–45° N latitude range (Table 1 and Fig. 1) and mostly represent two different trophic regimes, defined by the classical spring bloom (the most north-western basin) and an intermittently bloom (D’Ortenzio and Ribera, 2009).

The obtained Mg/Ca ratios have been compared with the isotopically derived calcification temperatures based on the  $\delta^{18}\text{O}$  measurements performed also in *G. bulloides* from the same samples. This estimation was performed after applying the **Shackleton (1974) paleotemperature equation** and using the  $\delta^{18}\text{O}_{\text{water}}$  data published by Pierre (1999), taking always into consideration the values of the closer stations and from the top 100 m. The resulting Mg/Ca-SST data have been plotted together with those *G. bulloides* data points from North Atlantic core tops previously published by Elderfield

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and Ganssen (2000). The resulting high correlation ( $r^2 = 0.9$ ; Fig. 7a) strongly supports the dominant thermal signal in the Mg/Ca ratios of the central-western Mediterranean Sea. Thus, the new data set from the Mediterranean core tops improves the sample coverage over the warm end of the calibration and the resulting exponential function indicates 9.7 % sensitivity in the Mg uptake respect to temperature, which is in agreement with the described range in the literature (i.e., Elderfield and Ganssen, 2000; Barker et al., 2005; Patton et al., 2011). The new calibration obtained from the combination of Mg/Ca-SST data from the western Mediterranean Sea and Atlantic Ocean is:

$$\text{Mg/Ca} = 0.6788e^{0.0973T} \quad (1)$$

The Mg/Ca-SST signal of *G. bulloides* has been compared with a compilation of water temperature profiles of the first 100 m measured between 1945–2000 yr in stations close to the studied core tops (MEDAR GROUP, 2002). Although significant regional and interannual variations have been observed, the obtained calcification temperatures of our core top samples present the best agreement with temperature values of the upper 40 m during the spring months (April–May) (Fig. 7b). This water depth is consistent with that found by plankton tows in the Mediterranean (Pujol and Vergnaud-Grazzini, 1995) and with results from multiannual sediment traps monitoring in the Alboran Sea and the GoL where maximum percentages were observed just before the beginning of thermal stratifications (see Bárcena et al., 2004; Bosc et al., 2004; Rigual-Hernández et al., 2012). Although the available information about depth and seasonality distribution of *G. bulloides* is relatively fragmented, this **specie** is generally situated in intermediate or even shallow waters (i.e. Bé, 1977; Ganssen and Kroon, 2000; Schiebel et al., 2002; Rogerson et al., 2004; Thornalley et al., 2009). However, *G. bulloides* has been also observed at deeper depths in some western Mediterranean Sea sub-basins (Pujol and Vergnaud-Grazzini, 1995). Extended data with enhanced spatial and seasonal coverage are required in order to better characterise production, seasonality, geographic and distribution patterns of live foraminifers as *G. bulloides*. Nevertheless,

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the obtained core top data set offers a solid evidence about the seasonal character of the recorded temperature signal in the Mg/Ca ratio.

## 5.2 A regional stack for SST-Mg/Ca records

The obtained Mg/Ca-SST profiles obtained from our sediment records are plotted with the resulting common age model in Fig. 8. The average SST values for the last 2700 years are  $18 \pm 0.8^\circ\text{C}$ . All the temperature reconstructions show the warmest sustained period during the RP, approximately between 170 yr BCE to 300 yr CE, except core MIN2, since this record ends at the RP-DA transition. In addition, all the records show a general consistent cooling trend after the RP with several centennial scale oscillations. Maximum Mg/Ca-SST value is observed in core MR3.3 ( $21 \pm 0.7^\circ\text{C}$ ) during the RP (Fig. 8c) and the minimum is recorded in core MIN1 ( $15.3 \pm 0.9^\circ\text{C}$ ) during the LIA (Fig. 8e). The records present high centennial-scale variability. Particularly, during MCA some warm events reached SST comparable to those of the RP and lightly higher than the average of maxima SST ( $20.3 \pm 0.6^\circ\text{C}$ ), but they were far shorter in duration (Fig. 8). The highest frequency of intense cold events occurred during the LIA and, especially, the last millennia recorded the minima average Mg/Ca-SST ( $16.1 \pm 0.8^\circ\text{C}$ ). Four of the five records show a pronounced minima SST after year 1275 CE when occurred the onset of LIA. In base to the differentiated patterns in Mg/Ca-SST the LIA period has been divided into two subperiods, an early warmer interval (LIAa) and a later colder interval (LIAb) with the boundary located at 1540 yr CE.

One of the main difficulties of working with SST reconstructions for the last millennia is that the targeted climatic signal has often comparable amplitude to the internal noise of the records due to sampling and proxy limitations. In order to minimize this inherent random noise, all the studied records have been combined in a regional Mg/Ca-SST anomaly stack with the aim to detect the most robust climatic structures along the different records and reduce the individual noise. Firstly, each SST record was converted into a SST anomaly record in relation to its average temperature (Fig. 8f). Secondly, in order to obtain a common sampling interval all records were interpolated. Although

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interpolation was performed at 3 different resolutions, results did not differ substantially (Fig. 8g). Subsequently, we selected the stack that provided the best resolution offered by our age models ( $20 \text{ yr cm}^{-1}$ ) since it preserves very well the high frequency variability of the individual records (Fig. 8g).

5 The obtained stack represents in a clearer way the main SST features described earlier and allows to better identify the most significant features at centennial-time scale. The most abrupt cooling events are recorded during the LIAa ( $-1^\circ\text{C}$  in 100 yr) and the LIAb ( $-0.5^\circ\text{C}$  in 20 yr) while the most abrupt warming ( $1^\circ\text{C}$  in 20 yr) is detected during the beginning of MCA. Another rapid warming ( $0.9^\circ\text{C}$  in 100 yr) is also detected during the TP, around 6th century BCE. When the whole studied period is considered a a long term cooling trend of about  $-0.5^\circ\text{C}$  is observed, however if we focus on the last 1800 yr, since the RP maxima, the observed cooling trend was far more intense, of about  $-3.2^\circ\text{C}$ .

### 5.3 Oxygen isotope records

15 Oxygen isotopes measured on carbonates shells of *G. bulloides* ( $\delta^{18}\text{O}_c$ ) and their derived  $\delta^{18}\text{O}_{\text{SW}}$  after removing the temperature effect with Mg/Ca-SST records (see Sect. 3.5) are shown in Fig. 9.  $\delta^{18}\text{O}_c$  and their derived  $\delta^{18}\text{O}_{\text{SW}}$  profiles have been respectively stacked following the same procedure for the SST-Mg/Ca stack (see Sect. 5.2). In general terms, all the records present a high stable pattern during the whole period with a weak depleting trend, which is almost undetectable in some cases (i.e. core MIN1).

25 Average  $\delta^{18}\text{O}_c$  values are  $1.3 \pm 0.1\text{‰}$  (uncertainty is expressed with a  $1 \sigma$ ) and, in general, MR3 cores show lightly heavier values ( $1.4\text{‰}$ ) than MIN cores ( $1.2\text{‰}$ ). Lightest  $\delta^{18}\text{O}_c$  values ( $1.1\text{‰} \pm 0.1$ ) mostly occur during the RP, although some short light excursions can be also observed during the end of the MCA and/or the LIA. Heaviest values ( $1.6\text{‰} \pm 0.2$ ) are mainly associated with short events during the LIA, the MCA and over the TP/RP transition. A significant increase of  $\delta^{18}\text{O}_c$  values is observed at

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the LIA/IE transition, although a sudden drop is recorded at the end of the stack record (after 1867 yr CE), which could result from a differential influence of the records (i.e. MIN1) and/or extreme artefact (Fig. 9g).

After removing the temperature effect on the  $\delta^{18}\text{O}_c$  record, the remaining  $\delta^{18}\text{O}_{\text{SW}}$  record mainly reflects changes in E–P balance, thus resulting as an indirect proxy for sea surface salinity. The average  $\delta^{18}\text{O}_{\text{SW}}$  values obtained for the studied period are  $1.8\text{‰} \pm 0.2$  (uncertainty is expressed with a  $1\sigma$ ). Heaviest  $\delta^{18}\text{O}_{\text{SW}}$  values ( $2.2\text{‰} \pm 0.2$ ) are recorded during the RP when the longest warm period is also observed. Enhancements of the E–P balance ( $\delta^{18}\text{O}_{\text{SW}}$  heavier values) are coincident with higher SST (Fig. 11). Lightest values ( $1.3 \pm 0.3$ ) are **remarkable** during the onset and the end of the LIA and also during the MCA. A drop in the E–P balance has been obtained approximately from the end of LIA to the most recent years. The  $\delta^{18}\text{O}_{\text{SW}}$  stacked record show variations during the studied period ranged about  $\pm 0.2 \text{ 20 yr}^{-1}$  ( $\sim 0.8 \text{ PSU } 20 \text{ yr}^{-1}$ ; Fig. 9). The most significant changes in our  $\delta^{18}\text{O}_{\text{SW}}$  (salinity) stack record correspond to an increase around 1000 yr CE and the decrease observed at the end of the LIA.

