CONTRIBUTION TO THE LATE TRIASSIC GEOCHRONOLOGY BY MAGNETOSTRATIGRAPHIC CORRELATIONS BETWEEN TETHYAN MARINE SECTIONS AND THE NEWARK APTS
(CONTRIBUTO ALLA GEOCRONOLOGIA DEL TRIASSICO SUPERIORE TRAMITE CORRELAZIONI MAGNETOSTRATIGRIFICHE TRA SEZIONI MARINE TETIDEE E IL NEWARK APTS)

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“Die Erde kann zu einer bestimmten Zeit durchaus nur ein Aussehen
gehabt haben. Direkte Auskunft hierüber gibt sie nicht. Wir stehen ihr
gegenüber wie der Richter gegenüber einem Angeklagten, der jede
Auskunft verweigert, und haben die Aufgabe, die Wahrheit auf dem
Wege des Indizienbeweises zu ermitteln.”

“At a specified time the Earth can have just one configuration. But
the Earth supplies no direct information about this. We are like a
judge confronted by a defendant who declines to answer; and we must
determine the truth from the circumstantial evidence.”

(Alfred Wegener, 1920, Die Entstehung der Kontinente und
Ozeane/The Origin of Continents and Oceans)

“It strikes me that all our knowledge about the structure of our Earth
is very much like what an old hen would know of the hundred-acre
field in a corner of which she is scratching.”

(Charles Darwin, 1831, Letter to William Darwin Fox)
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THESIS OUTLINE

This PhD thesis is subdivided in six chapters.

**Chapter I** is the state of the art of the Late Triassic chronology. In this part I present the chronostratigraphic organization of the Late Triassic, the history of the Stages (Carnian, Norian and Rhaetian), a detailed explanation of the last options of Late Triassic geochronology, and the aim of my research.

**Chapter II** represents an introduction to the concepts of paleomagnetism (geomagnetism, magnetic properties of materials, etc.) and to the methods of analysis (instruments, techniques, evaluation of data reliability, etc.). I present the geological setting of the investigated areas in **Chapter III**, focused mainly on the Late Triassic, and a brief introduction to the stratigraphic sections studied for magnetostratigraphy.

**Chapter IV** is the discussion of the paleomagnetic data obtained during my PhD. I subdivided the chapter by the considered time intervals: Rhaetian and Carnian Stages. Within these subchapters, I report the results for each Tethyan marine section investigated, with the detailed description of litho- and biostratigraphy, the paleomagnetic data and their interpretation, and the discussion about the magnetostratigraphic correlations here proposed.

In **Chapter V**, I propose an updated Geomagnetic Polarity Time Scale (GPTS) for the Late Triassic, based on literature and on the magnetostratigraphic correlations between Tethyan marine sections and the Newark APTS presented in Chapter IV. In this Chapter, I propose new ages for the Stage boundaries derived from the GPTS.

**Chapter VI** is the conclusive part of the thesis. It includes a part regarding the magnetostratigraphy of the studied Tethyan marine sections, subdivided as Chapter IV (Rhaetian and Carnian subchapters), and a part about the proposal of GPTS.
ABSTRACT

Chronology of Late Triassic (last Epoch of Triassic Period) is still a debated question. Late Triassic is constrained by two U/Pb ages, one near the Ladinian/Carnian boundary (237.773±0.052 Ma; Alpe di Siusi/Seiser Alm, Italy) and the other at the Rhaetian/Hettangian boundary (201.36±0.17 Ma; Levanto, Peru). Unfortunately, any radiometric age constraints the Stage boundaries of Late Triassic. Many attempts to assign an age to the Stages have been made during the last 20 years, correlating marine sections (usually from Tethys) with the Newark Astrochronological Polarity Time Scale (Newark APTS). The ages obtained was sometimes very different, in particular for the Rhaetian, with a duration that varied from ~2 My to ~9 My depending from the correlation performed with the APTS. The options proposed in the Geological Time Scale 2012 introduced two different ages for both Rhaetian (~205.4 Ma and ~209 Ma) and Norian (~221 Ma and ~228 Ma). The Norian age of ~228 Ma seems coherent with many other correlations between marine sections and the APTS. In an effort to help resolving the issues of the Late Triassic chronology, selected Tethyan marine sections, characterized by a detailed biostratigraphy, have been analyzed for paleomagnetism. The investigation are focused on two main intervals: the Rhaetian and the Carnian. The Rhaetian have been chosen for the reasons explained before (confused chronology), the Carnian because the few magnetostratigraphic data covering this interval, in particular its middle part, must be integrated to obtain a continuous magnetostratigraphy of this Stage. The chosen sections are: Pignola-Abriola, Mount Messapion and the ODP Leg 122-Hole 761C for the Rhaetian; Pignola-2, Dibona, and ODP Leg 122-Holes 759B/760B for the Carnian. The magnetostratigraphy of these sections have been integrated with the data from other Tethyan sections in literature, obtaining a continuous magnetostratigraphy spanning the entire Late Triassic. This composite magnetostratigraphy of Tethyan section has been time-calibrated using the Newark APTS, linked to the composite through the statistical correlations with Pignola-2, Pizzo Mondello, and Pignola-Abriola. The so obtained Geomagnetic Polarity Time Scale (GPTS) has been used to assign an age to the events calibrated to the magnetostratigraphy of the Late Triassic, like the bioevents defining the Stage and substage boundaries, or climatic events as the Carnian Pluvial Event.
La cronologia del Triassico Superiore (l’ultima Epoca del Periodo Triassico) è attualmente materia di dibattito. Il Triassico Superiore e vincolato da due età radiometriche U/Pb, una in prossimità del limite Ladinico/Carnico (237.773±0.052 Ma; Alpe di Siusi/Seiser Alm, Italia) e l’altra al limite Retico/Hettangiano (201.36±0.17 Ma; Levanto, Perù). Purtroppo, nessuna altra età radiometrica vincola direttamente gli altri Piani del Triassico Superiore. Negli ultimi 20 anni sono stati fatti numerosi tentativi di assegnare delle età ai Piani, correlando sezioni marine (solitamente della Tetide) con il Newark Astrochronological Polarity Time Scale (APTS). Le età ottenute sono state talvolta molto diverse, in particolare per il Retico, con una durata variabile da ~2 My a ~9 My a seconda della correlazione con l’APTS. Le opzioni proposte nella Geologic Time Scale 2012 hanno proposto due età diverse sia per il Retico (~204.5 Ma e ~209 Ma) che per il Norico (~221 Ma e ~228 Ma). L’età di ~228 Ma per il Norico sembra coerente con molte altre correlazioni tra sezioni marine e l’APTS. Nel tentativo di contribuire alla risoluzione dei problemi della cronologia del Triassico Superiore, sono state effettuate delle analisi paleomagnetiche su una serie di sezioni marine Tetidee caratterizzate da una biostratigrafia dettagliata. Le indagini sono focalizzate su due principali intervalli di tempo: il Retico e il Carnico. Il Retico è stato scelto per le ragioni spiegate precedentemente (cronologia confusa), il Carnico perché i pochi dati di magnetostratigrafia, in particolare nel Carnico medio, devono essere integrati per ottenere un record completo di questo Piano. Le sezioni scelte per le analisi sono: Pignola-Abriola (Italia), Monte Messapion (Grecia) e Leg 122-Hole 761C dell’ODP (Australia) per il Retico; Pignola-2 (Italia), Dibona (Italia) e Leg 122-Hole 759B/760B dell’ODP (Australia) per il Carnico. I dati di magnetostratigrafia di queste sezioni sono stati integrati con dati da altre sezioni Tetidee in letteratura, ottenendo una magnetostratigrafia continua dell’interno Triassico Superiore. Tale composita è stata calibrata con la Newark APTS, usando come collegamento le correlazioni statistiche tra l’APTS e le sezioni di Pignola-2, Pizzo Mondello e Pignola-Abriola. La Scala-Tempo delle Polarità Geomagnetiche (Geomagnetic Polarity Time Scale – GPTS) così ottenuta è stata usata per assegnare un’età agli eventi calibrati con la magnetostratigrafia del Triassico Superiore, come ad esempio i bioeventi che definiscono i limiti dei Piani e dei sottopiani, oppure gli eventi climatici come il Carnian Pluvial Event.
Chapter I
STATE OF THE ART

1.1 LATE TRIASSIC
Late Triassic is the last Epoch of the Triassic Period. It lasted ~36 My and represents the 70% of the entire Triassic. It is subdivided in three Stages: Carnian, Norian and Rhaetian (as recognized by the Subcommission on Triassic Stratigraphy) (Fig. 1.1). The Carnian has been divided in two substages: Julian and Tuvalian, whereas the Norian has three substages: Lacian, Alaunian and Sevatian (Fig. 1.1).

Carnian, Norian and Rhaetian (and other Triassic Stages) have been originally defined with ammonoid-rich successions of the Northern Calcareous Alps (e.g. Mojsisovics, 1869), in the tentative to unbind the Triassic to the classical definition of von Alberti (1834). The original Triassic was defined on three subsequent continental and shallow-marine German succession: the Buntsandstein, the Muschelkalk and the Keuper. These three formations were diffused in all the southern Germany, but they are difficult to correlate beyond this area. This is why the ammonoid-defined Triassic Stages were preferred to the old definition of the Period.

The base of the of the Carnian Stage (base of the upper Triassic Series) is defined by a GSSP (Global Stratotype Section and Point) located in the Prati di Stuores/Stuores Wiesen section (Italy, Broglio Loriga et al., 1999; Mietto et al., 2012). The age of the Carnian base is placed at ~237 Ma, confirmed by the U/Pb age obtained from an ash-bed in Seiser Alm/Alpe di Siusi area of 237.773±0.052 Ma (Mietto et al., 2012), biostratigraphically constrained in upper Ladinian (”Anolcites” neumayiri/”Frankites” regoledanus ammonoid subzones of Protrachyceras Zone).

The upper limit of the Late Triassic is defined by the GSSP for the base of the Jurassic System (Hettangian Stage) in the Kuhjoch section (Austria, Hillebrandt et al., 2013). The base of Hettangian is dated at 201.36±0.17 Ma (U/Pb) from ash-beds around the FO of Psiloceras spelae in northern Peru (Schoene et al., 2010; Guex et al., 2012; Wotzlaw et al., 2014). The bases of the Norian and the Rhaetian Stages, instead, still not have a ratified GSSP. The candidates of GSSP for the Norian are the Pizzo Mondello (Italy; Muttoni et al., 2004) and the Black Bear Ridge sections (Canada; Orchard et al., 2007), and for the Rhaetian are the Steinbergkogel (Austria; Krystyn et al., 2007a, 2007b) and the Pignola-Abriola sections (Italy; Rigo et al., 2015).
Here follows a detailed description of the Late Triassic Stages and their history.

### 1.1.1 Carnian

The Carnian was named after the Kärnten (Carinthia) region of Austria, or from the near Carnian Alps, and it was originally associated to the Hallstatt Limestone beds bearing *Trachyceras* and *Tropites* ammonoids (Mojsisovics, 1869). The base of the Carnian was traditionally associated with the FO of ammonoids *Trachyceras* (*T. aon* in Tethys or *T. desatoyense* in North America), although the presence of these ammonoids is asynchronous and not global (e.g. Mietto and Manfrin, 1999). Broglio Loriga et al (1999) proposed the levels recording the FAD of ammonoid *Daxatina canadensis* at the Prati di Stuores/Stuores Wiesen locality (Dolomites, Italy) as the GSSP of the Carnian, then ratified in 2008 (Mietto et al., 2012). Other markers are the FO of conodont *Paragondolella polygnathiformis* and palynomorphs *Patinasporeites densus* and *Vallasporites ignacii*. The Carnian GSSP at Prati di Stuores is just above the base of a normal magnetozone (S2n in Broglio Loriga et al., 1999; Mietto et al., 2012) and a maximum flooding surface within sequence Car1 (*sensu* Gianolla et al., 1998a; named Lad3 in Hardenbol et al., 1998), recognized at least in Tethyan basins (Hardenbol et al., 1998). The Carnian Stage has been subdivided in three substages (Mojsisovics et al., 1895): Cordevolian, Julian and Tuvalian. In recent times, the Cordevolian has been normally included as lower

<table>
<thead>
<tr>
<th>Late Triassic</th>
<th>Age</th>
<th>Sub-Age</th>
<th>Tethyan Ammonoids</th>
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<tbody>
<tr>
<td>Hettangian</td>
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<td><em>Plioceras spelas</em></td>
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<td><em>Choristoceras marathi</em></td>
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<td><em>&quot;Choristoceras&quot; haureri</em></td>
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<td><em>Metaspathites spinosens</em></td>
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<td><em>Mesohimavaleites columbiae</em></td>
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<td><em>Cyrtopleurites bicornatus</em></td>
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<td><em>Juavrites magnus</em></td>
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<td><em>Guemdeites jandalus</em></td>
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<td><em>Anatropites spinosus</em></td>
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<td><em>Tropites subboulus</em></td>
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<td><em>Tropites dilleri</em></td>
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<td>Tuvalian</td>
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<td><em>Austrochryseras australicum</em></td>
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<td><em>Trachyceras aanoi</em></td>
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<td><em>Trachyceras aon</em></td>
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<tr>
<td>Carnian</td>
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<td><em>Daxatina canadensis</em></td>
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<td><em>Frankeites regoidanus</em></td>
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<td>Julian</td>
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<td>Ladinian</td>
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Julian. The boundary between Julian and Tuvalian is usually placed with the FO of ammonoids *Tropites* (*T. subullatus* in Tethys, *T. dilleri* in North America) (Fig. 1.1). This boundary is characterized by an important biotic turnover that includes ammonoids, conodonts and radiolarians (Tozer, 1984; Simms and Ruffell, 1989; Kozur and Bachmann, 2010; Mazza et al., 2010). A climate change occurs at the end of the Julian, known as Carnian Pluvial Event (Simms and Ruffel, 1989), Reingraben turnover (Schlager and Schöllnberger, 1974), or Middle Carnian Wet Intermezzo (Kozur and Bachmann, 2010). The decrease in carbonatic fraction in the oceanic basins and the deposition of shales in restricted basins have been interpreted as a rise of the Calcite Compensation Depth (Rigo et al., 2007). The CPE is intended as global, probably triggered by both paleogeographic/oceanographic variations, as well as volcanic events (emplacement of Wrangellia Large Igneous Province; Furin et al., 2006; Dal Corso et al., 2012).

