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The Global Stratotype Section and Point (GSSP) of the Serravallian Stage (Middle Miocene)

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The Global Stratotype Section and Point (GSSP) for the Base of the Serravallian Stage (Middle Miocene) is defined in the Ras il Pellegrin section located in the coastal cliffs along the Fomm Ir-Rih Bay on the west coast of Malta (35°54'50"N, 14°20'10"E). The GSSP is at the base of the Blue Clay Formation (i.e., top of the transitional bed of the uppermost Globigerina Limestone). This boundary between the Langhian and Serravallian stages coincides with the end of the major Mi-3b global cooling step in the oxygen isotopes and reflects a major increase in Antarctic ice volume, marking the end of the Middle Miocene climate transition and the Earth's transformation into an "Icehouse" climate state. The associated major glacio-eustatic sea-level drop corresponds with sequence boundary Ser1 of Hardenbol et al. (1998) and supposedly with the TB2.5 sequence boundary of Haq et al (1987). This event is slightly older than the last common and/or continuous occurrence of the calcareous nannofossil Sphenolithus heteromorphus, previously considered as guiding criterion for the boundary, and is projected to fall within the younger half of Chron C5ACn. The GSSP level is in full agreement with the definitions of the Langhian and Serravallian in their respective historical stratotype sections in northern Italy and has an astronomical age of 13.82 Ma.

Introduction and Motivation

The aim of this paper is to announce the formal ratification of the Global Stratotype Section and Point (GSSP) of the Serravallian Stage which, together with the preceding Langhian, constitutes the twofold subdivision of the Middle Miocene Subseries in the Global Standard Global Chronostratigraphic scale. Boundaries between global stages, the basic chronostratigraphic units, are defined by a GSSP for the younger stage. Their formal definition at a well defined

point in a continuous marine section facilitates communication among Earth Scientists as it permits to export the boundary as a timeline away from the GSSP, using multiple stratigraphic tools.

During the last decade, much progress has been made in the Neogene by defining GSSPs of the Zanclean (Van Couvering et al., 2000), Piacenzian (Castradori et al., 1998) and Gelasian (Rio et al., 1998) Stages of the Pliocene, and the Messinian and Tortonian Stages of the (Upper) Miocene (Hilgen et al., 2000a; Hilgen et al., 2005). The next logical step is to select and propose the GSSP for the next older stage in the Miocene, the Serravallian (Pareto, 1865). Unfortunately the historical stratotype section of the Serravallian consists of shallow-marine sediments (Vervloet, 1966) that are unsuitable for defining the GSSP. However the definition of the GSSP is greatly facilitated by the progress made in establishing an orbitally-tuned and integrated stratigraphic framework for the Middle-Upper Miocene both in the Mediterranean (Hilgen et al., 2000b, 2003; Caruso et al., 2002; Lirer et al., 2002; Sprovieri M. et al., 2002; Abels et al., 2005) as well as in the open ocean (Shackleton and Crowhurst, 1997; Westerhold et al., 2005; Holbourn et al., 2005; 2007).

Formal definition of Middle Miocene global chronostratigraphic units via their GSSPs is also timely in view of the current interest in major climate transitions and perturbations that occurred along the Cenozoic climatic deterioration from the Eocene "Greenhouse World" into the present "Icehouse World". In fact, one of the major changes in the climate system is termed the Middle Miocene climate transition that started from the Miocene climatic optimum around 16 to 15 Ma. The end of the transition is marked by the major Mi-3b isotope shift reflecting a significant increase in Antarctic ice volume and the final transition into the "Icehouse World" (Fig.1). In fact it is this isotope shift, or more accurately the end of this shift, that is taken as prime guiding criterion for the Serravallian GSSP rather than one of the more conventional biostratigraphic criteria that proved to be slightly diachronous between the Mediterranean and open ocean.

The Ras il Pellegrin section located along the west coast of Malta was selected as the best section for defining the Serravallian GSSP because it covers the critical time interval in a continuous deep marine succession suitable for integrated stratigraphic studies. In 2006, a formal proposal for defining the boundary (Hilgen et al., 2006) was unanimously accepted (86% quorum, all 18 votes positive, one with reservations) by the voting members of the Subcommittee on Neogene Stratigraphy (SNS). Official acceptance of the revised proposal by the International Commission on Stratigraphy (ICS)

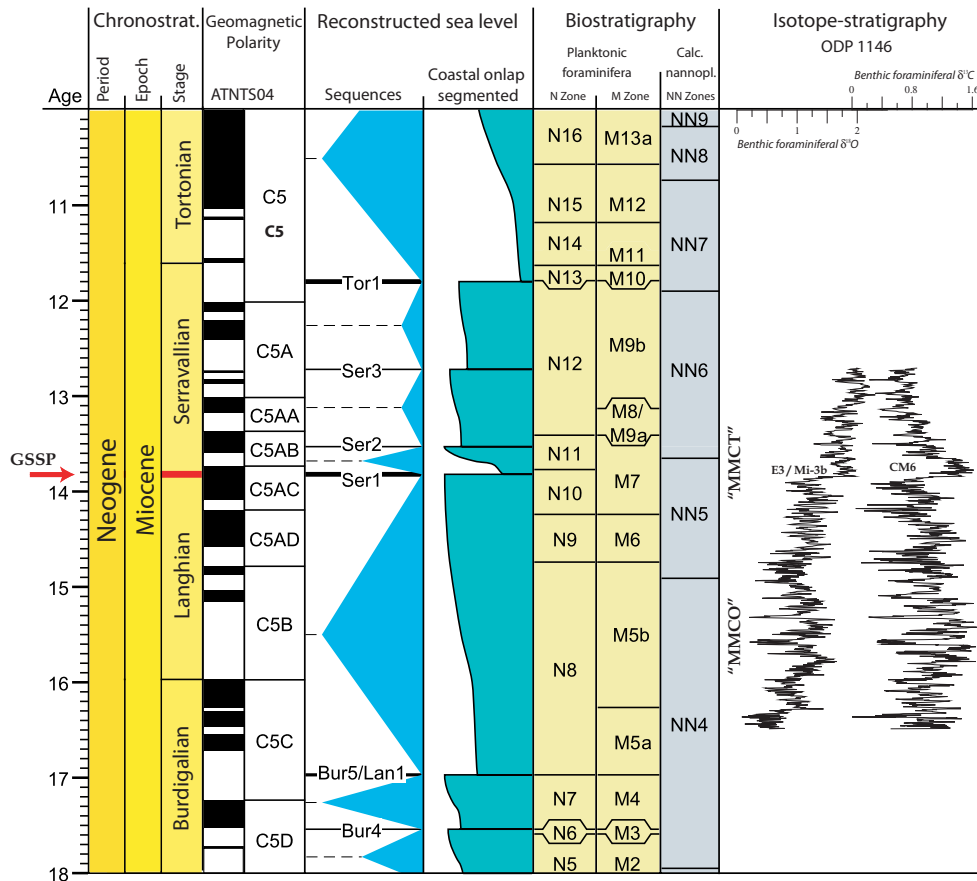


Figure 1. Summary of the Geologic Time Scale 2004 (GTS04) from 18 to 10 Ma, showing the geomagnetic polarity time scale of the ATNTS2004 (Lourens et al., 2004), the reconstructed sea level and segmented coastal onlap (after Hardenbol et al., 1998), the planktonic foraminiferal N and M zones (Blow, 1969 and Berggren et al., 1995), the calcareous nannofossil NN zones (after Raffi et al., 2006), and the benthic oxygen and carbon isotope records of ODP Site 1146 (after Holbourn et al., 2005; 2007). Biostratigraphic ages are from the Mediterranean if available, other ages are from ODP Sites 925 and 926. Figure modified from a chart produced with the Time Scale Creator program, available at: <http://stratigraphy.org/column.php?id=Time%20Scale%20Creator>. Serravallian GSSP is indicated by an arrow to the left.

(15 votes or 88% positive, one abstain and one negative) and ratification by the Executive Committee of the International Union of Geological Sciences (IUGS) were obtained later in 2006 and in 2007, respectively.

Serravallian Stage: a brief historical review

In this section we start with a brief outline of the original definition of the Serravallian Stage by Pareto in 1865, followed by a short description of the Serravallian historical stratotype of Vervloet (1966) and the position of the base of the Serravallian and top Langhian in their respective stratotypes.

Original definition of the Serravallian (Pareto, 1865)

The Serravallian Stage, named after the village of Serravalle Scrivia in northern Italy, was introduced by Pareto in 1865 (in english translation):

“Towards the upper part of the Langhian stage you can see the beginning of an alternation of greyish marls with beds of yellowish sands which start to show aspects of deposits that formed in a sea

less deep and less far from the coast. It is at the beginning of the alternations of greyish sandy marls and yellow sands that I place the lower limit of the third subdivision of the Miocene terrains, which is that of the Upper Miocene and which I name the Serravallian stage, after the village of Serravalle, where the stage is well developed and forms a range of high hills that extend to the west and east of the village and which represent a special geological aspect and composition”.

Note that Pareto (1865) considered the Serravallian as the third (upper) Miocene stage following the – now obsolete – Bormidian and the Langhian defined by him at the same time. He placed the Tortonian already in the Pliocene below the upper Pliocene Piacenzian (see Fig.1a in Hilgen et al., 2005).

