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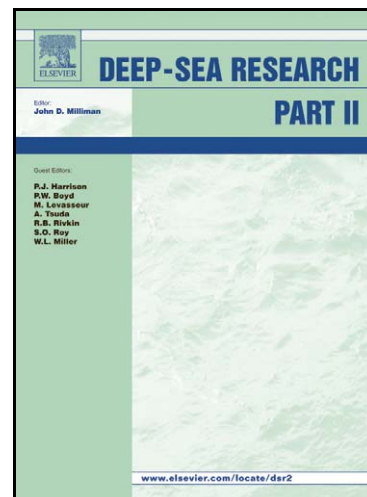
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Integrated stratigraphic reconstruction for the last 80 kyr in a deep sector of the Sardinia Channel (Western Mediterranean).

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### Abstract

*A quantitative analysis of planktonic foraminifera, coupled with petrophysical and paleomagnetic measurements and <sup>14</sup>C AMS calibrations, was carried out on a deep core recovered in the Sardinia Channel (Western Mediterranean Sea), during the CIESM Sub2 survey, providing an integrated stratigraphic time-framework over the last 80 kyr.*

*Significant changes in the quantitative distribution of planktonic foraminifera allowed the identification of several eco-bioevents useful to accurately mark the boundaries of the eco-biozones widely recognised in the Western Mediterranean records and used for large scale correlations.*

*Namely, 10 eco-biozones were identified based on the relative abundance of selected climate sensitive planktonic foraminiferal species. Sixteen codified eco-bioevents were correlated with the Alboran Sea planktonic foraminiferal data and four climatic global events (Sapropel S1, Younger Dryas, Greenland Isotope Interstadial 1, Greenland Isotope Stadial 2, Heinrich event H1-H6) were recognized.*

*The eco-bioevents together with the <sup>14</sup>C AMS calibrations concurred to define an accurate age model, spanning between 2 and 83 kyr cal. BP. The reliability of the age model was confirmed by comparing the colour reflectance (550 nm%) data of the studied record with the astronomically*

*tuned one of the Ionian sea (ODP-Site 964). A mean sedimentation rate of about 7 cm/kyr was evaluated including three turbidite event beds that were chronologically constrained within the relative low stand and lowering sea level phases of the MIS 4 and MIS 3. The deep sea sedimentary record includes a distinct tephra occurring at the base of the core which dates 79 ka. The paleomagnetic data provide a well-defined record of the characteristic remanent magnetization that may be used to reconstruct the geomagnetic paleosecular variation for the Mediterranean back to 83 kyr cal. BP.*

Key words: integrated stratigraphy, late Neogene marine record, eco-bio-events, reflectance 550 nm %, Sardinia Channel, Western Mediterranean.

## **1. Introduction**

Since climate excursions recorded in Northern Hemisphere in the Greenland GISP and GRIP ice cores (GRIP members, 1993) over the last 100 kyr had more or less synchronous effects in the Mediterranean area, many researches have focused on Mediterranean marine cores, with the aim to detect their intensity and the impact on the marine environment. During the last glacial period the Mediterranean region experienced rapid modifications in hydrographic conditions in response to fast climatic excursions, known as Heinrich events (HE) and Dansgaard-Oeschger (D-O) Stadials (cold) and Interstadials (warm) (Heinrich, 1988; Dansgaard et al., 1993). In particular, Rohling et al. (1998) and Cacho et al. (1999, 2000) have proved that the millennial scale D-O and HE directly control the winds and precipitation system on the Northern Mediterranean basin. Even during the Holocene the principal climatic events and oscillations of the Northern Hemisphere were clearly traceable in different sectors of the Mediterranean Basin sedimentary records (Cacho et al., 1999, 2000, 2001; Rohling et al., 2002; Sprovieri et al., 2003, 2006; Perez-Folgado et al., 2003, 2004; Geraga et al., 2005).

A very detailed outline of the paleoenvironmental changes and their control on marine communities, calibrated by several independent proxies (tephra, sapropel,  $^{14}\text{C}$  geochronology), is available for the Mediterranean area (Buccheri et al., 2002; Ducassou et al., 2007; Emeis et al., 2003; Narcisi and Vezzoli, 1999; Lourens, 2004; Principato et al., 2003; Rohling et al., 2003; Sangiorgi et al., 2006; Sprovieri et al., 2003; 2006, and reference therein). Several codified eco-bioevents, if clearly detected in marine records, can be used as tie points to chronologically constrain the late Pleistocene-Holocene Mediterranean marine sequences. Nevertheless, even if many reference records are available from deep-sea sites, most of them span a short time interval and lack a high resolution detail of the paleo-environmental and paleo-ecological changes before 40 kyr. Recently, Perez-Folgado et al. (2003, 2004) carried out a high-resolution study of the tuned ODP-Site 977, located in the western part of the Alboran Sea, and identified several planktonic foraminiferal eco-bioevents occurred during the marine isotope stages (MIS) 1 to 5. These eco-bioevents represent the best tool to correlate deep marine records from different Mediterranean sites.

Many recent studies emphasize the challenge when studying deep sea records to establish a reliable chronology even for the deposition of turbidites (Walker, 1992; Beaudouin et al., 2004; Ducassou et al., 2007) and underline the utility to support conventional dating methodologies with different constraints. It is widely accepted that one of the main factor controlling and enhancing turbidite deposition along deep sea fan is the fall and lowstand of sea-level, whereas sea-level rise and highstand phases reduce terrigenous supply to deep sea systems (Walker, 1992; Richards et al., 1998; Normark et al., 1998).

The CIESM core C08 is located at the mouth of the Bizerte Canyon in the Sentinelle Valley in a key position of paleoceanographic and geological significance (Fig. 1). The Sardinia Channel connects the Alboran to the Tyrrhenian Basin and offers a stratigraphic record with the potential to link the eco-stratigraphic and paleoceanographic observations between the Western, the Central and Eastern Mediterranean late Pleistocene Holocene marine records (Perez-Folgado et al., 2004;

Sbaffi et al., 2004; Geraga et al., 2005; Asioli et al., 2001; Ariztegui et al., 2000). In fact, a portion of the Modified Atlantic Water (MAW) coming from the Strait of Gibraltar (Bryden and Kinder, 1991), diverges from the part that enters the Eastern Mediterranean and flows through the Sardinia Channel into the Tyrrhenian Sea along the northern Sicilian coast (Milot, 1987), forming a secondary circulation gyre. The circulation system in this sector of the Tyrrhenian Sea is counter-clockwise with the Levantine Intermediate Water (LIW) inflows lapping on the northern Sicilian coast and the outflow occurring along the eastern Sardinia coast (Pinardi and Masetti, 2000 and references therein).

The core site is even in a strategic position to check the efficiency of a submarine canyon in driving density flow to the deep sea environment (see next paragraph), even if not directly connected to any emerged sector nor to continental shelf areas (Reading and Richards, 1994). Thus the possibility that such a type of canyon would form a fan can be evaluated, even verifying the significance and the timing of the turbidite deposition.

The aim of this study is to provide a record of integrated stratigraphic data spanning back to about 80 kyr, relatively to a deep basin area, based on eco-biozones,  $^{14}\text{C}$ -dated ages, Event Stratigraphy, lithostratigraphy, petrophysical properties and paleomagnetic measurements. Furthermore, the reliability of the reflectance parameter 550 nm % is evaluated as an independent correlation tool for tuning marine records.