#### 5.4 Alkenone-SST records

The two alkenone ( $\text{U}_{37}^{K'}$ )-derived SSTs of MIN cores were already published in Moreno et al. (2012), while the records from MR3 cores are new (Fig. 10). The four Alkenone-SST records show a similar general cooling trend during the studied period and they have also been integrated in a SST anomaly stack (Fig. 10g). The whole cooling trend is of about  $-1.6^\circ\text{C}$  when the whole studied period is considered and about  $-2^\circ\text{C}$  since the maximum SST recorded in the RP. Previous studies have interpreted the Alkenone-SST signal in the western Mediterranean Sea as an annual average (Ternois et al., 1996; Cacho et al., 1999a, b; Martrat et al., 2004). The average Alkenone-SST for the studied period (last 2700 yr) is  $17.2 \pm 0.2^\circ\text{C}$  (uncertainty in average values is expressed with a  $1\sigma$ ), which is in substantial agreement with the annual mean corresponding to a Balearic site ( $18.7 \pm 1.1^\circ\text{C}$ ) according to the integrate values of the upper

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50 m (Ternois et al., 1996; Cacho et al., 1999a) of the GCC-IEO database that covers January 1994–July 2008.

The alkenone temperatures ranged between  $16.0 \pm 0.8^\circ\text{C}$ , core MIN2 during the LIAa, and  $18.4 \pm 0.8^\circ\text{C}$ , core MR3.3 during the MCA). Values near the average of maxima SST ( $18.1 \pm 0.2^\circ\text{C}$ ) are observed more frequently during TP, RP and MCA, while temperatures during the onset of MCA and LIA show many values closer to the average of minima SST ( $16.2 \pm 0.1^\circ\text{C}$ ). The most abrupt coolings ( $-0.3^\circ\text{C } 20 \text{ yr}^{-1}$ ) are observed at the end of the XX century ( $-0.8^\circ\text{C } 100 \text{ yr}^{-1}$ ) and the end of the MCA, while the highest warming rates ( $+0.3^\circ\text{C } 20 \text{ yr}^{-1}$ ;  $+0.5^\circ\text{C } 100 \text{ yr}^{-1}$ ) are recorded during the MCA.

## 5.5 Mg/Ca vs. Alkenone SST records

The mean Alkenone-SST values are about  $1^\circ\text{C}$  colder than those from the Mg/Ca-SST reconstruction ( $17.2 \pm 0.2$  vs.  $18.0 \pm 0.8^\circ\text{C}$ ). This difference cannot be attributed to the different habitat depth since alkenones should reflect the surface photic layer ( $< 50 \text{ m}$ ), while *G. bulloides* has the capability to develop in a wider and deeper environment (Bé, 1977; Pujol and Vergnaud-Grazzini, 1995; Ternois et al., 1996; Sicre et al., 1999; Ganssen and Kroon, 2000; Schiebel et al., 2002; Rogerson et al., 2004; Thornalley et al., 2009). Consequently this proxy difference should be associated with the growing season of the signal carriers.  $U_{37}^{k'}$ -SST records in the western Mediterranean Sea have been interpreted to represent mean annual SST (i.e. Cacho et al., 1999a; Martrat et al., 2004) but seasonal variations in alkenone production could play an important role in the  $U_{37}^{k'}$ -SST values (Rodrigo-Gámiz et al., 2014). Considering that during the summer months the Mediterranean Sea is a very stratified and oligotrophic sea, it should be expected reduced alkenone production during this season (Ternois et al., 1996; Sicre et al., 1999; Bárcena et al., 2004; Versteegh et al., 2007; Hernández-Almeida et al., 2011). This observation is further supported by the results from sediment traps located in the GoL showing very low coccolith fluxes during the

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summer months (Rigual-Hernández et al., 2013), while they show higher values during autumn, winter and spring, reaching maximum values at the end of the winter season, during SST minima. In contrast, high fluxes of *G. bulloides* are almost restricted to the upwelling spring signal, when coccolith fluxes have already started to decrease (Rigual-Hernández et al., 2012, 2013). This different growth season can explain the proxy bias in the SST reconstructions, with colder SST recorded by the alkenones.

Both Mg/Ca-SST and  $U_{37}^{K'}$ -SST records show a consistent cooling trend during the studied period, which since the RP maxima is of about  $2^{\circ}\text{C}$  in the alkenones record and  $3.2^{\circ}\text{C}$  in the Mg/Ca record. This enhanced Mg/Ca-SST variability is also reflected in the short term oscillations, at centennial time scale, which are better represented in the Mg/Ca record with oscillations over  $1^{\circ}\text{C}$ , while in the alkenone record are mostly shorter than  $0.5^{\circ}\text{C}$ . This enhanced Mg/Ca-SST variability could be also attributed to the highly restricted seasonal character of its signal, which purely reflects SST changes during the spring season. However, the coccolith signal integrates a wider time period from autumn to spring and, consequently, changes associated with specific seasons become more diluted in the resultant averaged signal.

The detailed comparison of the centennial SST variability recorded by both proxy stacks consistently indicates a puzzling antiphase (Fig. 11b and c). Although the main trends are consistently parallel in both alkenone and Mg/Ca proxies ( $r = 0.4$ ;  $p$  value = 0) as has been noted in other regions, short-term variability appears to have an apposite character. This unexpected outcome is a firm evidence of the relevance of the seasonal variability in the climate evolution and would indicate that extreme winter coolings were followed by a more rapid and intense spring warmings. Nevertheless, regarding the low amplitude of several of these oscillations, often close to the error of the proxies, this observation needs to be probed with further constraints as a solid regional feature.



## 6 Discussion

### 6.1 Climate patterns during the last 2.7 kyr

Changes in SST in the Minorca region have implications in the surface air mass temperature and moisture source regions that would determine air mass trajectories and ultimately precipitation regime in the Western Mediterranean Region (Millán et al., 2005; Labuhn et al., 2015). Observations of recent data have identified SST as a key factor in the development of torrential rain events in the Western Mediterranean Basin (Pastor et al., 2001), being able to act as a source of potential instability of air masses that transit over these waters (Pastor, 2012). In this line, the combination of SST reconstruction with  $\delta^{18}\text{O}_{\text{SW}}$  can provide a light to analyse the connection between thermal changes and moisture export from the central-western Mediterranean Sea during the last 2.7 kyr.

The older period recorded by our records is the so-call Talaiotic Period (TP), which corresponds to the Ancient Ages as the Greek Period in other geographic areas. Both studied SST proxies are consistent showing maximum SST at 600 yr BCE and a general cooling trend reaching minimum values by the end of the period ( $\sim 120$  yr BCE), synchronously with a reduction in the E–P rate occurred (Fig. 11a–c). Very few other records exist from this time period to compare these trends at regional scale.

One of the most outstanding features in the two SST-reconstructions, particularly in the Mg/Ca-SST stack is the warm SST that dominated especially during the second half of the RP (150–400 yr CE). The onset of the RP was relatively cold and a  $2.1\text{ }^{\circ}\text{C}$  warming occurred during the first part of this period ( $0.8\text{ }^{\circ}\text{C } 100\text{ yr}^{-1}$ ). This SST evolution from colder to warmer conditions during the RP is consistent with the isotopic record from the Gulf of Taranto (Taricco et al., 2009) and peat reconstructions from north-western Spain (Martínez-Cortizas et al., 1999), and to some extent to SST proxies in the SE Tyrrhenian Sea (Lirer et al., 2014). However none of these records indicate that the RP was the warmest period of the last 2 kyr. Other records from higher latitudes such as Greenland (Dahl-Jensen et al., 1998), North Europe (Esper et al., 2014), North

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Atlantic Ocean (Bond et al., 2001; Sicre et al., 2008), speleothem records from North Iberia (Martín-Chivelet et al., 2011) and even the multiproxy PAGES 2K reconstruction from Europe, suggest a rather warmer early RP than late RP and, again, none of these records highlights the roman times as the warmest climate period of the last 2 kyr. Consequently, these very warm RP conditions recorded in the Minorca Mg/Ca-SST stack appears to have a very regional character and suggest a rather heterogeneous thermal response along the European continent and surrounding marine regions.