After its introduction, the Serravallian was soon abandoned in favour of the Helvetian introduced earlier by Mayer-Eymar (1858; see Fig.1a in Hilgen et al., 2005). But the term Serravallian was revived after it was realized that the type Helvetian was time-equivalent with the Burdigalian (Regional Committee on Mediterranean Neogene Stratigraphy (RCMNS) congress, Vienna, 1959). A proposal was presented – and accepted – at the RCMNS congress in Bratislava (1975) to incorporate the Serravallian

in the Standard Chronostratigraphic Scale as the second upper subdivision of the Middle Miocene, above the Langhian and below the Tortonian (see Fig.1b in Hilgen et al., 2005). The Serravallian has been consistently used as a global stage in all published standard geological time scales afterwards (e.g., Harland et al., 1989; Berggren et al., 1995; Gradstein et al., 2004; Lourens et al., 2004) (see Fig. 1c in Hilgen et al., 2005). Nevertheless, the term was rarely used outside the Mediterranean probably as a consequence of different interpretations proposed for the position of the Langhian-Serravallian boundary relative to the Geomagnetic Polarity Time Scale (GPTS) (Rio et al., 1997). However, since 1997, there was general consensus to use the *Sphenolithus heteromorphus* Last Occurrence (LO) as primary guiding criterion for defining the boundary (Rio et al., 1997).

Serravallian stratotype section (Vervloet, 1966)

After the revival of the Serravallian, a stratotype section was designated by Vervloet in the Scrivia valley near the village of Serravalle Scrivia in agreement with the type section of his Serravalle Sandstone Formation (Vervloet, 1966; see also Boni and Selli, 1971). More specifically, the stratotype section is located in the western

Tertiary Piedmont Basin (Italy) near Serravalle Scrivia along the left bend of the lower course of the Scrivia river in the province of Alessandria, approximately 28 km SE of that city.

The Serravalle Sandstone near Serravalle Scrivia represents a succession of shallow marine shelf deposits most likely punctuated by hiatuses (Caprara et al., 1985; Ghibaudo et al., 1985; Gnaccolini, 1989). Therefore an alternative stratotype section was proposed by Boni (1967), and a deep water parastratotype by Cita and Premoli Silva (1968) to overcome the scarcity of planktonic foraminifera and the shallow-marine character of the stratotype section of Vervloet (1966). Nevertheless it is the latter section that is unanimously considered as the historical stratotype of the Serravallian.

The planktonic foraminifera in the historical stratotype were studied by Vervloet (1966), Cita and Blow (1969), Foresi (1993), Miculan (1994) and Fornaciari et al. (1997a), and are poorly preserved and not very age diagnostic. The timing of the type Serravallian was therefore partly based on the study of other reference sections in the area. The calcareous nannofossils of the Serravallian stratotype were investigated by Müller (1975) and, in more detail, by Fornaciari et al. (1997a; see below and also Rio et al., 1997).

Timing of the base of the Serravallian and top of the Langhian stratotype

In the stratotype section, the base of the Serravallian is located just below the Last Occurrence (LO) of the calcareous nannofossil *Sphenolithus heteromorphus* and above the First Common Occurrence (FCO) of the calcareous nannofossil *Helicosphaera walbersdorfensis* and the First Occurrence (FO) of the planktonic foraminifer *Orbulina universa* (Fig. 2) (Fornaciari et al., 1997a; Rio et al., 1997). The latter two events are virtually coincident with the top of the Langhian in its historical stratotype (Fig. 2) (Rio et al., 1997; Fornaciari et al., 1997b). The Langhian-Serravallian boundary is thus bracketed by the *O. universa* FO and *H. walbersdorfensis* FCO below, and the LO of *S. heteromorphus* above (Fig. 3).

In the Mediterranean, the *O. universa* FO has been calibrated to the GPTS using the magnetobiostratigraphy of DSDP Site 372, and is associated with Chron C5ADn with an estimated age of 14.36 Ma (Abdul Aziz et al., 2008). Unfortunately, this event is seldomly reported from the open ocean, where instead the *O. suturalis* FO (i.e. the *Orbulina* datum), used as zonal marker in standard zonations (Blow, 1969; Berggren et al., 1995), has been calibrated. However, the calibration of *O. suturalis* FO in the Mediterranean (lowermost part of Chron C5ADn with an age estimate of 14.56 Ma) (Abdul Aziz et al., 2008) is not in agreement with that reported from the low-latitude open ocean by Berggren et al. (1995) (associated with Chron C5Bn.2n and an age of 15.1 Ma) and Lourens et al. (2004) (astronomical age of 14.74 Ma). The *H. walbersdorfensis* FCO is an event that is only well recognizable in the Mediterranean, where it is linked to Chron C5ACn with an age of 14.05 Ma (Abdul Aziz et al., 2008). The extinction level of *S. heteromorphus* is an excellent bioevent for global correlation. Note that the marked abundance decrease of *S. heteromorphus*, usually defined as LO (Last Occurrence) (e.g. Fornaciari et al., 1996; Rio et al., 1997; Raffi et al., 2006) is here indicated as L(C)O (Last Common and/or Continuous Occurrence) (see discussion in Di Stefano et al., 2008). The L(C)O of *S. heteromorphus* is a well calibrated event both in the Mediterranean and in the open oceans; it is associated with Chron

C5ABr (Abels et al., 2005; Abdul Aziz et al. 2008, and references therein) and has been astronomically dated at 13.654 Ma in the Mediterranean (Abels et al., 2005) and at 13.532 Ma in the Equatorial Atlantic Ocean (Backman and Raffi, 1997).

Therefore, following the concept that the top of a stage is defined by the base of the next younger stage, the LO of the calcareous marker species *S. heteromorphus* was commonly taken as primary guiding criterion for the Langhian-Serravallian boundary (Rio et al., 1997), even though the boundary was not yet formally defined.

Selecting the most suitable section and level for defining the Serravallian GSSP

In the Neogene, orbitally tuned cyclostratigraphies play an important role in addition to the conventional criteria outlined by ICS in the revised guidelines for establishing global chronostratigraphic standards (Remane et al., 1996). This extra criterion is added here because all ratified Neogene GSSPs are defined at lithological marker beds that are astronomically dated. In this way they are tied via first-order calibrations to the standard Neogene time scale which is underlain by astronomical tuning (Lourens et al., 2004). This implies that if other requirements are equal then cyclostratigraphy plays a decisive role in selecting the most suitable section and level for defining the Serravallian GSSP.

Selecting the guiding criterion for defining the boundary

According to the biostratigraphic data from the historical stratotype sections (Langhian and Serravallian), the Serravallian GSSP should best be defined at or close to the *S. heteromorphus* L(C)O. However, the astronomical age for this event is ~ 100 kyr older in the Mediterranean as compared with the equatorial Atlantic (13.654 vs 13.532 Ma; Backman and Raffi, 1997; Abels et al., 2005), rendering this event less suitable for defining the boundary. The major shift in the Middle Miocene marine oxygen isotope record (e.g., Abels et al., 2005) provides an alternative and more suitable guiding criterion. This shift towards heavier $\delta^{18}\text{O}$ values, labelled Mi-3b, has now been recorded in a number of deep-sea cores and marks a major step in the Middle Miocene cooling and Antarctic ice sheet build up, in fact reflecting the final step in the transition from a “Greenhouse” to “Icehouse” climate state over the past 50 myr (Woodruff and Savin, 1991; Miller et al., 1991, 1996; Flower and Kennett, 1993, 1994). This shift is a major truly global synchronous event, dated astronomically at 13.82 Ma (Abels et al., 2005; Holbourn et al., 2005). Taking this event as defining criterion for the Serravallian base would also meet the requirement that the boundary should be slightly older than the *S. heteromorphus* LO found in the basal part of the historical stratotype of the Serravallian and younger than the *O. universa* FO and the *H. walbersdorfensis* FCO found below the Serravallian base and coincident with the Langhian top in its historical stratotype (Figs. 2 and 3) (Rio et al., 1997; see also Odin et al., 1997).

Selecting the section and defining the boundary

As mentioned before, the historical stratotype is considered unsuitable to define the Serravallian GSSP as it contains shallow

