Marine response to climate changes during the last five millennia in the central Mediterranean Sea

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Abstract

We present a high-resolution paleoclimatic and paleoenvironmental reconstruction of the last five millennia from a shallow water marine sedimentary record from the central Tyrrhenian Sea (Gulf of Gaeta) using planktonic foraminifera, pollen, oxygen stable isotope, tephrostratigrapy and magnetostratigrapy. This multiproxy approach allows to evidence and characterize nine time intervals

associated with archaeological/cultural periods: Eneolithic (base of the core- ca. 2410 BC), Early Bronze Age (ca. 2410 BC – ca. 1900 BC), Middle Bronze Age - Iron Age (ca. 1900 BC – ca. 500 BC), Roman Period (ca. 500 BC – ca. 550 AD), Dark Age (ca. 550 AD – ca. 860 AD), Medieval Climate Anomaly (ca. 860 AD – ca. 1250 AD), Little Ice Age (ca. 1250 AD – ca. 1850 AD), Industrial Period (ca. 1850 AD – ca. 1950 AD), Modern Warm Period (ca. 1950 AD – present day). The reconstructed climatic evolution in the investigated sedimentary succession is coherent with the short-term climate variability documented at the Mediterranean scale.
By integrating the planktonic foraminiferal turnover from carnivorous to herbivorous-opportunistic species, the oxygen isotope record and the pollen distribution, we document important modification from the onset of the Roman Period to the present-day. From ca. 500 AD upwards the documentation of the cooling trend punctuated by climate variability at secular scale evidenced by the short-term $\delta^{18}O$ is very detailed. We hypothesise that the present day warm conditions started from the end of cold Maunder event. Additionally, we provide that the North Atlantic Oscillation (NAO) directly affected the central Mediterranean region during the investigated time interval.

Key words: planktonic foraminifera; oxygen stable isotope; pollen; tephrostratigraphy; magnetostratigraphy; Tyrrhenian Sea; Mediterranean Sea.

1. Introduction

Over the last millennia, the Mediterranean Sea was affected by very significant shifts in climate (i.e., Luterbacher et al., 2012; Maselli and Trincardi 2013; Büntgen et al., 2016). The most common phases may be correlated with the major archaeological subdivisions found in the literature: Roman Period, Dark Age, Medieval Climate Anomaly and Little Ice Age (i.e., Luterbacher et al., 2012; Büntgen et al., 2016 and references therein).

However, it is worth noting the consensus, over the last two millennia, concerning the used and chronology of terms Little Ice Age (LIA), Medieval Climate Anomaly (MCA) and Dark Age (i.e., Luterbacher et al., 2012 and references therein). This consensus is basically related to Pages 2k Network research activities (mostly tree ring data, i.e., Büntgen et al., 2016 and references therein). Recently, marine data also contributed to recognition of these events (i.e., Lirer et al., 2014; Holmgren et al., 2015; Cisneros et al., 2015). No agreement exists about the climatic variability during the second half (first 400 years AD) of the Roman Period (Table 1). In fact, despite of the data available for this part of the Roman Period (i.e., Moreno et al., 2012; Grauel et al., 2013a; Lirer et al., 2014; Goudeau et al., 2015; Cisneros et al., 2015; Gogou et al., 2016), other factors as local overprint (i.e., Cisneros et al., 2015), a no uniform response of climate signals amongst the various basins (Gogou et al., 2016) and the sensitivity of the biotic and abiotic proxies, can suggest different paleoclimatic interpretation (i.e., Roman Warm Period / Roman Humid Period). The climatic subdivision results more complicated for the last 3000 years BC because this period has less chronological datasets.

The link between the most significant paleoenvironmental changes and the climate phases has been recently documented in different marine and continental archives (e.g., Piva et al., 2008; Jalut et al., 2009; Roberts et al., 2011; Lirer et al., 2014; Goudeau et al., 2015; Sadori et al., 2015; Gogou et al., 2016). These correlations are very useful to solve the land-sea interactions (synchronicity of proxy-events) and evaluate the impact of climate changes in societal organizations. In fact, recent studies (i.e. Magny et al., 2013; Holmgren et al., 2015; Sadori et al., 2015; Büntgen et al., 2016) suggested a continuous interaction between climate changes and the modifications of human societies and adaptive strategies.

The complicated land–sea interactions and the presence of steep mountain ridges close to the coast help to explain the spatial heterogeneity of climate in the Mediterranean region and represent a well-known problem for the correct simulation of its climate (Corte-Real et a., 1995; Lionello, 2012). However, it is largely accepted that the North Atlantic Oscillation (NAO), El Nino–Southern Oscillation (ENSO) and the Atlantic Multi-decadal Oscillation (AMO) represent the major factors in climate and oceanic variability in the Mediterranean region (Malanotte-Rizzoli et al., 2014 and reference therein). Previous research has also identified the NAO as one of the dominant atmospheric mode controlling the temporal evolution of precipitation and temperature in the Mediterranean area (Lòpez-Moreno et al., 2011) even if the interpretation of the NAO effect in the Mediterranean region is somewhat controversial, making crucial the need of fossil archive investigations. Planktonic foraminifera are commonly used as paleoenvironmental proxies in paleoceanographic investigations since they respond to changes of the environmental parameters of the water masses where they live (Bè and Tolderlund, 1971; Bè, 1977; Fairbanks et al., 1980; Hemleben et al., 1989; Ravelo et al., 1990; Le and Shackleton, 1994; Kucera et al., 2005). Furthermore, high-resolution studies performed in different basins of Mediterranean pointed out that the distributional pattern of some planktonic foraminiferal species are also useful for regional correlation (i.e., Sprovieri et al., 2003; Piva et al., 2008a; Budillon et al., 2009; Rouis-Zargouni et al., 2010; Lirer et al., 2013).

Within this framework, the high sedimentation rates characterizing the northern (Di Bella et al., 2014) and southern Tyrrhenian Sea (Budillon et al., 2005; Sacchi et al., 2009; Lirer et al., 2013, 2014), Gulf of Taranto (Grauel et al., 2013a; Taricco et al., 2015; Goudeau et al., 2015) and Adriatic Sea (Oldfield et al., 2003; Piva et al., 2008), make the Mediterranean area an ideal archive to investigate paleoclimate changes at decadal and secular scale over the last millennia.

To contribute to a better understanding of late Holocene paleoclimate changes in the Mediterranean area, this work presents data collected in the central Tyrrhenian Sea (Gulf of Gaeta). We used planktonic foraminifera and pollen data with oxygen stable isotope signal.

Based on radionuclides analyses, tephrostratigraphy and oxygen isotope stratigraphy we offer a record at decadal time resolution. In addition, we provide the comparison of our data with the NAO index curve in order to document the link of the Mediterranean region with the North Atlantic climate change conditions.

2. Study area

The Mediterranean is an elongated and semi-enclosed basin, with an anti-estuarine circulation pattern forced by the negative hydrological balance and the density gradient with the Atlantic Ocean (Robinson and Golnaraghi, 1994) where evaporation exceeds precipitation (Bergamasco et al., 2010). At Gibraltar, Atlantic water inflows in the surface layer with temperature T = 15°C and salinity S = 36.2 psu and becomes Modified Atlantic Water (MAW) along its path to the Eastern basin. In the bottom layer Mediterranean water, the Levantine Intermediate Water (LIW) with T = 13.5°C and S = 38.4 ‰ psu outflows. The transformation of MAW into LIW occurs through surface heat loss and evaporation specifically in the Levantine basin. Mediterranean has an overall resulting mean heat loss in the range of 3–7 W/m^2 (i.e., Bergamasco et al., 2010).

The Tyrrhenian Sea is the deepest major basin in the western Mediterranean (Astraldi and Gasparini, 1994) and is connected to the other Mediterranean sub-basins through the Corsica Channel in the north and the Sardinia Channel in the south. The circulation is overall cyclonic triggered by the MAW entering off the northern Sicilian coast and establishing a northward current along the western Italian coast (Krivosheya and Ovchinnikov, 1973; Millot, 1987; Artale et al., 1994; Pierini and Simioli, 1998). According to De Pippo et al., (2003–2004) the circulation pattern of the Tyrrhenian Sea, which influences the Gulf of Gaeta, has a cyclonic vortex that interacts with the superficial (down to 10 m depth) and the intermediate (from 10 to 100 m depth) water layers (Bonomo et al., 2014).

The Gulf of Gaeta is also strongly influenced by the presence of the Volturno River, the longest river in southern Italy (175 km) with an estimated mean discharge of 40 $m^3 s^{-1}$, and a 1550 km^2 catchment basin (Iermano et al., 2012).