According to the  $\delta^{18}\text{O}_{\text{SW}}$ -stack the RP seems to be accompanied by an increase in the E–P ratio (Fig. 11a) as also has been observed in some close regions as Alps (Holzhauser et al., 2005; Joerin et al., 2006). But a lake record from Southern Spain indicates relatively high levels when  $\delta^{18}\text{O}_{\text{SW}}$  stack indicates the maximum in E–P ratio (Martín-Puertas et al., 2008). This information is not necessarily contradictory, since enhanced E–P balance in the Mediterranean could induce enhanced precipitation in some of the regions, but more detailed geographical information should be required to really evaluate such situation.

After the RP, during the whole DMA and until the MCA, Mg/Ca-SST stack shows a  $2^\circ\text{C}$  cooling ( $-0.3^\circ\text{C } 100\text{yr}^{-1}$ ), which is of  $0.4^\circ\text{C}$  in the case of the Alkenone-SST stack; E–P rate is also decreasing. This trend is in contrast with the general warming trend interpreted in speleothem records from the North Iberia (Martín-Chivelet et al., 2011) or the transition towards drier conditions discussed from Alboran recods (Nieto-Moreno et al., 2011). SST proxies from the Tyrrhenian Sea show a coherent cooling tend with our SST records after the second half of the DMA and the Roman IV cold/dry phase described by Lirer et al. (2014) can be tentatively correlated (Fig. 11). This cooling phase is also documented in  $\delta^{18}\text{O}_{G. ruber}$  record of Gulf of Taranto by Grauel et al. (2013). The heterogeneity of the signal in the diferent proxies and regions reveals the dificulty to characterise the climate variability during these short periods and reinforce the need of better geographical coverage of individual proxies.

Frequently, the Medieval Period is described as a very warm period in numerous regions in the Northern Hemisphere (Hughes and Diaz, 1994; Mann et al., 2008; Martín-

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Chivelet et al., 2011), but an increasing number of studies are questioning the existence of such a “warm” period (i.e. Chen et al., 2013). Minorca SST-stacks also indicate variable temperatures and it does not stand as a particular warm period within the last 2 kyr (Fig. 11). A significant warming event is centred at 900 yr CE and a later cooling with minimum values at about 1200 yr CE (Fig. 11). Higher variability is found in Greenland record (Kobashi et al., 2011) while a early warm MCA and posterior cooling is also observed in temperature reconstructions from Central Europe (Büntgen et al., 2011) and also the European multi-proxy 2k stack for PAGES 2K Consortium (2013). But all these proxies agree in indicating overall warmer temperatures during the MCA than during the LIA. At the MCA/LIA transition a progressive cooling and a change in cyclic oscillation before and after the onset of LIA are visible. This transition is considered the last rapid climate change (RCC) of Mayewski et al. (2004).

In the context of the Mediterranean Sea, lake, marine and speleothem proxies suggest drier conditions during the MCA than during the LIA (Moreno et al., 2012; Chen et al., 2013; Nieto-Moreno et al., 2013; Wassenburg et al., 2013). Looking to the  $\delta^{18}\text{O}_{\text{SW}}$  stack, several oscillations are observed during the MCA and LIA but any clear differentiation between the MCA and LIA can be inferred from this proxy, indicating that these reduced precipitation also involved reduced evaporation in the basin without altering the E–P balance recorded by the  $\delta^{18}\text{O}_{\text{SW}}$  proxy. The centennial scale variability detected in both the Mg/Ca-SST stack and  $\delta^{18}\text{O}_{\text{SW}}$  stack reveal that higher E–P conditions existed during the warmer intervals (Fig. 11a and c).

The LIA stands as a period of high thermal variability according to the Mg/Ca-SST stack and, in base to these records, two substages can be differentiated, a first one when SST oscillations were larger and average temperatures warmer (LIAa) and a second one with shorter oscillations and colder average SST (LIAb). We suggest that LIAa interval could be linked to the Wolf and Spörer solar minima and LIAb corresponds to Maunder and Dalton cold events. In agreement with previous observations (i.e. Vallefuoco et al., 2012), this correlation suggests a global climatic.

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Furthermore, the two LIA substages are also present in the Greenland record (Kobashi et al., 2011). The intense cooling drop ( $-1.0^{\circ}\text{C } 100\text{yr}^{-1}$ ) at the onset of the LIAb is in agreement with the suggested coolings of  $0.5$  and  $1^{\circ}\text{C}$  in the Northern Hemisphere (i.e. Matthews and Briffa, 2005; Mann et al., 2009). The described two steps within the LIA are clearer in the Mg/Ca-SST stack than in the Alkenone-SST stack; this is also the case of the alkenone records in Alboran Sea (Nieto-Moreno et al., 2011) and may be consequence of the general reduced SST variability detected by these proxies (see Sect. 5.5).

In terms of humidity, the LIA is described as a period of increased runoff according to the Alboran record (Nieto-Moreno et al., 2011). The available lake level reconstruction from South Spain also reveals a progressive increase after the MCA, reaching a maximum during the LIAb (Martín-Puertas et al., 2008). Different records of flood events in the Iberia Peninsula also report a significant increase of extreme events during the LIA (Barriendos et al., 1998; Benito et al., 2003; Moreno et al., 2008). These conditions are consistent with the described enhanced storm activity over the GoL for the LIA (Sabatier et al., 2012). These conditions could account for the enhanced humidity transport towards the Mediterranean Sea that could produce the reduced E–P ratio detected in the  $\delta^{18}\text{O}_{\text{SW}}$  particularly for the LIAb (Fig. 11a and c).

The end of the LIA and onset of the IE is marked in the Mg/Ca-SST stack with a warming phase of about  $1^{\circ}\text{C}$  and less pronounced in the Alkenone-SST stack. This initial warm climatic event is also documented in other Mediterranean regions (Taricco et al., 2009; Marullo et al., 2011; Lirer et al., 2014) and Europe (PAGES 2K Consortium, 2013), which is coincident with a Total Solar Irradiance (TSI) enhancement after Dalton Minima. The two Minorca SST stacks show a cooling trend by the end of the record, which does not seem coherent with the instrumental atmospheric records. In Western Mediterranean, warming has been registered in two main phases: from the mid-1920s to 1950s and from the mid-1970s onwards (Lionello et al., 2006). The Minorca stacks do not show such a warming although they do not cover the second period of warming. Nevertheless, according to instrumental data from the upper layer on the Western

Mediterranean since the beginning of the XX century, no warming trends were detected before the 1980s (Vargas-Yáñez et al., 2010).

## 6.2 Climate forcing mechanisms

The general cooling trend observed in both Mg/Ca-SST and Alkenone-SST stacks presents a good correlation with the summer insolation evolution in the North Hemisphere, which actually dominates the annual insolation balance ( $r = 0.2$  and  $0.8$ ,  $p$  value  $\leq 0.007$ , respectively) (Fig. 12). This external forcing has already been proposed to control major SST trends for the whole Holocene period in numerous records from Northern Hemisphere (i.e. Wright, 1994; Marchal et al., 2002; Kaufman et al., 2009; Moreno et al., 2012). Also summer insolation seems to influence significantly the isotope variability ( $r = 0.4$ ,  $p$  value = 0) as has been suggested in the study of Ausín et al. (2015), among others. Nevertheless, another forcing needs to account for the centennial-scale variability of the records.

Solar variability has frequently suggested as a primary driver of the Holocene millennial-scale variability (i.e. Bond et al., 2001). Several oscillations can be observed in the TSI record (Fig. 12a) whose correlation with the Mg/Ca-SST and Alkenone-SST stacks are low, since most of the major drops in TSI does not correspond to SST cold events; although in the case of the Alkenone-SST stack some degree of correlation exists between the two records ( $r = 0.5$ ,  $p$  value = 0). Nevertheless, TSI does not seem to be the primer driver of the centennial scale SST variability in the studied records.

Furthermore, one of the major drivers of Mediterranean inter-annual variability in the Mediterranean region is the NAO (Hurrell, 1995; Lionello and Sanna, 2005; Mariotti, 2011). High state of the NAO produces high pressure over the Mediterranean Sea inducing an increment of the E–P balance and reduces sea level over several sectors of the Mediterranean Sea (Tsimplis and Josey, 2001). During these positive NAO periods, winds over the Mediterranean enhance their north direction, overall salinity increases and formation of dense deep water masses is reinforced as the water exchange through the Corsica channel while the arrival of north storm waves decreases

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(Wallace and Gutzler, 1981; Tsimplis and Baker, 2000; Lionello and Sanna, 2005). The effect of NAO on Mediterranean temperatures is more ambiguous. Changes during the last decades does not show significant variability with NAO (Luterbacher, 2004; Mariotti, 2011) although some studies suggest an opposite response between the two basins with cooling responses in some eastern basins and warming in the western during positive NAO conditions (Demirov and Pinardi, 2002; Tsimplis and Rixen, 2002). Although still controversial, some NAO reconstructions on proxy-records start to be available for the studied period (Lehner et al., 2012; Olsen et al., 2012; Trouet et al., 2012; Ortega et al., 2015). The last millennia are the best-resolved period and that allows a direct comparison with our data to evaluate the potential link to NAO.