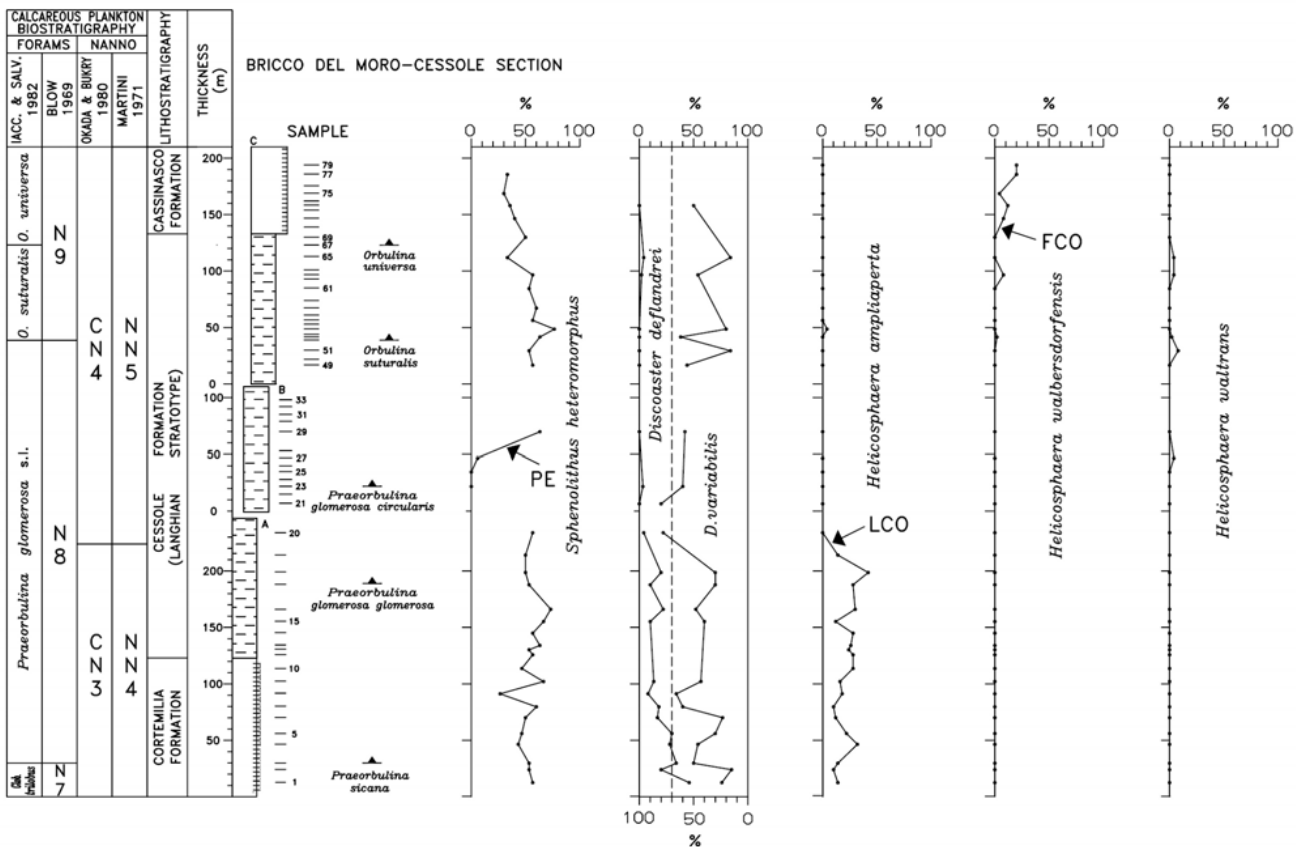
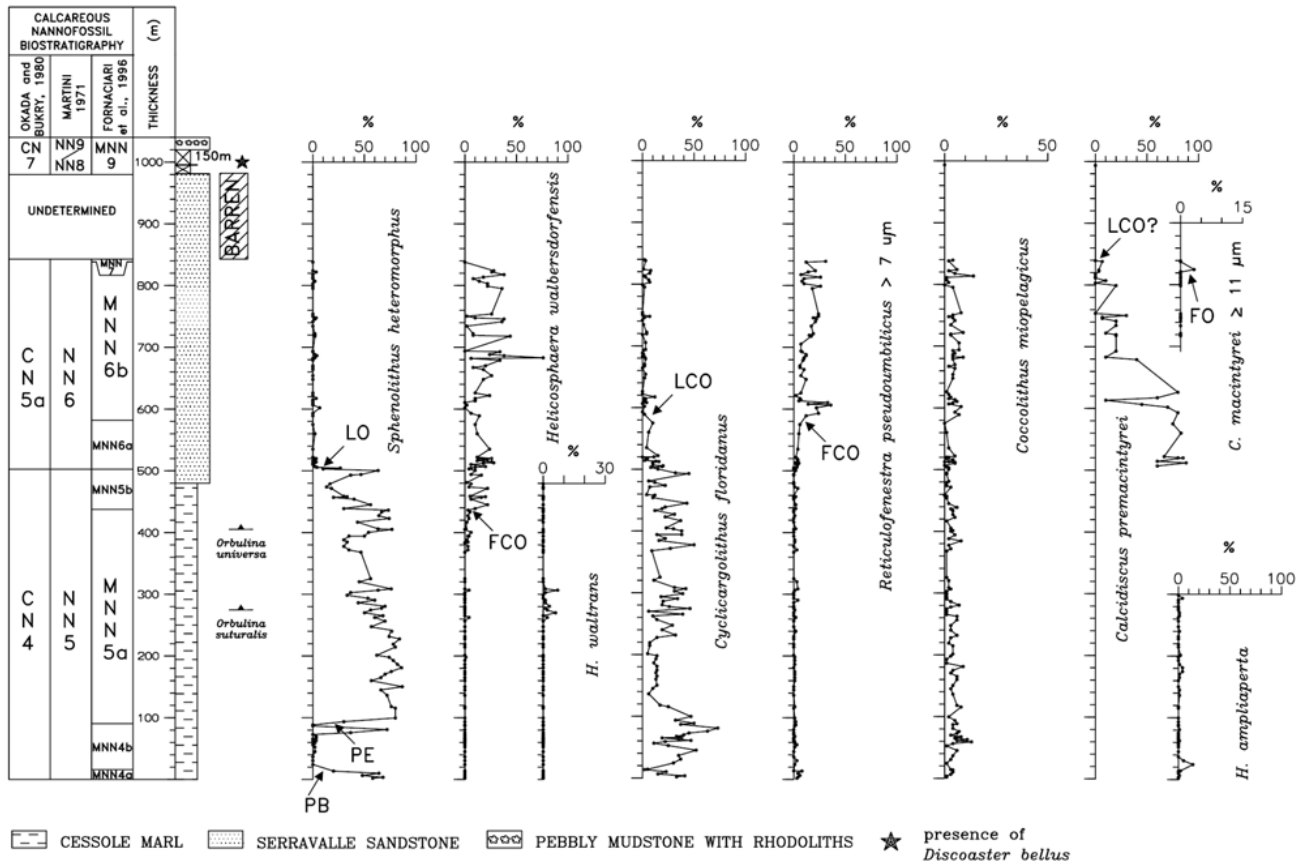


Figure 2. Modern calcareous plankton biostratigraphic data of the Langhian (Bricco del Moro-Cessolo) and Serravallian historical stratotypes (from Fornaciari et al., 1997a,b; Rio et al., 1997).

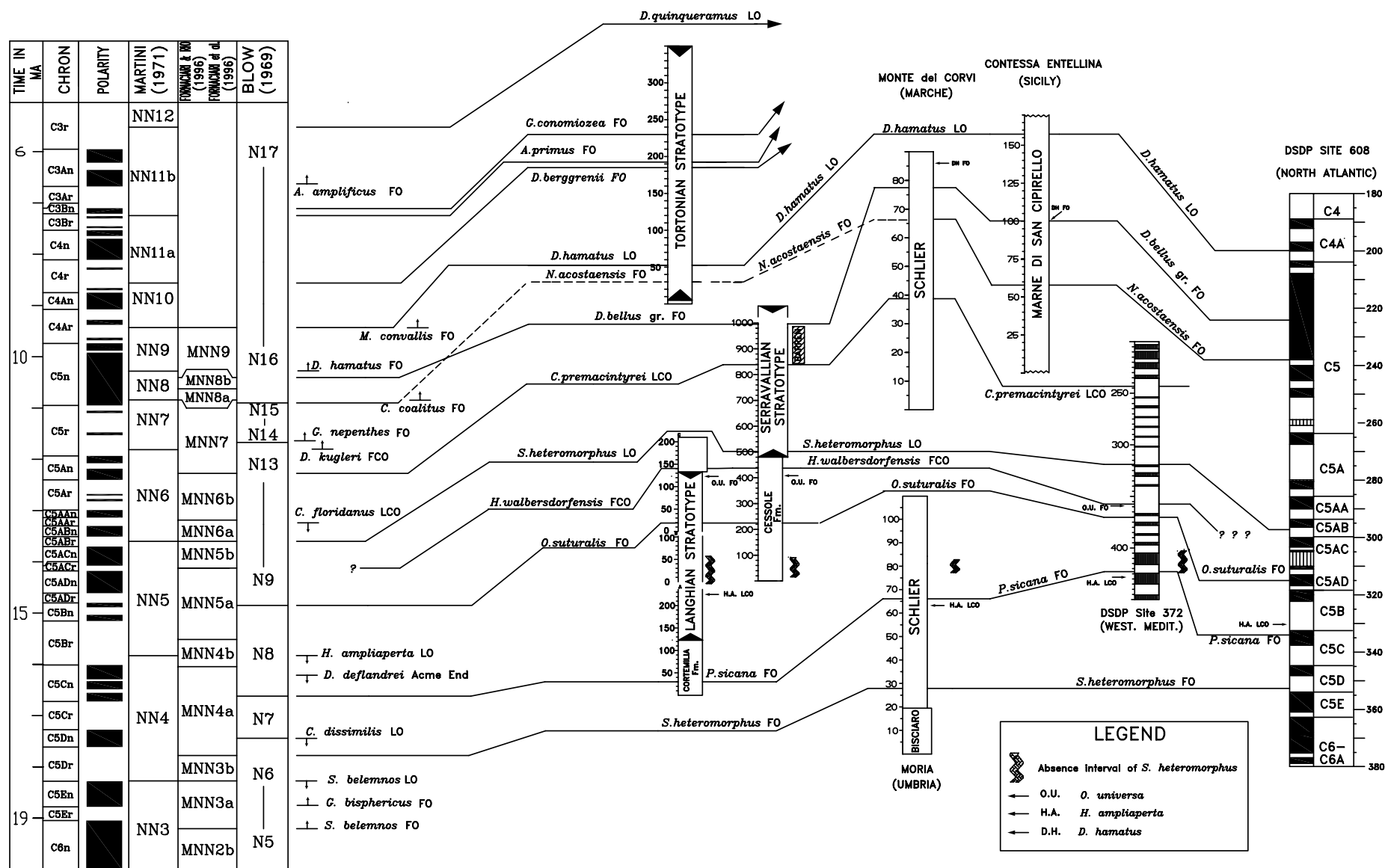


Figure 3. Overview of calcareous plankton correlations between the Langhian, Serravallian and Tortonian historical stratotype sections, and other Mediterranean sections and DSDP Site 608 in the North Atlantic. Calcareous plankton biochronology for the Mediterranean and low-latitude ocean is shown to the left. The Geomagnetic Polarity Time Scale is that of Cande and Kent (1995). Biochronologic data are of Berggren et al. (1985), Backman et al. (1990) and Raffi et al. (1995) (from Rio et al., 1997).

marine sediments unsuitable for detailed biostratigraphic and paleomagnetic studies and astronomical dating. In addition, the succession may contain hiatuses and preservation of foraminifera is poor thus preventing the construction of meaningful isotope records. However, continuous and cyclic deep marine sections that might be much more suitable for defining the boundary are found near Ancona (Italy), and on Tremiti Islands (Italy) and Malta-Gozo.