## **2. Geological setting**

The core C08 was collected in the Sentinelle Valley of the Sardinia Channel, 30 nautical miles from the mouth of the Bizerte Canyon, halfway from Sicily, Sardinia and Tunisia (Fig.1), during the cruise CIESM Sub2 onboard the R/V Urania in December 2005 ( $38^{\circ}38.5364'\text{N}$ ,  $10^{\circ}21.5576'\text{E}$  - 2370 m below sea level). In this area a 400 km long submerged sector of the Apennine-Maghrebian branch of the Alpine orogen separates the Tyrrhenian (Plio-Pleistocene in age) and the Algero Provençal (Miocene in age) oceanic basins. This sector of the chain was not completely fragmented during the opening of the basins (Masclé et al., 2004). Due to the relatively minor post-orogenic

extension and the good preservation of morpho-structural features, the Sardinia Channel is an important area for the reconstruction of the geodynamic evolution of the Western Mediterranean sector and was recently investigated through submersible surveys (Masclé et al., 2001a; Masclé et al., 2004). The triangular shaped valley is bounded by a NE-SW oriented Median Ridge on its north-western side and by the South Cornaglia slope on the south-eastern one. The south-western sector of the Sentinelle Valley receives the sedimentary contribution of the Bizerte Canyon, which engraves the Tunisian Plateau and the south margin of the Sentinelle Bank (Masclé et al., 2001b). The canyon head appears disconnected from the Tunisian shelf margin and extends over an area of about 1000 km<sup>2</sup> at an average depth of about 500 m (Fig. 1). It represents a particular type of canyons since it is not fed by an emerged areas or by a fluvio-deltaic systems (Reading and Richards, 1994; Prins et al., 2000; Kenyon et al., 2002;), but it drains a wide submarine plateau.

### **3. Material and methods**

The gravity corer entered the uppermost portion of a sedimentary sequence made up with lateral continuous parallel and thin reflectors (Sartori et al., 2001), possibly related to hemipelagic and turbiditic deposition. It recovered about 5.40 m of hemipelagic mud interlayered with three fine to medium sand turbidite layers of increasing thickness towards the top of the core (Fig.2).

#### *3.1 Physical property measurements*

The physical properties of the core were measured at 1 cm step in a fully automated GEOTEK Multi-Sensor Core Logger (MSCL), in the petrophysical laboratories of IAMC in Naples (Italy). The MSCL system includes a Bartington MS2E Point sensor, to measure the low-field magnetic susceptibility (MS) with a spatial resolution of 0.4 cm and a Minolta Spectrophotometer CM 2002 which records at 0.8 cm step, the percentage of reflected energy (RSC) at 31 wavelengths in 10-nm steps, over the visible spectrum (from 400 to 700 nm). MS and RSC measurements were taken on the archive half, ~1 hr after the core had been split. The split core was covered with cling film to protect the glass cover of the Minolta aperture while measuring. Both measurements, log plotted,

were visually compared in order to detect similar trends and tentatively group the data according to their physical properties and corresponding stratigraphy (Wolf-Welling et al., 2001).

### *3.2 Paleomagnetic analysis*

For the paleomagnetic study, 1-m long u-channel specimens were sampled from archive halves of the CIESM 08 core. Only the turbiditic sandy interval at 182-208 cm depth was not suitable for u-channel sampling due to its coarse-grained and loose lithology. The u-channels, sampled from the archive sections of the core, were measured in the magnetically shielded room of the paleomagnetic laboratory at the Istituto Nazionale di Geofisica e Vulcanologia in Rome, using an automated pass through 2-G Enterprises DC SQUID cryogenic magnetometer system. For each u-channel the natural remanent magnetization (NRM) was measured at 1 cm step. It is emphasized however that, due to the intrinsic response functions of the SQUID sensors, remanence measurements may be considered truly independent only every ca. 5 cm. The NRM was progressively demagnetized by the alternating field (AF) in nine steps up to a maximum field peak of 100 mT, by translating the u-channel through a set of three perpendicular AF coils at a speed of 10 cm/s. The remanence was measured at the same 1-cm spacing after each demagnetization step. Demagnetization data from u-channels were analyzed on orthogonal vector projections and the paleomagnetic data obtained generally provided straightforward demagnetization diagrams, indicating that the sediments carry an almost single-component NRM. In fact, during the AF demagnetization a complete removal of any drilling overprint and/or laboratory induced remanences was achieved at low AF peaks (10 mT), and the characteristic remanent magnetization (ChRMs) of the sediments was clearly identified and determined by principal component analysis (Kirschvink, 1980). For each u-channel the uppermost and lowermost few centimeters were disregarded for the paleomagnetic analysis, to avoid any deflection to the remanence direction due to disturbances that may have been introduced during sampling. As stated before, no paleomagnetic data were collected for the coarse grained level at 181-207 cm depth, and the NRM demagnetization diagrams did not allow a clear identification of a ChRM at 137-151 cm and 309-317 cm depth intervals.

### 3.3 Planktonic assemblage analysis

The analysis of planktonic foraminifera was conducted on 216 samples. Sampling spacing was 2 cm from the top of the core down to the base. Each wet sample of about 20 g was dried at 50° C and washed over sieves with mesh-width size of 63 microns. Quantitative planktonic analyses were carried out on the fraction >125µm. The adopted taxonomic units were those reported by Jorissen et al. (1993), Capotondi et al. (1999). According to Jorissen et al. (1993) and Capotondi et al. (1999), we introduced some supraspecific categories (which remain unchanged even under bad preservation conditions), reducing the number of species actually occurring in the planktonic foraminiferal assemblages.

### 3.4 <sup>14</sup>C Radiocarbon calibrations

About 10 mg of *Globigerina bulloides* and *Globorotalia inflata* were picked from three samples in the first 2 m beneath the sea floor (at the core top, at 0.44 and 1.43 mbsf) and were AMS radiocarbon dated at the Centre for Isotopic Research on Cultural and Environmental Heritage (CIRCE) radiocarbon laboratory, Caserta, (Italy). The CIRCE Accelerator Mass Spectrometry (AMS) system, based on the 3MV 9SDH-2 Pelletron accelerator, provides a mean overall precision of 0.63% (Passariello et. al, 2007). All radiocarbon dates were calibrated using the CALIB 5.0 Program and a mean  $\Delta R$  value of 48 +/- 21 yr (calculated among six of the Tyrrhenian Sea, from the marine data base of Stuiver and Reimer, 1993). The AMS RC age at 1.430 mbsf (21590 +/- 120 a) close to the lower limit of the calibration curves was not used for the age model.

## 4. Results

### 4.1 Lithostratigraphy and petrophysical properties

The sediment consists for the 95 % of hemipelagic mud, ranging in colour from reddish and ochre to light, olive and dark grey, and for the remaining 5 % of turbiditic sand layers (Fig.2).

The uppermost 0.50 m consists of a reddish mud pervasively oxidized, then about 0.11 m of alternating dark grey and ochre laminated mud, and further below, 0.05 m of fine laminated ochre



mud occurs. From about 0.66 m down to 2.40 m olive grey mud occurs with rare dark patches probably due to bioturbation, while from 2.40 m to the bottom of the core an alternation of several mud intervals occur, whose thickness range from centimeters to decimeters, showing different shades from light grey to dark grey. Three turbidite sand beds are interlayered in the mud sequence and occur from 1.82 to 2.08 m (T1), from 3.10 to 3.17 m (T2), and from 4.33 to 4.34 m (T3). They constitute 5% of the whole stratigraphic thickness of the core and are all characterized by low values of magnetic susceptibility. The turbidite layer T1 is marked by a sharp erosive contact and consists of a thin layer of oxidized sand, and then of a massive fine to medium sand with high percentage of shell fragments; the upper boundary is sharp and the grain size populations comprised between fine sand and clay, that usually pertain to “b, c, d, e” divisions of the classical Bouma sequence (Bouma, 1962), are missing, as evidenced by the abrupt decrease of MS and Grape density values, which are a function of grain size and lithology. A sharp contact marks the onset of the turbidite layer T2 which contains a thin layer of well sorted dark fine sand passing to a massive layer of well sorted bioclastic fine sand; also in this case the upper boundary is sharp and the passage from massive and structureless sand to the hemipelagic mud is abrupt. Turbidite layer T3 has well defined sharp boundaries and starts at the base with a thin layer of dark sand passing to light grey bioclastic sand. The thickness of the turbidites increases upward, but no sand layers occur in the uppermost 0.18 m of the core. A discrete fine ash layer, dark stained and characterized by diffuse boundaries occurs at 5.08 m and is marked by a MS peak value of  $37 \text{ SI} \times 10^{-5}$ . This tephra layer consists mainly of white pumices, dark green to yellowish glass shards, and few dark scorias, while lithic elements are scarce (Stella Tamburrino, pers. comm. 2007).