The correlations between our Minorca temperatures stacks with NAO reconstructions (Fig. 12) are relatively low in the case of Mg/Ca-SST ( $r = 0.3$ ,  $p$  value  $\leq 0.002$ ) and not significant in the Alkenone stack, indicating that this forcing is probably not the driver of the main trends in the records, although several uncertainties still exist about the long NAO reconstructions (Lehner et al., 2012). Nevertheless, when a detailed analysis is done focussing on the more intense negative NAO phases, those bellow 0 (Fig. 12), they mostly appear to correlate with cooling phases in the Mg/Ca-stack. Although some inconsistencies exist as it is the case of the oldest negative NAO event, which in the TP period corresponds with a warm event in the Mg/Ca-SST stack (Fig. 12). The frequency of these negative events is particularly high during the LIA, and mostly during its second phase (LIA<sub>b</sub>) when the coldest intervals of our SST-stacks occurred. When the last centuries are compared in detail with the last NAO reconstruction based on several different proxy records of annual resolution and tested with some model assimilations (Ortega et al., 2015), it can be observed a consistent pattern of variability with our  $\delta^{18}\text{O}_{\text{SW}}$  reconstruction, with high (low) isotopic values dominating during positive (negative) NAO phases (Fig. 13). This picture is coherent with the described increase in the E–P balance during high NAO phases described for the last decades (Tsimplis and Josey, 2001), which would also contribute to the concentration of the  $^{18}\text{O}$  in the Mediterranean waters. The SST stacks also suggest some degree o

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correlation between warm SST and high NAO values (Fig. 12) but a more coherent picture is observed when the SST-records are compared to the AMO reconstruction: warm SST dominated during high AMO values (Fig. 14). This picture of salinity changes related to NAO and SST to AMO has actually been also described in base to the analysis of last decades data (Mariotti, 2011; Guemas et al., 2014) and confirms the complex but tied response of the Mediterranean to atmospheric and marine changes over the North Atlantic Ocean.

The pattern of high  $\delta^{18}\text{O}_{\text{SW}}$  when dominant positive NAO conditions occurred should indicate a reduction in the humidity transport over the Mediterranean region as a consequence of the high atmospheric pressure conditions (Tsimplis and Josey, 2001). To test this hypothesis, the  $\delta^{18}\text{O}_{\text{SW}}$  stack and the NAO reconstruction is compared to a proxy interpreted to reflect storm intensity over the GoL (Fig. 13), also linked to increased storm activity in the Eastern North Atlantic (Sabatier et al., 2012). Periods of increased/decreased storm activity in the GoL correlate indeed with low/high values in the  $\delta^{18}\text{O}_{\text{SW}}$  supporting that during negative NAO conditions North European storm waves can more frequently arrive into the Mediterranean Sea (Lionello and Sanna, 2005), contributing to the reduction of the E–P balance (Fig. 13). This data comparison would also support that during these enhanced storm periods, cold SST conditions would dominate in the region as has been previously suggested (Sabatier et al., 2012).

The fact, that previous intense negative NAO events such as that during the TP (550 yr BCE) correlates with a intense warm event in the Mg/Ca-SST stack and high  $\delta^{18}\text{O}_{\text{SW}}$  values. It may indicates that not all the negative NAO phases may have identical atmospheric configuration modes and thus impact over the western Mediterranean Sea as has already been described in base to in recent NAO observations (Sáez de Cámara et al., in proof, 2015). Nevertheless, NAO reconstructions during this early period of the record need still to be reproduced by different proxies. In this regard, the maximum SST temperatures and  $\delta^{18}\text{O}_{\text{SW}}$  recorded during the RP (100–300 yr CE) may suggest the occurrence of persistent positive NAO conditions, which would also be consistent with a high pressure driven drop in relatively sea level as has been re-

constructed in the north-western Mediterranean Sea (Southern France) ( $-40 \pm 10$  cm) (Morhange et al., 2013).

It is interesting to note that during the DMA a pronounced and intense cooling event is recorded in the Mg/Ca-SST stack at about 500 yr CE. Several references document in the scientific literature the occurrence of the so-called dimming of the sun at 536–537 yr CE (Stothers, 1984). This event, in base to ice core records, has been able to be linked a tropical volcanic eruption (Larsen et al., 2008). Tree-ring data reconstructions from Europe and also historical documents indicate the persistence during several years (536–550 yr CE) of what is described as the most severe cooling across the Northern Hemisphere during the last two millennia (Larsen et al., 2008). Despite the limitations derived from the resolution of our records, Mg/Ca-SST stack record may have caught this cooling and that would prove the robustness of our age models.

## 7 Summary and conclusions

The review of new core top data of *G. bulloides*-Mg/Ca ratios from the central-western Mediterranean Sea together with previous published data support a consistent temperature sensitivity for the Mediterranean samples and allows to refine the previously calibrations. The recorded Mg/Ca-SST signal from *G. bulloides* is interpreted to reflect April–May conditions from the upper 40 m layer. In contrast, the Alkenone-SST estimations are interpreted to integrate a more annual averaged signal, although biased toward the winter months since primary productivity during the summer months in the Mediterranean Sea is extremely low. This more averaged signal of the Alkenone-SST records may explain why they present more smoothed oscillations in comparison to the Mg/Ca-SST records.

After the careful construction of a common chronology for the studied multicores, in base to several chronological tools, the individual proxy records have been joined in an anomaly-stacked record to allow a better identification of the more solid patterns and structures. **Both Alkenone and Mg/Ca-SST stacks show a consistent cooling trend**

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over the studied period and since the Roman Period maxima this cooling is of about  $2^{\circ}\text{C}$  in the alkenones record and  $3.2^{\circ}\text{C}$  in the Mg/Ca record. This cooling trend seems to be consistent with the general lowering in summer insolation. This general cooling trend is punctuated by several SST oscillations at centennial time scale, which represent: rather warm times during Talaiotic Period (TP); maximum SST dominated during most of the Roman Period (RP); a progressive cooling during Dark Middle Ages (DMA); pronounced variability during Medieval Climate Anomaly (MCA) with two intense warming phases reaching warmer SST than during Little Ice Age (LIA); and very unstable and rather cold LIA, with two substages, a first one with SST oscillations larger and average temperatures warmer (LIAa) and a second one with shorter oscillations and colder average SST (LIAb). The described two stages within the LIA are clearer in the Mg/Ca-SST stack than in the Alkenone-SST record. Comparison of Mg/Ca-SST and  $\delta^{18}\text{O}_{\text{SW}}$  stacks indicates that warmer intervals have been accompanied by higher Evaporation–Precipitation (E–P) conditions. The E–P balance oscillations over each defined climatic period during the last 2.7 kyr suggest variations in the thermal change and moisture export patterns in the central-western Mediterranean.

The comparison of the Minorca SST-stacks with other paleoclimatic records from Europe suggests a rather heterogonous thermal response along the European continent and surrounding marine regions. Comparison of the new Mediterranean records with the reconstructed variations in Total Solar Irradiance (TSI) does not support a clear connection with this climate forcing. Nevertheless, changes in the North Atlantic Oscillation (NAO) and Atlantic Multidecadal Oscillation (AMO) seem to have exerted a more relevant role controlling climate changes in the region. The negative NAO phases appear to correlate mostly with cooling phases in the Mg/Ca-stack, although this connection is complex and apparently clearer during the most intense negative phases. Nevertheless, when the comparison is focussed in the last 1 kyr, when NAO reconstructions are better constrained, a more consistent pattern arises, with cold and particularly fresher  $\delta^{18}\text{O}_{\text{SW}}$  values (reduced E–P balance) during negative NAO phases. A picture of enhanced southward transport of European storm tracks during this period would be

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coherent with the new data and previous reconstructions of storm activity in the GoL. Nevertheless, the SST-stacks seem to present a more tied relation to AMO during the last four centuries (the available period of AMO reconstructions): warm SST dominated during high AMO values. These evidences would support a close connection between Mediterranean and North Atlantic oceanography for the last 2 kyr.