The continental shelf of the Gulf of Gaeta is the seaward extension of the Garigliano and Volturno coastal alluvial plains filled by Plio-Quaternary clastic and volcaniclastic deposits (de Alteriis et al., 2006); it is bounded by Cape Circeo to the north, Ischia and the Gulf of Naples to the south, and the Pontine Islands to the west; it narrows from NW to SE (from some tens of kilometres to kilometres north of Ischia) (de Alteriis et al., 2006).

3. Material and Methods

3.1 Core Lithology

The present study focused on the composite marine sequence of the core SW104-C5 (40°58'24,993"N, 13°47'03,040"E), 108 centimetres below the sea floor (cmbsf) length, and core C5 (40°58’24.953’N, 13°47'02,514"E), 710 cmbsf length, recovered in the Gulf of Gaeta, at 93 m water depths during the oceanographic cruise AMICA2013 (Fig. 1). The correlation between cores SW104-C5 and C5 is based on the identification of a Vesuvius tephra layer in the magnetic susceptibility record (Fig. 2) (see chapter 4.2 for details).

The studied interval (the first 452 cm composite depth) is characterized by light grey hemipelagic sediments (Fig. 2).

3.2 Planktonic foraminifera

Planktonic foraminiferal analysis was based on 346 samples_collected at 1.3 cm spacing.

Samples were oven dried at 50°C and washed using a 63 µm mesh sieve. Quantitative planktonic foraminiferal analysis was carried out on the fraction >90 µm to avoid the juvenile specimens (Vallefuoco et al., 2012).

The main ecological interpretations used in this work follow, especially, Hemleben et al. (1989) and Pujol and Vergnaud-Grazzini (1995) and are summarised in Table 2. Some planktonic species have been grouped as follows: *Orbulina* spp. includes *Orbulina universa* and *O. suturalis*; *Globigerinoides quadrilobatus* includes *G. trilobus*; *G. ruber* includes *G. gomitulus*; *Globigerina bulloides* includes *G. falconensis*; *Globigerinella siphonifera* includes *G. calida*. Analysis discriminated also between left and right coiling of *Globorotalia truncatulinoides*, *Globorotalia inflata* and *Neogloboquadrina pachyderma*. Planktonic foraminiferal species are plotted in percentages of the total assemblage.

Globorotalia scitula and *Neogloboquadrina pachyderma* were lumped together as indicators of cool water conditions; in particular, *G. scitula* is generally associated with cool water (Bè and Hutson, 1977; Hemleben et al., 1989), *N. pachyderma* is a deep-dwelling species living close to or below the thermocline (Capotondi et al., 2006). Also *G. glutinata* and *T. quinqueloba* are summed together in interpreted as proxy of the productivity in the sub- surface waters (Cita et al., 1977; Corselli et al., 2002; Geraga et al., 2008; Jonkers et al., 2010) (Tab. 2).

In order to characterize environmental changes we also plotted the distribution pattern of the herbivorous-opportunistic planktonic foraminiferal species (*T. quinqueloba*, *G. glutinata*, *G. bulloides*) and carnivorous ones (*G. ruber*, *G. quadrilobatus*, *Orbulina* spp., *G. siphonifera*) based on their ecological requirements reported in Table 2.

3.3 Radionuclides ^{210}Pb and ^{137}Cs

The chronology for the uppermost 60 cmbsf is based on sedimentation rate estimated by ^{210}Pb and ^{137}Cs. The ^{210}Pb and ^{137}Cs analysis were carried out at ISMAR-CNR Bologna, following the procedures reported in Bellucci et al. (2007). Alpha spectrometry of ^{210}Po was used for ^{210}Pb determinations, assuming secular equilibrium between the two isotopes. Supported ^{210}Pb activities were obtained from the constant values at depth in the core, where ^{210}Pb and ^{226}Ra were considered to be in radioactive equilibrium. Excess ^{210}Pb was calculated by subtracting the supported ^{210}Pb activity from the total ^{210}Pb activity.

The sediment accumulation rate was calculated using the constant flux–constant sedimentation model (CF–CS) (Sanchez-Cabeza and Ruiz-Fernández, 2012). To validate the ^{210}Pb-derived accumulation rates, ^{137}Cs activities were measured via gamma spectrometry using coaxial intrinsic germanium detectors.

3.4 *Magnetostratigraphy*

As an additional check on the age model, a paleomagnetic inclination study was conducted using u-channel samples extracted from the centre of the C5 core and analysed at 1 cm intervals. Analysis was carried out at the INGV paleomagnetic laboratory in Rome. The u-channels were progressively demagnetized by alternating-field treatment at 5, 10, 15, 20, 25, 30, 40, 50, 60, 80, and 100 mT, and the remaining magnetization measured at each using a 2G Enterprises cryogenic magnetometer. The resulting demagnetization paths were analysed using the PuffinPlot software of Lurcock and Wilson (2012) to apply principal component analysis (Kirschvink, 1980), providing paleomagnetic inclination values for comparison with regional geomagnetic reference curves.

3.5 Oxygen Stable Isotope

Oxygen and Carbon isotope analyses were carried out on about ten specimens of the planktonic foraminiferal species *G. ruber* alba variety. Analyses were performed at the geochemistry laboratory of the IAMC-CNR (Naples, Italy) with an automated continuous flow carbonate preparation Gas BenchII device (Spötl and Vennemann, 2003) and a ThermoElectron Delta Plus XP mass spectrometer. Acidification of samples was performed at 50 °C. Every 6 samples, an internal standard (Carrara Marble with $\delta^{18}O$ = -2.43‰ versus VPDB and $\delta^{13}C$ = 2.43‰ vs. VPDB) was run and every 30 samples the NBS19 international standard was measured. Standard deviations of carbon and oxygen isotope measures were estimated at 0.1 and 0.08‰, respectively, on the basis of ~200 samples measured three times. All the isotope data are reported in δ‰ versus VPDB.

3.6 Pollen

Pollen analysis was carried out on 86 samples collected in the upper 480 cm of the SW104-C5-C5 composite core. They were chemically treated with HCl (37%), HF (40%) and NaOH (20%), following the standard procedure proposed by Fægri et al. (1989). Pollen concentration values were estimated by adding *Lycopodium* tablets to known weights of sediment (Stockmarr, 1971). Pollen grains were identified by means of a light microscope at 400 and 630 magnifications, with the help of both pollen morphology atlases (Reille, 1992, 1995, 1998; Beug, 2004) and the reference collection at the Laboratory of Palaeobotany and Palynology of Sapienza University of Rome. The main percentage sum is based on terrestrial pollen excluding pollen of aquatics and non-pollen palynomorphs (fungal and algal spores, as well as microscopic fragments of various organisms found in the pollen slides). Excluding aquatics, spores and other non-pollen palynomorphs (NPPs), an average number of ca. 200 pollen grains per sample were counted. These represent a statistically reliable number to undertake a reconstruction of the vegetation history, especially considering the low pollen concentrations (1300-7000 grains/gram of sediment) and the fraction of analysed sample (often >10%).

The results of pollen analysis are presented as a summary diagram including: the record of total pollen concentration, the records of cumulative percentages of conifers (mostly represented by *Pinus*, *Juniperus*, and *Abies*), riparian trees (*Alnus, Salix, Populus,* and *Tamarix*), deciduous trees (mostly deciduous *Quercus*, *Corylus*, *Fagus*, *Ostrya*/*Carpinus orientalis*, *Carpinus betulus* and *Ulmus*), evergreen trees and shrubs (evergreen *Quercus,* Ericaceae, *Phillyrea,* and *Pistacia*), anthropogenic indicators (including *Castanea*, *Olea,* and other cultivated and anthropocore plants such as *Juglans*, *Vitis, cereals,* etc.), the record of Arboreal Pollen (AP) percentages. The group "other herbs" includes all the remaining herbaceous taxa. We decided to count *Olea* and *Castanea* in the anthropogenic indicator, instead of evergreen and deciduous trees respectively, because their trend in the pollen record seems mostly determined by human activity.

The distribution of the modern vegetation of the Gulf of Gaeta borderlands appears to be strongly related to both the inland orographic complexity and the vicinity of the sea, being influenced by insolation, altitude, moisture availability and soil (Blasi et al., 2014 for a recent bioclimatic classification of the area). *Sclerophyllous shrublands* and *Quercus ilex* woodlands generally dominate the coastal promontories and the south-facing slopes at low altitudes (ca. 0-600 m), while mixed evergreen/deciduous and deciduous forest formations are more frequent at higher altitudes, favoured by orographic humidity. In the limestone massif of the Ausoni and Aurunci mountains, for example, *Quercus pubescens* woodland is mostly distributed on the footslopes, whereas *Quercus cerris* woodland dominates the bottom of the intra-montane karst plateaus. The north facing slopes of these mountains are rich in *Carpinus orientalis* and *Ostrya carpinifolia* woods, located in the hilly and montane zone respectively. The highest altitudes of the montane zone are covered by *Fagus sylvatica* forests (Di Pietro, 2011). In the volcanic district of Roccamonfina, chestnut cultivations represent the main element of land cover (Catalano et al., 2010; Croce and Nazzaro, 2012). Conifer forests have a restricted patchy distribution in the land bordering the Gulf of Gaeta; including coastal and inland *Pinus* plantations (Croce and Nazzaro, 2012). The agricultural areas, with arable lands and permanent orchards and olive groves, extensively cover plains and foothill zones.