**1.1.2 Norian**

The name Norian derives from the Roman province of Noria, including the area of Hallstatt (Austria), where Mojsisovics (1869) defined the Stage with strata containing the ammonoid *Pinacoceras metternichi*. Originally, Mojsisovics considered the Norian as older than Carnian, but he recognized the error and change the name of Norian in Juvavian, and calling “Norian” the Hallstatt limestones below the Carnian strata (Mojsisovics, 1892). This created confusion, and Bittner (1892) propose to maintain the orginal name Norian for the strata younger than Carnian and to refer to the older strata as Ladinian.

In North America the beginning of the *Stikinoceras kerri* ammonoid Zone is considered coeval to the base of the Norian (in Tethys is approximated by the *Guembelites jandianus* Zone; Fig. 1.1). The FO of conodont *Metapolygnathus* ex gr. *M. echinatus* approximates the base of the *S. kerri* Zone (Orchard, 2010), and is also approximated with the FO of the bivalves *Halobia austriaca* and *Halobia beyrichi* (McRoberts, 2007). The FO of *M. echinatus* and *H. austriaca* appear associated also in Tethys (e.g. Balini et al., 2010). GSSP candidates for the Norian Stage are Pizzo Mondello in Italy (Muttoni et al., 2001, 2004; Nicora et al., 2007; Balini et al., 2010; Mazza et al., 2012) and Black Bear Ridge in British Columbia (Orchard et al., 2001; Orchard, 2007; McRoberts, 2007). In both the candidate stratigraphic sections, the GSSP corresponds to the level bearing the FAD of conodont *Metapolygnathus echinatus* and is approximated by the FO of bivalve *Halobia austriaca*. In Pizzo Mondello section the Norian base is approximated also by the Last Occurrence
Halobia lenticularis (Balini et al., 2010). Pizzo Mondello section is accompanied by magnetostratigraphy, with the base of the Norian corresponding with the top of a reverse magnetozone (PM4r; Muttoni et al., 2001, 2004) and just above a positive $\delta^{13}$C$_{\text{carb}}$ shift (Nicora et al., 2007; Muttoni et al., 2014). Norian is subdivided in three sub-stages: Lacian, Alainian and Sevatan. The Lacian/Alainian boundary is usually defined with the base of the Cytropleurites bicrenatus ammonoid Zone in Tethys (Fig. 1.1). The Alainian/Sevatan boundary corresponds to the beginning of the Gnomohalorites cordilleranus ammonoid Zone in North America and Sagenites quinquepunctatus ammonoid Zone in Tethys (Fig. 1.1). The Sevatan creates problems in defining the base of the Rhaetian. In Tozer (1994) the Rhaetian was not recognized and the Sevatan included the ammonoid Zones traditionally attributed to Rhaetian (Choristoceras marshi and Ch. haueri Zones). In Krystyn et al. (2007b) the ammonoid Zones Paracochloceras suessi/Sagenites reticulatus, usually referred to upper Sevatan (uppermost Norian), have been included in the Rhaetian, concurrent with the FO of conodont Misikella posthernsteini.

1.1.3 Rhaetian

The Rhaetian Stage has been defined first by von Gümbel (1859) as “Rhätische Gebilde”, with the first occurrence of bivalve Rhaetivcula contorta in the Kössen Beds (Austria). The name comes both from the Rhätische Alps and the Roman province of Rhaetium. The Rhaetian was included as the last Stage of the Late Triassic by Mojsisovics et al. (1895), defined as Avicula contorta Zone.

Since then, the Rhaetian has been recognized as a Stage (e.g. Pearson, 1970; Ager, 1987), considered a substage of the Norian (e.g. Tozer, 1967; Silberling and Tozer, 1968; Zapfe, 1974; Palmer, 1983) or the first Stage of the Jurassic (with the name of Bavarian; Slavin, 1961, 1963). However, most of the specialists in Triassic stratigraphy (e.g. Kozur and Mock, 1974; Gazdicki et al., 1979; Krystyn 1980, 1990) considered the Rhaetian a Stage of the Late Triassic. In Zapfe (1983) is presented the last report of the International Geoscience Programme (IGCP) 4, in which the Rhaetian is formally included in the Triassic timescale, but only for the Tethyan realm. Only in 1991 the Rhaetian was officially considered a Stage by the Subcommission on Triassic Stratigraphy (STS). Although the Rhaetian is now widely accepted, it lacks of a formal marker defining its base. Kozur (1973) place the base of the Rhaetian with the first appearance (FAD) of the conodont Misikella posthernsteini, well distributed in the Tethys but rare in the western margin of Pangea (i.e. North America). The FAD of M. posthernsteini is considered almost
coeval with the *Proparvicingula moniliformis* radiolarian Zone (e.g. Kozur, 2003; Giordano et al., 2011; Rigo et al., 2015). Following this last correlation between conodonts and radiolarians, the FAD of *M. posthernsteini* in North America is younger than in Tethys, and the Norian/Rhaetian boundary is approximated by the FAD of the conodont *Epigondolella mosheri* morphotype A (Carter and Orchard, 2007). Recently, the Task Group for Rhaetian Stage placed the base of the Rhaetian with the First Occurrence (FO) of *M. posthersteini* (Krystyn, 2010), also suggesting the following secondary markers:

- the FO of the ammonoid *Paracochloceras suessi* and genus *Cochloceras* (Fig. 1.1)
- the Last Occurrence (LO) of ammonoid genus *Metasibirites* (Fig. 1.1)
- the FO of conodont *E. mosheri* morphotype A
- the FO of the radiolarian *P. moniliformis* and the beginning of the *P. moniliformis* Zone
- the LO of large *Monotis* bivalves, except for the dwarf species (only in Tethys; McRoberts et al., 2008)
- just below a prominent change of magnetic polarity after a long prominent normal magnetozone to a thinner reversal (as seen in many sections of Tethys).

Although ammonoid biostratigraphy is still considered one of the best tools for high-resolution geochronology and correlations in the Triassic (Balini et al., 2010), Norian and Rhaetian ammonoids are documented mostly locally. Conodonts and radiolarians are instead more common and normally used for correlations and geochronological definitions in the Triassic. The use of *Misikella posthernsteini* as a marker of the Norian/Rhaetian boundary (NRB) is widely used, in particular in Tethyan sections. This bioevent has been identified in both the GSSP candidates for the Rhaetian Stage (Steinbergkogel section, Austria; Pignola-Abriola section, Italy).

The Steinbergkogel section (Krystyn et al., 2007a, 2007b) is a pelagic succession of condensed limestone with conodonts, ammonoids and bivalves. The section was named STK-A to differentiate it to the parallel section STK-B+C. In Steinbergkogel three possible markers have been proposed: FO of *Misikella hernsteini* (Krystyn et al., 2007a, 2007b), corresponding to the FO of *Mockina mosheri* morphotype A and close to the FO of ammonoids *Tragorhacorecas* and *Rhaetites*; FAD of *Misikella posthernsteini* (Krystyn et al., 2007a, 2007b), coeval to the base of *Proparvicingula moniliformis* radiolarian Zone and the FO of *Paracochloceras*
suessi; FAD of ammonoid Vandaites (Krystyn et al., 2007b), in association with FO of Cyclocellites and Choristoceras and LO of Sagenites, Dionites and Pinacoceras. The Steinbergkogel section is accompanied by magnetostratigraphy in both STK-A and STK-B+C.

The Pignola-Abriola section (Rigo et al., 2015) is a pelagic succession of cherty limestones with conodonts, radiolarians and bivalves. The section is accompanied by a detailed magnetostratigraphy (presented here in Chapter 4.1.1 and in Maron et al., 2015) and δ^{13}C_{org} chemostratigraphy (Rigo et al., 2015). The identification of the NRB in Pignola-Abriola section have been proposed with a major negative δ^{13}C_{org} spike (~30%) occurring ~0.5 m below the FAD of Misikella posthernsteini and the beginning of Proparvicingula moniliformis radiolarian Zone.

1.2 LATE TRIASSIC GEOCHRONOLOGY

The Late Triassic is chronologically constrained at its lower (Ladinian/Carnian) and upper boundary (Rhaetian/Hettangian). In uppermost Ladinian a U/Pb age of 237.773±0.052 Ma (Mietto et al., 2012) was obtained from an ash-bed in Rio Nigra section (Alpe di Siusi/Seiser Alm), biostratigraphically constrained in upper Longobardian (Protrachyceras ammonoid Zone, “Anolcites” neumayri/“Frankites” reoledanus subzones), the last substage of the Ladinian. Using this biostratigraphic data, Mietto et al. (2012) approximated the age of the Ladinian/Carnian boundary at ~237 Ma. The age assigned to the Rhaetian/Hettangian boundary (201.31±0.18 Ma) was obtained from U/Pb datings (Schoene et al., 2010; Guex et al., 2012) in ash-beds around the FO of ammonoid Psiloceras spelae (considered a marker of the Hettangian base; e.g. Guex et al., 2004; Morton and Hesselbo, 2008; Hillebrandt et al., 2013), in a section near Levanto (northern Peru). This age has been recalculated using new tracer calibration by Wotzlaw et al. (2014), obtaining an age for the Rhaetian/Hettangian boundary of 201.36±0.17 Ma. Thus, the base and the top of the Late Triassic are well constrained by radiometric ages, and the correspondent GSSPs, which are the base of the Carnian and base of the Hettangian stages, have been ratified. Instead, the bases of the Norian and the Rhaetian lack of a clear definition. In facts, no radiometric datings (U/Pb, Ar/Ar, etc.) have been provided for those sections characterized by clear biostratigraphic markers of the Stage boundaries, with the only exception of the U/Pb age of 230.91±0.33 Ma, obtained from the Aglianico ash-bed in the Pignola-2 section (Italy, Furin et al., 2006). Hence, there were no possibilities to time calibrate Stages and substages, until in the 1990s the detailed magnetostratigraphic study of the
Newark Basin sedimentary sequence (eastern USA) by Kent et al. (1995) paved the way to a new approach in defining the chronology of the Late Triassic.

1.2.1 The Newark Astrochronological Polarity Time Scale (Newark APTS)
The Newark Supergroup is a continental sedimentary/volcanic sequence deposited during the Late Triassic/earliest Jurassic (Cornet and Olsen, 1985) in a series of rift basins in eastern North America, originated during the initial rift of Pangea. The Newark sedimentary sequence is partially exposed, and most of the sequence comes from a series of cores drilled in different part of the Newark Basin, the most extended rift basin of the Newark Supergroup. The Late Triassic portion is represented by lacustrine/alluvial sediments, organized in three formations: Stockton Fm., Lockatong Fm., and Passaic Fm (Olsen, 1980; Olsen et al., 1996) (Fig. 1.2). The Stockton Fm is dominated by fluvial sequences, with brownish and red conglomerates and arkoses, passing to mudstone in the upper part. The Lockatong Fm is instead typical of lacustrine environment, with prevalent gray mudstone. The Passaic Fm is characterized by red mudstone with sandstone and conglomerate, indicating a filling of the lacustrine basin. Van Houten (1964, 1969, 1980) recognized a hierarchy of lacustrine cycles in the Lockatong Fm, ascribing them to astronomical control of climate. Olsen (1986) recognized these cycles also in the Passaic Fm. Four types of cycles have been identified: Van Houten cycle (~20 kyr; precession cycle), short modulating cycle, McLaughlin cycle (~404 kyr; eccentricity cycle), long modulating cycle. The McLaughlin cycles (McLaughlin, 1933) have the strongest expression in the sedimentary sequence and they have been considered as stratigraphic members (e.g. Kent et al., 1995; Olsen et al., 1996). Later, at the 52 McLaughlin cycles of the Lockatong and Passaic Fms (Olsen et al., 1996), 8 new cycles of the same type have been identified in the upper Stockton Fm (Raven Rock Member) (Kent and Olsen, 1999; Olsen and Kent, 1999) (Fig. 1.2). The Late Triassic sedimentary sequence of the Newark Supergroup is topped by three basalt flows, Hettangian in age (Orange Mt. Basalts, Preakness Basalts, Hook Mt. Basalts) (Fig. 1.2), alternated with two sedimentary formations (Feltville and Towaco Fms), and overlaid by the sedimentary Boonton Fm. The basalt flows have been ascribed to the emplacement of the CAMP – Central Atlantic Magmatic Province (Marzoli et al., 1999, 2004).