The lithologic features of the cored sediment are clearly detected by the petrophysical properties measured with a 1 cm step: the volume magnetic susceptibility (MS) traces even thin tephra layers (Iorio et al., 2004), the GRAPE density is a compelling proxy for lithology and grain size (Wolf-Welling et al., 2001) and the reflectance at 550 nm % wavelength provides a quantitative evaluation of variations in sediment colour and could serve as a powerful correlation tool between

long distance core sites (Peterson et al., 2000). For instance, the meter composite depth of ODP-Sites records (ODP- Leg 154, Curry et al., 1995), is composed on the variation in natural gamma rays, reflectance and magnetic susceptibility parameters, to identify sedimentary patterns and to calibrate marine records to the standard variation of astronomical parameters (Lourens et al., 1998; Palike et al., 2006).

The three turbidites are well evidenced by low MS values (T1 and T2 measuring less than  $5 \text{ SI} \times 10^{-5}$ , T3 of  $12 \text{ SI} \times 10^{-5}$ ) and by GRAPE density value (close to  $2.1 \text{ gr/cc}$ ), but are not highlighted by RSC. The RSC parameter traces well the changes in shade colour throughout the core, and in particular it points out those occurring at the core top (due to the pervasive seabed sediment oxidation) and in the deepest 1.4 m (Fig. 2). Along this latter interval the wide oscillations in the RSC parameter, ranging between 15% and 30%, correspond to analogous oscillations in MS values.

#### *4.2 Paleomagnetism*

The main paleomagnetic results are plotted in Fig 2, which shows the stratigraphic trend of the NRM intensity, the ChRM inclination and declination and the maximum angular deviation (MAD). The NRM is weak and unstable in the topmost 0.15 m of the core and will not be discussed in the following. The data from below 0.15 m show that NRM intensity is relatively high and characterized by a limited variation, with values oscillating between 0.6 and  $2.3 (10^{-2} \text{ A/m})$ . The ChRM inclination oscillates from  $30^\circ$  to  $60^\circ$  with an average value of  $48^\circ$  that is about  $10^\circ$  lower than the expected value ( $58^\circ$ ) at the latitude of the coring site. Such a kind of shallow bias, or inclination error, has been found in laboratory redeposition experiments and in some modern natural sediments. The sedimentary inclination shallowing is a well known process linked to the deposition of magnetic particles and subsequent sediment compaction (see Tauxe and Kent, 1984; 2004; Arason and Levi, 1990; Deamer and Kodama 1990; Tan et al., 2003; Tan and Kodoma, 2003). The piston core was not azimuthally oriented, but a fiducial mark ensured the reciprocal orientation of each section from which u-channels were collected. The ChRM declination of each u-channel was

rotated to bring the core average at 0°. Under this transformation the ChRM declination oscillates between -30° and + 30°, which is in the range of the expected geomagnetic secular variation. The principal component analysis of the demagnetization data shows a MAD always well below 3.3° along the core, which indicates a well defined paleomagnetic direction, except a positive peak, up to a maximum of 7.5°, measured in the 0.47 – 0.48 m interval. This interval corresponds to the largest oscillations in the ChRM inclination and declination values, suggesting that some uncompensated lithological factor affects the paleomagnetic signal in that stratigraphic interval.

#### 4.3 Quantitative distribution of planktonic foraminifera

A number of 13 species or groups of species were distinguished: *G. bulloides* (including extremely rare specimens of *G. falconensis*); *G. quadrilobatus* (including *G. trilobus* and very subordinate *G. sacculifer*); *G. ruber* (including white and pink varieties, this latter extremely rare), *G. elongatus* (very rare) and *G. gomitolus* (very rare); *Globorotalia truncatulinoides* sl. right (very rare) and left coiling; *G. tenellus* (rare); *Neogloboquadrina pachyderma* right and left coiling (extremely rare); *Globigerinita glutinata*; *Orbulina universa*; *Turborotalita quinqueloba*; *Globigerinatella siphoniphera* (rare) including *G. calida* (very rare); *Globoturborotalita rubescens* (rare).

The planktonic foraminifera, characterized by modern assemblages, are abundant and well-preserved and the percentages of foraminiferal fragments are very low and do not alter the composition of the planktonic assemblage.

The long-term trend in planktonic foraminifera reveals that the faunal composition of the studied interval does not show drastic changes in the abundance patterns (Fig.3). In particular, among the taxa that have a continuous distribution patterns, *G. bulloides*, *G. ruber*, *G. inflata* left coiled, *G. scitula* right coiled, *G. glutinata*, *N. pachyderma* right coiled and *T. quinqueloba* show long-term oscillation (trend) superimposed on short-term fluctuations possibly related to high-frequency climatic oscillations (Fig.3).

Among the planktonic species having discontinuous distribution, *G. quadrilobatus*, *O. universa*, *G. truncatulinoides* left coiled, *G. tenellus*, *G. rubescens* only occasionally reach significant percentages (Fig.3).

## 5. Discussion

### 5.1 Planktonic foraminiferal eco-biozonation

Significant changes in quantitative distribution of the planktonic foraminifera species allowed several authors (Asioli et al., 1999, 2001; Capotondi et al., 1999; Casford et al., 2002; Principato et al., 2003; Sprovieri et al., 2003; Ducassou et al., 2007) to define eco-biozones useful for fine subdivisions of the stratigraphic record. The eco-biozone boundaries are characterized by temporary appearance or disappearance and/or evident abundance peaks of different taxa. In the present work, we refer the stratigraphic record to eco-biostratigraphic classification of Sprovieri et al. (2003), slightly modified. At present, the eco-biostratigraphic classification of Sprovieri et al. (2003) proposes 9 eco-biozones over the last 23 kyr. In this work, we propose to mark the base of eco-biozone 9 with the strong increase of *G. inflata*, occurring in the Mediterranean area at about 30 ka (Ducassou et al., 2007; Geraga et al., 2005), and to extend the eco-biozone 10 back to ~80 kyr. Actually, using the quantitative distribution pattern of the most abundant planktonic foraminifera species counted throughout the C08 core, we identified 10 evident eco-biozones from top to bottom (Fig. 3).

The uppermost part of the studied record attributed to the eco-biozone 1, which bases at 0.08 m, is characterized by a decrease in abundance of *G. quadrilobatus* and the end of *G. truncatulinoides* right coiled. This attribution is also supported by a  $^{14}\text{C}$  – AMS calibration, which dates 2.18 ka cal. BP (Fig. 3; Tab. 1).

The eco-biozone 2 is defined by the concomitant abundance of *G. quadrilobatus* and *G. truncatulinoides* left and right coiled and by low abundance values of *N. pachyderma* right coiled in

the lower part. Besides, the strong increase of *G. truncatulinoides* left coiled marks the base of the eco-biozone (Fig. 3).

The short interval represented by eco-biozone 3, whose base is at 0.31 m, is marked by the end of the micropaleontological signature of sapropel S1 and is characterized by low abundance values of *T. quinqueloba*, *G. quadrilobatus*, and *G. truncatulinoides* left coiled.

The eco-biozone 4 corresponds to the time interval of sapropel S1 deposition, although no lithological evidence was found, but for colour shades (Fig. 2). *G. ruber* oscillations allowed a reliable identification of the faunal signature of the climatic events associated to the deposition of sapropel S1 (Fig.3). In particular, two distinct peaks in *G. ruber* mark the two short-term warm oscillations (S1a and S1b, Sprovieri et al., 2003), separated by a cold phase between them (Fig.3). The event is supported by a  $^{14}\text{C}$ -AMS datum at 0.44 mbsf (within the cold phase) with an age of 8.79 ka cal. BP (Tab.1) as well as an increase in abundance of *G. tenellus* and *G. quadrilobatus* (Sprovieri et al., 2003).

The eco-biozone 5 is defined by the concomitant occurrence of *G. ruber* and *G. inflata*, by a distinct peak of *G. truncatulinoides* left coiled and absence of *N. pachyderma* right coiled and very low value of *T. quinqueloba*. The eco-biozone 6 is marked by the absence of *G. ruber* and *G. inflata* left coiled, by a distinct peak of *T. quinqueloba* and of *N. pachyderma* right coiled. This eco-biozone corresponds to the Younger Dryas event, according to Sprovieri et al., (2003). The eco-biozone 7 is defined by the increase in abundance of *G. ruber* and *G. inflata*, by the absence of *T. quinqueloba* and by a distinct peak of *G. quadrilobatus* and corresponds to the warmer (interstadial) GI-1. According to Sprovieri et al., (2003) in the eco-biozone 8 the persistent high abundance of cold species permit correlation of this interval with the GRIP GS-2 period. In particular this eco-biozone is dominated by *T. quinqueloba*, *N. pachyderma* right coiled, *G. scitula* and by absence of *G. inflata* left coiled and rare *G. ruber*. The base of this eco-biozone approximates to the base of MIS 2 (Perez-Folgado et al., 2003) (Fig. 3).