3.7 Tephrostratigraphic analysis

Tephra recognition was driven by the occurrence of peaks in magnetic susceptibility signal coupled to the inspection of washed sediments (>63 μm fraction) used for planktonic foraminifera analysis. Cryptotephra samples, used for chemical analyses, were labelled with C5/ followed by an alphanumerical code pointing to the depth in cmbsf of the very base of the deposit (C5/53; C5/319; C5/403; C5/414 and C5/437).

For each sample, at least 30 juvenile fragments were then embedded in epoxy resin and suitably polished for microprobe analysis. In situ Energy Dispersive Spectrometric (EDS) analyses were performed on glass shards and loose minerals using JEOL JSM-5310 SEM at CISAG (Centro Interdipartimentale di Servizio per Analisi Geomineralogiche) of University of Federico II Napoli through Oxford Instruments Microanalysis Unit, equipped with an INCA X-act detector. Operating conditions were 15 kV primary beam voltage, 50-100 mA filament current, 50 sec acquisition time with variable spot size. Correction for matrix effect was performed using INCA version 4.08 software that used the XPP correction routine, based on a Phi-Ro-Zeta approach. Primary calibration was performed using international mineral and glass standards USMN reference samples according to the following scheme: Anorthoclase 133868 for Si and Na, Microcline 143966 for Al and K, Fayalite 85276 for Mn, Anorthite 137041 for Ca, Hornblende 143965 for Fe, Mg and Ti, Scapolite 6600-1 for Cl, Apatite 104021 for P. Precision and accuracy were assessed using the rhyolitic glass USMN 75854 as secondary standard. Mean precision was <5% for SiO_2, Al_2O_3, K_2O, CaO and FeO, and around 10% for the other elements.

4. Chronology

4.1 Radionuclides ^{210}Pb and ^{137}Cs

The ^{210}Pb activity-profile in composite core SW104-C5-C5 records an exponential decline with depth (Fig. 3), suggesting a constant sedimentation accumulation in the topmost part of the core. Using this profile, the sedimentation rate was calculated back to 60 cm bsf, applying the CF-CS model (Sanchez-Cabeza and Ruiz-Fernández, 2012). A mean sedimentation accumulation rate of 0.46 cm/yr was obtained, defining an age of 1885 AD at 60 cm to the sediment surface (Fig. 3). The ^{137}Cs activity is low, as previously reported in the Gulf of Salerno by Vallefuoco et al. (2012), but shows a clear trend, detectable from 34.5 cmbsf (Fig. 3). The peaks at 30.5 cmbsf and at 23.5 cmbsf, associated with 1945 AD (beginning of nuclear testing) and to 1963 AD (maximum ^{137}Cs fallout from nuclear testing), respectively, have been used as two independent tie-points for the construction of the age-depth profile (Fig. 3).

4.2 Tephrostratigraphy

Five cryptotephras were recognised along the record and they consist mainly of pumice, glass shard and minor scoria fragments, with rare lava lithic clasts and variable amounts of loose crystals. Three tephra layers (sampled at 437, 414 and 403 cm bsf), almost entirely made up of fresh glass, are interbedded within the stratigraphic interval from 437 cm bsf to 401 cm bsf and are mostly characterised by volcanic materials and minor bioclastic fragments, which suggest a continuous period of volcanoclastic input into the sedimentary system (Tab. 3).

The results of chemical analyses are reported in Table 4 as average values of individual point analyses for each sample recalculated to 100% water free. Individual chemical data points are given in Online Supplementary Materials. The analysed tephras have a wide range of composition ranging from phono-tephrites (tephra C5/53) to tephri-phonolites/latites (C5/319 and C5/414), trachytes and trachy-phonolites (C5/403, C5/414 and C5/437) according to TAS (Total alkali/Silica; Le Maitre,

2005) classification diagram (Fig. 4). Chemical features clearly indicate a provenance from the currently active volcanoes of the Neapolitan area (Ischia Island, Campi Flegrei, and Vesuvius).

4.2.1 Tephra C5/58

This is the youngest pyroclastic deposit, found in the core between 57-58 cm bsf Its lithological features and the occurrence of leucite bearing scoriae, phono-tephritic in composition, are typical of Somma-Vesuvius deposits younger than 79 AD (Santacroce et al., 2008). Stratigraphy, ^{210}Pb and ^{137}Cs results clearly indicate for this tephra layer an emplacement slighlty younger than 1885 AD (see above). Taking into account this chronological constrain along with the good chemical match between the studied sample and proximal deposits (Tab. 4), we relate tephra C5/58 to the 1906 eruption, the final phase of which produced fine ash fragments singularly spread towards the west of the volcano (Mastrolorenzo et al., 1993; Barsotti et al., 2015) and possibly affecting the core site area. The occurrence of these deposits in the Gaeta Bay represents the first finding of the post-1631 Vesuvius products in marine settings north of Naples Bay.

4.2.2 Tephra C5/319

The composition of glasses straddles the boundary between latites and tephri-phonolites (Fig. 4). They show a TiO_2 content mostly exceeding 0.8%, which represents a threshold that discriminates latites erupted at Ischia island from those erupted at Campi Flegrei (CF) in the last 4000 years (Fig. 5). Few analytical points display high Al_2O_3 values (ca. 20 %) for these types of rocks (Tab. 4) and this chemical feature was recently reported by D'Antonio et al. (2013) for a number of latitic deposits outcropping on the island. During the last 4000 years several close in time low VEI events took place with a dispersal area of products restricted to in narrow sectors around the vent (de Vita et al., 2010). Among them, the Vateliero eruption (VI-IV cent. BC, Tab. 3), occurred in the south-eastern sector of Ischia, was characterised by a sustained column phase emplacing a well sorted pumice fallout with a maximum thickness of 1 m in the vent area (Unit EUB2; de Vita et al., 2010). The good chemical match between tephra C5/319 and the glass fractions of the Vateliero deposits (D'Antonio et al., 2013), allows us to infer this land-sea correlation (Tab. 4). The correlation of this tephra with the Vateliero volcanic deposits is also supported by the occurrence in the study core of this tephra layer just above the acme end of *G. quadrilobatus* (see paragraph 4.3), dated at 2.7 ka BP by Lirer et al. (2013). As for the 1906 tephra, the occurrence of tephra C5/319 at this site represents the first finding of Vateliero products in a marine setting, thus enlarging the previously known dispersal.

4.2.3 Tephra C5/403, C5/414, C5/437

Glass fragments of tephras C5/403 define two slightly different compositions in the phonolite field, tephra C5/414 has a bimodal composition since it associates a minor tephri-phonolitic/latitic population to the prevailing trachy-phonolitic one, and C5/437 shows a homogeneous trachy-phonolitic composition (Fig. 4). Chemical features of the analysed tephras are typical of Campi Flegrei (CF) products erupted during the last 5000 years (Smith et al., 2011). In particular, the bimodality of tephra C5/414 allows us to correlate this deposit to the Astroni3 event (Tab. 4) which is the only terrestrial counterpart showing such a peculiar chemistry for that time period (Smith et al., 2011). Taking into account this strong chemistry-supported correlation, the Astroni3 tephra can be considered a good chronological marker in the studied succession at 4098-4297 years BP (Smith et al., 2011) and it helps to temporally constrain tephras C5/403 and C5/437, which are otherwise characterised by a barely distinguishable glass chemistry.

In order to find a possible counterpart for tephra C5/403, we set its chemistry against that of Campi Flegrei deposits younger than Astroni3 and, according to SiO_2vsCaO variation diagram, a tentative correlation could be proposed with Astroni6 products (4297-4192 years BP, Smith et al., 2011) characterised by a comparable large variability (Fig. 6a). However, the modelled age for tephra C5/403 ranges from ca. 3939 years BP to 4100 years BP (see section 4.3) in good agreement also with the age of Capo Miseno event obtained from both proximal (3700±500 years; Di Renzo et al., 2011) and offshore (3904±60 cal. years BP; Sacchi et al., 2014) deposits. Moreover, the most representative chemical composition of C5/403 (16 out of 20 individual point data) is also well comparable to the chemistry of Capo Miseno glasses (Tab. 4). The co-occurrence of this tephra layer with the base of *G. quadrilobatus* acme event (see section 4.3), dated at 3.7 ka BP (Lirer et al., 2013), strongly supports the correlation with Capo Miseno event.