Kent et al. (1995) performed a detailed magnetostratigraphic investigation on the cores from the Newark Basin stratigraphic sequence, covering more than 5000 m from the Feltville Fm to the base of the Stockton Fm. The final composite of Newark
magnetostratigraphy included 45 magnetozones from E1r to E23n, where E23n have been extended to Boonton Fm following the magnetostratigraphy of the Newark outcrops of Witte et al. (1991). Later, Kent and Olsen (1999) and Olsen and Kent (1999) considered the E23n.2r as a magnetozone, changing its name to E23r. As a consequence, the normal magnetozone above E23r was named E24n (Fig. 1.2). The

Figure 1.2: Newark Astrochronological Polarity Time Scale (APTS). From the left: formations and lithostratigraphy, cyclostratigraphy (McLaughlin cycles, 404 ky), magnetostratigraphy, paleolatitude of the cores, vertebrates and palynomorphs biostratigraphy. (from Olsen and Kent, 1999)
Newark magnetostratigraphy has been calibrated with the cyclostratigraphy by Kent et al. (1995), and then refined by Kent and Olsen (1999) and Olsen and Kent (1999), obtaining an Astrochronological Polarity Time Scale (APTS: or Astronomically-calibrated geomagnetic Polarity Time Scale, following the nomenclature of Olsen et al., 2011). Considering that the cyclostratigraphy starts from the Rock Raven Mb (upper Stockton Fm), the time calibration of the remaining part of the Stockton Fm was approximated by extrapolation of the average accumulation rate of the Raven Rock Mb (Olsen and Kent, 1999).

The potential of the Newark APTS as a geochronological tool is huge. Fundamentally is a Geomagnetic Polarity Time Scale (GPTS), but the biostratigraphic-based chronostratigraphy is not clearly defined. In the Newark APTS, palynomorphs assemblages (e.g. Cornet, 1977, 1993; Fowell et al., 1994) are the mainly markers for the Stage boundaries, in association with conchostracans (Kozur and Weems, 2005, 2007, 2010). A great problem of palynomorphs is provincialism, as argued by Cornet (1993), Cornet and Olsen (1985), and Fowell and Olsen (1993). For example, taxa typical of Tethys are absent in parts of the boreal realm, regardless of age (Hochuli et al., 1989; Mørk et al., 1992; Hochuli and Vigran, 2010). Olsen et al. (2011) justify the provinciality of palynomorphs with the different climates between Newark area (tropical arid) and the alpine and central Europe (tropical humid to temperate), at least for Norian and Rhaetian, also recalling the provincialism of *Rhaetipollis germanicus*, typical of Rhaetian in Europe and extended to early Carnian in Siberia (likely corresponding to geographic North Pole in Late Triassic; e.g. Kozur and Weems, 2007). About conchostracans, the definition of Stages based on this continental fossils is limited by the lack of relations with marine fossils marking the same events, and normally considered as the main markers for Stage boundary definition due to their greater areal extension (e.g. the markers of Rhaetian Stage proposed by STS do not include continental taxa; Krystyn, 2010).

**1.2.2 Newark APTS vs. marine sections: solving the Late Triassic geochronology**

These problems restrict the direct use of the Newark APTS as GPTS, so since 1995 many authors have attempted a correlation between marine sections (in which the chronostratigraphy is well defined by fossils or other events) and the APTS. Obviously, correlating a marine section with a continental section using biostratigraphy is a hard endeavor, in particular if palynomorphs are excluded because of their provinciality. An easy way out is offered by the magnetostratigraphy, which potentially allows global correlations. However, the numerous attempts of magnetostratigraphic correlation
between marine sections (mostly of Tethyan realm) and the Newark APTS led to
different (sometimes completely discordant) interpretations of the Late Triassic
geochronology. Gallet et al. (2000) reconstructed a composite magnetostratigraphy
of the Norian in Tethys, combining the magnetostratigraphy of Kavalaani (Turkey,
Gallet et al., 2000), Kaur Tepe (Turkey, Gallet et al., 1993), Scheiblkogel (Austria,
Gallet et al., 1996), and Bolücektasi Tepe (Turkey, Gallet et al., 1992) sections.
This Tethyan composite, calibrated with conodonts, have been visually compared
with the Newark APTS, but any correlation was performed for a lacking of part of
the Alaunian magnetostratigraphy and for the condensation of the considered
marine sections, which can also preclude the detection of small reversals, as stated
by Gallet et al. (2000). The earliest true correlation attempts between a Tethyan
marine section and the Newark APTS was performed by Muttoni et al. (2001), with
the basinal section of Pizzo Mondello (Italy), and by Chennell et al. (2003), with
the Silická Brezová section (Slovakia). In Muttoni et al. (2001) the Norian base was
not clearly defined in Pizzo Mondello, and the Carnian/Norian boundary (CNB)
was placed inside the E14r-E15r interval of the APTS, around ~213-216 Ma, near
(2002) integrate the composite Tethyan magnetostratigraphy by Gallet et al. (2000)
with the magnetostratigraphy of Pizzo Mondello (Muttoni et al., 2001) to cover
the Tuvalian missing in Bolücektasi Tepe (Gallet et al., 1992). Then a correlation
with the Newark APTS was attempted, placing the CNB at the top of magnetozone
E6r, at ~229 Ma. This correlation have been improved by Gallet et al. (2003), that
confirm the position of the CNB in Newark E6r (Krystyn et al., 2002) and assign
a Norian age to the Newark APTS up to E22n. Gallet et al. (2003) recalculated
the Newark APTS assigning an age of ~200 Ma to the Jurassic base (Pálfy et al.,
2000; Courtillot and Renne, 2003), considered coeval to the base of the Orange Mt.
Basalts. In this case the Rhaetian falls at ~202 Ma (lasting ~2 My) and the CNB
at ~227 Ma. However, this recalculation is not valid anymore after the radiometric
age of 201.36±0.17 Ma for the base of the Hettangian stage (Wotzlaw et al., 2014).
Chennell et al. (2003) compared the more extended section of Silická Brezová
(Tuvalian to Sevatan, calibrated with conodonts) with Pizzo Mondello (Tuvalian-
Lacian), and then attempt a correlation with the Newark APTS. The greater extension
of the Silická Brezová magnetostratigraphy allowed a more precise correlation with
the APTS, placing the CNB within E7r, at ~227 Ma, resulting older than the CNB
proposed for the Newark APTS (e.g. Kent et al., 1995; Kent and Olsen, 1999). The
magnetostratigraphic correlations with the Newark was performed exclusively on
a visual base, comparing the pattern of reversals and chosing the correlation that fit best. Muttoni et al. (2004) extended the magnetostratigraphy of the Pizzo Mondello section up to the Sevatan. Having a continuous magnetostratigraphy from the end-Carnian to the end-Norian, Muttoni et al. (2004) compare the thickness of the Pizzo Mondello magnetozones to the duration of the Newark magnetozones, “moving” the Pizzo Mondello magnetostratigraphy along the entire APTS two magnetozones at time (to have always a comparison between magnetozones of the same polarity) and taking into account all the possible correlations. This kind of correlation is based on the assumption that in a basin the sedimentation rate is almost stable, so the thickness of the magnetozones is a proxy of time. Then the numerical thickness/duration comparison was evaluated with a linear regression and a Student t-test, and the correlations having the higher t value were considered the most reliable. Of 16 possible correlations, two resulted the most statistically reliable (significant at 95%). From these two options (Option #1 and Option #2), two age models have been derived to evaluate average sedimentation rates. Of the two options, preferred #2 has the highest t-value and smoother average values of sedimentation rates. Following this option, the CNB (refined by new conodont investigations) was placed within Newark magnetozone E7r at ~227-228 Ma, confirming the Norian age proposed by Channell et al. (2003).

The investigation of St. Audrie’s Bay section (Hounslow et al., 2004) provided a magnetostratigraphy for the upper Norian/Rhaetian (calibrated with microfloral and palynomorphs) that show discontinuity around the Norian/Rhaetian boundary (NRB) for the presence of two unconformities. The NRB is approximated by the FO of dinoflagellate cyst *Rhaetogonyalux rhaetica*, at the top of the Blue Anchor Fm and at the base of magnetozone SA5n. Hence, the Rhaetian is characterized by a zone of prevalent normal polarity. The correlation with the Newark APTS constraint the Rhaetian in the E22-E23 interval.

Gallet et al. (2007) correlate the section of Oyuklu (Turkey), calibrated with conodonts, with the Tethyan composite of Gallet et al. (2003), constructing a Sevatan-Rhaetian Tethyan composite that was correlated with the Newark APTS. This correlation follows the one from Gallet et al. (2003), where the Rhaetian is about 2 My long, from E22r to E23r. The Rhaetian of Oyuklu is mostly normal polarity, as well as in St. Audrie’s Bay (Hounslow et al., 2004), and that seems to confirm the correlation of Gallet et al. (2007), although in both sections part of the Rhaetian is missing. The absence of a correspondent magnetozone in Newark for Oyuklu magnetozone J- led Gallet et al. (2007) to conclude that a small part of the
Rhaetian in the Newark APTS is missing. The duration for the Rhaetian of ~2 My is based on the Triassic/Jurassic boundary (TJB) of Kozur and Weems (2005), placed inside the Preakness Basalts of Newark. Gallet et al. (2007) consider the possibility to include the Sevatian 2 as part of the Rhaetian (after Ogg, 2004). This means to include the Misikella posthernsteini-bidentata Zone in the Rhaetian. In this case the Rhaetian base is placed in upper E21n, obtaining a duration of this Stage of ~4.5 My (Gallet et al., 2007), considering the TJB by Kozur and Weems (2005). Using the TJB of Wotzlaw et al. (2014) the Rhaetian of Gallet et al. (2003) should last ~1 My (“Traditional Rhaetian”) or ~3.2 My (including Sevatian 2). The idea of a missing Rhaetian in the Newark succession was early proposed by Van Veen (1995) and Kozur and Weems (2005) on a biostratigraphic basis. In the Newark, Cornet (1977) found Triassic pollens just below the TJB defined by Classopollis meyeriana, whereas these taxa disappear in the basal Rhaetian of Europe (Van Veen, 1995). Using conchostracans, Kozur and Weems (2005) show how in the Newark Supergroup the strata just below the Orange Mt Basalts reveal a quick transition from typical upper Sevatian conchostracans (Shipingia olseni) to typical upper Rhaetian assemblages. Following these interpretations, the Rhaetian in Newark should be condensed in a small interval. Anyway, the well-known provinciality of palynomorphs and the poor information about Rhaetian conchostracans make these hypotheses debatable.

Combining the marine sections with a well-calibrated magnetostratigraphy, Hounslow and Muttoni (2010) construct a time-calibrated composite of the Triassic magnetostratigraphy (Geomagnetic Polarity Time Scale – GPTS). The Late Triassic portion of the GPTS was partially calibrated using the Newark APTS (in particular with the Rhaetian). The comparison between the Norian to Rhaetian GPTS with the Newark APTS produced three options (Fig. 1.3):

- Option A follows the proposal of Muttoni et al. (2004), suggesting the NRB within E17 in the Newark APTS (as in Olsen and Kent, 1999), similar to the correlation of Channell et al. (2003). This option assumes an incomplete magnetostratigraphy at the E13-E14 interval.
- Option B follows the correlations of Krystyn et al. (2002), Gallet et al. (2003) and option 1 of Gallet et al. (2007). This option implies large sedimentation rates changes around the Norian/Rhaetian boundary. In Gallet et al. (2003) the NRB is placed in the E21r, while in Hounslow and Muttoni (2010) is placed around E19-E20.
- Option C follows the correlations of Hounslow et al. (2004) and option
2 of Gallet et al. (2007). This option is consistent with the hypothesis of a lack of Rhaetian sediments in the Newark APTS (e.g. Van Veen, 1995; Gallet et al., 2007). Following Hounslow et al. (2004), the NRB should fall within the E22-E23 interval but the previous biostratigraphic investigation of Orbell (1973), which place the FO of *Rhaetagonyalux rhaetica* lower in the Blue Anchor Fm, was not considered. Moreover, the *R. rhaetica* in alpine Tethyan sections (Krystyn et al., 2007b) is typical of upper-middle Rhaetian, suggesting a lower position of the NRB in St. Audrie’s Bay, in the SA4r or lower.

Although any option is preferred to another, Option A seems the one less affected by stratigraphic inconsistencies (e.g. large variations of sedimentation rates,
inconsistencies in biostratigraphic markers, etc.).