The eco-biozone 9 is characterized by the concomitant absence of *G. inflata* left coiled and *G. ruber*, by low abundance value of *N. pachyderma* right coiled and high abundance of *T. quinqueloba* and *G. scitula* right coiled. The eco-biozone 10 is clearly marked by the progressive downward increase in abundance of *G. inflata* left coiled and *G. ruber* and the progressive decrease of *G. scitula* right coiled, *T. quinqueloba*, *N. pachyderma* right coiled and *N. dutertrei* right coiled (Fig.3). No distinctive or drastic events in the planktonic faunal patterns are visible towards the base of the studied record but only short-term oscillation in *G. bulloides* (B3-B11 eco-bioevents; the adopted sampling resolution does not allow the recognition of *G. bulloides* B8 eco-bioevent), *G. inflata* left coiled (I3-I5 eco-bioevents), *T. quinqueloba* (Q4-Q9 eco-bioevents) and *N. pachyderma* right coiled (P5-P7 eco-bioevents), clearly associated to the millennial climate oscillations occurring in the last 80 kyr (Perez-Folgado et al., 2004). According to Perez-Folgado et al. (2004) close to the *G. bulloides* eco-bioevents B7 and B10 are placed at the MIS 4/MIS 3 and MIS4/MIS 5a transition, respectively. Finally, according to Perez-Folgado et al. (2004) the lowermost part of the studied record lies within the eco-bioevent B11 and within the uppermost part of MIS 5b (Fig. 3).

## 5.2 Age model

The identified foraminiferal marker events, regarded as to reflect major changes in oceanographic conditions and already recognised for the Central and Western Mediterranean (Jorissen et al., 1993; Capotondi et al., 1999; Asioli et al., 2001; Saffi et al., 2004; Sprovieri et al., 2003; Perez-Folgado et al., 2004; Geraga et al., 2005) were used, in combination with two  $^{14}\text{C}$  – AMS data (Tab.1), to constrain the age of core CIESM-C08 and strengthen the correlations between the Mediterranean sites. In particular, we used the age model proposed by Perez-Folgado et al. (2003, 2004) for the Alboran Sea, to recognize the top and base of the eco-bioevents recorded in the core C08, the age model proposed by Asioli et al. (2001) for the Adriatic Sea, to identify the Younger Dryas and the base of the Greenland isotope interstadial 1 (GI-1) and the age model of Andersen et al. (2006) for the NGRIP record to distinguish the Greenland isotope stadial 2 (GS-2) (Tab. 1).

During the recovery of the core, a small amount of material was supposedly lost at sea. This occurrence is clearly supported by the  $^{14}\text{C}$  – AMS dating of 2.18 ka cal. BP at the top of the core and by comparing the planktonic foraminiferal contents of the studied record with those reported by Sprovieri et al. (2003) for ODP-Site 963 (Sicily Channel). This correlation confirms that the sedimentation of the last 2000 years is not preserved in CIESM C08 core.

The turbidite layers T1, T2 and T3 have been taken into account to construct the age model curve and to estimate the sedimentation rate. A second-order polynomial is needed to describe the age-depth relationship for the studied record, indicating an average sedimentation rate of  $\sim 7$  cm/kyr from the base to the top and four main excursions (Fig.4).

In order to confirm the reliability of the proposed age model, a three step validation process have been performed. First, the visual comparison, in time domain, between *G. bulloides* distribution pattern of Perez-Folgado et al. (2004) for the Alboran Sea and the patterns in core CIESM C08 (Fig. 4), which confirmed the tuning accuracy.

The second step consisted of the comparison of the colour reflectance record at 550 nm (%) of the core CIESM C08, with the record of the ODP-Site 964 (Fig. 5), drilled in the Ionian Sea at 3650 mbsl and astronomically calibrated (Lourens, 2004). Using the proposed age-depth profile (Fig. 4), the colour reflectance record of the studied core was plotted against time by a cubic spline interpolation method. This comparison (Fig.5) shows that the large-scale reflectance fluctuations in core CIESM C08 not only have similar pattern to those reported in the Ionian sea record but also encompass absolute values in the same range ( $\pm 10$  nm). Lourens (2004) used the reflectance parameter to detect sapropel and tephra layers in the Ionian Sea (Mediterranean area), which in turn, are tuned to the insolation curve of Laskar et al. (2004). According to Hilgen et al. (2003 and references therein), the Neogene marine successions (on land and in ODP-sites) in the Mediterranean area are often characterized by the cyclic recurrence of sapropels layers and sedimentary patterns, clearly forced by variation in astronomical parameters (Hilgen, 1991a, 1991b; Hilgen et al., 1999, 2003; Hilgen and Krijgsman, 1999; Lourens et al., 1996a, 1996b, 1998

and references therein). Therefore it is not surprising that the cyclic patterns in the colour reflectance records relative to two adjacent deep environmental Mediterranean areas are similar. Remarkable is also the strong similarity between the colour reflectance signature of the sapropel S1 equivalent, recorded in the studied core, with the sapropel S1 colour reflectance signature in the Ionian basin. Nevertheless, some of the minimum peaks of the ODP-Site 964 reflectance record, related to the occurrence of tephra layers, are not present in the studied core. This is probably due to the different spatial distribution of the tephra layers in the Mediterranean. On the whole, the good visual correlation obtained between the two records supports the validity of the age model based on the identified eco-bioevents and  $^{14}\text{C}$  AMS calibrations.

The third control step consisted in the comparison of the *G. ruber* distribution pattern chronology with the  $\delta^{18}\text{O}$  NGRIP ice core record. The ecological features of *G. ruber* associated to warm and oligotrophic surface waters has been established in several oceanographic settings (Hemleben et al., 1989; Pujol and Vergnaud-Grazzini, 1995; Watkins et al., 1996; Zaric et al., 2005). Moreover, several authors (Sprovieri, 1991; Sanvoisin et al., 1993; Sprovieri et al., 2003; 2006) confirmed the utility of relative abundance fluctuations of *G. ruber* as recorders of climatic variability and as for age model proxies. The *G. ruber* and  $\delta^{18}\text{O}$  NGRIP ice core records exhibit a remarkable agreement, with the identification of the Heinrich events (H1 to H6) and of the Younger Dryas (YD) in the studied record which further support the reliability of our tuning (Fig. 6).

The observed strong link between  $\delta^{18}\text{O}$  of Greenland ice cores, a proxy for Greenland air temperature, and the thermal character of surface water masses of Mediterranean area has been previously established in several papers and usually explained by teleconnective phenomena (Rohling et al., 1998; Cacho et al., 1999, 2000; Saffi et al., 2004; Combourieu Nebout et al., 2002; Moreno et al., 2002; Rohling et al., 2002).

The proposed age model was also adopted to plot paleomagnetic data against time (Fig. 7). We note that the time interval spanned by the core encompasses three known geomagnetic excursions (the Mono Lake, dated at  $33\pm 1$  ka; the Laschamp, dated at  $41\pm 1$  ka and the Norwegian-Greenland Sea,



dated at  $61 \pm 2$  ka; see Lund et al., 2006). Nevertheless in the paleomagnetic record of CIESM C08 core there is no clear marked oscillation which can be ascribed to any of such excursions. The lack of the paleomagnetic evidence for geomagnetic excursions in this record may be due to a relatively low sedimentation rate (cfr. 5.1, Fig. 4), that may have resulted in a prolonged time interval for the lock-in of the natural remanence of the sediments (Roberts and Winklhofer, 2004; Sagnotti et al., 2005), beside the response functions of the SQUID sensors in the magnetometer (i.e. Oda and Shibuya, 1996; Guyodo et al., 2002). Despite the suggested smoothing of the paleomagnetic signal, the well defined paleomagnetic data of core CIESM-08 provide an original source of information to improve the confidence of geomagnetic paleosecular variation reconstructions in the Mediterranean for the last 83 kyr.