Tephra C5/437 (9 cm below Astroni3) is the most prominent along the record and characterised by a large amount of fresh glass with different morphologies (Tab. 3). In the SiO_2vsCaO variation diagram we compared its composition with that of Astroni1 and Agnano Monte Spina (4153-4345 years BP and 4482-4625 years BP, respectively; Smith et al., 2011) glasses and a fair agreement can be observed with the latter ones (Fig. 6b). Lithology and thickness (~14 cm) of the deposit support the correlation of tephra C5/437 with the eruption of Agnano Monte Spina (AMS) which represents the highest magnitude event of this time span. The correlation of this tephra with the AMS volcanic deposits is also supported by the occurrence in the study core of this tephra layer just above the strong drop in abundance of *G. truncatulinoides* left coiled (Fig. 7) dated at 4.5 ka BP by Lirer et al. (2013). The products of the AMS eruption have been recently found in the Salerno Bay (Amato et al., 2012; Lirer et al., 2013), although they are mainly spread towards the eastern sector of the volcano (de Vita et al., 1999) and frequently found in the marine archives of the Adriatic Sea (Zanchetta et al., 2011 and references therein).

4.3 Age model

The age model has been constructed starting by radionuclides ages (^{210}Pb activity-depth profile and ^{137}Cs activity) for the last ca. 150 years and tephrostratigraphy of five tephra layers recorded in the study core [Vesuvius (1906 AD), Vateliero-Ischia (2.4-2.6 ka BP), Capo Miseno (3.9 ka BP), Astroni3 (4.1-4.3 ka BP), Agnano M. Spina (4.42 ka BP)] (Tab. 5). In addition we considered the following planktonic foraminiferal events: i) the abundance peak of *Globorotalia truncatulinoides* left coiled (1718 ±10 yr AD, Lirer et al. 2013; 2014); ii) the acme interval of *Globigerinoides quadrilobatus*, (base 3.7 ± 0.048 ky BP top 2.7 ± 0.048 ky BP, Lirer et al., 2013) (Tab 5). These

bioevents are documented in the different basins of the Mediterranean Sea and well time-constrained (Sprovieri et al., 2003; Di Bella et al., 2014; Cisneros et al., 2015). The chronology of the stratigraphic interval between the acme base of *G. truncatulinoides* left coiled and the top of *G. quadrilobatus* acme interval, has been obtained thought the tuning of the $\delta^{18}O_{\textit{G. ruber}}$ record with the same signal from core C90 (Gulf of Salerno, south Tyrrhenian Sea, Lirer et al. 2013; 2014) (Fig. 7). The good visual comparison between the $\delta^{18}O_{\textit{G. ruber}}$ signal from the study site (with data from south Tyrrhenian Sea (Lirer et al., 2013, 2014), Gulf of Taranto (Grauel et al., 2013a) and eastern Mediterranean (Schilman et al., 2001) (Fig. 8) increases our confidence on the robustness of the our age model.
Linear interpolation between the tie-points used for constructing of the age-depth profile shows a progressive decrease in sedimentation rate from the top down to the base core (Fig. 7).

4.4 Magnetostratigraphy

The paleomagnetic inclinations were used to confirm the age model by comparison with a reference curve. In the absence of a sufficiently detailed reference curve for Italy or the Tyrrhenian Sea, we used data generated for the sampling location by the SHA.DIF.14k model of Pavón-Carrasco et al. (2014). The temporal resolution of the geomagnetic model is lower than that of our data, but sufficient to check the age model on centennial and longer time scales.
We conducted the comparison by using the age tie points listed in Table 5, and tuning the age-depth transformation between the tie-points using the Match software of Lisiecki and Lisiecki (2002). The age-inclination curve for Core C5 is shown in Figure 9, with the reference curve included for comparison. The uppermost 60 cm of the core are omitted from this analysis, since the sediment was too liquid for reliable paleomagnetic measurement; soft sediment deformation is probably also responsible for the abnormally low inclination near the top of the measured core. There is good agreement between the major features of C5 inclination record and the reference curve, confirming the strength of the constructed chronology. Slight divergences can be explained in part by limitations on the accuracy of the model data at this location.

5. Results

5.1 Oxygen isotope analysis

The oxygen isotope signal measured out on the planktonic foraminiferal species *G. ruber* alba variety, from the last five millennia varies between 0.87‰ to -1.91‰ with a mean value of -0.12‰ (Fig. 10). $\delta^{18}O_{G.\ ruber}$ signal shows from the base of the core up to 200 AD a gentle shift from 0.86‰ to 0.12‰ (Fig. 10). This long interval is characterised by five distinct lower $\delta^{18}O_{G.\ ruber}$ values centred at 2600 BC, 1800 BC, 1600 BC, 1400 BC, and 1200 BC (Fig. 10).

During the last two millennia, $\delta^{18}O_{G.\ ruber}$ signal shows an increase in frequency and amplitude oscillations than the previous three millennia BC (Fig. 10). In particular, high $\delta^{18}O_{G.\ ruber}$ values (mean values of 0.5 ‰) are documented at 500 AD and 800 AD, between 1300-1600AD and 1700-1800 AD and at 1900 AD (Fig. 10). Low $\delta^{18}O_{G.\ ruber}$ values (mean values of -0.5 ‰) are detected at 350 AD, 600 AD, 1200 AD, 1650 AD, and at 1850 AD (mean values of -1 ‰), and at 1980 AD (mean values of -1.5 ‰) (Fig. 10).

5.2 Planktonic foraminifera

Planktonic foraminifera are abundant, well preserved and with mostly a very thin test. *G. ruber* alba and *G. elongatus* show a progressive decreasing trends from base core to 400 AD (Fig. 10). Upwards, *G. elongatus* is still present to 1550 AD (Fig. 10) with very low abundance (<2%), while from 1550 AD to present day it is almost absent (Fig. 10). Conversely, *G. ruber* alba shows two increasing trends, one starts at 400 AD and a second one at 1550 AD (Fig. 10). The onset of this latter abrupt increase in abundance (from 20 to 70 %) fits with the virtual absence of *G. elongatus* (Fig. 10).

T. quinqueloba and *G. glutinata* display a progressive increase in abundance to ca. 1650 AD. From 1650 AD to 1900 AD, these taxa show a strong decrease in abundance while in the uppermost part of the core (from ca. 1900 AD to present day) they go through a significant percentage increase up to

60% (Fig. 10). *G. quadrilobatus* shows a peak in frequency from 700 BC to ca. 1750 BC reaching percentages of 25% (Fig. 10).

The other planktonic foraminiferal species show percentages ranging between 0.1 and 25% and only occasionally reached significant frequencies. *G. ruber* pink variety, *G. truncatulinoides* and *G. inflata* left coiled have a scattered distribution pattern and they are continuously present only from ca. 350 AD to present day. It is noteworthy the significant peak in percentage of *G. truncatulinoides* (17%) and *G. inflata* left coiled (5%) during the Maunder phase (Fig. 10).

G. siphonifera and *Orbulina* spp. are present with low percentages along the core and occasionally shows distinct peaks in frequency (Fig. 10). *G. scitula* and *N. pachyderma* right coiled group generally show very low percentages through the entire investigated interval. (Fig. 10).

5.3 Pollen analysis

The main vegetation features profiled by the pollen record over the last 5000 years suggest a landscape characterized by mixed evergreen and deciduous oak-dominated woodlands, showing major changes in both structure and floristic composition (Fig. 11). Between 3000 and 900 cal. BC (Fig. 11) the forest composition shows the prevalence of evergreen elements, recorded in high frequencies (>45%). Between 900 and 100 cal. BC a clear decrease in evergreen trees and shrubs is accompanied by an increase in herbs (max. 46%), enriched by xeric taxa like *Artemisia*, thus indicating an opening of the forest vegetation (Fig. 11). During the last two millennia, a fluctuating trend of broadleaved trees is recorded. The forest cover expanded, between ca. 100 BC and 800 AD, mostly due to a general increase in deciduous taxa (up to 42%), while after 800 AD the landscape experienced a major new opening, resulting from a clear increase in herbaceous taxa and a further decline of broadleaved trees, especially evergreen taxa (Fig. 11). In particular, the AP pollen record shows two forest drops, from 800 to 1100 AD and from 1600 and 1850 AD, intermixed by a new moderate increase in arboreal vegetation from 1100 to 1600 AD (Fig. 11). Finally, the last two centuries are characterized by a new arboreal vegetation expansion (67%), mostly related to an increase in conifers dominated by *Pinus* (Fig. 11).

This last time interval are also marked by high frequencies of anthropogenic pollen indicators (up to 24%), highlighting an undeniable influence exerted by human activities on the natural environment (Fig. 11).