Recently, Hüsing et al. (2011) re-sampled the Steinbergkogel section, GSSP candidate for the Rhaetian Stage (Krystyn et al., 2007a, 2007b). The new magnetostratigraphy have been correlated with other Tethyan sections, including Pizzo Mondello (Muttoni et al., 2004). The astronomical cycles recognized in Pizzo Mondello (~1.75 My, Hüsing et al., 2011) was well correlated with the same cycles in the Newark APTS, confirming the proposed correlation of Steinbergkogel through Pizzo Mondello. The NR, traced using the FO of conodont *Misikella hernsteini* instead of the younger *M. posthernsteini*, is correlated to the top of E16n, at ~209.8 Ma. This correlation is in agreement with Kent et al. (1995), Channell et al. (2003), and Muttoni et al. (2004).

Recently, the Geological Time Scale 2012 (Gradstein et al., 2012) propose two different options for the Stages of the Late Triassic (Fig. 1.4), based on the correlation between a magneto-biostratigraphic sequence (in which Stage boundaries are defined by ammonoid Zones) from the Tethys, and the Newark APTS (for both options, Carnian age is ~237 Ma and Hettangian age is ~201.3):

- **Long-Tuvalian option**: This option consider a hiatus of ~5 My in the Newark Supergroup, following the Late Triassic chronology proposed by Lucas et al. (2012). In the Newark Supergroup, strata just below the Orange Mt Basalts (oldest CAMP lava flows in the Newark Basin) show a quick transition from typical upper Sevatian (late Norian) conchostracans (*Shipingia olseni*) to typical upper Rhaetian assemblages (Kozur and Weems 2005, 2007, 2010). Anyway, conchostracans of lower Rhaetian are poorly described and the distribution of *S. olseni* could cover this interval up to the appearance of *Euestheria brodieana* (upper Rhaetian; Kozur and Weems, 2005). Lucas et al. (2012) justified this theory assuming that the Rhaetian should cover a large reversal polarity period, whereas in the Newark the conventional Rhaetian is mostly of normal polarity. Introducing the gap in the Newark APTS, the part of the magnetozone E24n included in the Passaic Fm is assigned to the uppermost Norian at ~206.5 Ma. Introducing the gap in the Rhaetian of the Newark Supergroup, the APTS has been recalculated and the Ladinian-Carnian boundary (~237 Ma; Mietto et al., 2012) corresponds to the Newark magnetozone E2r. The Rhaetian is placed at 205.5, apparently on the assumption that the Rhaetian base should be ~1 Myr younger than the Norian E24n (~206.5 Ma). The Carnian/Norian boundary is here placed at the top of E13r, at ~221 Ma, in agreement with the CNB placed with
palynomorphs (Olsen and Kent, 1999), and with the correlation Option #1 with the Pizzo Mondello section of Muttoni et al. (2004). Therefore, duration of the Stages are: Carnian ~16 My, Norian ~15.6 My and Rhaetian ~4.1 My long.

Figure 1.4: Long-Tuvalian and Long-Rhaetian options from the Geological Time Scale 2012 (Gradstein et al., 2012)
- **Long-Rhaetian option**: In this option, stratigraphy of the Newark Supergroup is considered as continuous. In facts, there is no lithostratigraphic evidences of a gap in the upper Passaic Formation (Gradstein et al., 2012) and coeval successions with cyclostratigraphy show same time span between the first CAMP basalt and the last reversal (~20 kyr) (Deenen, 2010; Deenen et al., 2010). The age of the Carnian-Norian boundary is assigned to the base of E7n (Channell et al., 2003; Muttoni et al., 2004), around 227-228 Ma. The base of the Rhaetian is placed at 209.5 Ma, at the base of E17r in the Newark, following Option A of Hounslow and Muttoni (2010) and the similar proposal of Hüsing et al. (2011). Summarizing, Carnian is ~8.6 My, Norian is ~18.9 My and the Rhaetian is ~8.2 My long.

The Long Rhaetian have been considered for the summary figures of the Geological Time Scale 2012. The choice is conditioned by the U/Pb radiometric ages from volcanic tuffs in Nicola Group (British Columbia) that constraint the lower/middle Norian boundary at ~224 Ma (Diakow et al., 2011), an age inconsistent with the Long Tuvalian option.

Concluding, Late Triassic is constrained by radiometric ages at the top (201.36±0.17 Ma; Wotzlaw et al., 2014) and at its base (237.773±0.052 Ma; Mietto et al., 2012). Stage boundaries as Carnian/Norian and Norian/Rhaetian lacks of a formally accepted numerical age, because of many different proposals of correlations between the marine sections and the Newark APTS. To refine the Late Triassic geochronology further investigations are needed, focusing on biostratigraphic well-calibrated marine sections, which offer the possibilities to recognize clearly the markers of Late Triassic Stages and substages. The study of magnetostratigraphy in marine sections, as seen before, represent a valid tool to improve the Late Triassic chronology by the correlation with the Newark APTS.

### 1.3 A NEW PROPOSAL FOR LATE TRIASSIC CHRONOLOGY

The presence of (at least) two different geochronological options for the Late Triassic (Gradstein et al., 2012) lead to hardly assign an age or duration to the events occurring during this time period. To find a way out from this impasse is necessary to improve the integrated stratigraphy of the Late Triassic. Considering the importance of the Newark APTS as geochronological tool, the investigation of magnetostratigraphy is fundamental to attempt a correlation with stratigraphic succession recording clearly the chronological key marker.

To give a contribution to the Late Triassic chronology I investigate for
magnetostratigraphy a series of selected Tethyan sections covering the Norian-Rhaetian interval and the middle Carnian stage. The choice of these two intervals is strategical. The base of the Rhaetian is still debated so it is necessary to analyze the magnetostratigraphy of a section in which the Stage boundary is clearly defined and then attempt a correlation with the Newark APTS. The Pignola-Abriola section (Italy) is a perfect candidate because of a detailed conodont biostratigraphy, in association with radiolarians. The section was also investigated for organic carbon chemostratigraphy, obtaining a $\delta^{13}C_{org}$ curve. Moreover, I investigate the Rhaetian/ Hettangian section of Mount Messapion (Greece) and the Norian/Rhaetian sequence of ODP Leg 122 - Hole 761C (Australia).

Also the Carnian Stage shows similar issues regarding its chronology. The Ladinian/Carnian boundary has been defined with a GSSP (Mietto et al., 2012) and the proposals of Carnian/Norian based on marine fossils are considered reliable (e.g. conodonts, ammonoids), but it is still debated the discrepancy with the Norian based on continental fossils (palynomorphs, conchostracans, vertebrates; Lucas et al., 2012). The Carnian/Norian boundary has a detailed magnetostratigraphy, which unfortunately is missing for the rest of the Carnian, characterized instead by the few sections affected by condensation or discontinuities like faults and disconformities/unconformities (e.g. Bolücektasi Tepe, Gallet et al., 1992). To improve the magnetostratigraphy of the middle Carnian I decided to investigate the Pignola-2 section in Southern Italy and the Dibona section in Northern Italy, both covering the middle Carnian, around the Julian/Tuvalian boundary. I also investigated the Carnian/Norian stratigraphic sequences of ODP Leg 122 - Holes 759B/760B (Australia).
PART A
ROCK MAGNETISM

2.1 THE GEOMAGNETIC FIELD
As known, the Earth is wrapped by a magnetic field that protects the planet from the energy-charged particles of the solar wind. This “geomagnetic” field originates inside the Earth’s Fe-Ni core, from the convection currents in the liquid outer core. The type of interaction is explainable by a self-exciting magnetohydrodynamic dynamo model. A self-exciting dynamo needs a moving electrical conductor (the

![Diagram of the geomagnetic field](image1)

Figure 2.1: a) model of the Geomorphic Axial Dipole (modified from McElhinny, 1973); b) model of the Inclined Geocentric Dipole (modified from McElhinny, 1973); c) non-dipolar geomagnetic field. Magnitude and direction of the field is represented by arrows (scale in the lower-right corner) (from Butler, 1992; modified from Bullard et al., 1950)
outer core) that generates a magnetic field, which in turn provide to the conductor the energy required to maintain the necessary rotation (by the Lorentz force) to feed the magnetic field itself. However, part of the energy is loss by electrical resistivity in the conductor and this energy must be supplied to keep the self-exciting dynamo active. Considering that the Earth has not been “roasted” yet by the solar wind, some kind of energy supply must be present. The most likely source of energy has been identified in the progressive cooling of the Earth’s outer core, which can provide the estimated $10^{13}$ W necessary to sustain the geomagnetic field. Theoretically, the geomagnetic field is approximated to a Geocentric Axial Dipole (GAD) (Fig. 2.1a), generated by a single dipole in the center of the planet and with the magnetic axis parallel to the rotation axis. Indeed, the real morphology of the geomagnetic field is quite different and the magnetic axis is inclined of about 11.5° respect to the rotation axis (Inclined Geocentric Dipole) (Fig. 2.1b). Assuming the IGD model as real, the inclined dipole axis must be coincident with the dip axis of the magnetic field lines, but this is not completely true. The Eccentric Dipole model (a dipole diverged from the center of the Earth) best fits the magnetic field in about the 80% of the Earth’s surface, but in some areas the model does not fit. To explain this anomaly, a non-dipolar component of the magnetic field has been invoked (Fig. 2.1c), originating at the interface between the outer core and the lower mantle.

2.2 MAGNETIC PROPERTIES OF MINERALS
Most of the rocks in the Earth are characterized by magnetic properties and are able to retain a magnetization if they have been subjected to a magnetic field. The magnetic features of a rock are related to the content of magnetic minerals, which are characterized by different behaviors.

2.2.1 Behavior of the magnetic minerals
Magnetic minerals have been divided in three major categories: diamagnetic, paramagnetic and ferromagnetic minerals. Diamagnetic minerals do not acquire any magnetization if subjected to an external magnetic field, while the paramagnetics acquire a magnetization until the magnetic field is removed. Ferromagnetic minerals retain the magnetization even after the magnetic field is removed, showing a complicated behavior and at least three subcategories that are: ferromagnetic *sensu stricto*, antiferromagnetic and ferrimagnetic minerals. Ferromagnetic minerals *s.s.* are characterized by magnetic moments in minerals that are oriented parallel to the applied field (Fig. 2.2a), acquiring a strong residual magnetization. Minerals having
an antiferromagnetic behavior have magnetic moments oriented in both parallel and anti-parallel position respect to the applied field (Fig. 2.2b), with a resultant magnetization equals to zero; antiferromagnetic behavior pass to paramagnetic over the Néel temperature ($T_N$). In ferrimagnetic minerals the magnetic moments are oriented parallel and anti-parallel too, but there is a dominant direction of magnetization (Fig. 2.2c); most of the magnetic minerals are ferrimagnetic. A ferromagnetic mineral become paramagnetic over the Curie temperature ($T_C$) and return ferromagnetic under the same temperature. There are also some special behaviors, like canted-antiferromagnetic and superparamagnetic.

A canted-antiferromagnetictic mineral differs from an antiferromagnetic s.s. for a non-zero magnetization, due to canted magnetic moments that induce a preferred direction of the bulk magnetization vector.

Superparamagnetic minerals can acquire a magnetization from an applied magnetic field, but are characterized by a very short relaxation time ($t_s$, the period of time after which a magnetic mineral lose the acquired magnetization), so they lose their magnetization quickly; $t_s$ is depending on temperature (Fig. 2.3) and its variation is characteristic for every magnetic mineral, so there are superparamagnetic materials at room-temperature and ferromagnetic minerals that become superparamagnetic at very high temperature (called blocking temperature, $T_B$).

Figure 2.2: coupling of magnetic moments in a) ferromagnetic (s.s.), b) antiferromagnetic, c) ferrimagnetic minerals (from Butler, 1992).

Figure 2.3: log plot of relaxation time of a single-domain magnetite as function of temperature. $T_B$ is the blocking temperature (from Butler, 1992).
2.2.2 Main ferromagnetic minerals

Principal ferromagnetic minerals are the Fe-enriched phases of titanomagnetite and titanohematite solid solutions. Titanomagnetite is a solid solution between magnetite ($\text{Fe}_3\text{O}_4$) and ulvospinel ($\text{Fe}_2\text{TiO}_4$), where ulvospinel is paramagnetic at room-temperature (due to Ti-enrichment) and magnetite is ferromagnetic ($T_C=580^\circ\text{C}$, $J_s=4.5\times10^5 \ \text{A/m}$) (Fig. 2.4a). Titanohematite is a solid solution between hematite ($\alpha\text{Fe}_2\text{O}_3$) and ilmenite ($\text{FeTiO}_3$), where hematite is canted-antiferromagnetic (sometimes ferromagnetic s.s. due to structural defects) with a $T_N=680^\circ\text{C}$ ($J_s=2\times10^3 \ \text{A/m}$) and ilmenite is paramagnetic at room-temperature (Fig. 2.4b). Minor ferromagnetic minerals are Fe-oxhydroxides (e.g. goethite), maghemite and Fe-sulphurs (e.g. pyrrhotite).

\[\text{Fe}_3\text{O}_4 + x\text{TiO}_2 \rightarrow (\text{Fe}_3-x\text{Ti}_x)\text{O}_4\]  

Figure 2.4: saturation magnetization and Curie temperature of the titanomagnetite series (a) and titanohematite series (b) (from Butler, 1992; modified from Nagata, 1961, and Stacey and Banerjee, 1974)

2.2.3 Magnetic domains

The acquired magnetization is also function of the grain structure. Crystals of magnetic minerals are characterized by magnetic domains, indicating the distribution of magnetic charge on the surface of the grain. In a single-domain (SD) crystal magnetic charge is equally distributed on the surface (Fig. 2.5a) and the acquired magnetization ($J$) is equal to the maximum acquirable magnetization (saturation magnetization, $J_s$). In a multi-domain (MD) grain there are more domains with different orientation and the magnetic charge is not equally distributed on the surface of the crystal (Fig. 2.5b); as a consequence, $J$ is lower than $J_s$.