### *5.3 Ages and provenance of turbidite events*

The size population of grains, the grain fabric, the high content in bioclasts (gastropod, bivalve and echinoderm debris) and the features of the surfaces bounding T1, T2 and T3 turbidites lead us to infer a distant source of transported material, since it seems to be remobilized from high productivity areas. Both the slope of the median Ridge, and the southern Cornaglia slope can be ruled out as possible source areas for this bioclastic sand rich density currents, since their top is located respectively at about 1300 and 1000 mbsl. Namely, Sartucya 6 diving survey (Masclé et al., 2001a) showed that the base of the southern slope of the median Ridge (investigate from 2270 to 1940 mbsl) is draped with mud, shaped by current bed-forms, while the upper slope (investigated from 1990 to 1640 mbsl) is characterized by conglomerates and sandstone layers outcropping from the mud, then volcanic rocks. The lithology pertaining to the southern Cornaglia slope has been described during the Sarcya 2 submersible diving (Masclé et al., 2001a), which evidenced the occurrence of pelagic mud along the plain (surveyed from 2500 to 2250 mbsl) and cemented coarse material in correspondence of the steep slopes (from 2250 to 2060 mbsl). Although the occurrence of several canyons along the Southern Cornaglia slope was evidenced by the bathymetric surveys (Brocard, 2001), nevertheless they enter the Sentinelle Valley seaward to the core site and thus they

possibly fed a deepest sector of the basin. Thus, we have to infer that the Bizerte Canyon may have acted as the main conduit to transport the bioclastic sand from the high productivity areas of the Tunisian plateau (Fig. 1); then the flows had to cover more than 50 km before settling. As highlighted by many authors over the years (Walker, 1967; Normark et al., 1984; Reading and Richards, 1994) gentle gradient slopes or pre-existing slope conduits can drive very efficient density currents (Damuth and Flood, 1984; Richards et al., 1998) able to cover long distances in relatively “instantaneous” time and to “segregate the original grain populations into distinct and relatively well-sorted facies types with distances” (Mutti et al., 1999). The beds occurring in C08 core and in particular T1 and T2, could correspond to the facies tract F7 in Mutti et al. (1999), composed predominantly of medium to fine grained well sorted sand. In fact in this model, which associates the horizontal grain size partition of the deposit with the different degrees of flow efficiency, the F7 facies tract consists of medium to fine sand overlying a mm thick traction carpet that accounts for the development of an erosion surface at the base. This model may explain why T1 and T2 turbidites lack parallel and ripple lamination and pelitic divisions.

The turbidite layers in CIESM core C08 are confined in the lower 4 m of the core and their thickness increases upward peaking at T1 (fig.2) and pointing to a general regressive trend. Through the age model scheme of CIESM C08 core here proposed, it is possible to date T layers respectively at 28 ka cal. BP (T1), 48 ka cal. BP (T2) and 63 ka cal. BP (T3) (fig.5), thus during the MIS 4 and MIS 3. The emplacement of T1 event bed caused the removal of an undefined thickness of hemipelagic mud corresponding to a time span of about 4 kyr (Fig. 4).

Plotting their inferred age on the sea level variation curve relative to the last 100 kyr (Waelbroeck et al., 2002; Imbrie et al., 1984), a strong correlation of T2 and T3 turbidite event beds is evidenced with two relative sea level low stand phases, and of T1 with the falling stage of sea level that led to the maximum sea level low stand at about 20 ka BP (fig.8). This observation seems to agree with the most accepted stratigraphic models of deep sea deposition (Normark et al., 1998). Thus, the part of the basin fan intercepted by the CIESM core C08 was active and fed with bioclastic sand

deposition during the relative sea level minimum and increased its transport efficiency following the sea level lowering. Nevertheless the CIESM core C08 does not record any further event of sand deposition during the maximum sea level low stand, relative to the MIS 2. A rapid starvation in detritus supply occurred in this area starting from 28 kyr and this sector of the fan fossilized below a drape of hemipelagic mud. It is reasonable to suppose that during this phase, this sector of the fan acted mainly as a bypass area. In this hypothesis any possible sand flow would have been deposited basinward moving possibly through the channel showed in fig.1b.

## 6. Conclusions

The multidisciplinary study of core C08, recovered from the deep sector of the Sardinia Channel, based on planktonic foraminiferal assemblages, petrophysical (MS, Grape density, 550 nm % reflectance) and paleomagnetic (NRM Intensity, ChRM Inclination, ChRM Declination) data, provides an integrated stratigraphic reference record for the Western Mediterranean Sea that spans back for about 83 kyr. The most important eco-bioevents widely used for large scale correlation in the Western Mediterranean area were recognized providing a detailed correlation with the eco-stratigraphic reconstruction proposed by Perez-Folgado et al. (2003, 2004) for the Alboran Sea.

In particular, the relative abundance of selected climate sensitive planktonic foraminiferal species have led us to identify several eco-bioevents in *G. bulloides* (B3-B11 eco-bioevents), *G. inflata* left coiled (I3-I5 eco-bioevents), *T. quinqueloba* (Q4-Q9 eco-bioevents) and *N. pachyderma* right coiled (P5-P7 eco-bioevents). According to Perez-Folgado et al. (2003, 2004), the documented short-term oscillations in the planktonic foraminiferal fauna are clearly associated to the stadial/interstadial excursions occurring over the last 80 kyr and allowed the identification in core C08 of the S1, YD, GI-1 and GS-2 climatic events in the last 23 kyr. Furthermore, the comparison between the  $\delta^{18}\text{O}$  NGRIP ice core record with *G. ruber* oscillation of core C08, suggests that the Heinrich events (H1 to H6) and the Younger Dryas (YD) are even recorded.

The eco-bioevents chronology combined with  $^{14}\text{C}$  – AMS data were used to define a detailed age model which was compared by means of reflectance parameters to the astronomically tuned age

model proposed for the Ionian Sea ODP-Site 964 (Lourens, 2004). The similarity between the two reflectance records plotted vs time, validated the age model of the studied record especially in the time intervals between 2-25 kyr and 60-83 kyr. This methodology, if confirmed with further evidences, proved a powerful tool for reliably correlating marine records between comparable deep sea environment settings.

The sector of the Sentinelle Valley intercepted by the CIESM core C08 has been sporadically fed by sand turbidite flows, likely driven along the Bizerte Canyon from the northern sector of the Tunisian Plateau, during relative sea level minimum and sea level drop of MIS 4 and 3. This sector of the basin was reached by three sand deposition events of increasing thickness in the time interval from 64 to 28 kyr cal. BP following a regressive trend. Starting from about 28 kyr this part of the fan was deactivated and fossilized beneath a carpet of hemipelagic mud at a sedimentation rate of about 7 cm/kyr.

The combined logging of sedimentological, petrophysical and paleomagnetic data of core CIESM-C08, integrated with the ecobiozone stratigraphy could provide an important source of information useful to improve the confidence of correlations in the Mediterranean for the last 83 kyr.

In this framework, future efforts will be focused on the comparison of the ChRM directions obtained for the core with the other available European geomagnetic paleosecular variation records, as well as on the reconstruction of a relative geomagnetic paleointensity curve, to establish a Mediterranean paleomagnetic reference record to be used as an original dating tool for coeval sequences of this region.