6. Discussion

6.1 Paleoenvironmental reconstruction in the central Tyrrhenian Sea

An integrated dataset based on planktonic foraminifera, pollen, tephrostratigrahy and oxygen isotopes analysis performed on a marine sediment core collected in the central Tyrrhenian Sea allows identifying nine paleoclimatic intervals during the past five millennia, with decadal resolution. Based on our age model, these intervals correspond to the recent archaeological/cultural periods: Eneolithic, Early Bronze Age, Middle Bronze Age - Iron Age, Roman Period, Dark Age, Medieval Climatic Anomaly, Little Ice Age, Industrial Period and Modern Warm Period (Fig. 10, 11).

6.1.1 Eneolithic: base of the core to ca. 2410 BC/base core – ca. 4360 BP

Generally warm-water conditions dominate during this interval as documented by the high percentages in frequency of *G. ruber* alba, *G. elongatus* and *Orbulina* spp..

$\delta^{18}O_{G.ruber}$ values associated with maximum abundance of *G. elongatus* evidence an increase in temperatures at ca. 2600 BC (Fig. 10). In addition, the strong decrease in frequency of *T. quinqueloba* and *G. glutinata*, assumed as proxies of seasonal flooding events (Vallefuoco et al., 2012), suggests an environment setting with a reduced river runoff. Our reconstruction is in agreement with the paleoenvironmental scenario proposed by Piva et al. (2008) in the Adriatic Sea at time of the warm event W4-Copper Age.

The pollen record documents a slight and slow opening of the forest vegetation especially in the broadleaved evergreen taxa (Fig. 11), suggesting a progressive establishment of a more arid climate in the third millennium BC (Di Rita and Magri, 2012).

6.1.2 Early Bronze Age: ca. 2410 BC to ca. 1900 BC/ca. 4360 BP – ca. 3850 BP

The beginning of the Early Bronze Age at ca. 2410 BC (Fig. 10) corresponds to a turnover between carnivorous and herbivorous-opportunistic planktonic foraminifera marks.

The high abundance of *G. ruber*, G. *siphonifera* and *G. elongatus* in the lower part of this interval until ca. 2400 BC reflects warm summer conditions (Fig. 10). The highest frequencies of *G. ruber* in the Mediterranean are generally reported at the end of the summer (Pujol and Vergnaud-Grazzini, 1995) or in fall (Bàrcena et al., 2004). At the same time, the occurrence of *G. truncatulinoides* and *G. glutinata* (Fig. 10) suggests the prevalence of cold, well-mixed, nutrient-rich waters in winter (Sprovieri et al., 2003). A similar climatic reconstruction was reported in the Adriatic Sea and in the north-eastern Ionian Sea (Piva et al., 2008; Geraga et al., 2008) evidencing an atmospheric connection among the different basins of the Mediterranean Sea.

From ca. 2300 BC to ca. 2050 BC, strong increase in *T. quinqueloba* abundance, associated with the $\delta^{18}O_{G.ruber}$ signal enrichment (from 0‰ to 0.5‰) (Fig. 10) reflects the cold event correlable with the so-called '4.2 kyr event' observed at a global scale (from North America, through the Middle East to China; and from Africa, parts of South America, and Antarctica, Mayewski et al., 2004; Staubwasser and Weiss, 2006; Walker et al., 2012).

During the Early Bronze Age, clear evidence of the "4.2 kyr" deforestation event is also documented in the pollen record (Fig. 11), where the process of landscape opening, starting at ca. 2700 BC and reaching a maximum at around 2200 BC, mostly affected the evergreen vegetation (Fig. 11). Our data are consistent with other pollen records in the Central Mediterranean (south of 43° N), showing a deforestation process (ca. 2500-1900 BC) that had a great impact on the evergreen forests cover (Sadori and Narcisi, 2001; Di Rita and Magri, 2009; Tinner et al., 2009; Di Rita et al., 2011).

6.1.3 Middle Bronze Age-Iron Age: ca. 1900 BC to ca. 500 BC/ca. 3850 BP – ca. 2450 BP

The base of the Middle Bronze Age-Iron Age period at 1900 BC is marked by a dominance of herbivorous-opportunistic planktonic foraminifera (Fig. 10). This phase is basically characterized by the acme interval (from ca. 1752 BC to ca. 750 BC) of planktonic foraminifer *G. quadrilobatus*. This species is indicative of warm, oligotrophic surface waters in summer (Hembleben et al., 1989; Pujol and Vergnaud-Grazzini, 1995). Their co-occurrence with *G. ruber* alba and *Orbulina* spp. suggests oligotrophic conditions (at least in summer) during the Middle Bronze Age-Iron Age period. At the same time, the high abundance percentages of the herbivorous-opportunistic species *T. quinqueloba* and *G. glutinata* indicates high productivity surface waters, strong seasonality and the presence of continental runoff. This latter condition is also documented by Zolitschka et al. (2003) in Lake Steisslingen and Lake Holzmaar in Germany, where a period of increasing runoff, between 850 BC and 750 BC, was related to wetter climatic conditions (prolonged suppression of radial tree growth).

During the Middle Bronze Age-Iron Age period, we identify two sub-intervals:

i) from 1850 BC to 1450 BC, there is an increase in abundance of warm water taxa *G. quadrilobatus*, *G. siphonifera* and *Orbulina* spp. (Pujol and Vergnaud-Grazzini, 1995) associated with *G. elongatus* increase and lower values of the $\delta^{18}O_{G.ruber}$ signal (Fig. 10). We interpret these paleoproxies as the result of a warm phase chronologically correlated with the warm event W3 (Late Bronze Age) reported by Piva et al. (2008) in the Adriatic Sea;

ii) at ca. 1050 BC, high values in the $\delta^{18}O_{G.ruber}$ signal (Fig. 10) result time equivalent to the A1-cold spell event, marked by a decrease in alkenone SST and the C2 cold event (Iron Age) reported in the Adriatic Sea (Sangiorgi et al., 2003; Piva et al. 2008) and in eastern Mediterranean Sea (Rohling et al. 2002).

From 1900 BC to 900 BC, pollen data suggest a forested landscape, with a prevalence of evergreen trees and shrubs (Fig. 11). A similar vegetation pattern is observed in many coastal and inland pollen sites of the central Mediterranean region (Di Rita and Magri, 2009, 2012 and references therein), especially those affected by the aridification processes at 4.2 ky BP, where a phase of increased evergreen vegetation (ca. 2000-900 BC) is clearly recorded after the deforestation. This forest recovery may have been influenced by the establishment of a generally stable humid and warm climate phase, whose regional signature seems reflected also in the pollen record and in the *G. quadrilobatus* acme interval (Fig. 10) of the study core.

Between ca 900 BC and 500 BC a new drop in forest vegetation is recorded, coupled with an increase in *Artemisia* and other xerophytes (Fig. 11). This process may be the effect of new dry climate phase probably induced by the 2.8 ky BP event (Bond event 2) (Bond et al., 2001). This short climate change, was also documented by a decrease in arboreal pollen percentages in the central Mediterranean region (Joannin et al., 2012; Azuara et al., 2015), associated with a decrease in magnetic solar activity (van Geel et al., 2000; Martìn-Puertas et al., 2012).

6.1.4 Roman Period: ca. 500 BC to ca. 550 AD/ca. 2450 BP – ca. 1400 BP

The lowermost part of Roman period is characterized by the occurrence of *G. ruber* pink variety and high abundance of *G. ruber* alba up to ca. 150 BC (Fig. 10), suggesting warmer climatic conditions with respect to the late Roman Period. This information fits the reconstructed Sea Surface Temperatures in the central-western Mediterranean Sea (Martínez-Cortizas et al., 1999; Lirer et al., 2014; Cisneros et al., 2015), documenting consistent marine thermal responses to climatic changes during this time interval.

Between 300 BC and 100 BC the abrupt increase of the *G. scitula* - *N. pachyderma* group (Fig. 10), is interpreted as the result of the winter cold phase between 350 BC and 100 BC reported in the historical documents describing frozen Tiber in Rome (Lamb, 1977).

The pollen data in the first 500 years of the Roman period shows relatively open condition, as suggested by appreciable values of herbaceous taxa, including *Artemisia* (Fig. 11).