Grain size influence the development of single or multiple domains, in particular small grains are characterized by single domains easily than large grains, which form usually multi-domains for energetic convenience. Is difficult to understand the exact grain size limit between SD and MD grains. In facts, a particular interval of grain size shows intermediate features between SD and MD. This situation refers to the pseudo-single-domain (PSD) grains. The PSD grains have a small number
of domains inside and they also show a high coercivity and a relevant time stability of the magnetization. Most of the times, PSD grains are important carriers of remanent magnetism. Moreover, isotropic crystals are easily magnetizable than anisotropic ones, and anisotropic minerals are simply magnetizable along the long axis than the short one.

2.3 ACQUIRING THE MAGNETIZATION

Magnetization in situ of the rocks is the sum of the magnetization induced by the actual geomagnetic field and the Natural Remanent Magnetization (NRM). The NRM is necessary for paleomagnetic investigations and is recorded in rocks, contrary to the induced magnetization that is not preserved. The NRM results from the combination of a primary NRM (recorded during rock formation) and secondary NRM (acquired after the rock formation). The primary NRM is acquired mainly in three ways: cooling from high temperatures (ThermoRemanent Magnetization), deposition of magnetic minerals in detritus (Detrital Remanent Magnetization) and sometimes by growth of magnetic minerals below the $T_c$ (Chemical Remanent Magnetization). A secondary NRM can be recorded in many ways, mainly by application of a relative weak magnetic field for a long time (Viscous Remanent Magnetization), chemical alteration of ferromagnetic minerals (secondary CRM) or by application of a very strong magnetic impulse (Isothermal Remanent Magnetization).

ThermoRemanent Magnetization (TRM)

The TRM, typical in magmatic rocks, is acquired during the cooling of a magnetic mineral subjected to a magnetic field, when the temperature of the system decreases below the $T_c$. In particular, TRM becomes stable below the $T_y$, at the passage between superparamagnetic and ferromagnetic behavior. The acquisition of the TRM is not instantaneous for the whole rock, but is acquired gradually when each phase reaches the $T_b$. The stability of TRM during time is strongly related to the grain size: small crystals have high probability to develop single domains, while large crystals are probably characterized by a multi-domain structure.
Detrital Remanent Magnetization (DRM)

The DRM is acquired during the deposition of sedimentary rocks. The DRM recorded by orientation of magnetic minerals in the moment of deposition is called depositional DRM (dDRM), while the DRM recorded successively to the deposition and before the lithification is called post-depositional DRM (pDRM). The classic model of DRM acquisition consider a complete, rapid alignment of the grains along the magnetic field direction, but is not what happens in reality. In many experiments (e.g. Verosub, 1977) grains showed a lower degree of alignment than expected and a shallowing in inclination respect to the magnetic field (Fig. 2.6a). In effect, during the deposition, grains rotate while laying down on the bottom (Fig. 2.6b). This phenomenon should induce an error in the recorded inclination, but the observations in nature are not coherent. Considering this, most of the recorded DRM is actually pDRM. The experiments of Verosub (1977) show that in water-rich sediment grains orient themselves exactly along the applied field (Fig. 2.6c) and that in the presence of a polarity inversion the grains in a 10-20 days old sediment rotate to align themselves to the magnetic field. The pDRM is not the only contribution to the whole DRM, but dDRM is also present. The dDRM/pDRM ratio is function of many factors, mainly:

- grain size: small grains remain suspended in the water column longer than large grains, increasing the pDRM contribution.
- bioturbation: sediments reworked by biological activities record a pDRM.
- sedimentation rate: slow rates enhance the pDRM, increasing the residence

![Figure 2.6: a) inclination of DRM (I_p) related to the inclination of the applied magnetic field (I_H) (from Buter, 1992; modified from Verosub, 1977). b) inclination shallowing of DRM due to rotation of elongate grains (m is magnetic moment, H is the applied magnetic field). c) inclination of pDRM against inclination of the applied magnetic field (H); vertical bars are the confidence limits on pDRM inclination, straight line represent the perfect agreement between pDRM and H inclinations. From Butler (1992).](image-url)
time of the magnetic particles in the high water content zone of the sediments. As mentioned before, grain size has also effects on the quantity of multi-, single- or pseudo-single-domain grains. Fine silt or clay has more SD (and PSD) particles than a coarse sand or silt, which are full of MD grains. SD and PSD grains retain a higher magnetization than MD grains, and the record is more stable. Moreover, coarse grains are more sensitive to mechanical energies that override the magnetic alignment of the particles, and the high permeability of coarse sediments enhance the chemical alteration of ferromagnetic minerals. For these reasons, fine-grained sedimentary rocks are preferable to coarse-grained for paleomagnetic analyses.

**Chemical Remanent Magnetization (CRM)**

If magnetic minerals form after chemical reaction and are subjected to a magnetic field, the acquired magnetization is called Chemical Remanent Magnetization (CRM). The magnetization is acquired during the growth of the crystals, and become stable after the crystals reach the blocking volume, allowing the formation of stable single domains. During alteration of minerals, if the crystal structure did not change (i.e. magnetite to maghemite), the direction of the CRM is controlled by the previous NRM. The acquisition of CRM can occur also from precipitation of magnetic minerals from a solution, or for postdepositional alteration. If the alteration occurs soon after the deposition, the CRM could be regarded as primary (i.e. oxidation reactions with formation of hematite; dehydration of goethite in hematite). Otherwise, if the alteration occurs long after deposition, the CRM is regarded as secondary (i.e. diagenetic production of sulfides).

**Viscous Remanent Magnetization (VRM)**

Magnetic minerals subjected to a weak magnetic field (i.e. the geomagnetic field) for a long time can acquire a secondary magnetization called Viscous Remanent Magnetization (VRM). Acquisition of VRM has a logarithmic growth in time, so normally is dominated by recent geomagnetic field. Moreover, the VRM is susceptible to the temperature and its intensity increase for high temperature. As argued before, the VRM is substantially related to time and SD grains with a short relaxation time are probably carrier of VRM. Different is the behavior of PSD and MD grains, where thermal energy is necessary to acquire the VRM, activating the domain walls motion. Magnetization of PSD and MD grains interacts with the applied field facilitating the domain walls motion, and the magnetization increases along the direction of the applied magnetic field. Furthermore, the lower is the coercivity of the PSD/MD grains, the higher is the VRM acquired.
Isothermal Remanent Magnetization (IRM)

After the rock is formed, strong magnetic fields (induced by natural or artificial sources) can remagnetize rocks completely, obliterating every remanent magnetism acquired before. This kind of remagnetization is called Isothermal Remanent Magnetization (IRM) and is acquired at room temperature. Rare in nature (acquired normally by effect of lightnings hitting the outcrops), IRM represents a valid analysis tool to reveal the main magnetic carrier in samples. Methods of IRM analyses are explained later.

PART B
METHODS

2.4 UNDERSTANDING THE ROCK MAGNETIC PROPERTIES

Rocks are made by an assemblage of minerals and some of them are magnetic. The magnetic properties of a rock sample reflect its content in magnetic minerals and are dominated by the most abundant phase. Knowing the properties of a magnetic mineral is fundamental in paleomagnetism, because it influences the preservation of the original magnetization during time. The reactions of magnetic minerals subjected to different conditions (external magnetic fields, heating, etc.) are unique, and many analytic techniques have been developed to quantify these reactions.

2.4.1 Analysis of Natural Remanent Magnetization (NRM)

To reveal the Natural Remanent Magnetization originally acquired during the formation of the rock, oriented samples are needed. Samples cored on the field are oriented using an orienter tool and a compass, then the orientation is marked on the sample with a line (Fig. 2.7) along z-axis, normally plunging into the outcrop. Samples can be obtained also from oriented blocks, where the azimuth is derived adding 180° to the dip of the block, and the hade is 90-dip angle (Fig. 2.8). Most of

![Figure 2.7: example of oriented core. Direction information are azimuth (the angle between the horizontal projection of the x-axis and the North) and hade (the angle from vertical). From Butler (1992)
the instruments fits with 1-inch diameter cores, 1 inch long. The so-oriented samples are then demagnetized progressively. The magnetization retained by the magnetic minerals in the rock is removed by heating (thermal demagnetization) or by application of an alternate gradient magnetic field. The “soft” magnetic minerals (e.g. pyrrhotite, goethite, Ti-rich magnetite) are removed by low temperatures and fields, isolating the magnetization recorded by “hard” minerals (e.g. magnetite, hematite). For each step of demagnetization, the magnetometer detects the direction and the intensity of the magnetization vector, then plotted into an end-point vector diagram (Zijderveld, 1967) (Fig. 2.9).

The diagram shows the projection of the magnetization vector on the horizontal (full symbols; Fig. 2.9a) and vertical (empty symbols; Fig. 2.9b) planes. When the magnetization component identified as characteristic of the sample (Characteristic NRM, ChRM; Fig. 2.10a) is isolated (Fig. 2.10b), the mean direction and intensity of this component are calculated using a Principal Component Analysis (PCA) (Fig. 2.10c). Once determined the direction of the characteristic component, the datum is interpreted paying attention to the values of azimuth and inclination. If the direction is unusual (e.g. north/up or south/down vectors in northern hemisphere, instead of north/down or south/up), orientation data of the samples should be checked and, in case, corrected or rejected.

**Evaluation of paleomagnetic stability**

Once derived the ChRM from a set of samples, is difficult sometimes to be sure that this component is primary or have been acquired lately. The application of paleomagnetic stability tests could
provide information about the origin and timing of the acquired magnetization. Here are presented the most common tests.

Fold test
This test evaluates the time of the ChRM acquisition related with folding (obviously, this test does not work on homoclinal sections). If the ChRM is acquired before folding, the directions from the opposite limbs of the fold are dispersed, but they should converge after the application of the structural correction (Fig. 2.11). If the site pass the fold test (original ChRM) the structural corrected directions should be more clustered than before; on the contrary, if the fold test is failed, the corrected directions are more dispersed.

Figure 2.11: example of the fold and conglomerate tests. Arrows in fold limbs and conglomerate clasts are the magnetization directions. The random distribution of the directions in conglomerate suggest the acquisition of a primary ChRM, as well as the alignment of the directions in fold after structural correction. (from Butler, 1992)

Conglomerate test
If in a conglomerate the clasts acquired a ChRM before the deposition of the conglomerate (in the source rock), the ChRM directions should be randomly distributed (the site passed the conglomerate test) (Fig. 2.11). If the direction are consistent, the test is failed, which means that the ChRM have been acquired after the deposition of the conglomerate. This test usually works on interbedded
conglomerate to test the stability of the ChRM of the source rock in the same site.

**Reversal test**

As seen before (Chapter 2.1) the geomagnetic field is approximated by an axial dipole; this is true for a time-averaged geomagnetic field, both for normal or reverse polarity periods. Consequently, in a site the mean directions of primary normal and reverse components should differ of ~180°. Secondary components acquired later deviate the angular difference between the mean normal and reverse directions, resulting non-antiparallel (Fig. 2.12). If the reversal test is passed, it means that the secular variation of the geomagnetic field is well averaged and the ChRM is coherent with a primary acquisition. On the contrary, if the test is failed probably the ChRM is strongly contaminated by secondary magnetization, or the paleomagnetic data are not sufficient to average the secular variation of the geomagnetic field. This test is applicable to every dataset because any particular geological setting is required (as folding or conglomerates).

![Figure 2.12: black arrows indicate the expected antiparallel configuration of the mean directions. In presence of a secondary component of magnetization (light gray arrows), the resultant mean directions (dark gray arrows) are non-antiparallel (from Butler, 1992). More the secondary component is relevant, less are the possibilities to pass the reversals test.](image)

2.4.2 Isothermal Remanent magnetism (IRM): techniques

Magnetic characteristics of a rock are depending on the content in magnetic minerals. Each magnetic mineral has a different behavior if subjected to an induced magnetic field or if heated. By the application of different artificial IRM (Isothermal Remanent Magnetism) techniques is possible to understand the particular magnetic properties of a rock sample as a reflection of the dominant phases of magnetic minerals.