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Events	mbsf	Age (Ka cal. BP)	References
AMS <sup>14</sup> C –	0	2,18	this work
top S1	0,360	6,5	Casford et al. (2002) (Aegean Sea)
AMS <sup>14</sup> C –	0,440	8,8	this work
base S1	0,470	9,8	Casford et al. (2002) (Aegean Sea)
top YD	0,570	11,5	Asioli et al. (2001) (Adriatic Sea)
base YD	0,700	12,6	Asioli et al. (2001) (Adriatic Sea)
base GI-1	0,850	14,8	Asioli et al. (2001) (Adriatic Sea)
base GS-2	1,350	20	Andersen et al. (2006) (Greenland Ice core)
AMS <sup>14</sup> C –	1,430	25,1	this work (not used in the age model)
Base B2	1.540	22	Perez-Folgado et al. (2003) (Alboran Sea)
Top B3	1.620	25.3	Perez-Folgado et al. (2004) (Alboran Sea)
Base B3	1.720	27	Perez-Folgado et al. (2003) (Alboran Sea)
Top B4	2,360	36,50	Perez-Folgado et al. (2004) (Alboran Sea)
Base B4	2,690	39.50	Perez-Folgado et al. (2004) (Alboran Sea)
Top B5	3.040	47,00	Perez-Folgado et al. (2004) (Alboran Sea)
Base B5	3,200	49,00	Perez-Folgado et al. (2004) (Alboran Sea)
Top B6	3,570	51,50	Perez-Folgado et al. (2004) (Alboran Sea)
Base B6	3,740	53,10	Perez-Folgado et al. (2004) (Alboran Sea)
Top B7	3,875	56,70	Perez-Folgado et al. (2004) (Alboran Sea)
Base B7	4.075	60,20	Perez-Folgado et al. (2004) (Alboran Sea)
Top B8	4.125	62.70	Perez-Folgado et al. (2004) (Alboran Sea)
Base B8	4.275	64.30	Perez-Folgado et al. (2004) (Alboran Sea)
Top B9	4.430	64,90	Perez-Folgado et al. (2004) (Alboran Sea)
Base B9	4.610	67,70	Perez-Folgado et al. (2004) (Alboran Sea)
Top B10	4,680	71,10	Perez-Folgado et al. (2004) (Alboran Sea)
Base B10	4,910	74,20	Perez-Folgado et al. (2004) (Alboran Sea)
Top B11	5.080	79,00	Perez-Folgado et al. (2004) (Alboran Sea)
Base B11	5.210	81,30	Perez-Folgado et al. (2004) (Alboran Sea)

Table 1.

### Captions to figures

Fig.1 – A) Location map of CIESM Core C08. Bathymetry from The General Bathymetric Chart of the Oceans (GEBCO, 1997); B) bathymetric detail of core site, close to a channel South-North oriented (green arrow).

Fig.2 - The core CIESM 08: photography, lithologic log, petrophysical properties curves (magnetic susceptibility, Grape density, reflectance 550 nm %), paleomagnetic measurements curves (Normalized intensity, ChRM inclination, ChRM declination, MAD) plotted against depth (cm below sea floor). In the paleomagnetic curves each dot represents a single measurement taken at 1-cm spacing. NRM: natural remanent magnetization; ChRM: characteristic remanent magnetization determined by principal component analysis on the demagnetization data; MAD: maximum angular deviation. The piston core was not azimuthally oriented, but a fiducial mark ensured the reciprocal orientation of each section from which u-channels were collected. The declination of each u-channel was rotated to bring the average value at 0°.

Fig.3 – Relative abundance of selected planktonic foraminifera from core CIESM C08 plotted vs depth (m bsf). Eco-biozones 1 to 10 are pointed out, according to the scheme of Sprovieri et al. (2003), slightly modified. Grey bands show the selected eco-bioevents proposed by Perez-Folgado et al. (2003, 2004). Event Stratigraphy according to GRIP scheme. MIS and age scale according to Perez-Folgado et al. (2003, 2004). Associated to *G. ruber* curve, two grey bands interbedded with the black one correspond to the position of sapropel 1 equivalent with the two interval S1a and S1b. The two banded areas indicate the position of the turbidite layers.

Fig.4 – From left to right: comparison in time domain between the distribution pattern of *G. bulloides* from ODP-Site 977 (Perez-Folgado et al., 2004) and the studied core CIESM C08 (the red curve represents a 3-points average). The grey bands and the labels B1 to B11 are from Perez-Folgado et al., (2004). Age-Depth profile and Sedimentation Rate of core

CIESM C08. The adopted tie-points by eco-bioevents and by  $^{14}\text{C}$  – AMS data are shown respectively with black boxes and grey boxes.

Fig. 5- Comparison in time domain of colour reflectance data of core CIESM C08 (black curve, with 3 points average, red dotted curve) with reflectance data after Lourens (2004) for ODP-Site 964 (thin black curve) in the Ionian Sea. The numbers I1-I6 indicate tephra layers in ODP-Site 964 and the code 1 and X indicate the position of the sapropel 1 and X, respectively.

Fig. 6- Distribution pattern of *G. ruber* (thin black line) with 3 points average (red line) of core CIESM C08 plotted versus  $\delta^{18}\text{O}$  NGRIP (NGRIP members, 2004) record, with 7 points average, in time domain. Labels H1 to H6 indicate the position of Heinrich events and the label YD the position of the Younger Dryas.

Fig. 7 – Downcore plots of paleomagnetic data. The NRM intensity plotted versus stratigraphic depth is shown aside to the main paleomagnetic data (NRM, inclination and declination of the characteristic remanent magnetization) plotted versus age, accordingly to the independently derived age model discussed in this study.

Fig.8 – The inferred age of turbidite sand beds is plotted on curves of sea level variations over the last 100 kyr (from Antonioli et al., 2004, modified). T2 and T3 turbidites occur during two relative lowstand phases of the MIS 4 and MIS 3, while the T1 turbidite falls during the last dropping phase of sea level of MIS 3.

#### Table caption

Tab. 1- Tie-points used for age-depth profile of core C08. The *Globigerina bulloides* eco-bioevents B3 to B11 are coded according to Perez-Folgado et al. (2003, 2004).