The Monte dei Corvi section located near Ancona (central Italy) was designated to formally define the Tortonian GSSP (Hilgen et al., 2003, 2005). The section can be extended downward in the La Vedova and Monte dei Corvi High Cliff sections, where the *O. universa* FO, the *H. walbersdorfensis* FCO and the *S. heteromorphus* LO are found following an unexposed interval (Montanari et al., 1997). The critical interval around the Mi-3b event is covered by a landslide along the beach but it is exposed high in the cliffs where it is very difficult to reach. Moreover, the poor preservation of the calcareous plankton hampers the construction of a reliable isotope record.

Cyclic successions of middle Miocene age are also exposed on the islands of San Nicola and Cretaccio (Tremiti Islands). The younger part is found on San Nicola where it covers the interval from 13.7 to 11.1 Ma. This part of the succession has been astronomically tuned (Lirer et al., 2002; Abels et al., 2005) and is suitable for isotope studies. The older part of the succession is exposed on Cretaccio and ranges from approximately 15.7 to 14.3 Ma (from the Acme Beginning of *Paragloborotalia siakensis* up to the *Helicosphaera waltrans* Last Common Occurrence; Di Stefano et al., 2008). Consequently the critical interval is not exposed on Tremiti islands.

The critical interval is present in the open marine succession exposed on the islands of Malta and Gozo in the part of the succession that ranges from the Upper Globigerina Limestone into the Blue Clay Formation. The Ras il Pellegrin section along the Fomm Ir-Rih Bay on the west coast of Malta contains this interval and is selected because of its excellent exposures and distinct sedimentary cyclicity. The section was studied in detail by Italian and Dutch research teams (Bellanca et al., 2002; Bonaduce and Barra, 2002; Foresi et al., 2002; Sprovieri M. et al., 2002; Abels et al., 2005). The section proved suitable for astronomical tuning and the final age for the main shift in $\delta^{18}\text{O}$ (Abels et al., 2005) was in excellent agreement with the age for the same event in the open ocean (Holbourn et al., 2005; Westerhold et al., 2005).

The Ras il Pellegrin section was selected as it was the only section in which the boundary interval is exposed with certainty, apart from the difficult to reach La Vedova high cliff section. The section is demonstrably continuous, tectonically undisturbed, excellently exposed and easily accessible, and contains sedimentary cycles that allow the section to be tuned across the critical interval.

The Serravallian GSSP at Ras il Pellegrin (Malta)

The Serravallian GSSP was proposed and is now formally defined at the boundary between the Globigerina Limestone and Blue Clay Formations in the Ras il Pellegrin section on Malta. First we will shortly describe the geological setting and stratigraphic succession of Malta and Gozo before we go into the relevant details of the Ras il Pellegrin section itself.

Geological setting and stratigraphic succession

The classical Oligocene-Miocene succession of the Maltese

islands was deposited in the Maltese Graben System delineated by NW-SE and ENE-WSW trending faults. This system developed on the African foreland of the Sicilian Apennine-Maghrebian fold and thrust belt, as part of a series of extensional basins during the Miocene Quaternary (De Visser, 1991; Dart et al., 1993).

The marine succession of limestones and marls more or less retained its original horizontal bedding orientation despite extensional faulting. It is classically divided into five formations: Lower Coralline Limestone (of late Oligocene age), Globigerina Limestone (Aquitania-Langhian), Blue Clay (Serravallian), Greensand (Serravallian-Tortonian), and Upper Coralline Limestone (Tortonian) (Felix, 1973; Pedley, 1975; following earlier studies by e.g. Spratt, 1843 and Murray, 1890). The stratigraphically highest unit of the sedimentary succession on the Maltese islands are Quaternary deposits consisting of sands and conglomerates with interbedded paleosols. All boundaries are conformable except for the contact between the Greensand and the Upper Coralline Limestone Formation and the contact with the Quaternary unit. The Globigerina Limestone is divided into three members (Lower, Middle and Upper) separated by two phosphate nodule beds (Felix, 1973; Pedley, 1975). These phosphate beds were studied in detail by Pedley and Bennett (1985) and mark significant hiatuses in the succession (Theodoridis, 1984).

The boundary between the Globigerina Limestone and Blue Clay Formations is long known to roughly coincide with the Langhian-Serravallian boundary because the *S. heteromorphus* LO, until recently considered the prime guiding criterion for the boundary, occurs several meters above the formation boundary (Theodoridis, 1984; de Visser, 1991). The boundary interval is well exposed both on Gozo (Marsalforn: de Visser, 1991) and Malta (Fomm Ir-Rih Bay). The formation boundary is not sharp but marked by a transitional bed (Transitional Zone of Felix, 1973, p.30), reaching a thickness of about 1.80 m at Fomm Ir-Rih. We placed the formation boundary at the top of the transitional bed because, in most sections, this interval resembles the lithological expression of the Globigerina Limestone more closely.

The coastal cliffs around the Fomm Ir-Rih Bay and nearby Gnejna Bay contain excellent exposures of the Oligocene to Miocene succession of limestones and marls. The local succession was studied in detail by Felix (1973) and subsequently by Pedley (1975), Giannelli and Salvatorini (1972, 1975) and Jacobs et al. (1996). Several years ago, an Italian research team put considerable effort in establishing an integrated stratigraphy, including an astronomical tuning, of the Ras il Pellegrin section situated in the SW facing cliffs on the NE side of Fomm Ir-Rih Bay (Sprovieri M. et al., 2002; Foresi et al., 2002); this section contains the best exposures of the Upper Globigerina Limestone and Blue Clay Formation. The most recent study of the section is that of Abels et al. (2005) who included a magnetostratigraphy and revised the previously published astronomical tuning of the Blue Clay.

The Serravallian GSSP was proposed and formally defined at the Globigerina Limestone - Blue Clay Formation boundary in the Ras il Pellegrin section. Background studies summarized below also include information from parallel sections located both on Malta and on nearby Gozo (Giannelli and Salvatorini, 1972; 1975). It is important to realize that ICS guidelines (Remane et al., 1996) indicate that GSSPs should preferably not be placed at major changes in lithofacies. Nevertheless we selected this level because it closely corresponds with the major Mi-3b isotope event that can be recognized

worldwide and marks the – end of the – so-called Middle Miocene climate transition (e.g., Holbourn et al., 2005; Westerhold et al., 2005; Raffi et al., 2006).

The Ras il Pellegrin section

The Ras il Pellegrin section is located some 20 km west of Valetta town and exposed in coastal cliffs along the Fomm Ir-Rih Bay on the west coast of Malta (Fig. 4). The actual section is located in the SW facing cliffs on the NE side of Fomm Ir-Rih Bay at 35°54' 50" North Latitude and 14°20' 10" East Longitude. The entire section contains the middle Globigerina Limestone up to the Upper Coralline