**IRM acquisition curve**

A magnetic field is applied increasing progressively the intensity of the field and the acquired magnetization is recorded after each step of treatment. When the sample reaches the maximum magnetization, it is saturated (Fig. 2.12). The minimum magnetic field necessary to reach saturation is the saturation field, which is diagnostic for magnetic minerals (e.g. magnetite saturates around 500 mT).
Backfield curve

This technique is useful for the determination of the coercivity, or the minimum magnetic field necessary to delete completely the acquired magnetization of a mineral, after complete saturation. The magnetic coercivity measures the resistance of a ferromagnetic mineral to complete demagnetization. High-coercivity minerals (magnetically hard) are more difficult to demagnetize, or remagnetize, than low-coercivity minerals (magnetically soft). The experiment consists in saturate the sample by the application of a high magnetic field (usually > 500 Mt), and then apply a gradient magnetic field until the total magnetic moment reaches zero (Fig. 2.13). Coercivity ($H_c$) is usually measured in A/m.

Hysteresis loop

Once a ferromagnetic mineral is magnetized, it will not spontaneously come back to zero magnetization once the applied magnetic field is removed. It needs an external magnetic field in opposite direction to drive the magnetization to zero. Applying a gradient magnetic field to a mineral, the magnetization traces a closed curve called hysteresis loop (Fig. 2.14a). The magnetic hysteresis represents the lag between the magnetization curve (after application of a magnetic field) and the relaxation curve (once the magnetic field is removed), due to the presence of magnetic domains that need energy to be reoriented. The magnetization retained at zero-field is called remanence. In rock samples (representing a mixture of magnetic minerals),
the hysteresis loop reflects the content in SD, MD and PSD grains. SD grains start orienting along the magnetic field (Fig. 2.14b) direction until saturation, where they are all oriented (Fig. 2.14c). Once the field is removed, the magnetic moments orient along the long axis nearer to the magnetic field direction (remanence) (Fig. 2.14d). An opposite field is required to erase the magnetization (Fig. 2.14e). In MD grains, the application of an external field causes the growth of domains with magnetization parallel to the field. If the field is strong enough, the domain walls are destroyed and the grains saturate. Removing the field the domain walls regenerate, returning to initial conditions, but for the presence of imperfections in the grains they retain a weak remanence. The field required to drive the domain walls to original position is weak, reflecting the low coercivity force of MD grains. The PSD grains have a small number of domains, showing a behavior more similar to SD than MD.

Three-axial IRM
Magnetic minerals lose completely their magnetization if they reach the Curie temperature (Tc; for ferromagnetic s.s., ferromagnetic and canted-antiferromagnetic minerals) or the Néel temperature (TN; for antiferromagnetic minerals). Tc and TN are characteristics for each mineralogical phase and the three-axial IRM technique (Lowrie, 1990) has been developed to detect these temperatures. Samples are magnetized along X,Y,Z axes, saturating one axis (usually Z) at high magnetic force (2-2.5 T) and then magnetizing the others with lowest fields (e.g. X: 400 Mt; Y: 120 Mt). Then, the samples are thermally demagnetized increasing progressively.
the temperature. After each step of temperature, the samples are analyzed in a magnetometer. The intensity of the resulting magnetic vector is then decomposed in X, Y and Z components (Fig. 2.15). The demagnetization curve reveals which category (magnetic hard or soft minerals) are dominant and at which temperature the samples lose magnetization, as a reflection of the magnetic minerals contained in the rock.

**Thermomagnetic curve**

In a constant external magnetic field, a sample subjected to a variation of temperature changes its magnetic moment. The decay curve of temperature represents the change of behavior (from ferromagnetic to paramagnetic or superparamagnetic) of the magnetic minerals inside the rock sample, depending on temperature (Fig. 2.16). If the analysis is performed in air, the curve could show the presence of neo-formed minerals by thermal oxidation of the original mineralogical phases (e.g. pyrite in magnetite, goethite in hematite, etc.), with increase of magnetization until the $T_c$ (or $T_N$) of the neo-formed mineral is reached. If the new phase is stable, the magnetization at room temperature results higher than the starting magnetization.

**Figure 2.15:** example of three-axial IRM. Three different demagnetization paths reflect different applied fields along x, y and z-axes. Along z-axis the sample is saturated, then other two weaker fields are applied along x and y to orient low-coercivity minerals. Then the sample is thermal demagnetized, and the path give information about the $T_c$ and $T_N$ of the magnetic minerals in the sample. (from Tauxe, 1998)

**Figure 2.16:** example of thermomagnetic curve. Magnetization is calculated in relation with the mass of the sample. The curve show an increase in magnetization at ~400°C, due to the presence of neo-forming minerals. This minerals decay at ~580°C, suggesting that it could be magnetite. The temperature of neo-formation is typical of oxidation of pyrite. (from Hoffman et al., 2011)
2.4.3 Instruments

Alternate Gradient Field Magnetometer (AGM)
The AGM is based on the vibrating resonance of a magnetic material subjected to a DC (direct current) field and a small AC (alternating current) field gradient. The sample is put on a flexible stick, and then between two electromagnets generating a gradient magnetic field. The resonance vibrations are detected by a piezoelectric transducer (bimorph), connected to the sample/stick system, and then amplified. The sensibility of the instrument is attested at $\sim 10^{10}$-10$^{11}$ Am², and the required mass for a single sample is very low (around 50 g). Example of AGM in Fig. 2.17.

Cryogenic Superconducting Quantum Interference Device (SQUID) Magnetometer
The SQUID magnetometer is made of a superconducting loop with one (RF-SQUID) or two (DC-SQUID) Josephson junctions inductively coupled with a series of superconducting coils (second-order gradiometer). To act as superconductors, the coils are maintained at a temperature of 4K using liquid Helium. The two Josephson junction are connected each other by a non-conducting (or low conducting) material. The coils are electrified ($I$), and if exposed to a magnetic field they are charged of an induced current $I_s$ (Josephson effect). The induced current has the same direction of the original current in one coil (total current $I_t = I/2 + I_s$), and opposite in the other one ($I_t = I/2 - I_s$). When the total currents in the coils exceed the critical current ($I_c$), an electric tension originates in the system. This tension is then transferred to a detector and translated into the induced magnetic field. In Fig. 2.18 a DC-SQUID cryogenic magnetometer.
Thermal Demagnetizer

The thermal demagnetizer is a shielded cylindrical oven where the samples are progressively demagnetized by application of heat (Fig. 2.19). When a mineralogical phase reaches the Curie temperature (for all ferromagnetic materials) or the Néel temperature (for antiferromagnetic materials), the contribution of this phase in the total magnetism of the sample disappears. When the “softer” phases are removed (i.e. goethite, pyrrhotite, high-Ti titanomagnetite), the “hardest” remains (i.e. hematite, magnetite), revealing the original magnetism of the rock (usually preserved by hard minerals).

Alternating Field (AF) Demagnetizer

The AF demagnetizer works with the application of a decaying alternating magnetic field to a sample. After each step of treatment, the sample is cleaned of any remanent magnetization of coercivity less than the peak of intensity of the applied field. A magnetic field is applied to the sample along the X, Y and Z orthogonal axes, randomizing the mobile magnetic domains. The applied field is decaying, so the amplitude of each half-cycle of the applied AF is smaller than the previous one. During each half-cycle, the domains with coercivities less than the applied field orient themselves along the field. A small percentage of these domains have a coercivity greater than the intensity of the following half-cycle, remaining fixed in the acquired direction. In this way, equal number of domains are oriented in positive and negative direction, resulting in a zero total field. Example of AF demagnetizer in Fig. 2.20.
Curie Balance

The Curie Balance is a simple mechanic magnetic detector designed and built for the first time by Pierre Curie in 1895. Charles Cheneveau performed the most important improvement of the balance in 1930, introducing a torsion arm suspended by a fine wire from the torsion head. In the balance, the sample is placed on a torsion arm (Curie-Cheneveau balance) and exposed to a non-uniform magnetic field generated by an electromagnet, displacing the sample from its originary position. The balance detect the torque/moment necessary to bring the specimen to the starting position. The magnitude of the torque is typical of a phase or a class of magnetic minerals. Using the Curie balance coupled to a furnace (example in Fig. 2.21) is possible to determine the behavior of the samples during heating or following a series of heating/cooling cycles.

Impulse Magnetizer

The Impulse Magnetizer generates a magnetic field for the artificial magnetization of samples. A capacitor is charged until the necessary tension is reached, then the tension is released in a coil generating a magnetic field inside it (Ampère’s Law). The sample is placed inside the coil (or just in front of it) and magnetized along the axis parallel to the coil. Example of impulse magnetizer in Fig. 2.22.
Susceptibility Meter

This instrument is necessary to measure the capacity of a material of being magnetized if subjected to an external magnetic field. The measurement can be static (for analysis of bulk susceptibility) or dynamic (to determine the anisotropy of susceptibility) with the sample rotating along three orthogonal axes x, y and z. The sample is subjected to a non-uniform field of determined frequency, and then the instrument detects the intensity of the acquired magnetization. Example of susceptibility meter in Fig. 2.23.

2.5 USE OF THE PALEOMAGNETIC DATA

The data obtained from the analysis of the NRM provide the directions of the magnetic vector acquired by rocks during their formation. In the case of sedimentary rocks, it represents the magnetization acquired during the deposition and is normally in line with the direction of the magnetic field during that period. This information can be used to define the period of polarity inversions occurred during a particular period and recorded in a sedimentary sequence, or to extrapolate paleogeographic data from the position of the geomagnetic pole during the studied time interval.

2.5.1 Magnetostratigraphy

The sequence of paleomagnetic directions obtained by NRM analysis reflects the direction of the geomagnetic field in that time interval. For the northern hemisphere, northward declination and downward declination indicate normal polarity, whereas southward declination and upward inclination indicate reverse polarity. Obviously, in the southern hemisphere the situation is the opposite. A Virtual Geomagnetic Pole (VGP) can be derived from each direction datum using spherical trigonometric formulas (Fig. 2.24), and represents the position of the paleomagnetic pole in a location at one point in time (formulas are the same for each kind of magnetic pole).
To determine a VGP position \((Lat_p; Long_p)\), we need the paleomagnetic direction \((Inc_m; Dec_m)\) and site coordinates \((Lat_s; Long_s)\). First, we need to obtain the magnetic colatitude \((p)\), or the great-circle distance of the pole from our site:

\[
p = \tan^{-1}\left(\frac{2}{\tan Inc_m}\right)
\]

Then we derive the VGP latitude:

\[
Lat_p = \sin^{-1}(\sin Lat_s \cos p + \cos Lat_s \cos Dec_m)
\]

The difference of longitude between the pole and our site is given by \(\beta\) (positive to east):

\[
\beta = \sin^{-1}\left(\frac{\sin p \sin Dec_m}{\cos Lat_p}\right)
\]

Now, there are two kind of derivations for the pole longitude:

If: \(\cos p \geq \sin Lat_s \sin Lat_p\), then: \(Long_p = Long_s + \beta\)

If: \(\cos p < \sin Lat_s \sin Lat_p\), then: \(Long_p = Long_s + 180^\circ - \beta\)

The latitude of a VGP places the paleomagnetic pole in the northern or southern hemisphere, indicating respectively normal and reverse polarity. Graphically, black and white bars represent the sequence of geomagnetic polarity inversions respectively for normal and reverse polarity. A zone characterized by the same polarity is called magnetozone. Usually, inside a magnetozone some brief polarity

Figure 2.24: paleopole determination. \(S\) is the studied site, \(P\) the paleopole, and \(M\) the direction of the geomagnetic field. Magnetic colatitude is calculated from paleomagnetic inclination \((Inc)\) and then used to determine the paleopole latitude \((Lat)\). Lat, and magnetic declination \(Dec\), are used to calculate \(\beta\) (the longitudinal difference between paleopole and site). \(\beta\) is used to calculate the paleopole longitude \(Long_p\) (modified from Butler, 1992)
inversions occur and they are defined as submagnetozones. A magnetozone is named using an abbreviation followed by the sequence number and the polarity (“n” for normal, “r” for reverse; e.g. ID-1n or ID-1r); submagnetozones are named adding an extension to the name of the magnetozone following the same criteria (e.g. ID-1n.1r, ID-1n.2n, etc.). A sequence of magnetozones related to a stratigraphic succession is called magnetostratigraphy. The duration of a polarity inversion is not fixed, so the pattern of a sequence of polarity inversions is nearly unique. This is why magnetostratigraphy is a powerful tool to perform correlation between stratigraphic sections, considering also that the global range of the paleomagnetic record.