Fig. 1

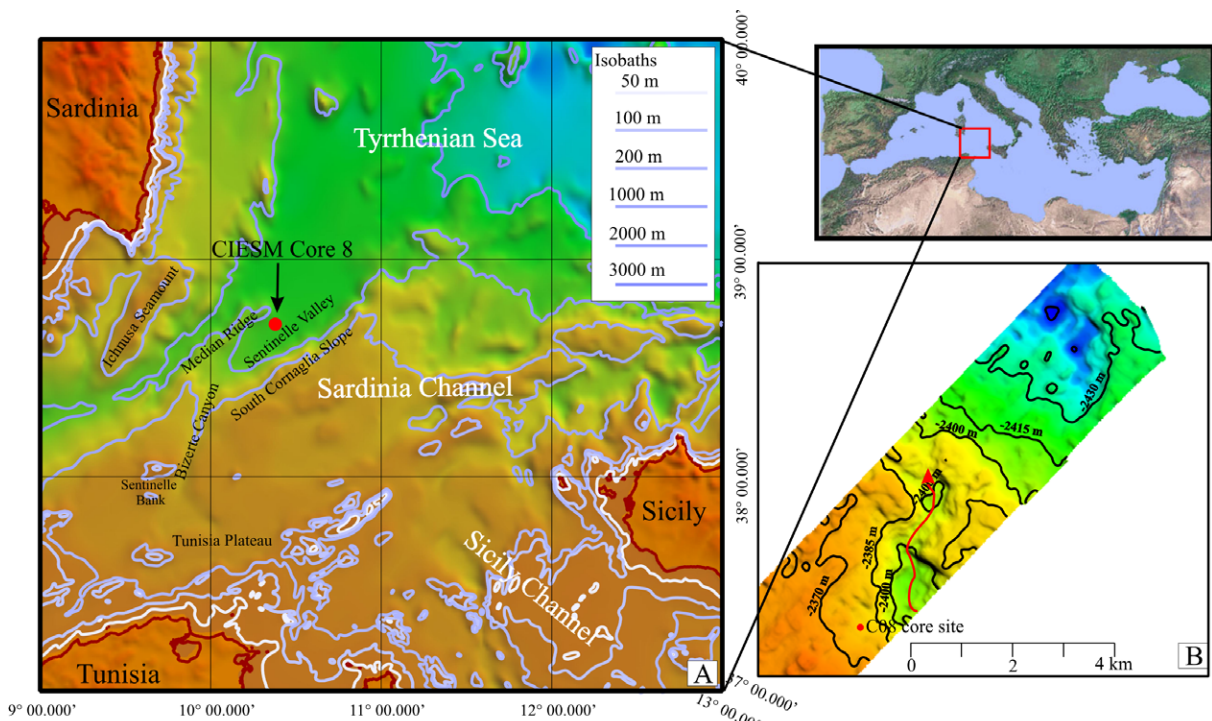
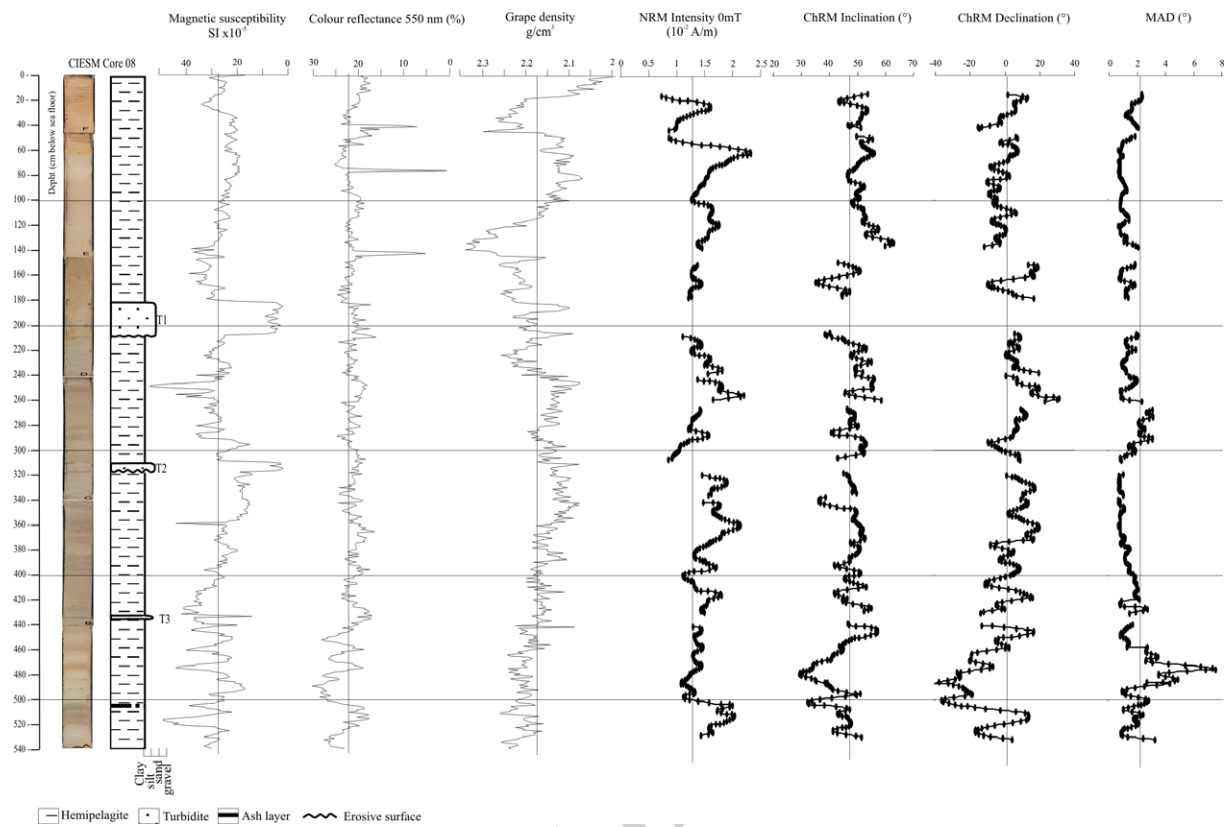
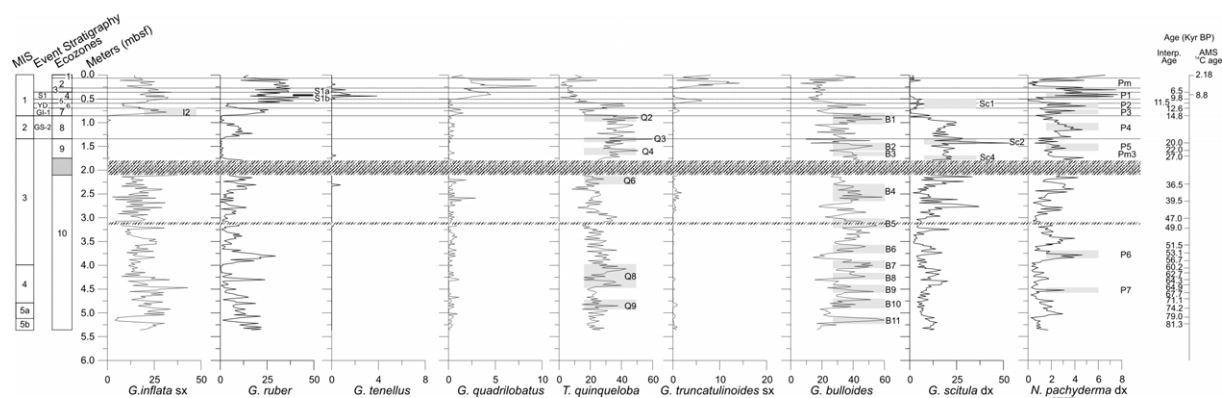


Fig. 2



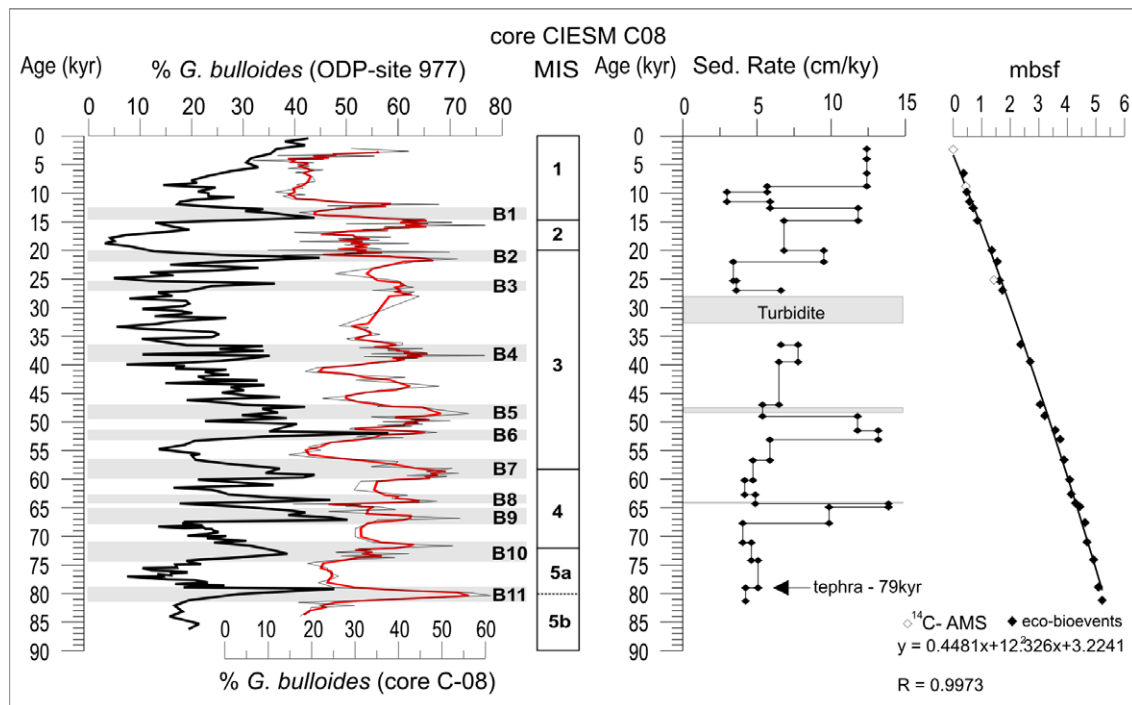
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Fig. 3