During the upper part of the Roman Period (at ca. 200 AD), the $\delta^{18}O_{G.ruber}$ record shows a change in frequency and amplitude oscillations that corresponds to an increase in abundance (up to 80%) of herbivorous-opportunistic planktonic foraminifera (Fig. 10). This $\delta^{18}O_{G.ruber}$ signature allows to identify three cold intervals, previously documented by Lirer et al. (2014), which can be correlated with the solar activity (Roman I, Roman II and Roman III) (Fig. 10). These phases are characterized by a slightly decrease in abundance of high surface water planktonic foraminiferal indicators *T. quinqueloba* and *G. glutinata*, probably associated to the low temperatures and absence of runoff. Moreover, the observed distributional pattern of *G. ruber* pink variety, the occurrence of *G. scitula* – *N. pachyderma* group and the strong decrease in abundance of *T. quinqueloba* - *G. glutinata* group (Fig. 10) suggests relatively wetter condition during summer and cold-dry ones during winter only during the Roman III event.

6.1.5 Dark Age: ca. 550 to ca. 860 AD/ca. 1400 BP – ca. 1090 BP

The early Dark Age (from 550 to 750 AD) is characterized by warm climatic conditions documented by $\delta^{18}O_{G.ruber}$ values (-0.5 ‰) and the increase of warm water species *G. ruber*, *G. siphonifera* and *Orbulina* spp. (Fig. 10). These species are currently very abundant in the Tyrrhenian Sea, especially in the Gulf of Naples (De Castro Coppa et al., 1980; Pujol and Vergnaud-Grazzini, 1995; Sprovieri et al., 2003).

In the upper part of this interval (from 750 to 860 AD), the $\delta^{18}O_{G.ruber}$ signal associated with maximum abundance of the cold *G. scitula* - *N. pachyderma* group and a decrease in warm water species, shows a cooling event corresponding to the Roman IV Period (Fig. 10). This event, documented also in the Salerno Gulf (southern Tyrrhenian Sea) by Lirer et al. (2014), agrees with the $\delta^{18}O$ records of the Taranto Gulf (Grauel et al., 2013) and Adriatic Sea (Piva et al., 2008) (Fig. 8), but shows some differences with the western Mediterranean Sea (Cisneros et al., 2015), probably due to the different resolution scale.

Between 400 and 600 AD, a decline in the values of trees suggests a reduction of forest cover (Fig. 11). This was facilitated by a rapid decrease in both evergreen and deciduous broadleaved taxa, as well in conifers, although a marked expansion of *Castanea*, among anthropogenic indicators, contributed to keep high the tree percentage values. This complex forest dynamics may be explained by the influence of cooler climate and decreased humidity (cf. Dark Ages Cold Period), which may have had a significant impact on the development of the natural tree populations, already cleared by man in favour of chestnut forestry.

The dryness, characterizing the Dark Age, is correlable with decreased humidity in the western Mediterranean (Nieto Moreno et al., 2011), evidenced by forest cover regression episodes (Jalut et al., 2000, 2009; Combourieu-Nebout et al., 2009), a decrease in river activity in southern Europe (Magny et al., 2002; Macklin et al., 2006), cooling events in the Balearic Basin (Frigola et al., 2007) and lower lake levels in southern Spain (Carriòn, 2002).

6.1.6 Medieval Climate Anomaly (MCA): ca. 860 to ca. 1250 AD/ca. 1090 BP – ca. 700 BP

The transition between the Dark Age and the Medieval Climate Anomaly (MCA) Period is associated to the transition from carnivorous to herbivorous–opportunistic planktonic species that culminate at 1220 AD, when the carnivorous taxa dominate (Fig. 10).

Several authors (e.g., Lamb, 1977; Jones et al., 2004; Mann et al., 2009; Büntgen and Tegel, 2011) described this interval as a relatively stable and warm period. The planktonic foraminifera from this time interval document a general temperate climate condition as testified by the coexistence of warm and cold species in the planktonic foraminiferal assemblage (Fig. 10). In addition, a reduction in abundance of *G. ruber* alba from ca. 1000 to ca. 1100 AD associated with a slightly increase of *G. ruber* pink, seems to suggest less temperate and humid conditions. During this short time interval, the decrease in deciduous trees and a considerable parallel increase in herbaceous taxa document a rapid and significant opening of the vegetation landscape. This expansion of herbaceous communities may have been favoured by an oscillation towards a more arid and cool climate. We associate this feature with the “Medieval cold” phase corresponding to the $\Delta^{14}C$ Medieval minimum (Stuiver and Braziunas, 1988) (Fig. 11).

Similar dry condition are also documented in the Iberian Peninsula (Moreno et al., 2012) based on multiproxy evidence (e.g. lake levels decrease, presence of xerophytic and heliophytic vegetation, low frequency of floods, major Saharan eolian fluxes, and less fluvial input to marine basins) and in the Alboran Sea basin (Nieto Moreno et al., 2013).

Between 1220 and 1250 AD, a marked shift in $\delta^{18}O_{G.ruber}$ signal towards negative values, associated with a strong increase in *G. ruber* abundance, document the warmest interval (Medieval Warm Period) occurring during the MCA (Fig. 10). At that time, the increase in abundance of *G. truncatulinoides* (Fig. 10) suggests the presence of a deep mixed layer during winter.

During this short time interval, pollen data show a new forest recovery, mostly related to an increase in both deciduous and evergreen arboreal taxa, suggesting a climate change toward more warm and humid condition that favoured the growth of broadleaved populations (Fig. 11).

This climate feature may be correlated with the abrupt increase of ∼1–1.5 °C in the SST profile occurring around 980 AD in the North Icelandic (Sicre et al., 2008).

6.1.7 Little Ice Age Period (LIA): ca. 1250 to 1850 AD/ ca. 700 BP – ca. 100 BP

The MCA–LIA transition is the last global-scale Rapid Climatic Change (RCC) event reported in the Holocene by Mayewski et al. (2004) and is recognizable also in the Mediterranean marine records (e.g., Piva et al., 2008; Incarbona et al., 2010, Lirer et al., 2014; Goudeau et al., 2015). This transition is marked by an important change in nutrient availability in water column (Lirer et al., 2014), documented by the change from carnivorous to herbivorous-opportunistic planktonic foraminifera taxa (Fig. 10). The high-resolution $\delta^{18}O_{G.ruber}$ data allowed us to identify, within the Little Ice Age, four climatic oscillations related to solar activity: Wolf, Spörer, Maunder and Dalton cold events (Fig. 10), also already documented in the Gulf of Salerno (Lirer et al., 2014).

The Wolf and Spörer climatic phases are characterised by a shift to cooler conditions suggested by the increase of cool water planktonic foraminiferal species *G. inflata*, *G. truncatulinoides* and *G. scitula* - *N. pachyderma* group, by a decrease of the warm water taxon *G. ruber* (Fig. 10) and by an increase of high productivity surface waters taxa (*G. glutinata - T. quinqueloba*) (Fig. 10). The increase in regime productivity agrees with the results obtained in the Sicily Channel and in the Adriatic Sea (Piva et al., 2008; Incarbona et al., 2010; Siani et al., 2013) and in the western Mediterranean Basin (Nieto Moreno et al., 2011). This might indicate a larger southward extension of the westerlies leading to increase in precipitation and thus an enhanced outflow of the rivers during this interval.

Around 1600 AD a shift of $\delta^{18}O_{G.ruber}$ signal from 0.5 ‰ to -1 ‰ and the increase of warm-water taxa *G. siphonifera* and *G. quadrilobatus* marked the warm interval between Spörer and Maunder (Fig. 10).

The onset of the Maunder is characterized by a strong increase of abundance of *G. truncatulinoides* and *G. inflata* (Fig. 10) suggesting the presence of a deep mixed layer during winter. This oceanographic feature can be induced by the increase in winds intensity probably due to Atmospheric blocking events. Atmospheric blocking events are mid-latitude weather systems where a quasi-stationary high-pressure system, located in the Northeast Atlantic, modifies the flow of the westerly winds by blocking or diverting their pathway (Moffa Sanchez et al., 2014).

Blocking is accompanied by cold winter temperatures in Western Europe; the climatological maximum in winter blocking days is located over Western Europe, with a secondary maximum over Greenland (Häkkinen et al., 2011). Variability of atmospheric blocking over years to several decades shows correlation with the ocean surface temperature and with significant changes in Atlantic Ocean circulation, mediated by wind-stress curl and air-sea heat exchange (Häkkinen et al., 2011).

The Maunder Minimum (MM) represents the coldest period of the Little Ice Age, an interval of reduced solar activity. Within the MM, the Late Maunder Minimum was a period of persistent extremely cold winters in Europe (Barriopedro et al., 2008). Barriopedro et al. (2008) relate the particular cooling recorded in Maunder in the Northern Hemisphere to events of Atlantic blocking. Moffa Sanchez et al. (2014) shows that small-scale atmospheric patterns in the North-East Atlantic, such as atmospheric blocking events as part of the east Atlantic pattern or polar mesoscale storms, may considerably contribute to driving North Atlantic surface circulation.

At the end of the LIA, around in 1850 AD, a strong change in the pattern of carnivorous and herbivorous planktonic foraminifera is recorded (Fig. 10).