*Acknowledgements.* Cores MINMC06 were recovered by HERMES 3 cruise in 2006 on R/V *Thethys II* and HER-MC-MR3 cores were collected by HERMESIONE expedition on board of R/V *Hespérides* in 2009. This research has financially been supported by OPERA (CTM2013-48639-C2-1-R). We thank Generalitat de Catalunya Grups de Recerca Consolidats grant 2009 SGR 1305 to GRC Geociències Marines. We are grateful to M. Guart (Dept. d'Estratigrafia, Paleontologia i Geociències Marines, Universitat de Barcelona), M. Romero, T. Padró and J. Perona (Serveis Científico-Tècnics, Universitat de Barcelona) and J.M. Bruach (Departament de Física, Universitat Autònoma de Barcelona) for their help with the laboratory work, D. Amblàs for his collaboration with the artwork of maps and to Paleoteam for the unconditional support. S. Giralt and M. Blaauw are acknowledged for their help. M. Bermejo benefited from a fellowship of the University of Barcelona.

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**Table 1.** Core tops taken into account in the calibration's adjustment.

Core	Location	Latitude	Longitude	Mg/Ca (mmol mol <sup>-1</sup> )	$\delta^{18}\text{O}_c$
TR4-157	Balearic Abyssal Plain	40°30.00' N	4°55.76' E	3.36	0.53
KTB-34	Cat-Bal Sea (Balears)	40°27.17' N	3°43.38' E	4.44	1.05
ALB1	Alboran Sea (WMed)	36°14.31' N	4°15.52' W	3.20	0.80
ALBT1	Alboran Sea (WMed)	36°22.05' N	4°18.14' W	3.44	0.65
ALBT2	Alboran Sea (EMed)	36°06.09' N	3°02.41' W	3.63	0.57
ALBT4	Alboran Sea (EMed)	36°39.63' N	1°32.35' W	3.72	0.93
ALBT5	Alboran Sea (EMed)	36°13.60' N	1°35.97' W	3.38	0.64

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**Table 2.** Radiocarbon dates and calibrated ages, these last one are expressed in years Before Common Era (BCE) and Common Era (CE). MR3.3 dates are presented for the first time in this study. Cores were analysed at the NOSAMS/Woods Hole Oceanographic Institution, USA (OS) and at Direct AMS Radiocarbon Dating Service, USA (D-AMS).

Laboratory Code	Core	Comp. Depth (cm)	$^{14}\text{C}$ ages	Cal years BCE/CE ( $2\sigma$ )
OS-67294	MIN1	7–7.5	$895 \pm 35$	1411–1529 CE
OS-67296		19–19.5	$2010 \pm 35$	304–544 CE
OS-67291	MIN2	11–11.5	$845 \pm 35$	1440–1598 CE
OS-67297		18–18.5	$1190 \pm 35$	1170–1312 CE
OS-67324		25–25.5	$1540 \pm 25$	804–989 CE
OS-67323		28.5–29	$1840 \pm 30$	520–680 CE
D-AMS 004812	MR3.3	3.5–4	$938 \pm 25$	1383–1484 CE
OS-87613		6.5–7	$1270 \pm 35$	1063–1256 CE
OS-87614		12–12.5	$1420 \pm 30$	911–1085 CE
OS-87615		16–17	$1900 \pm 30$	438–621 CE
D-AMS 004811		20–21	$2350 \pm 29$	88 BC–107 CE
OS-87619		24–25	$2620 \pm 25$	388 BCE–214 BCE

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**Table 3.** Summary of records analysed and methods utilized in age models.

Core	Records analysed	Age model
MIN1	Mg/Ca-SST, $U_{37}^{k'}$ -SST, $\delta^{18}O$	$^{14}C$ , $^{210}Pb$ , $^{137}Cs$ , software-simulations, SST-tuning
MIN2	Mg/Ca-SST, $U_{37}^{k'}$ -SST, $\delta^{18}O$	$^{14}C$ , $^{210}Pb$ , software simulations, SST-tuning
MR3.1A	Mg/Ca-SST, $\delta^{18}O$	$^{210}Pb$ , SST-tuning, geochemical chronostratigraphy, foraminiferal assemblage
MR3.1B	Mg/Ca-SST, $\delta^{18}O$ , Geochemical composition	SST-tuning, geochemical chronostratigraphy
MR3.2	Geochemical composition	$^{210}Pb$ , geochemical chronostratigraphy
MR3.3	Mg/Ca-SST, $U_{37}^{k'}$ -SST, $\delta^{18}O$ ,	$^{14}C$ , software-simulations, SST-tuning, foraminiferal assemblage

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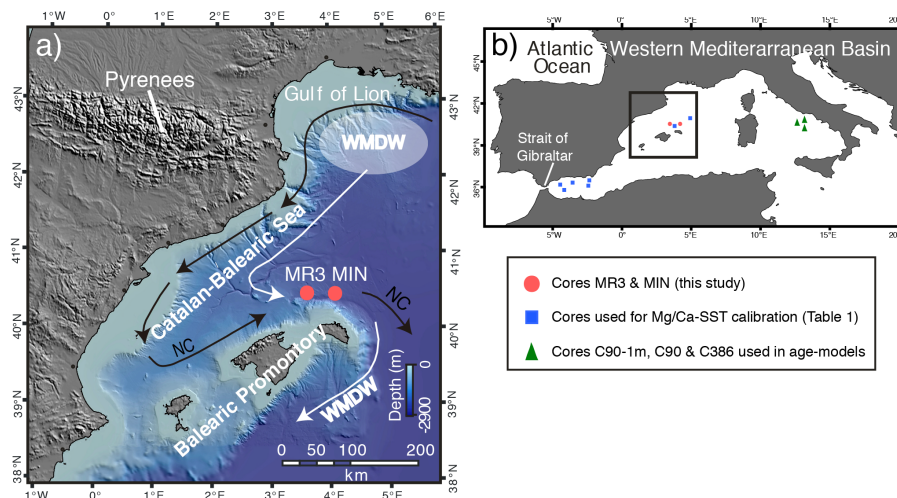
**Table 4.** Mean accumulation rates, years covered and mean time resolution of all cores according to final age-depth models.

Core	Mean acc. rate (cm ky <sup>-1</sup> )	Spanning time (yr)	Mean time resolution (yr cm <sup>-1</sup> )
MIN1	14	2528	83
MIN2	25	1538	48
MR3.3	17	2443	78
MR3.1A	15	2635	95
MR3.1B	16	2706	98
MR3.2	15	1797	102



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**Figure 1.** Location of the studied area. **(a)** Central-western Mediterranean Sea: cores MIN and MR3 effect of this study (red dots) with relevant features of surface (NC: Northern Current) and deep water circulation (WMDW: Western Mediterranean Deep Water). **(b)** Cores used in age-models development from the Tyrrhenian Sea (green triangles) (Lirer et al., 2013) and cores used in Mg/Ca-SST calibration from the Western Mediterranean Basin (blue squares).