2.5.2 Paleogeography
As seen in the previous section, is possible to derive the position of a magnetic pole from the magnetic direction. Using the mean-site direction and the same formulas of the VGP we can obtain the mean-site paleomagnetic pole. The mean-site paleomagnetic pole represents the position of the pole on the long time, eliminating the fast-changing (~3000 yr.) non-dipolar component of the geomagnetic field. A site-mean direction is usually associated to a circular confidence limit $\alpha_{95}$. The $\alpha_{95}$ is transformed into an ellipse of confidence about the paleopole position. Along the great circle connecting the site to the pole, the semi-axis of the ellipse ($dp$) is calculated by:

$$dp = \alpha_{95} \left( \frac{1 + 3 \cos^2 p}{2} \right)$$
The semi-axis perpendicular to the pole to site great circle \((dm)\) is given by:

\[
dm = \alpha_{95} \left( \frac{\sin p}{\cos Inc_m} \right)
\]

In case of multiple site-mean poles, the paleomagnetic pole can be derived using a Fisher statistic, with a circular confidence limit \(A_{95}\).

The mean paleomagnetic inclination \((Inc_m)\) can be used to determine the paleolatitude \((Lat_0)\) of the site:

\[
Lat_0 = \tan^{-1}\left( \frac{\tan Inc_m}{2} \right)
\]

Once determined the site-mean paleomagnetic pole \((PP)\), is possible to calculate the rotation of the site \((S)\) against a stable reference pole \((RP)\) of the same period (Fig. 2.25). First, we need the distance \(S-RP\) \((pr)\), \(S-PP\) \((pp)\) and \(PP-RP\) \((s)\), forming a spherical triangle on Earth surface:

\[
pr = \cos^{-1}\left[ \sin Lat, \sin Lat_s + \cos Lat, \cos Lat, \cos(\text{Long}_r - \text{Long}_s) \right]
\]

\[
pp = \cos^{-1}\left[ \sin Lat, \sin Lat_p + \cos Lat, \cos Lat, \cos(\text{Long}_s - \text{Long}_p) \right]
\]

\[
s = \cos^{-1}\left[ \sin Lat, \sin Lat_p + \cos Lat, \cos Lat, \cos(\text{Long}_r - \text{Long}_p) \right]
\]

Considering the rotation \((R)\) as the angle at apex \(S\) of the triangle we can apply the law of cosines, solving for \(R\):

\[
R = \cos^{-1}\left( \frac{\cos s - \cos pp \cos pr}{\sin pp \sin pr} \right)
\]
Chapter III
GEOLOGICAL FRAMEWORK OF THE CONCERNED AREAS

For my investigations of Late Triassic magnetostratigraphy, I considered four main areas: the Lagonegro Basin (Southern Italy), the Dolomites (Northern Italy), the Wombat Plateau (South-Western Australia) and the Pelagonian Domain (Greece).

3.1 LAGONEGRO BASIN (SOUTHERN ITALY)

3.1.1 Structural framework
The Lagonegro area is located in the Southern Apennines, near Potenza (Basilicata region, Southern Italy). The geological record is represented by the sediments of the Lagonegro Basin, active since the Late Permian to the Miocene (Scandone, 1967; Finetti, 1982, 2005; Catalano et al., 2001; Ciarapica and Passeri, 2002, 2005; Argnani, 2005; Rigo et al., 2012a). The Lagonegro Basin belonged to the Ionian Ocean, a western branch of the Tethys (Fig. 3.1). The Ionian Ocean originated in the Late Permian/Early Triassic as continental rifting and probably evolved in ocean rifting (Catalano et al., 2001) during the Ladinian, coeval to the deposition of the radiolarites of the Monte Facito Formation (Ciarapica and Passeri, 2002, 2005).

The Lagonegro area is included in the southern part of the Apennines mountain range, an east-vergent fold and thrust belt spanning through the Italian peninsula, from the Po valley to the Ionian Sea. The sedimentary sequences of the Lagonegro Basin are some of the tectonic units imbricated between the Apula platform (on

Figure 3.1: position of the Lagonegro Basin in the Tethys (a) and paleogeographic reconstruction of Lagonegro Basin (b; from Ciarapica and Passeri, 2005)
footwall) and the Apenninic platform (on headwall) in the Southern Apennines (Fig. 3.2). The Lagonegro Basin sequence was subdivided in two tectonic units (Scandone, 1967, 1972, 1975): the Lagonegro Unit I, including the more distal sequences, and the Lagonegro Unit II, represented by the more proximal facies. The Lagonegro Lower Sequence (Mostardini and Merlini, 1986; Ciarapica and Passeri, 2005) includes, from the bottom, the Monte Facito Formation (Permian to Middle Triassic), the Calcari con Selce (Middle to Late Triassic), the Scisti Silicei (Late Triassic to Late Jurassic) and the Flysch Galestrino (Late Jurassic to Early Cretaceous) (Fig. 3.2). The Lagonegro Upper Sequence, always detached from the Lower Sequence (Mostardini and Merlini, 1986), includes the Flysch Rosso (Early Cretaceous to Early Miocene), the Flysch Numidico (Early Miocene to Middle Miocene) and the Flysch Irpini (Middle Miocene to Late Miocene).

### 3.1.2 Lithostratigraphy of the Late Triassic sequence

Focusing on the Late Triassic, the Carnian to Norian/Rhaetian is dominated by hemipelagic carbonatic sedimentation (Calcarei con Selce), increasing in siliceous sedimentation around the Norian/Rhaetian boundary (Intervallo di Transizione-Transitional Interval; Miconnet, 1983) and passing to the Scisti Silicei in the Rhaetian/Hettangian (Fig. 3.3).

The Calcarei con Selce is made of limestones and marly limestones, usually silicized towards the end of the succession, with chert nodules and bends, interbedded with dark, greenish or reddish shales. Limestones are mainly mudstone-wackestone (sometimes packstone) with radiolarians and bivalves (e.g. Rigo et al., 2012a). Thick banks of grey calcarenite are common in proximal successions, sometimes silicized, made of materials coming from near carbonate platforms, such as benthic foraminifera and fragments of echinoderms (e.g. Bazzucchi et al., 2005; Bertinelli et al., 2005). Micritic sediments might have been exported from neighbour platforms by currents in the nepheloid layer (Passeri et al., 2005), even if a major contributor...
up to ca. 60 % to deep-water late Rhaetian carbonate hemipelagic lime-mudstones in western Tethys could have been derived from planktonic calcareous nanofossil (e.g. Prinsiosphaera) (Di Nocera and Scandone, 1977; Bown, 1987; Preto et al., 2012). Miconnet (1983) suggest a submarine delta as the main depositional environment of the carbonatic sequences of the Lagonegro. In facts, inside calcarenite is common to find fining-upward layers, similar to the Ta layers of a Bouma turbidity sequence. The boundary with the underlying Monte Facito Fm is not well defined and considered diachronous, from the upper Ladinian (Mietto and Panzanelli Fratoni, 1990; Rigo et al., 2007) to the lower-middle Carnian (Miconnet, 1983; Marsella et al., 1993).

The passage to the overlying Scisti Silicei is gradual, with the increase of terrigenous sediments, with few cherty limestone layers and abundant red shales, red cherts and radiolarites (Transitional Interval; Amodeo, 1999). To explain this change of sedimentation many causes should be invoked: a decrease in carbonate supply from the nearby platforms, a deepening of the basin below the Calcite Compensation Depth (CCD), a rise of the CCD due to upwelling (Passeri et al., 2005), a bloom of siliceous organisms due to an increased nutrient supply (i.e. for volcanic activity; Giordano et al., 2011). Also the base of the Scisti Silicei is diachronous, from late Norian to Rhaetian (Amodeo, 1999; Rigo et al., 2015).

Figure 3.3: Stratigraphy of the Lagonegro Basin sequence

<table>
<thead>
<tr>
<th>PERIOD</th>
<th>Stage</th>
<th>Formations</th>
</tr>
</thead>
<tbody>
<tr>
<td>JURASSIC</td>
<td>Hettangian</td>
<td>Scisti Silicei</td>
</tr>
<tr>
<td></td>
<td>Rhaetian</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Norian</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Carnian</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Ladinian</td>
<td>Mt. Facito Fm</td>
</tr>
</tbody>
</table>

Because of the difficulties to identify a sharp boundary between the Calcari con Sele and the Scisti Silicei, Miconnet (1982) introduced the “Transitional Interval” as the upper part of the Calcari con Sele, characterized by an increase of red radiolaritic intercalations, typical of the overlying Scisti Silicei (Amodeo 1999; Bertinelli et al. 2005a; Passeri et al. 2005; Reggiani et al. 2005; Rigo et al. 2012b). The Transitional Interval is included in the Calcari con Sele, as not official member (Amodeo, 1999). The base of the Transitional Interval is marked by a 2.5-4 m-thick interval of red shales (Amodeo 1999; Bertinelli et al. 2005b; Reggiani et al. 2005; Rigo et al. 2012b), Sevatian 1 in age (late Norian - Mockina bidentata Zone) (Rigo et al. 2005, 2012b). Deeper investigations on the Transitional Interval, mainly on
biostratigraphy (e.g. De Wever and Miconnet 1985; Bertinelli et al. 2005; Passeri et al. 2005; Reggiani et al. 2005; Rigo et al. 2005, 2012b), revealed atypical features in the Pignola-Abriola section, while it is easily recognized in the Mt. Volturino section.

3.1.3 Sections of interest
For my investigations I chose two sections from the Lagonegro Basin: the Pignola-Abriola section (Norian/Rhaetian) and the Pignola-2 section (Carnian). The Pignola-Abriola section is a continuous succession of pelagic cherty limestone with interbedded shaley levels (Calcarí con Selce), characterized by a detailed conodonts and radiolarians biostratigraphy. This section includes the Norian/Rhaetian boundary, well defined by the FO of conodont Misikella posthernsteini and supported by the concurrent radiolarian biostratigraphy (base of the Proparvicingula moniliformis Zone). Moreover, the low Conodont Alteration Index (CAI = 1.5; Giordano et al., 2010) indicate that the succession is preserved by thermal stress, that make the section suitable for paleomagnetic analyses. These characteristics make the Pignola-Abriola section very useful for my attempt of magnetostratigraphic integration of the Rhaetian.

The Pignola-2 section is a succession of pelagic cherty limestone (Calcarí con Selce) covering the middle Carnian, at the passage from Julian to Tuvalian substages. The section is characterized by a detailed conodonts and palynomorphs biostratigraphy and by a low CAI (1.5; Rigo et al., 2007). The section records the Carnian Pluvial Event as a 5 m-thick green clayey and radiolaritic interval. Radiometric U/Pb age of 230.91±0.33 Ma was obtained from one of the ash-beds inside this green interval (Aglianico ash-bed; Furin et al., 2006). Considering the radiometric and biostratigraphic constraints, the paleomagnetic investigation of this section could provide a valid contribution to the magnetostraigaphy of the Carnian.

3.2 DOLOMITES (NORTHERN ITALY)

3.2.1 Structural Framework
The Dolomites are a mountain group covering the areas of Belluno, Trento and Bolzano (Northern Italy). Their name derives from the French geologist Déodat de Dolomieu, discoverer of the mineral dolomite in the Adige Valley, on the western Dolomites. The Dolomites rose rapidly during the Pliocene-Pleistocene because of the Alpine orogeny, and have been modelled and eroded quickly during the Pleistocene
glacial phases. Dolomites are located in the central part of the Southern Alps, a south-vergent belt not affected by metamorphism (e.g. Castellarin and Doglioni, 1985). This area is limited northward by the Insubric Line (dextral transpressive) and southward by the Valsugana Thrust (e.g. Doglioni, 1987; Castellarin et al., 1998). As the rest of the Alps, the Dolomites was part of the African continental margin in Tethys (Ziegler, 1988; Dercourt et al., 1993). Sedimentary record covers the Permian to Cretaceous interval and the slight deformation affecting the area did not disrupt most of the sedimentary frameworks (De Zanche et al., 1993; Gianolla et al., 1998a). The Late Triassic is well preserved and represented by carbonatic sedimentation in platform and basin in the very early part, that leads to a flatten topography in late Carnian and Norian (Fig. 3.4), with the development of anoxic intraplatform basins in the Norian and Rhaetian (e.g. De Zanche et al., 1993; Gianolla et al., 1998a; Neri et al., 2007).

![Figure 3.4: Triassic stratigraphic sequence in Dolomites. (from Gianolla et al., 1998a)](image)

### 3.2.2 Lithostratigraphy of the Late Triassic sequences

The Late Triassic sequences in central Dolomites represent the recovery of the carbonate sedimentation after the Ladinian crisis due to intense volcanism (e.g. De Zanche et al., 1993).

Cassian platforms (e.g. Leonardi, 1968; Bosellini, 1984; De Zanche et al., 1993; Bosellini et al., 2003) are intensively dolomitized, but the depositional macrostructures (geometry, different paleoenvironments) are still well recognizable (e.g. Gianolla et al., 2008). Information about the characteristics of the platform margins derives from the deposits and the olistoliths in the coeval San Cassiano Formation (e.g. Gianolla et al., 2008). The platform margin was probably made of biogenic...
reefs (boundstone dominated by microbialitic automicrite, with colonial corals) cross-bounded with calcarenitic shoals. The internal lagoon shows peritidal cycles with subtidal fine-grained carbonates alternated to stromatolites, tepees and pisolitic beds of supratidal environment.