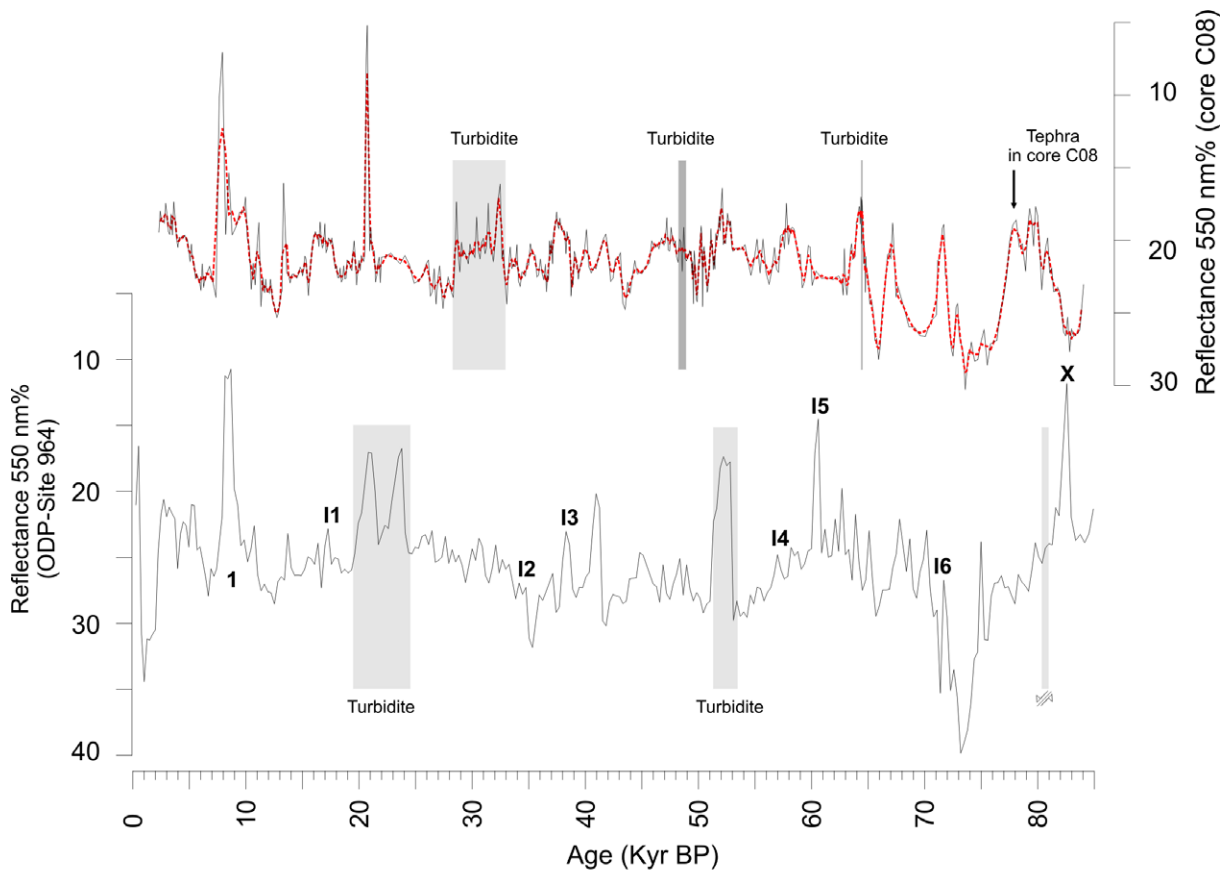
Accepted manuscript

Fig. 4



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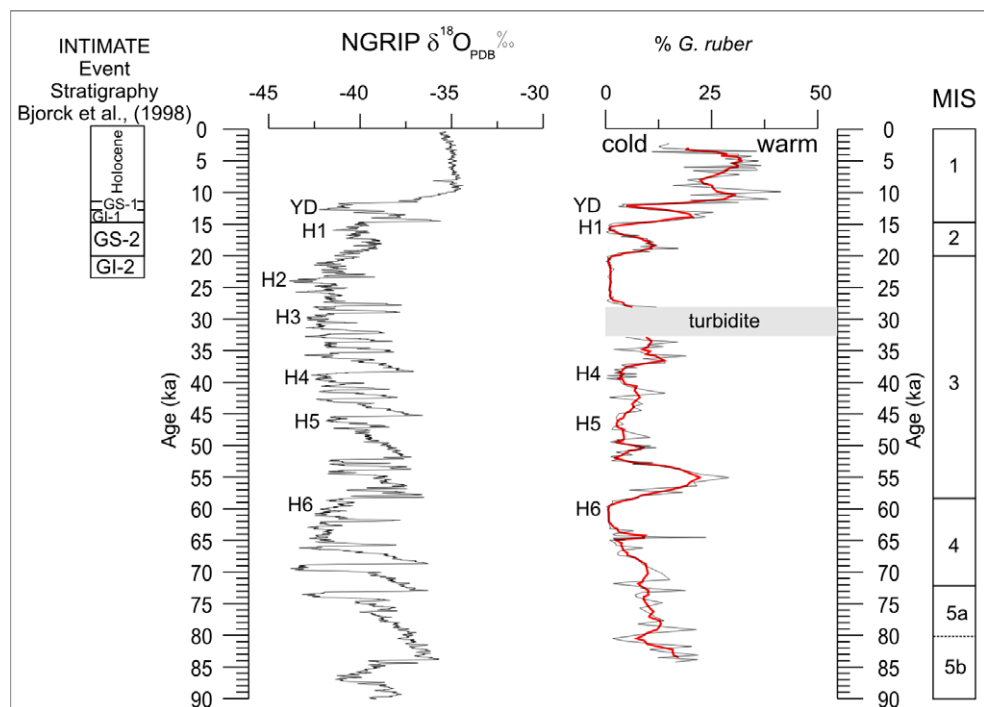
Fig. 5



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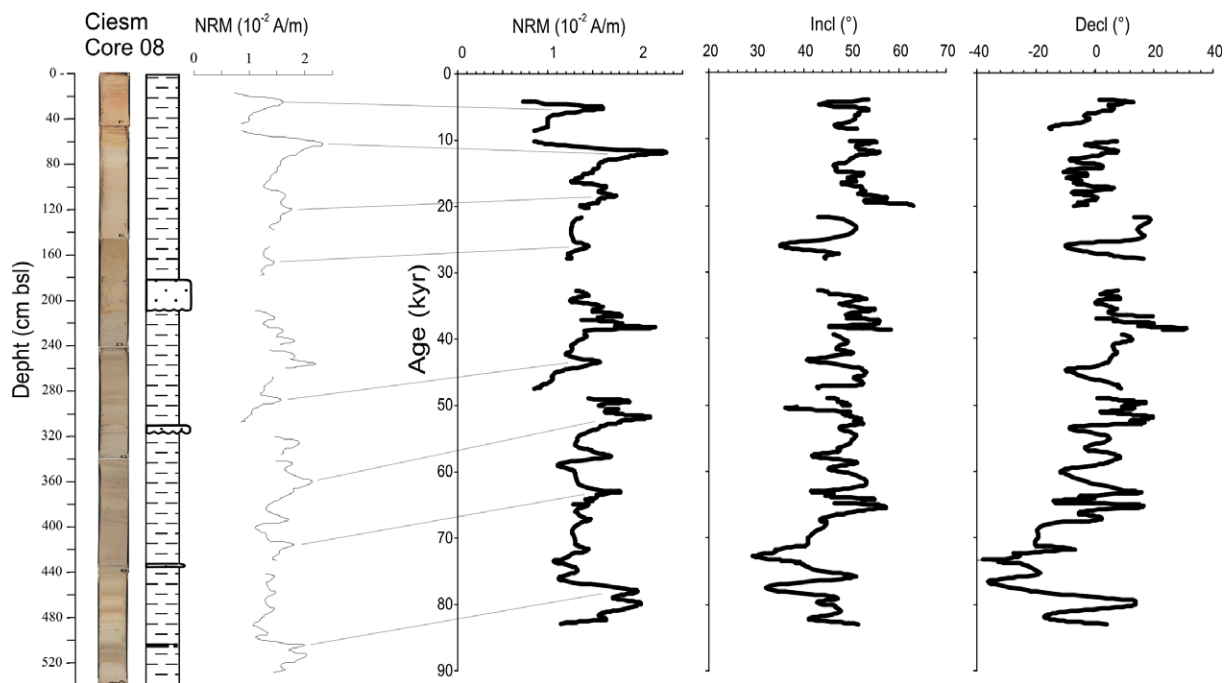


Fig. 6



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Fig. 7



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Fig. 8