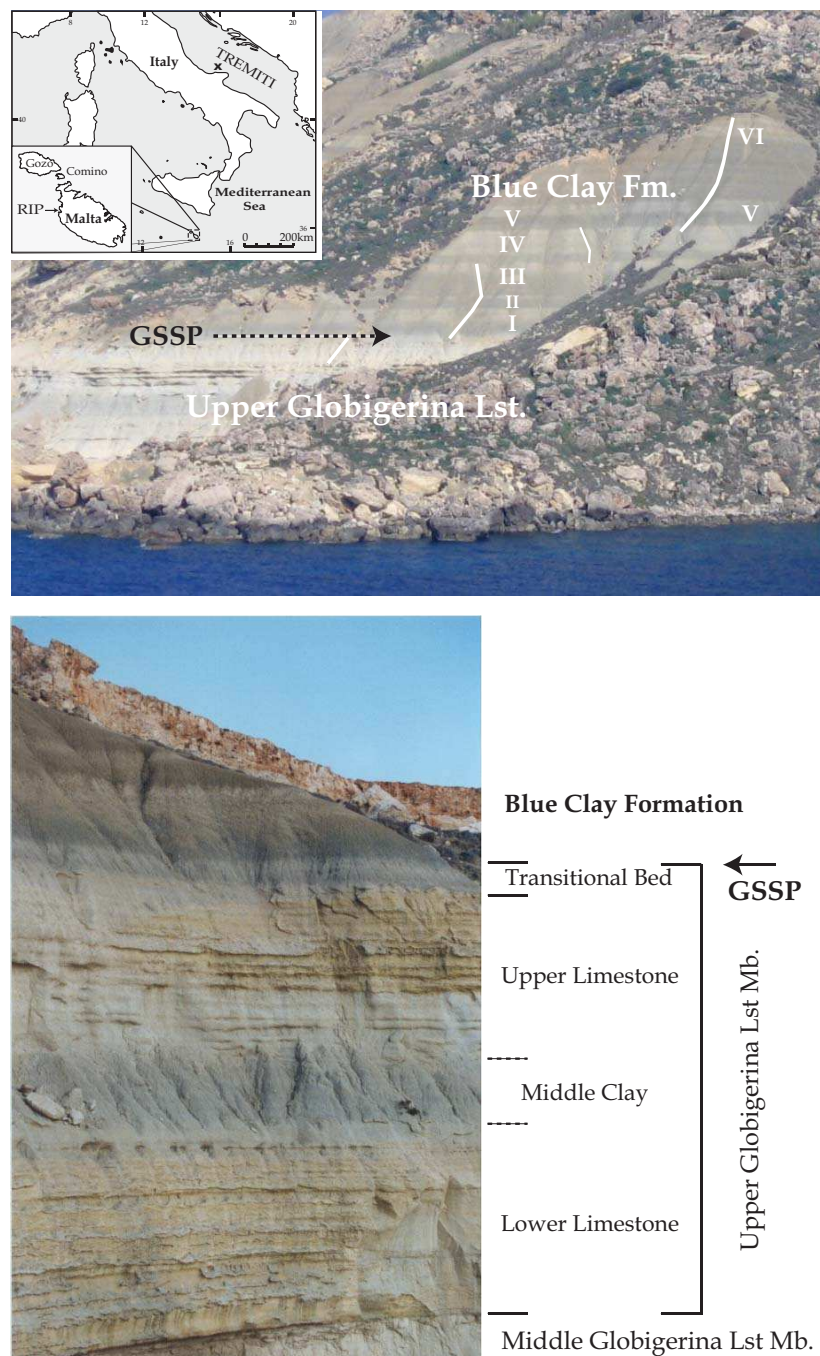


Figure 4. Location (a) and photographs (b,c) of the Ras il Pellegrin section on Malta (partly after Abels et al., 2005). Limestone intervals numbered I to VI following Sprovieri et al. (2002).

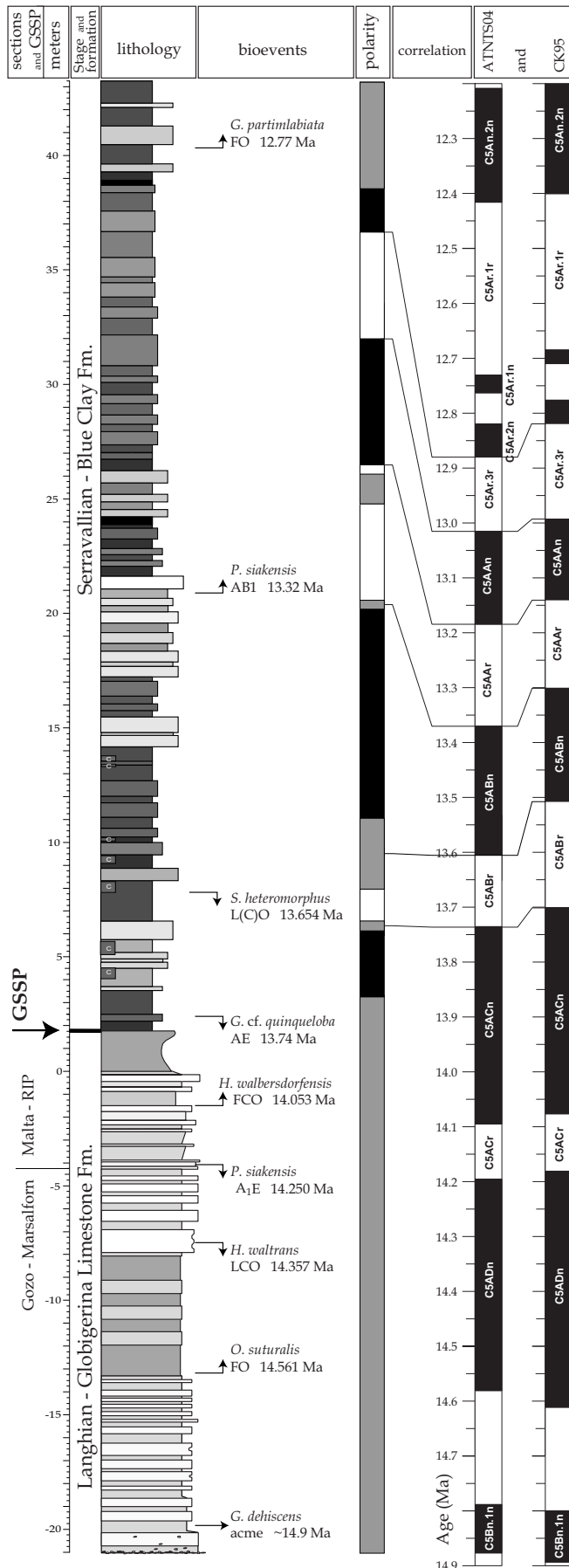
Limestone and was selected for stratigraphic studies because of its excellent exposures and distinct sedimentary cyclicity (Fig. 4). A transitional bed separates the yellowish marly limestones of the Globigerina Limestone from the softer greyish clayey marls of the Blue Clay.

The Blue Clay at Ras il Pellegrin reaches a thickness of less than 70 m and reveals a distinct and characteristic pattern of alternating homogeneous grey and white coloured marls (Figs. 4 and 5). The presence of two sapropels and several levels with chondritic trace fossils point to occasional anoxic or dysoxic bottom water conditions (Fig. 5). Finally, volcanic minerals including biotite were found at a single level around 40.45 m pointing to an ashfall in the younger part of the Blue Clay. The expression of the cycles as observed in the field was verified by geochemical analysis, in particular the Ca and Ca/K ratio (Fig. 6; Abels et al., 2005).

Calcareous nannofossil biostratigraphy

Calcareous nannofossils are abundant and their preservation is generally good to excellent in the Blue Clay and somewhat less good in the Globigerina Limestone. Calcareous nannofossil biostratigraphic studies of the Globigerina Limestone and Blue Clay on Malta and Gozo were carried out by Theodoridis (1984) and Mazzei (1985). Fornaciari et al. (1996) studied the Gnejna Bay section located several km north of the Ras il Pellegrin section, while Foresi et al. (2002) concentrated on the Blue Clay part of Ras il Pellegrin only. Combining the results of these studies, the following events are recorded – in stratigraphic order – in the upper member of the Globigerina Limestone and the Blue Clay: *H. waltrans* LO, and *H. walbersdorfensis* FCO in the Upper Globigerina Limestone, and *S. heteromorphus* L(C)O, *Cyclicargolithus floridanus* LCO, *Reticulofenestra pseudoumbilicus* FCO, *Calci-discus macintyreii* FO, and *C. premacintyreii* LCO in the Blue Clay (Fig. 5; Fig. 4 in Foresi et al., 2002).

This succession of events is essentially the same as found in other Mediterranean sections such as Monte dei Corvi (Montanari et al., 1997; Hilgen et al., 2003) and DSDP Site 372 (Di Stefano et al., 2003; Abdul Aziz et al., 2008; Di Stefano et al., 2008). The GSSP thus postdates the *H. walbersdorfensis* FCO and preceeds the *S. heteromorphus* L(C)O which together delimit zone MNN5b in terms of the standard Mediterranean zonation (Fornaciari et al., 1996; Raffi et al., 2003) and bracket the Serravallian base in the historical stratotype (Rio et al., 1997). The latter event in addition marks the MNN5-6 zonal boundary (Fornaciari et al., 1996; Raffi et al., 2003). In the open ocean, the GSSP level preceeds the NN5-6 zonal boundary of the standard zonation of Martini (1971) and the CN4-5 zonal boundary of the Okada and Bukry (1980) zonation. Similar to the MNN5-6 boundary in the Mediterranean, these zonal boundaries are defined by the *S. heteromorphus* L(C)O. Note however that the defining bioevent is slightly younger in the low



latitude open ocean (13.532 Ma; Backman and Raffi, 1997) than in the Mediterranean (13.654 Ma; Abels et al., 2005).