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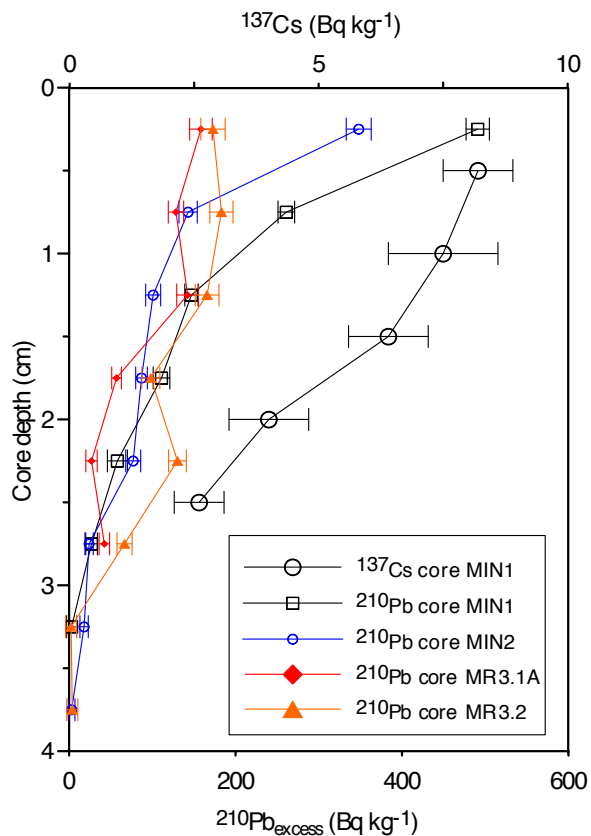
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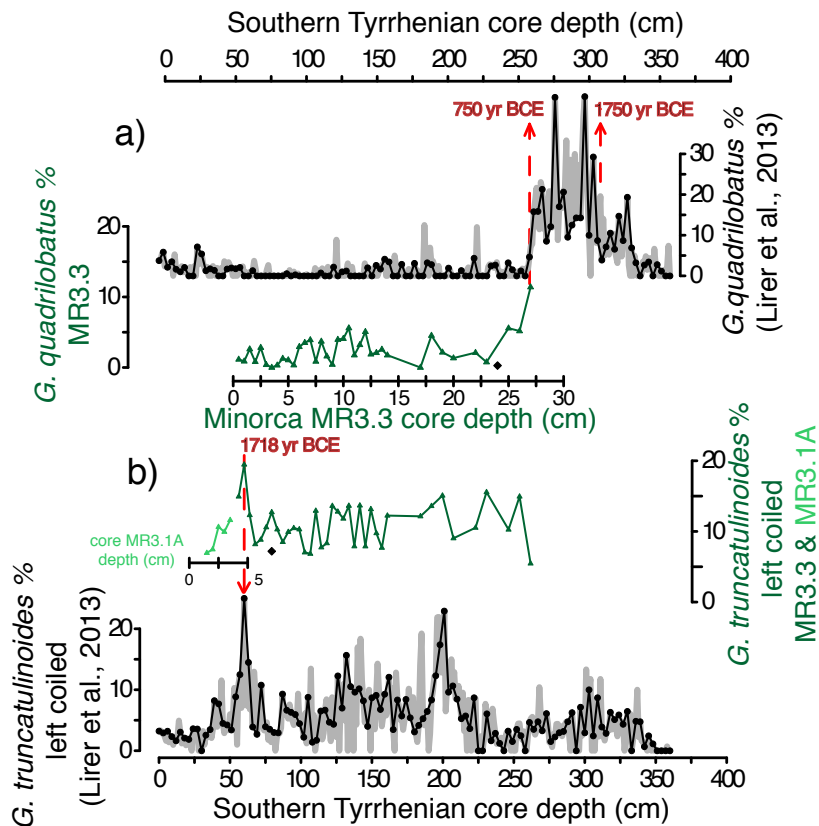
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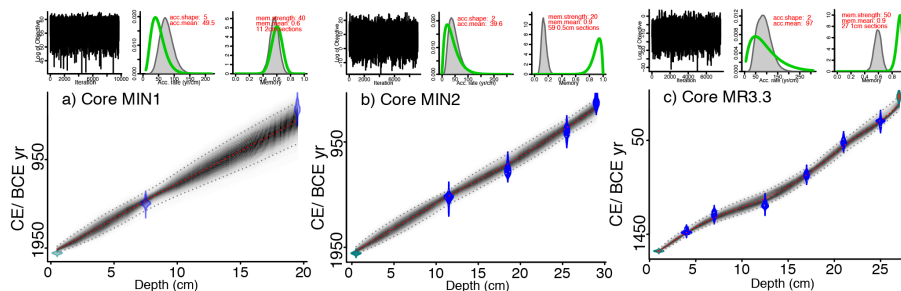
**Figure 2.** Excess  $^{210}\text{Pb}$  ( $\text{Bq kg}^{-1}$ ) profiles for cores MIN1, MIN2, MR3.1A and MR3.2 and also  $^{137}\text{Cs}$  concentration profile for core MIN1. Error bars represent  $1\sigma$  uncertainty.



**Figure 3.** Comparison among the quantitative distribution patterns of (a) *G. quadrilobatus* and (b) *G. truncatulinoides* left coiled with core MR3.3 (dark green plot) and data from a the composite core (C90-1 m, C90 and C836 cores) studied in the southern Tyrrhenian Sea (Lirer et al., 2013), expressed as 3 point average and with the grey area corresponding to the entire record. The two tie points used in age models (dashed red line) correspond to 1718 yr CE and 750 yr BCE. Black diamonds show <sup>14</sup>C dates from core MR3.3.

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**Figure 4.** Age–depth models based on Bayesian accumulation simulations (Blaauw and Christen, 2011): **(a)** core MIN, **(b)** MIN2 and **(c)** MR3.3. The three upper plots in each core show the stable MCMC run achieved (left), the prior (green line) and posterior (grey) distributions of the accumulation rates (middle), and the prior (green line) and posterior (grey) distributions of the memory (right). Each main graphic represents the age–depth model for each core (darker grey indicates more probable calendar ages) based on the prior information, the calibrated radiocarbon dates (purple symbols), sample year for cores MIN (blue symbols) and biostratigraphical dates from core MR3.3 (red symbols).

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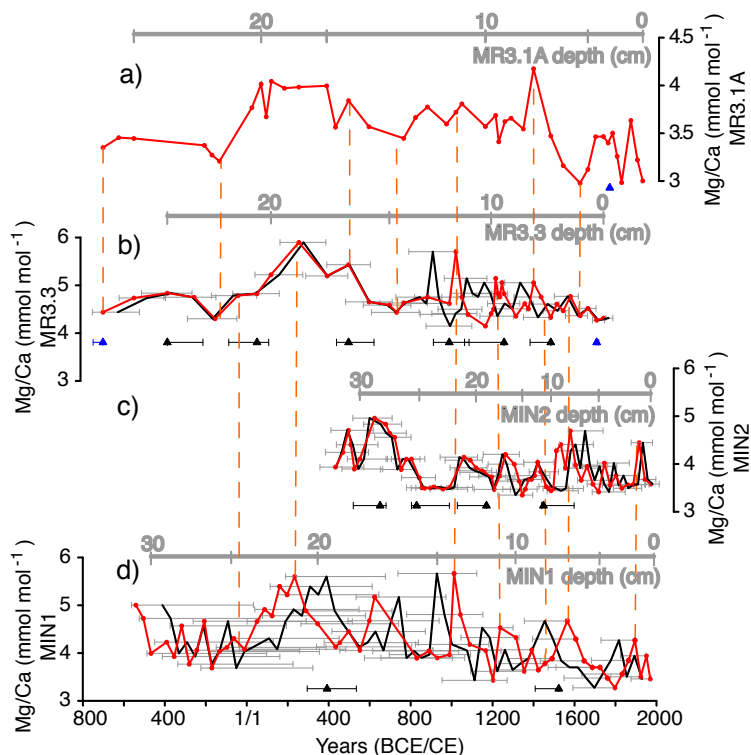
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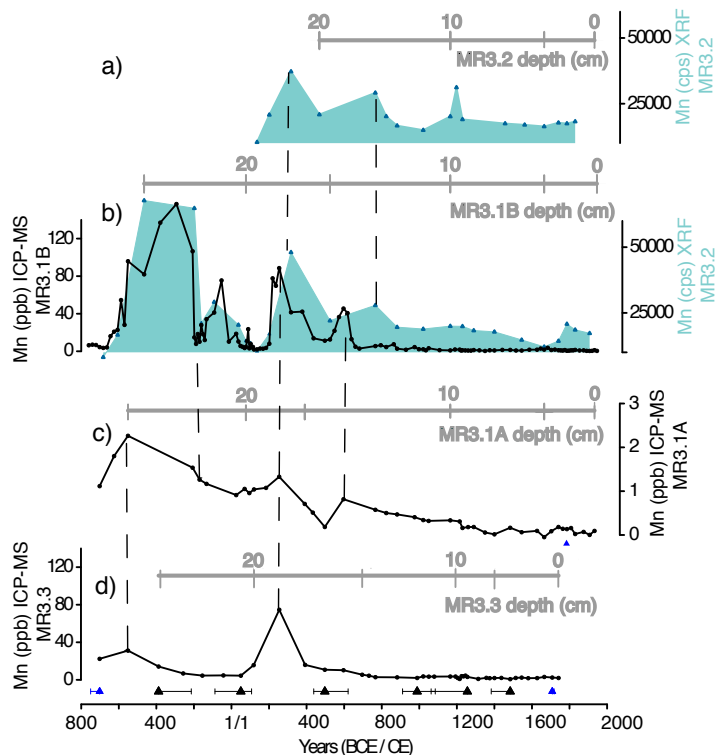
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**Figure 5.** Main procedures of multi-proxy chronostratigraphy performed with Mg/Ca records for cores: **(a)** MR3.1A, **(b)** MR3.3, **(c)** MIN2 and **(d)** MIN1. Final age-depth models are plotted in red. Black plots and grey error bars correspond to Bayesian accumulation age-depth models. Triangles represent to  $^{14}\text{C}$  dates (black) and biostratigraphical dates based on planktonic foraminifera (blue), and they are shown below the corresponding core and with their associated  $2\sigma$  errors. Depths in relation to the final age model can be observed above its corresponding core. Vertical dashed lines (orange) indicate tie points between the different Mg/Ca records.