The pollen record shows a new significant decrease in forest cover (Fig. 11). This process started at ca. 1300 AD and ended ca. 1850 AD, in good agreement with the chronological evidence of the LIA interval. The deforestation process seems to have particularly affected the broadleaved taxa, whose curves show the lowest percentage values of the entire sequence between 1650 and 1750 AD, in correspondence with the Maunder Minimum (Fig. 11). The cool climate associated with this minimum of solar activity may have affected the development of many arboreal taxa populations, except conifers that show a moderate increase (Fig. 11).

6.1.8 Industrial Period: ca. 1850 – 1950 AD

The Industrial Period is characterized by an increase of warm water species *G. quadrilobatus* and *G. ruber* that reached here their maximum percentages (Fig. 10). The coexistence of these taxa indicates the presence of the mixed layer in the water column (Spooner et al., 2005).

The onset of Damon (approximately 1900 AD) is marked *G. ruber* pink and *G. elongatus* peak suggesting a warm conditions confirmed by the absence of cold planktonic foraminiferal taxa *G. truncatulinoides* and *G. inflata* (Fig. 10). This record is consistent with studies performed in the Gulf of Taranto that report an increase in temperature at the beginning of the 20th century (Taricco et al., 2009; Grauel et al. 2013a).

The dominance of herbivorous-opportunistic planktonic foraminiferal species and the $\delta^{18}O_{G.ruber}$ signature can be interpreted as the occurrence of a humid climatic phase during the entire Industrial Period from approximately 1850 AD to 1940 AD as reported by Nieto Moreno (2012) in the western Mediterranean region. A clear trend toward humid conditions is also clearly reflected by the pollen record, which highlights a general arboreal forest development, mostly due to conifers and evergreen broadleaved populations (Fig. 11).

6.1.9 Modern Warm Period: 1950 AD to the present day

The onset of the Modern Warm Period is characterized by a strong increase in *G. glutinata* and *T. quinqueloba* abundances suggesting high surface productivity (Fig. 10). At same time, *G. ruber* alba shows an abrupt decrease in abundance and $\delta^{18}O_{G.ruber}$ signal records the most prominent negative excursion (Fig. 10) of the last two millennia (from 0.5 ‰ to -1.5 ‰). These features have been previously documented by Lirer et al. (2014) in the Salerno Gulf (south Tyrrhenian Sea), suggesting a possible human overprint on global warm climate condition during the last 50 years. During this period, Vallefuoco et al. (2012) documented, in the marine record of Gulf of Salerno (South Tyrrhenian Sea), also a rapid increase of benthic foraminifer *Bulimina aculeata*, suggesting high productivity conditions (Corliss, 1985; Mackensen and Douglas, 1989; Jorissen et al. 1992; Sen Gupta and Machin-Castillo 1993; Rathburn and Corliss, 1994). We speculate that these micropaleontological features, with an increase in organic matter flux associated to strong increase in planktonic foraminifera related to high-surface water productivity and coexistence of oligotrophic condition [abundance increase of *G. quadrilobatus* in the study core and in the Gulf of Salerno (Lirer et al., 2014)], could confirm a strong overprint from human activities. During this interval in fact the building of dams in central-southern Tyrrhenian coastal zones during the middle of last century that resulted in a strong reduction of coarse- grained materials, which caused a change in sediment size and a possible change in nutrient supply, as hypothesized by Vallefuoco et al. (2012). The human impact on natural ecosystems is also visible in the arboreal vegetation which is dominated by *Pinus* and reflects extensive plantation of pine forests in the coastal areas of the Gulf of Gaeta (Fig. 11). In addition, the nearby territory was increasingly occupied by cultivations of *Olea*, *Vitis*, cereals and hemp, confirming intensive land exploitation for agriculture (Fig. 11).

6.2 North Atlantic Oscillations (NAO) index vs marine multiproxy in the central Mediterranean Sea

Atmospheric circulation patterns in the northern hemisphere influence climate variability in the Mediterranean region (Jalut et al., 1997, 2000; Combourieu Nebout et al., 2002; Goy et al., 2003; Roberts et al., 2012; Fletcher et al., 2012). The North Atlantic Oscillations (NAO) is the most important mode of variability in the atmospheric circulation over the North Atlantic, with considerable influences winter temperature/precipitation throughout the Eurasian continent and eastern North America (Greatbatch, 2000). However, the interpretation of the NAO effect in the Mediterranean area is ambiguous because the NAO influence is not stationary through time and space and also because it is a dominant mode only in winter (e.g., Xoplaki et al., 2008). In general, there is a high-pressure system over the subtropical region near the Azores, and a low-pressure situation over the subpolar region near Iceland (Wanner et al., 2001). Within this configuration, the NAO index gets positive in a stronger phase with a high-pressure gradient and negative in phases with a weaker pressure gradient (Brönnimann, 2005). In Italy, the NAO index modulates the winter precipitation (Brunetti et al., 2002; Tomozeiu et al., 2002; Caloiero et al., 2011; López-Moreno et al., 2011; Casanueva et al., 2014; Benito et al., 2015) following a opposite trend respect to the northern Europe (López-Moreno et al., 2011; Benito et al., 2015; and references therein). The NAO forcing has been shown also in the fossil marine sedimentary archives (i.e., Chen et al., 2011; Nieto-Moreno et al. 2013; Goudeau et al., 2015; Jalali et al., 2015) and consequently may be important to document this forcing also in the high-resolution shallow-water marine record of the central Tyrrhenian Sea.

The comparison between NAO index (Trouet et al., 2009; Olsen et al., 2012), $\delta^{18}O_{G.ruber}$ signal, planktonic foraminiferal herbivorous *versus* carnivorous taxa and pollen AP index shows important features useful to understand global forcing within this central area of the Mediterranean region (Fig. 12).

The positive NAO index from 2500 BC to about 900 BC is not interrupted by significant polarity changes, in contrast from 900 to 100 BC is characterized by a generally negative NAO index. The long term comparison between the NAO index and the $\delta^{18}O_{G.ruber}$ signal shows an overall antithetic correlation (even though peak-to-peak correlation is not possible due to the different resolution of the two proxies) during the last five millennia, with the exception of scattered parallelism between 0 and 200 AD, probably due to the local overprint (Fig. 12). This antithetic trend supports the NAO reconstruction of López-Moreno et al. (2011): when the NAO index is positive south Europe climate it is mild and dry; on the contrary, a negative index is associated with the reverse pattern.

Following the cooling phase related to the cold 2.8 kyr event (Bond cycle 2), from beginning of the Roman Period (ca.500 BC) upwards, the climate system displays a turnover *vs* a more positive NAO index associated with a long-term trend to lower $\delta^{18}O_{G.ruber}$ values and with a significant planktonic foraminiferal changes from carnivorous *vs* herbivorous-opportunistic species (Fig. 12). The herbivorous-opportunistic species are the dominant group over most of the last two millennia suggesting a strong connection with nutrient availability. In addition, from Dark Age (ca. 500 AD) upwards, the Mediterranean planktonic foraminiferal $\delta^{18}O$ data (Fig. 8) document a synchronous progressive long-term shift to more positive values (cooling trend) as recently documented by Cisneros et al. (2015) from SST stack of the Menorca basin.

This climate mode seems to change again around 1450 AD (mid Little Ice Age) when the NAO index starts to change to negative values and the $\delta^{18}O_{G.ruber}$ record shifts towards higher values (Fig. 12), suggesting the onset of the modern warm climate condition.

Pollen data, suggesting stable warm climate conditions between 1900 and 900 BC, seem to be in agreement with the aforementioned NAO index reconstruction by López-Moreno et al. (2011). In the last two millennia, distinguishing the influence of climate from human activity in pollen records is a very challenging task. However, some pronounced vegetation fluctuations, also reflected in the NAO index record, may be interpreted as mainly influenced by climate changes. In particular, during the Maunder minimum of the Little Ice Age, a phase of negative NAO, associated with cool climate, may have caused a major decrease of broadleaved forest cover, due both to its direct influence on tree growth and to increased human pressure on woodlands for firewood provision. A strong climate influence on vegetation may be also envisaged during the Medieval Climate Anomaly, when a rapid oscillation of the NAO index corresponds to a clear decrease in the arboreal vegetation (Fig. 12). When the NAO index reached its minimum, between 1000 and 1100 AD, the forest cover may have suffered a cooling of climate. Conversely, when the NAO index started to increase after 1100 AD, population of trees expanded, probably in response to the establishment of milder conditions (Fig. 12).