Planktonic foraminiferal biostratigraphy

Planktonic foraminifera are usually abundant and their preservation is good to excellent in the Blue Clay but somewhat less good in the Globigerina Limestone. The following events are distinguished - in stratigraphical order - in the Upper Member of the Globigerina Limestone: *Globoquadrina dehiscens* Acme (just above the pebble bed at the base of the member), *O. suturalis* FO (at the base of the clayey interval in the middle part of the member), and *Paragloborotalia siakensis* Acme_b End (A_bE). *Globigerina* cf. *quinqueloba* AE is located approximately 0.5 m above the top of the transition bed and hence the formation boundary and proposed GSSP (Fig. 5). This event is followed by the *G. peripheroronda* LO, a second acme (AB1) of *P. siakensis*, *P. partimlabiata* FO, and *P. mayeri* FCO (Fig. 5) (Foresi et al., 2002; Sprovieri R. et al., 2002; Abels et al., 2005). The same succession of planktonic foraminiferal events has also been reported from Tremiti Islands (Fig. 6) (Abels et al., 2005) and from DSDP Site 372 in the Balearic Basin (Foresi et al., 2003; Turco et al., 2003; Iaccarino et al., 2005; Abdul Aziz et al., 2008; Di Stefano et al., 2008). This good correspondence, especially when combined with the calcareous nannofossil data, indicates that the succession is continuous across the formation boundary. This is in agreement with field observations which did not reveal any evidence for a hiatus either. The observed succession of well-defined planktonic foraminiferal events around the proposed GSSP can be used to export the boundary to other marine sections in the Mediterranean. The proposed GSSP falls within the MMi5c Subzone (*G. praemendardii* - *G. peripheroronda* Subzone) of the Mediterranean zonal scheme of Di Stefano et al. (2008). In the open ocean, the GSSP falls within the standard low-latitude zones N10 (*G. peripheroacuta* Zone) of Blow (1969) and the (sub)tropical zone M7 (*G. peripheroacuta* Lineage Zone) of Berggren et al. (1995).

Magnetostratigraphy

A detailed paleomagnetic study of the Ras il Pellegrin section has been carried out by Abels et al. (2005). The natural remanent magnetization (NRM) intensity of the samples from the Globigerina Limestone was very weak and no reliable polarities were obtained. The Blue Clay showed much higher intensities and demagnetization reveals a clear subdivision into two components. In all the samples, the low-temperature (low-field) component is of normal polarity and represents viscous magnetite induced by the present-day field. The high-temperature (high-field) component is of dual polarity and was interpreted as the primary signal (ChRM). Plotting the ChRM directions resulted in a magnetostratigraphy for the Blue Clay part of the section which combined with the calcareous plankton biostratigraphy could be calibrated to the geomagnetic polarity time

Figure 5. Lithologic column, position and ages of main calcareous plankton bio-events, and magnetostratigraphy of the Ras il Pellegrin, and the calibration of the magnetostratigraphy to the geomagnetic polarity time scales of the ATNTS2004 (Lourens et al., 2004) and CK95 (Cande and Kent, 1995). "C" indicates presence of chondrite trace fossils (from Abels et al., 2005). Ages of bio-events partly based on Hilgen et al. (2003) and Abdul-Aziz et al. (2008).

scale and ranges from C5ACn up to C5Ar.2n (Fig. 5). This calibration reveals that the formation boundary between the Globigerina Limestone and the Blue Clay and, hence, the Serravallian GSSP falls within C5ACn. Unfortunately the uppermost part of the section studied did not produce a reliable magnetostratigraphy. The calibration is confirmed by the position of the *S. heteromorphus* L(C)O in C5ABr which is the same position as found at DSDP 42 Site 372 in the Balearic Basin (Abdul Aziz et al., 2008) and in the (adjacent) Atlantic Ocean (Backman et al., 1990; Olafsson, 1991).

Cyclostratigraphy and astrochronology

The Blue Clay at Ras il Pellegrin reveals a cyclic alternation on various scales. On a large-scale, six whitish coloured marly intervals are separated by intervals dominated by grey marls (Figs. 5 and 6). These intervals correspond to the large-scale cyclicity recognized by Sprovieri M. et al. (2002), John et al. (2003) and Abels et al. (2005). The small-scale cyclicity is less easy to distinguish in the field (Sprovieri M. et al., 2002; Abels et al., 2005). The Blue Clay part of

the section studied by Abels et al. (2005) contains 44 small-scale cycles with an approximate thickness of 1 m; these cycles can be recognized in the field and were labelled as a subdivision of the larger scale alternations (Fig. 5). Identification of both the large-scale and small-scale cycles was corroborated by geochemical data in particular Ca % and the Ca/K ratio (Fig. 6). First order age control provided by the calcareous plankton biostratigraphy and the magnetostratigraphy (Fig. 5) showed that the small-scale cycles fall within the precession frequency band of the spectrum and the large-scale cycles in the eccentricity band. Unfortunately the cycles themselves did not reveal sufficient characteristic detail to allow a straightforward tuning of the cyclicity to astronomical target curves (Sprovieri M. et al., 2002; Abels et al., 2005). The intercalation of chondrite trace levels and two sapropelitic layers suggests that the small-scale cycles are related to the sapropel and carbonate cycles normally found in deep marine sequences of the Mediterranean Neogene.

Sprovieri M. et al. (2002) used astronomical ages of a number of primary bio-events from Tremiti islands (Lirer et al., 2002) as starting

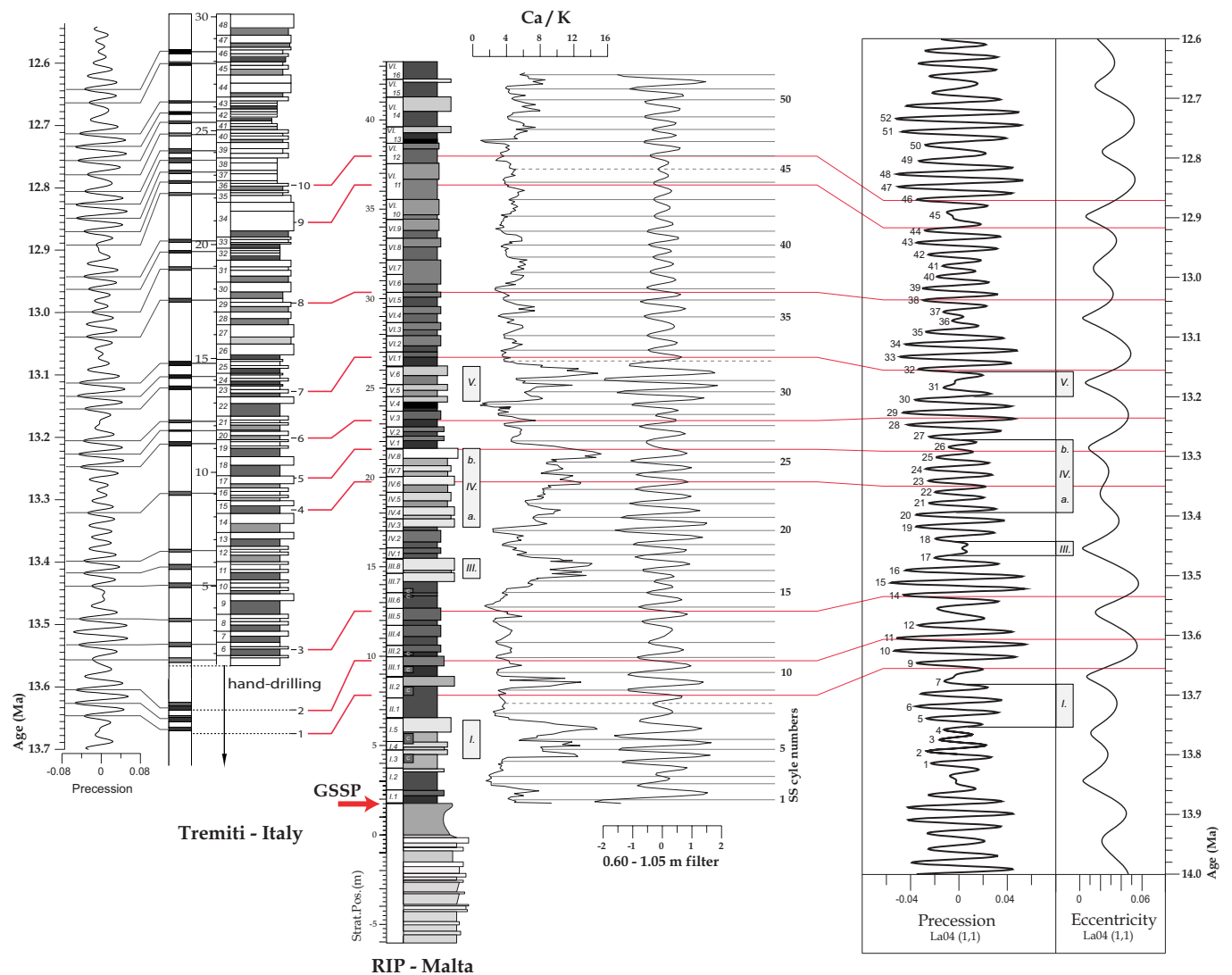


Figure 6. (Left) Tuning of red layers in the Tremiti section to precession (La2004_(1,1); Laskar et al., 2004). (Middle) Correlation of 10 bio-events between Tremiti and the Ras il Pellegrin section (solid lines): numbers 1 and 3 indicate the L(C)O of *S. heteromorphus* and LO of *G. peripheroronda*, respectively. (Right) Tuning of small-scale cycles in the Ras il Pellegrin section as also expressed in the Ca/K ratio and the filtered 0.6-1.05 m component in Ca/K to precession (La2004_(1,1)) (Modified after Abels et al. 2005).

point for their tuning to target curves derived from the La93 astronomical solution (Laskar, 1990; Laskar et al., 1993). They then employed cyclic variability in CaCO_3 content and *Globigerinoides* spp. as determined by spectral methods to establish a tuning to eccentricity and then to precession. Unfortunately the initial astronomical ages of the bio-events from Tremiti Islands (Lirer et al., 2002) proved to be incorrect due to complications in the stratigraphy (Hilgen et al., 2003). These problems were solved by studying different partial sections on Tremiti Islands and by incorporating Monte dei Corvi as parallel section (Hilgen et al., 2003). The latter section in combination with the adjusted cycle patterns on Tremiti islands were used to establish a more robust tuning and hence reliable astronomical age estimates for the calcareous plankton events (Hilgen et al., 2003; Abels et al., 2005). These improved ages were employed by Abels et al. (2005) as starting point for the tuning of the sedimentary cyclicity in the Ras il Pellegrin section, using the new numerical solution La2004 (Laskar et al., 2004). This tuning resulted in an age of 13.82 Ma for the formation boundary between the Globigerina Limestone and Blue Clay; the age of the base of the Transitional interval arrives at 13.86 Ma (Fig. 6). The tuning further points to a particular orbital configuration at times of the major shift in the Middle Miocene climate transition (see under stable isotopes).