**Figure 6.** Multi-proxy chronostratigraphy performed with Manganese profiles. Blue filled plots represents Mn profiles obtained by XRF Core-Scanner for cores (a) MR3.2 and (b) MR3.1B, respectively. Black plots shows Mn from trace elements analysed means of ICP-MS for cores (b) MR3.1B, (c) MR3.1A and (d) MR3.3. Vertical dashed lines indicate tie points of geochemical chronostratigraphy. Triangles represent to  $^{14}\text{C}$  dates (black) and biostratigraphical dates based on planktonic foraminifera (blue) and they are shown below the corresponding core and with their associated  $2\sigma$  errors.

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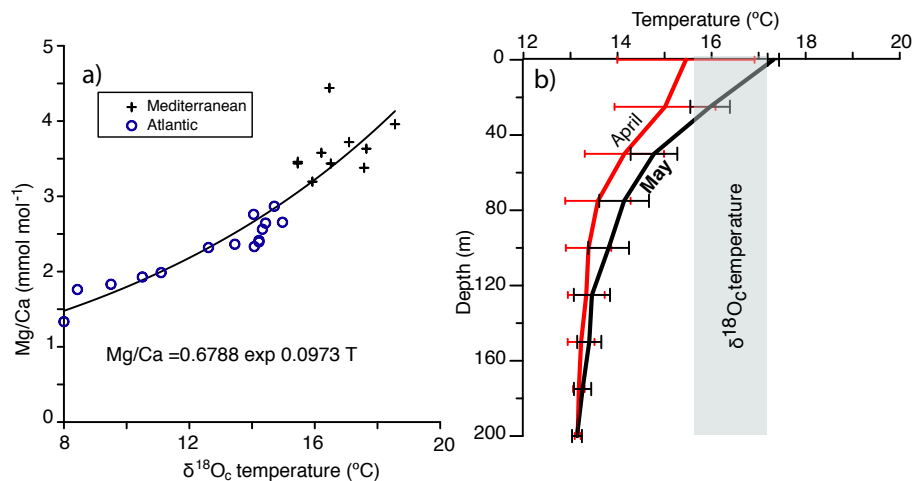
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**Figure 7.** (a) **Exponential function** and correlation obtained between  $\delta^{18}\text{O}_c$  temperatures and Mg/Ca for western Mediterranean Sea.  $\pm 0.7^\circ\text{C}$  is the standard error in calibrations on all the *G. bulloides* core tops utilized in this paper from the western Mediterranean Sea and it is consistent with  $\pm 0.6^\circ\text{C}$  obtained for the Atlantic Ocean in Elderfield and Ganssen (2000) and also  $\pm 1.1^\circ\text{C}$  in the same sp. culture data (Lea et al., 1999). (b) April (red) and May (black) temperature profiles of the first 100 m measured during years 1945–2000 in stations corresponding to the studied core tops (MEDAR GROUP, 2002). In grey is shown the  $\delta^{18}\text{O}_c$  average temperature of all cores.

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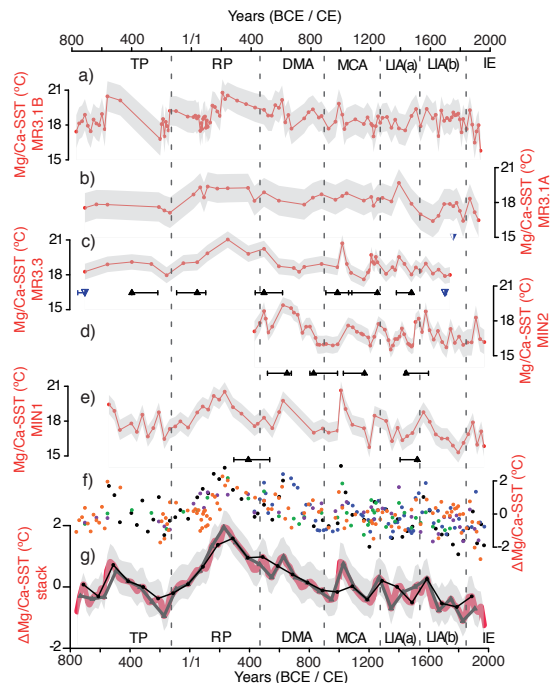
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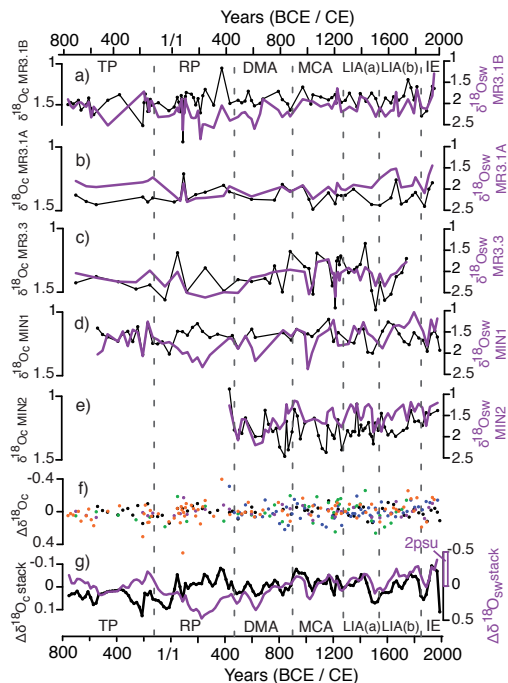


**Figure 8.** SST obtained means of analysis of Mg/Ca for cores: **(a)** MR3.1B, **(b)** MR3.1A, **(c)** MR3.3, **(d)** MIN2 and **(e)** MIN1. Grey-scalers integrate the reproducibility in Mg/Ca concentrations in each analysis and  $\pm 0.7^\circ\text{C}$ , which is the calculated standard error in *G. bulloides* core top calibrations for the central-western Mediterranean Sea developed in this paper. **(f)** All individual SST anomalies on their respective time step (MR3.1B: orange, MR3.1A: purple, MR3.3: green, MIN2: blue and MIN1: black dots). **(g)**  $20\text{ yr cm}^{-1}$  stacked temperature anomaly (red plot) with its  $2\sigma$  uncertainty (grey band). The  $80\text{ yr cm}^{-1}$  (black plot) and the  $100\text{ yr cm}^{-1}$  (grey plot) stacks are also shown. Triangles represent to  $^{14}\text{C}$  dates (black) and biostratigraphical dates based on planktonic foraminifera (blue) and they are shown below the corresponding core and with their associated  $2\sigma$  errors.



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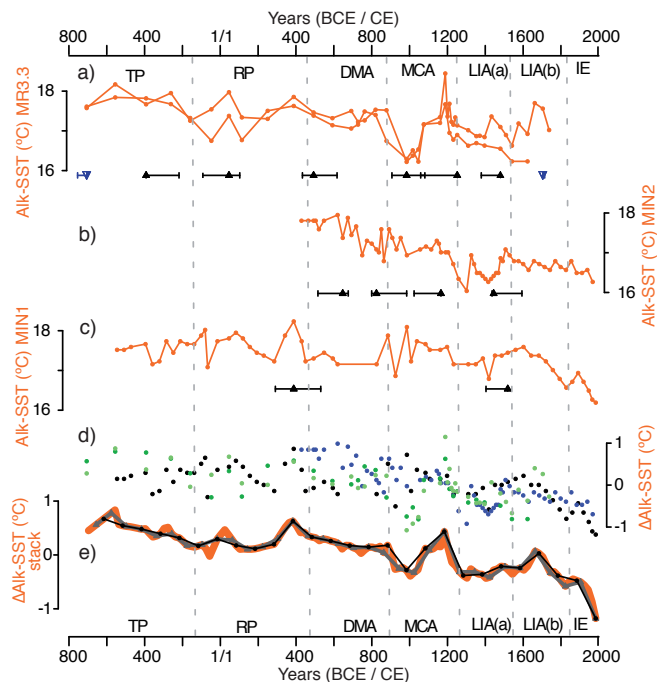
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**Figure 9.** Oxygen isotope measured on carbonates shells of *G. bulloides* ( $\delta^{18}\text{O}_C$ , in black) and their derived  $\delta^{18}\text{O}_{\text{SW}}$  (purple) for cores: **(a)** MR3.1B, **(b)** MR3.1A, **(c)** MR3.3 **(d)** MIN2 and **(e)** MIN1. **(f)** Individual  $\delta^{18}\text{O}_C$  anomalies on their respective time step. **(g)** Both respective anomaly stacked records and the equivalence between  $\delta^{18}\text{O}_{\text{SW}}$  and salinity, calculated according to Pierre (1999). It is estimated that the rise of one unit of  $\delta^{18}\text{O}_{\text{SW}}$  would amount to an enhancement of 4 practical salinity units.