7. Conclusions

In this study, we used a multi-proxy approach in order to investigate the paleoclimate variability in the central Mediterranean region during the late Holocene. The robust chronological control performed on radionuclides, tephrostratigraphy and isotopic stratigraphy allows a reconstruction at a century scales since 2900 BC. Nine main intervals have been identified and correlated with the archaeological/cultural periods: Eneolithic (base of the core- ca. 2410 BC), Early Bronze Age (ca. 2410 BC – ca. 1900 BC), Middle Bronze Age - Iron Age (ca. 1900 BC – ca. 500 BC), Roman Period (ca. 500 BC – ca. 550 AD), Dark Age (ca. 550 AD – ca. 860 AD), Medieval Climate Anomaly (ca. 860 AD – ca. 1250 AD), Little Ice Age (ca. 1250 AD – ca. 1850 AD), Industrial Period (ca. 1850 AD – ca. 1950 AD), Modern Warm Period (ca. 1950 AD – present day). Within these time intervals, our multiproxy record shows several short-term climate oscillations, adding new details to the records studied in different areas of the Mediterranean Basin (Alboran Sea, Gulf of Salerno, Gulf of Taranto, Adriatic Sea and Ionian Sea). Strong modification in climate system occurs from the onset of the Roman Period up to the present-day, recorded by long term trend and amplitude oscillations of the $\delta^{18}O_{G.ruber}$ signal, by the onset of main planktonic foraminiferal turnover from carnivorous to herbivorous-opportunistic species, and by the consistent fluctuations of the pollen records.

The good correspondence between the observed climate oscillations and recognized archaeological intervals underline the role exerted by climate change in determining rises and declines of civilizations.

In addition, the antithetic correlation between the NAO index and $\delta^{18}O_{G.ruber}$ signal suggests a global climate signature in the shallow water marine study record, and in particular, when the NAO index is positive south Europe climate it is mild and dry; on the contrary, a negative index is associated with the reverse pattern, suggesting a hemispheric-scale atmospheric connection.

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Figure Caption

Fig. 1- (a): Location map of the study area; (b): Bathymetric map of the study area with the location of the study core and the hydrographic grids of Garigliano and Volturno rivers.

Fig. 2 – From the left: Lithological log of core C5, magnetic susceptibility of C5 and SW-104-C5 cores. The dotted black line represents the correlation point between the two study cores.

Fig. 3 – Radionuclides analysis. ^{210}Pb and ^{137}Cs activity-depth profiles in core SW-104-C5 with position of tie points. The grey band indicates the stratigraphic position (cm bsf) of the tephra layer associated to Vesuvius volcanic event 1906 AD (this work).

Fig. 4 – Classification of the studied samples representative of tephra from core C5 according to the TAS (Total alkali/Silica diagram; Le Maitre, 2005).

Fig. 5 - TiO_2 vs. CaO diagram of tephras C5/319 and C5/414 found in core C5. The average compositional field of latitic products from Ischia and Campi Flegrei erupted in the last 4 kyrs are reported for comparison. Data from: (CF) Smith et al., (2011); (Ischia) D'Antonio et al., (2013).

Fig. 6 - a) SiO_2 *vs* CaO variation diagram for C5/403 tephra. The composition of possible proximal counterpart from CF are reported for comparison. Data from: (Astroni 4-5-6) Smith et al., 2011; (Capo Miseno), this study (supplementary materials) b) SiO_2 *vs* CaO variation diagram for tephra C5/437. The composition of possible proximal counterpart from CF are reported for comparison. Data from Smith et al. (2011).

Fig. 7 – From left side. Correlation between the composite core from Salerno Gulf (Lirer et al. 2013, 2014) and the composite core SW-104-C5-C5 (Gulf of Gaeta, this work), based as follows: a) the onset of nuclear activity (^{137}Cs activity); b) the micropaleontological events represented by the acme intervals of *G. truncatulinoides* and of *G. quadrilobatus* (grey bands), and the drop in abundance of *G. truncatulinoides* (black arrow); c) tephra layer associated to Agnano M. Spina volcanic event and d) $\delta^{18}O_{\textit{G. ruber}}$ patterns. In the middle: Age depth profile of composite core SW-104-C5-C5 (Gulf of Gaeta) with position of the tie-points used to calculate the sedimentation rate. The two red dotted lines represent the propagation of errors.

Fig. 8 – Comparison in time domain between $\delta^{18}O_{\textit{G. ruber}}$ signals from Gaeta Gulf (this work), Salerno Gulf (Lirer et al., 2013, 2014), Taranto Gulf (Grauel et al., 2013a) and $\delta^{18}O_{\textit{G. sacculifer}}$ data from Adriatic Sea (Piva et al., 2008).

Fig. 9 - Measured paleomagnetic inclination curve from core C5, compared with an inclination reference curve calculated from the SHA.DIF.14k geomagnetic model (Pavón-Carrasco et al., 2014) for the site location. The blue line is the measured inclination from the core; the green line is the reference curve; the red bars mark the tie-points which constrain the age model.

Fig. 10 –Comparison in time domain of planktonic foraminifera distribution patterns and $\delta^{18}O_{\textit{G.ruber}}$ data (red dotted line represent the raw data and thick red line is a 5-point moving average) with the position of the climatic phases (Eneolithic, Early Bronze Age, Middle Bronze Age-Iron Age, Roman Period, Dark Age, Medieval Climate Anomaly, Little Ice Age, Industrial Period, Modern Warn Period). We have plotted the distribution pattern of the herbivorous-opportunistic (*T. quinqueloba, G. glutinata* and *G. bulloides*) *vs* carnivorous foraminiferal species (*G. ruber* alba variety, *G.*

quadrilobatus, *Orbulina* spp. and *G. siphonifera*) to identify the climatic phases. The light grey and dark grey bands indicate the cold and warm phases, respectively.

Fig. 11 – Pollen diagrams and $\delta^{18}O_{G.ruber}$ data (thick black line represents a 5-point moving average) plotted *vs* time domain with the position of the identified climatic phases (Eneolithic, Early Bronze Age, Middle Bronze Age-Iron Age, Roman Period, Dark Age, Medieval Climate Anomaly, Little Ice Age, Industrial Period, Modern Warm Period). The grey bands represent the main climatic events detected by pollen data.

Fig. 12 –Comparison in time domain between the AP pollen data index, planktonic foraminiferal turnover (carnivorous *vs* herbivorous-opportunistic species), $\delta^{18}O_{G.\ ruber}$ signal (5 point moving average black line and 150 years moving average thick red dotted line) and NAO index (black line by Olsen et al., 2012; blue line by Trouet et al., 2009). The grey bands represent the identified climatic events. The labels 2.8 kyr event (B2) and 1.5 kyr event (B1) correspond to the position of Bond events 1 and 2 (Bond et al., 2001). The label Med. cold corresponds to Medieval cold. Modern WP= Modern Warm Period; Indust. Per. = Industrial Period. The black triangles with the number from 1 to 5 are the position of the identified tephra layers: 1- Vesuvius (1906 AD), 2- Vateliero-Ischia (2.4-2.6 ka BP), 3- Capo Miseno (3.9 ka BP), 4- Astroni3 (4.1-4.3 ka BP), 5- Agnano M. Spina (4.42 ka BP).

Table Caption

Tab. 1 – Table with ages and nomenclature of the climatic events documented in marine Mediterranean records for the last five millennia compared with the archaeological periods reported by Roberts et al. (2011). The acronym LBA corresponds to Late Bronze Age.

Tab. 2 - Ecological requirements for planktonic foraminifera from literature data.

Tab. 3 – Summary of tephra layers analysed in the last five ka record of composite core SW-104-C5-C5 (Gulf of Gaeta). a: age datum based on archaeological remains (de Vita et al., 2010); b: age datum from paleomagnetic measurements (Vezzoli et al., 2009); c: $^{40}Ar/^{39}Ar$ age (Di Renzo et al., 2011); d: ^{14}C calibrated age (Sacchi et al., 2014); e: modelled best age (Smith et al., 2011). Total Fe expressed as FeO. Abbreviation: av. is the number of analyses considered for the average (bold); s.d. is standard deviation (italics).

Tab. 4 - Averages and standard deviations (s) of the major-element composition of composite core SW-104-C5-C5 tephras and their proximal equivalents discussed in the text. All analyses recalculated water-free to 100%. Source of data used for comparison: (1906) this study; (Vateliero) D'Antonio et al., (2013); (Capo Miseno) this study; (Astroni3, Astroni6 and Agnano Monte Spina -AMS) Smith et al., (2011).

Tab. 5 – List of the tie-points used for the construction of the age model of composite core SW-104-C5-C5 (Gulf of Gaeta).

Supplementary materials:

Supplementary electronic material 1: Full analytical data of the individual SEM-EDS measurements for the glass fragments extracted from the single tephras of composite core SW-104-C5-C5 (Gulf of Gaeta) and for reference samples of Vesuvius 1906 and Capo Miseno eruption from Campi Flegrei analysed for this study.