Ar/Ar chronology

No radiometric age determinations are available for the Ras il Pellegrin section but a biotite containing ash layer has recently been discovered in the uppermost part of the Blue Clay in this section. However, an $^{40}\text{Ar}/^{39}\text{Ar}$ multigrain K-feldspar age of 13.81 ± 0.08 Ma was recently obtained for an ash layer at DSDP Site 372 in the western Mediterranean (Abdul Aziz et al., 2008); this age was calculated using an age of 28.02 Ma for the Fish Canyon sanidine dating standard. The ash layer is intercalated between the *H. walbersdorfensis* FCO below and the *S. heteromorphus* L(C)O above, and falls in the middle part of C5ACn. An astronomically calibrated age of 28.201 Ma (Kuiper et al., 2004; 2008) should be used for the FC sanidine for a direct comparison with the astronomical age of the GSSP, which results in a revised age of 13.90 Ma for the ash layer.

Sr-isotope stratigraphy

Sr-isotope data are available of authigenic phosphate peloids from the main phosphorite beds and carbonates of the Maltese islands (Jacobs et al., 1996). Using the regression of Hodell (1991), they range in age from 24.5 ± 0.74 Ma for the basal part of the Globigerina Limestone to 10.9 and 7.8 Ma for the Greensand and Upper Coralline Limestone, respectively. The Sr-isotope ages are in good agreement with the biostratigraphic ages, indicating that the samples are well preserved and that the Mediterranean and open ocean were well mixed with respect to the Sr-isotopes at that time (Jacobs et al., 1996).

Stable isotopes

Relatively low resolution benthic and planktonic stable isotope records were published from the Maltese succession by Jacobs et al. (1996). The data revealed a major excursion of +1 permille in $\delta^{13}\text{C}$ between 18 and 12.5 Ma, correlated to the Monterey carbon isotope excursion. This excursion precedes a benthic oxygen isotope shift to heavier values which started around 16 Ma; this shift is also recognized

in deep-sea records and is linked to the initiation of the Middle Miocene cooling associated with Antarctic ice build up (Woodruff and Savin, 1991; Jacobs et al., 1996). However the age model of Jacobs et al. (1996) is not correct because hiatuses evidenced by the calcareous plankton biostratigraphy around the two main phosphate pebble beds (Theodoridis, 1984) were not taken into consideration.

High-resolution isotope records of bulk carbonate were subsequently established for the (upper member of the) Globigerina Limestone and the Blue Clay in the Xatt-L' Ahmar and Ras il Pellegrin sections (Fig. 7) (John et al., 2003; Abels et al., 2005). These records revealed similar shifts to heavier $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values across the formation boundary correlated with the E3/Mi-3b and (transition to) CM6 of the global ocean, respectively. Planktonic and benthic isotope records of the Ras il Pellegrin section with a similar resolution are in progress.

The E3/Mi-3b event reflects rapid expansion of the Antarctic ice-sheet between 13.87 and 13.82 Ma (Fig. 7). The tuning suggest that this event and, hence, the Globigerina Limestone - Blue Clay formation boundary and GSSP is related to a prolonged interval of low seasonal contrast and cool Southern Hemisphere summers due to the combined effect of a prominent 1.2 Myr minimum in obliquity amplitude and 400- and 100-kyr minima in orbital eccentricity (Fig. 7) (Holbourn et al., 2005; Abels et al., 2005). Similar phase relations have been found for other major glacial oxygen isotope excursions in the Oligo-Miocene (Turco et al., 2001; Zachos et al., 2001; Wade and Pälike, 2004).

The abruptness and magnitude of the Mi-3b event are the more evident in high-resolution isotope records that have been established for different ODP sites located in different oceanic basins (e.g., Shevenell et al., 2004; Holbourn et al., 2005; Westerhold et al., 2005). In fact, the Serravallian GSSP as proposed at Ras il Pellegrin corresponds to the end of the isotope shift (see Abels et al., 2005; Fig. 7).

Sequence stratigraphy

Estimates for the glacio-eustatic sea-level lowering associated with the Mi-3b isotope event are in the order of ~60 m. It is therefore expected that the expression of this event is recognised in sequence stratigraphic records from all over the world. In fact it corresponds with sequence boundary Ser1 of Hardenbol et al. (1998) and supposedly with the TB2.5 sequence boundary of Haq et al. (1987). Mi-3b further correlates well with the hiatus between sequences Kw2c and Kw3 in the detailed sequence stratigraphic framework of the New Jersey passive continental margin and with the m3 reflector on the New Jersey slope, both of which were correlated to the Mi-3 and TB2.5 sequence boundary in the global cycle chart of Haq et al. (1987) by Miller et al. (1998). In this respect it is remarkable that the Mi-3b event does not coincide with a major sea-level fall in the global cycle chart of Haq et al. (1987; after time scale corrections) but with the relatively minor sequence boundary TB2.5, as suggested by Miller et al. (1998).

Correlation potential

Mediterranean and Europe

Integrated stratigraphic correlations of the Serravallian GSSP to other Mediterranean sections are straightforward and unambiguous. For this purpose both primary calcareous plankton events (S.

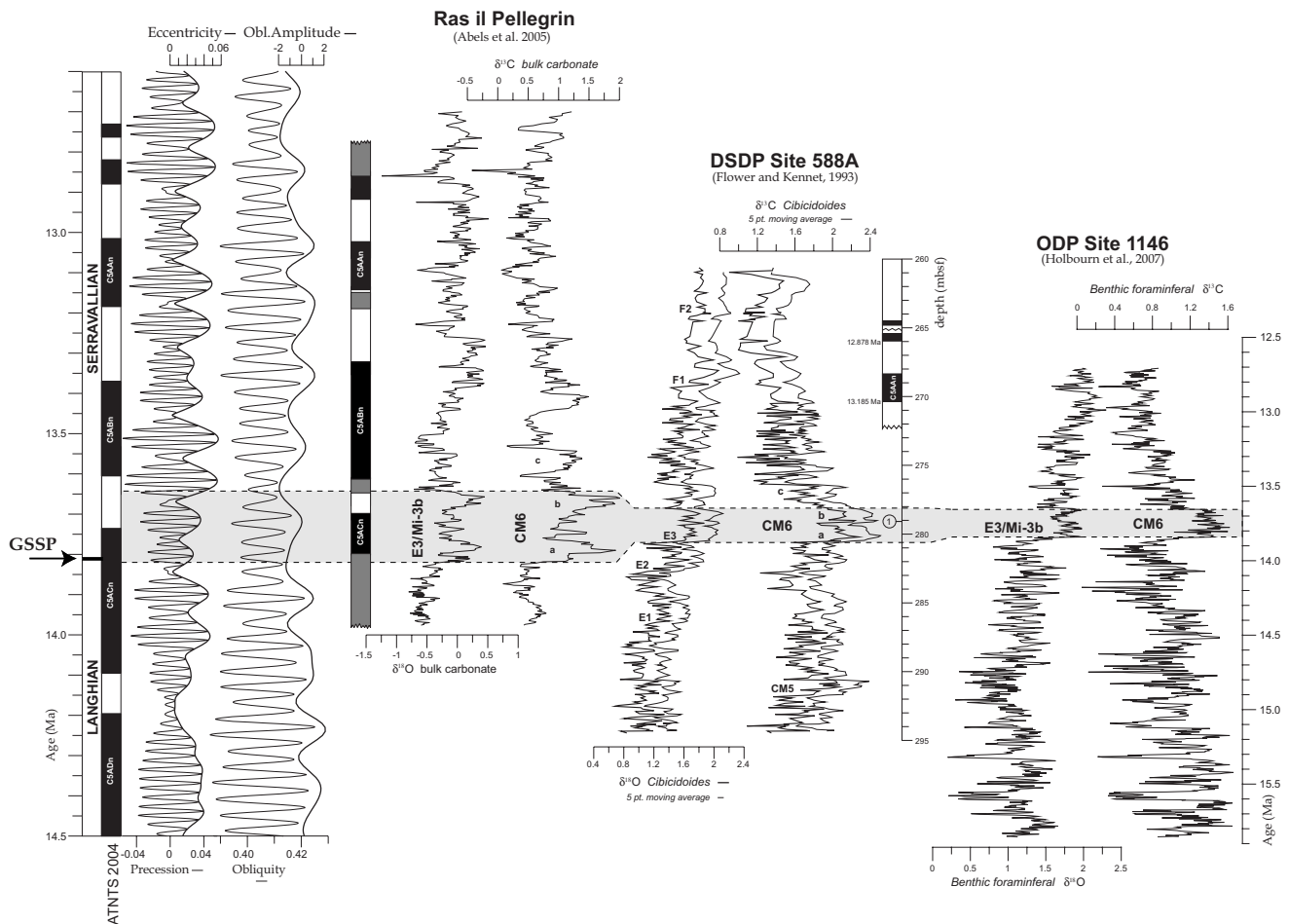


Figure 7. Magnetic polarity time scale of ATNTS2004, (Lourens et al., 2004) precession and its amplitude modulator eccentricity, obliquity and its amplitude modulator, the magnetostratigraphy of the Ras il Pellegrin section, and the tuned bulk carbonate isotope time series of that section and its correlation to the benthic isotope records of DSDP Site 588A (Flower and Kennett, 1993) and ODP Site 1146 (Holbourn et al., 2007). The record of Site 588A record is presented in the depth domain and of Site 1146 in the - astronomical-tuned - time domain. The magnetostratigraphy of Site 588A with corresponding ATNTS reversal ages are also indicated. Isotope events are labelled following Flower and Kennett (1993) and Miller et al. (1996). The shaded interval marks the CM6 isotope excursion to heavier values. Note that the Mi-3b event only refers to the main shift in the oxygen isotopes to heavier values at the base of CM6. The GSSP is thus defined at the end of this shift, which is the base of the grey shaded interval in this figure. (Modified after Abels et al. 2005).