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**Figure 10.** Alkenone temperature records from Minorca (this study) for cores: **(a)** MR3.3, **(b)** MIN2 and **(c)** MIN1. Triangles represent to  $^{14}\text{C}$  dates (black) and biostratigraphical dates based on planktonic foraminifera (blue) and they are shown below the corresponding core and with their associated  $2\sigma$  errors. **(d)** All individual alkenone derived SST anomalies on their respective time step (MR3.3: green, MIN2: blue and MIN1: black dots); **(e)**  $20\text{ yr cm}^{-1}$  stacked temperature anomaly (orange plot). The  $80\text{ yr cm}^{-1}$  (black plot) and the  $100\text{ yr cm}^{-1}$  (grey plot) stacks are also shown.

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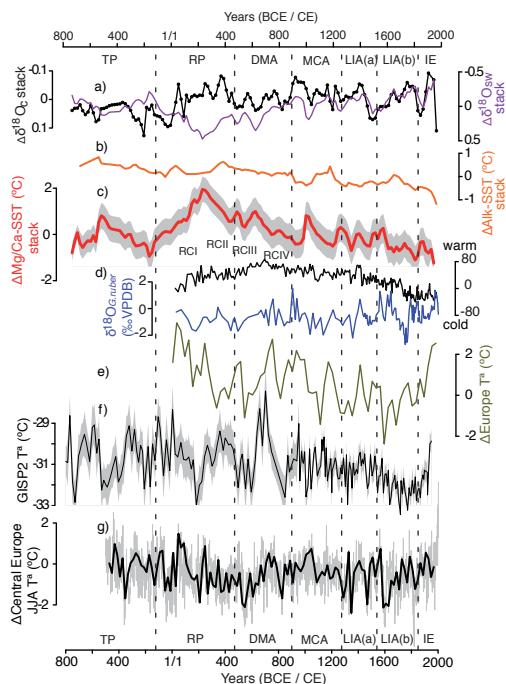
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**Figure 11.** Temperature and isotope anomaly records from Minorca (this study) and data from another regions. **(a)**  $\delta^{18}\text{O}_c$  and  $\delta^{18}\text{O}_{\text{SW}}$  Minorca stacks, **(b)** Alkenone-SST anomaly Minorca stack, **(c)** Mg/Ca-SST anomaly Minorca stack, **(d)** warm and cold phases and  $\delta^{18}\text{O}_{G. ruber}$  recorded by planktonic foraminifera from the southern Tyrrhenian composite core, respectively and RCI to RCIV showing roman cold periods (Lirer et al., 2014), **(e)** 30-year averages of the PAGES 2k Network (2013) Europe anomaly Temperature reconstruction, **(f)** Greenland snow surface temperature (Kobashi et al., 2011) and **(g)** Central Europe Summer anomaly temperature reconstruction in Central Europe (Büntgen et al., 2011).

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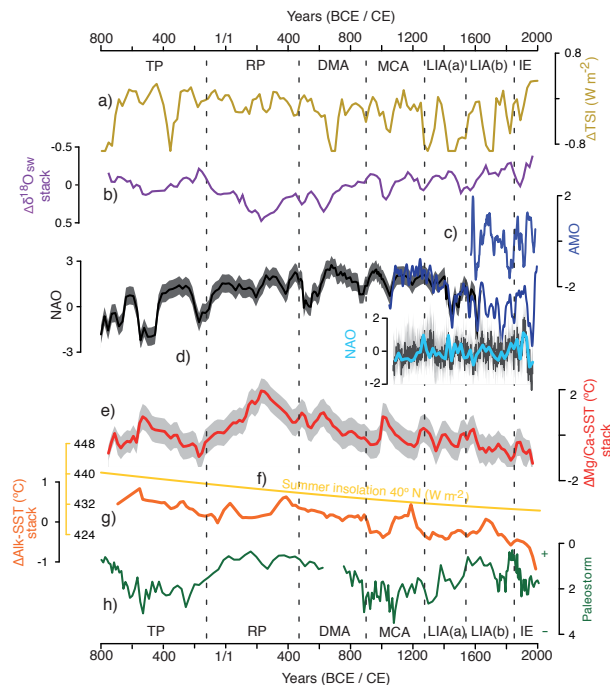
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**Figure 12.** Temperature and isotope anomaly records from Minorca (this study) and data from another regions and with external forcings: **(a)** Total Solar Irradiance (Steinhilber et al., 2009, 2012), **(b)**  $\delta^{18}\text{O}_{\text{SW}}$  Minorca stacks, **(c)** Atlantic Multidecadal Oscillation (AMO) (Gray et al., 2004), **(d)** North Atlantic Oscillation (NAO) reconstructions (Olsen et al., 2012; Trouet et al., 2009, and for the last millennium Ortega et al., 2015), **(e)** Mg/Ca-SST anomaly Minorca stack, **(f)** Summer Insolation at  $40^\circ\text{N}$  (Laskar et al., 2004), **(g)** Alkenone-SST anomaly Minorca stack and **(h)** Paleostorm activity in the Gulf of Lions (Sabatier et al., 2012).

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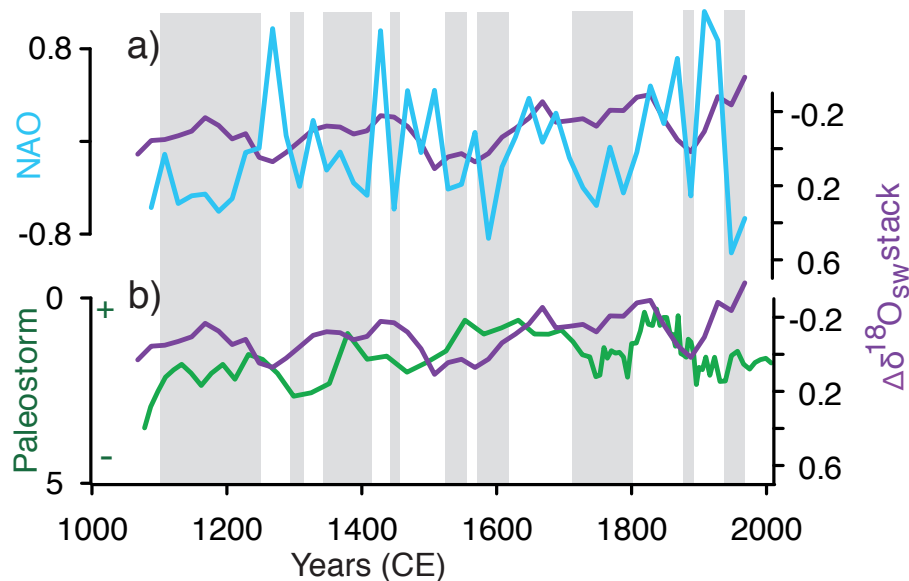
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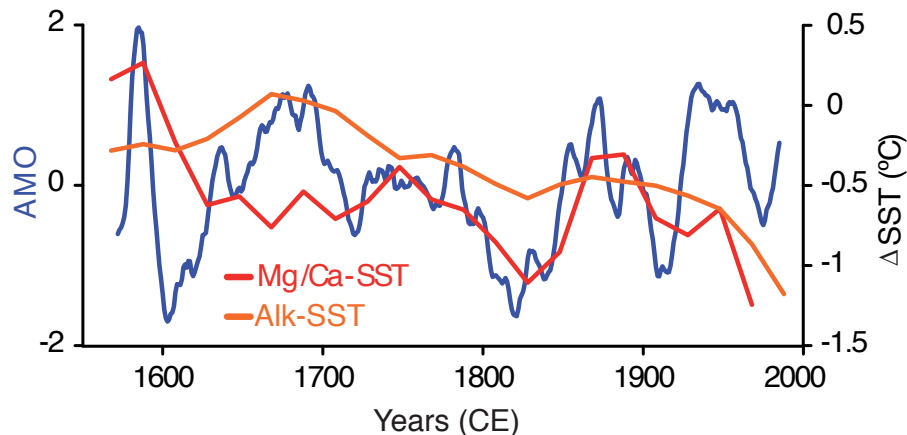
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**Figure 13.**  $\delta^{18}\text{O}_{\text{sw}}$  Minorca stack during the last millennium (age is expressed in years Common Era) plotted with (a) NAO reconstruction (Ortega et al., 2015) and (b) Paleostorm activity in the Gulf of Lion (Sabatier et al., 2012).

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**Figure 14.** Mg/Ca-SST and Alkenone-SST Minorca anomaly stacks during the last centuries plotted with AMO reconstruction (Gray et al., 2004).

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