heteromorphus L(C)O, *H. walbersdorfensis* FCO, *C. floridanus* LCO and *G. peripheroronda* LO), as well as secondary events such as the paracmes of *P. siakensis* and *G. cf. quinqueloba* AE can be used. The *O. universa* FO and *H. walbersdorfensis* FCO are considered reliable biostratigraphic marker events of regional importance for the Mediterranean Middle Miocene (e.g., Fornaciari et al., 1996; Rio et al., 1997; Raffi et al., 2006; Di Stefano et al., 2008).

As far as the continental record is concerned, the GSSP falls within zone MN6 and the Astaracian ELMA (European Land Mammal Age) in (central) Europe (Kempf et al., 1997). Diachroneity of faunal events plays an important role with regard to Spain where the GSSP coincides with the middle-late Aragonian and local zone E/F boundaries (Krijgsman et al., 1996).

Global

Stable isotope records of benthic foraminiferal carbonate especially in combination with a detailed magnetostratigraphy and calcareous plankton biostratigraphy provide the prime tool to

recognise the exact level of the boundary in the open ocean. The boundary coincides with oxygen isotope event Mi-3b of Miller et al. (1991; 1996) and E3 of Woodruff and Savin (1991), and carbon isotope excursion CM6 of Woodruff and Savin (1991); these events can readily be identified in stable isotope records from the open ocean marking the main shift associated with mid-Miocene cooling (e.g., Flower and Kennett, 1993). The abruptness and short duration (~50 kyr) of the Mi-3b isotope shift is particularly evident in high-resolution isotope records that have been generated from various ODP sites in different oceanic basins (e.g., Shevenell et al., 2004; Holbourn et al., 2005). The actual GSSP itself is defined at the end of the shift.

In the low-latitude open ocean the calcareous nannofossil events *S. heteromorphus* and *C. floridanus* LCOs occur above oxygen isotope event Mi-3b as in the Mediterranean, although the astronomical age for the *S. heteromorphus* L(C)O is slightly younger at Ceara Rise (13.523 Ma; Backman and Raffi, 1997) than in the Mediterranean (13.654 Ma; Abels et al., 2005). Nevertheless, the biostratigraphic reliability of the *S. heteromorphus* LO is well known on an almost global scale although the event has rarely been calibrated directly to

a reliable magnetostratigraphy and/or cyclostratigraphy. Regarding the astronomical calibration of planktonic foraminiferal events in the Equatorial Atlantic Ocean (in Lourens et al., 2004), the isotopic Mi-3b event occurs between the *G. peripheroacuta* FO and the *G. "praefohsi"* FO, very close to the *G. peripheroronda* LO, which is older than in the Mediterranean. The GSSP further falls within dinoflagellate zone D18 and radiolarian zone RN5 (Sanfilippo and Nigrini, 1998; see also Lourens et al., 2004). As far as diatoms are concerned, the GSSP falls within the *Denticulopsis simonsenii* Partial Range Zone in the northern part of the Atlantic sector of the Southern Ocean, in the *D. simonsenii* - *Nitzschia grossepunctata* Partial Range Zone in the southern part of the Atlantic sector of the Southern Ocean (Censarek and Gersonde, 2002) and in the *Coscinodiscus lewisianus* Zone in the (eastern) equatorial Pacific (Baldauf and Iwai, 1995)

The GSSP falls within the Barstovian NALMA (North American Land Mammal Age: Woodburne and Swisher, 1995; Alroy, 2002; Tedford et al., 2004) and coincides with the Colloncuran-Laventan SELMA (South American Land Mammal Age) (Fig. 8). Finally, the boundary falls within the younger part of Chron C5ACn (Fig. 8) (Abels et al., 2005). The associated major glacio-eustatic sealevel drop corresponds with sequence boundary Ser1 of Hardenbol et al. (1998) and supposedly corresponds with the TB2.5 sequence boundary of Haq et al (1987).

Conclusion

The formal definition of the base of the Serravallian represents an important next step towards the completion of the Standard Global Chronostratigraphic Scale for the Neogene. This scale is directly linked to the development of an astronomically dated integrated stratigraphic framework that underlies the standard Geological Time Scale for this interval of time.

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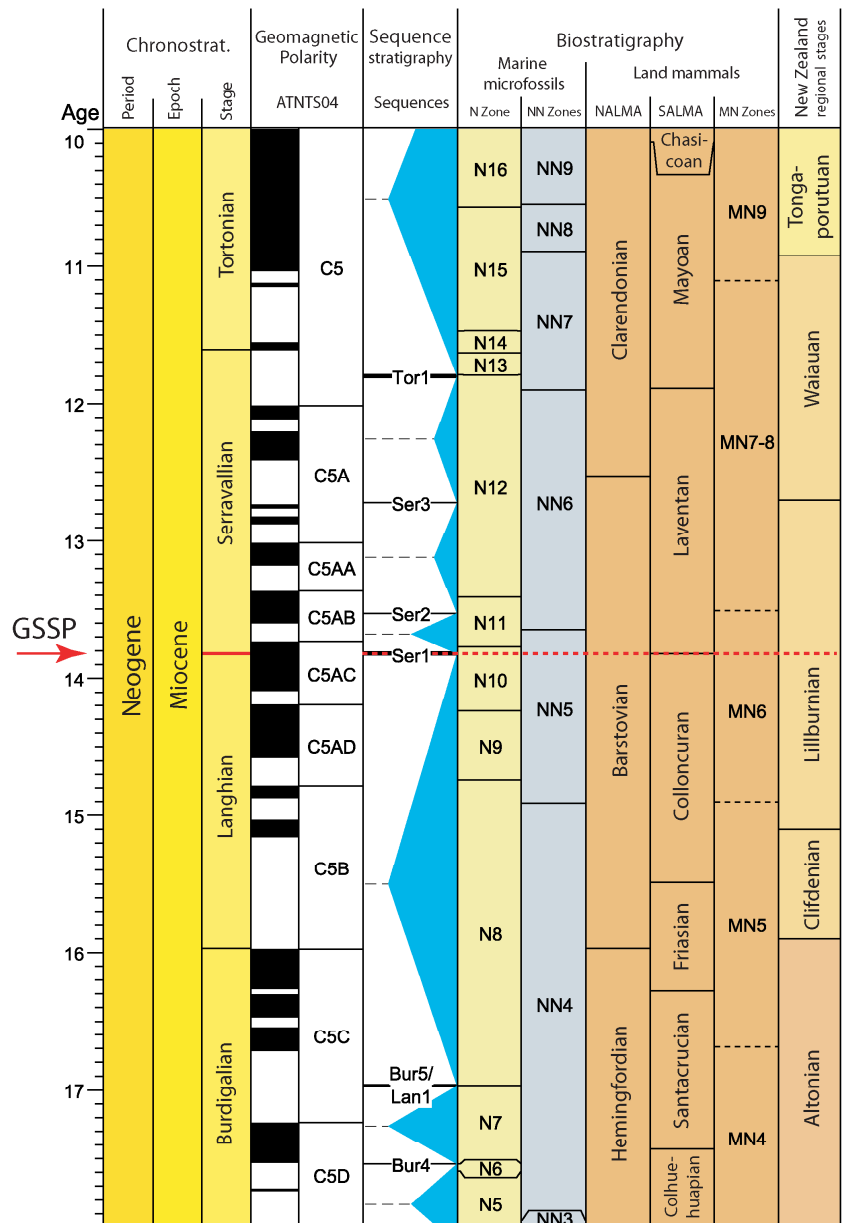


Figure 8. Chart showing the position and age of the Serravallian GSSP marked by a red arrow to the left relative to the geomagnetic polarity time scale (after Lourens et al., 2004), the sequence stratigraphic chart (after Hardenbol et al., 1998), the calcareous plankton N and NN zonal schemes, the North American and South American Land Mammal Ages (NALMA and SALMA, respectively) and the European MN Zonal scheme and the regional (st)ages of New Zealand. Ages of MN zonal boundaries are based on Central Europe following Kempf et al. (1997). Figure modified from a chart produced with Time Scale Creator program, see also caption to Figure 1.

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