The San Cassiano Formation is a basinal sequence of calcarenites, marls and volcanoclastic arenites (basal member) and a mostly carbonatic member in the upper part of the formation (De Zanche et al., 1993; Keim et al., 2006; Neri et al., 2007). The San Cassiano Formation is different from the underlying Wengen-La Valle Formation for the presence of carbonatic deposit transported from the coeval prograding Cassian platforms (i.e. oolitic-bioclastic turbidites). The quantity of volcanoclastic terrigenous mainly depends on the paleogeography: the higher contents of volcanoclastic sediments are near the old volcanic areas.

The overlying Heiligkreuz Formation (De Zanche et al., 1993; Preto and Hinov, 2003; Neri et al., 2007; Breda et al., 2009; Gattolin et al., 2013, 2015) represents the beginning of the Cassian Basins closure (Fig. 3.5). The decreased subsidence led to a strong regression in all the Cassian Basins, and the carbonatic/terrigenous coastline starts prograding fast (as suggested by the decreasing of the platforms slope angle). The Heiligkreuz Fm is in concordance with the San Cassiano Fm and in onlap to the Cassian platforms, sometimes overlapping the platform top in discoformity (in this cases the platform top shows karst erosion). The Heiligkreuz Fm is divided in three members (Neri et al., 2007) (Fig. 3.6):

- Borca Member: dolomitic limestone, arenitic dolostone and arenites, commonly interbedded with thin shale layers. Limestones dominate the base of the member, whereas the top is made of well-stratified dolostones with marly layers in peritidal cycles (with stromatolites), ending with paleosols.
- Dibona sandstones Member: conglomerates, sandstones and pelites, interbedded with limestones. Remains of plants are abundant, sometimes in thin coal layers.
- Lagazuoi Member: arenaceous dolostone, sandstones with carbonatic cement, oolitic-bioclastic dolomitic arenites/calcarenites.

The Travenanzes Formation (De Zanche et al. 1993, Neri et al. 2007, Breda and Preto 2011) deposited after the filling of the Cassian Basins in a low-gradient coastal environment (Breda and Preto, 2008) (Fig. 3.5). The Travenanzes Formation records the passage from a relative humid period (Heiligkreuz Formation) to an arid period. The contact with the underlying Heiligkreuz Formation is sharp, just above the Lagazuoi Mb. The Travenanzes Fm is made of mixed carbonatic-siliciclastic sediments, with marly levels at the base, passing to mainly siliciclastic sedimentations (shales, pelites, fine arenites) with levels of evaporites (mainly primary dolomite) indicating an arid lagoon environment (sabkha-type) (e.g. Breda et al., 2009; Breda and Preto, 2011). The passage to overlying Dolomia Principale is gradual, with an increase of carbonatic levels, dominated by dolomitic peritidal cycles of carbonate tidal-flat and shallow lagoon environments (e.g. Breda and Preto, 2011).

The Dolomia Principale is a very thick (max. 2000 m) sequence of dolostones and dolomitic limestones, outcropping from western Lombardy to western Slovenia (e.g. Bosellini and Hardie, 1988). The Dolomia Principale deposited in a shallow water environment, far away from coastline and controlled by constant subsidence persisting for several million years (in some areas for all the Norian). The sequence recorded peritidal cycles, made of grainstones with bivalves and gastropods passing to massive dolostones with *Megalodon* and *Worthenia* (subtidal facies), thin layers of stromatolitic dolostones (intertidal to supratidal facies), sometimes overlaid by a pedogenetic level with tepee structures (supratidal facies) (Gianolla et al., 1998a).
3.2.3 Section of interest
For paleomagnetic investigations I chose the Dibona section from the Dolomites, in order to check and integrate the magnetostratigraphy of the coeval Pignola-2 section. The Dibona section is a shallow-marine succession represented by limestones with fine to coarse-grained sandstones and shales (Heiligkreuz Fm), overlaid by limestones with interbedded shales that pass quickly to a shale-dominant sequence with interbedded limestones and evaporites (Travenanzes Fm). The section is characterized by palynomorphs biostratigraphy, comparable with the Pignola-2 record.

3.3 WOMBAT PLATEAU (NORTHWESTERN AUSTRALIA)

3.3.1 Geological setting
The Wombat Plateau is located in the Indian Ocean (~453 km from the northwestern Australia coast), and was cored during a ODP (Oceanic Drilling Program) campaign in 1988. The mission was named Leg 122 and divided in six sites: 759, 760, 761, 764 from the Wombat Plateau (16°44’21.78”S; 115°29’12.3”E), and 762, 763 from the Exmouth Plateau (20°44’47.57”S; 112°31’26.45”E) (Fig. 3.7). The sites 759, 760 and 761 cover a period from the Quaternary to the Lower-Upper

![Figure 3.7: geographic position of Wombat Plateau (Sites 759-760-761-764) and Exmouth Plateau (Sites 762 and 763). From Haq et al. (1990).](image)
Cretaceous, and then, after an unconformity, from Rhaetian to Carnian (Late Triassic). The variation in lithology reveals a passage between different environments, with facies typical of delta, shallow marine and tidal flat, where sedimentation is both carbonatic and siliciclastic. The Late Triassic portions of the Leg 122 sites have a sufficient recovery in the Carnian and early Norian sediments, whereas in the late Norian and Rhaetian part most of the material has been lost. Here are briefly presented the features of Sites 759, 760 and 761, which will be discussed in detail in the respective chapters.

3.3.2 Sections of interest
In the Wombat Plateau I chose three Sites for my paleomagnetic analyses: 759, 760 and 761. These three sites cover the upper Carnian to middle Rhaetian interval, which means a potential magnetostratigraphic record for the great part of the Late Triassic.

Site 759 (Carnian/Norian):
Shales and siltstones, with some intercalations of arenite and limestone, are representative of the lower part of the site (Carnian). Moving upward, the limestone levels increase and become dominant in the upper part of the Norian portion of the site. Limestones are mainly mudstones and wackestones, sometimes packstones, containing peloids, calcareous lithoclasts, skeletal fragments (mollusk shells, echinoids ossicles, green algae), oncos, and mudstone intraclasts. Also benthic foraminifers are present and often micritized. Most of the mudstones and wackestones show burrowing marks, responsible of the mottled color of these layers. Coal layers and coalified roots have been noticed in the terrigenous intervals. Diagenetic siderite (FeCO₃) is also present in the deepest cores (Carnian), in layers and surrounding the burrows.

Site 760 (Carnian/Norian):
This site is mostly siliciclastic, in particular in the Norian. The main lithology is black to dark gray claystones interbedded with dark greenish/gray clayey siltstones to silty sandstones, with occasional mollusk fragments, glauconite and coal. Fossiliferous limestones appear around the Carnian/Norian accompanied with siliciclastic dark claystones. Siliciclastic sediments become dominant in the Carnian, with dark silty claystones with minor greenish/gray silty sandstones. Molluscal shells and sideritic nodules are present.
Site 761 (Norian/Rhaetian):

In this site, limestones are concentrated in the youngest Rhaetian cores. White and pale brown limestones become dark grey/black in color in the lower Rhaetian, where rhythmical alternations with calcareous claystones are frequent. In the basal Rhaetian and upper Norian dark to black carbonaceous claystones are the main lithology, with subordinate crinoidal limestones in the Rhaetian. In the Norian silty claystones are also frequent, with coal levels.

3.4 PELAGONIAN DOMAIN (GREECE)

3.4.1 Geological setting

The Pelagonian Domain is an Early Permian to Late Jurassic sedimentary sequence, deposited on the metamorphic Pelagonia terrane, a Variscan crystalline basement of Late Carboniferous (Stampfli et al., 1998; Vavassis et al., 2000) and interpreted as a consequence of the rifting and spreading of the Maliac oceanic basin (De Bono et al., 2001). The Pelagonia was a part of the Africa/Apulian plate, along the south-western margin of the Tethys Ocean (e.g. Ciarapica and Passeri, 2002).

![Figure 3.8: Stratigraphic sequence of Pelagonian Domain (from Romano et al., 2008)](image)

Late Triassic is characterized by the emplacement of a wide carbonate platform that survived up to the Late Jurassic (e.g. Tataris et al., 1970; Celet and Ferrière, 1978; Celet et al., 1988), although in many areas is recorded until Early Jurassic (Pliensbachian) (e.g. Parginos et al., 2007) (Fig. 3.8). Large part of this platform has
been metamorphosed or intensely tectonized.

### 3.4.2 Section of interest

One of the few well-preserved Triassic/Jurassic portion of the platform is represented by the Mount Messapion section, located in the area of Chalkida, in the eastern part of central Greece. The section is a ~710 meters Late Triassic – Early Jurassic sequence of limestones, dolomitic limestones and dolomites (Fig. 3.9). The whole section is a pile of shallowing-upward peritidal cycles (Romano et al., 2008). The Mt. Messapion section has been subdivided in three units (A, B, C) based on sedimentary structures and fossil content (Fig. 3.9):

- **Unit A** – 70 meters: meter-scale shallowing upward peritidal cycles, bioclastic-intraclastic wackestones/packstones with rare Megalodontids (subtidal) passing gradually into laminated loferites (supratidal) with irregular and laminoid fenestrae.

- **Unit B** – 230 meters: peritidal cycles similar to those described in Unit A. The differences are the tepee structure inside the supratidal levels.

- **Unit C** – 410 meters: peritidal cycles with tepee structures, and with well-developed microbial intervals in the inter-supratidal part of the cycles. Unit C is subdivided in subunits C1 (180 meters) and C2 (230 meters). Subunit C2 is different from C1 for the absence of Megalodontids and bioturbation, and for the gradual enrichment in oolitic bioclastic grainstones-packstones, with benthic foraminifera, green algae, gastropods and bivalves.

For my paleomagnetic investigations I considered the interval between subunits C1 and C2, around the Triassic/Jurassic boundary.

![Figure 3.9: Stratigraphy of Mount Messapion section (from Romano et al., 2008). The bracketed bar represent the investigated area.](image)
Chapter IV
INVESTIGATION ON SELECTED TETHYAN MARINE SECTIONS

4.1 RHAETIAN

As explained in Chapter 1, Rhaetian is affected by uncertainties about its base, mainly due to the different (and mostly discordant) markers used to place the Norian/Rhaetian boundary in stratigraphic sections. The following Tethyan marine sections have been analyzed for magnetostratigraphy in order to contribute to the resolution of this problem.

4.1.1 Pignola-Abriola section (from Maron et al., 2015: GSA Bulletin, v. 127, p. 962-974; see Attached publications)

Here are presented new biostratigraphic, magnetostratigraphic, and chemostratigraphic data from the Pignola-Abriola section of Italy. This section

Figure 4.1: The Pignola-Abriola sections (A, B) are located in the southern Apennines, near Potenza (southern Italy). The main section (A) crops out on the western flank of Mount Crocetta, along the main road SP5 connecting the towns of Pignola and Abriola (40°33'23.50"N, 15°47'1.71"E), whereas the auxiliary subsection (B) crops out close to an unused railway tunnel located ~10 m below the SP5 road level (40°33'24.74"N, 15°46'59.59"E).
records the FAD of *Misikella posthernsteini*, occurring in the lower *Proparvicingula moniliformis* radiolarian zone (Giordano et al., 2010). I date these events by means of magnetostratigraphic correlation with the Newark APTS, while addressing in detail the taxonomic complexities vexing the use of the conodont *M. posthernsteini* as proxy for the Norian-Rhaetian boundary level. I also illustrate the occurrence of a prominent negative δ^{13}Corg excursion at meter level 44.5, ~0.5 m below the FAD of *M. posthernsteini* (within the base of the *P. moniliformis* zone), which serves as a useful geochemical proxy for the Norian- Rhaetian boundary level.

**Geological Setting**

The Pignola-Abriola section crops out on the western side of Mount Crocetta, along the road SP5 connecting the village of Pignola to Abriola (Potenza, southern Italy; Fig. 4.1, section A, coordinates: 40°33′23.50″N, 15°47′1.71″E). The road section is ~58 m thick (Fig. 4.2, left panel) and is complemented by an ancillary 7-m-thick subsection (Fig. 4.2, right panel) outcropping close to a unused railway tunnel located ~10 m below the SP5 road level (Fig. 4.1, section B, coordinates: 40°33′24.74″N, 15°46′59.59″E). The stratigraphic sequence is composed of the Calcari con Selce (i.e., Cherty Limestone) Formation, which was deposited in the Lago negro Basin, a branch of the western Tethys Ocean characterized by pelagic sedimentation since the Permian (Finetti, 1982, 2005; Catalano et al., 2001; Ciarapica and Passeri, 2002, 2005; Argnani, 2005; Rigo et al., 2012a). The Calcari con Selce Formation consists of thinly bedded cherty hemipelagic to pelagic limestones (mudstones, wackestones, and rare packstones), interbedded with shales and marls, with common radiolarians, conodonts, and sporadic bivalves. The lower part of the section is dominated by cherty limestones, often dolomitized, intercalated with very thin marls or clayey levels (Fig. 4.2). The upper portion is instead dominated by an alternation of silicified limestones and black to brown or greenish, thinly laminated shales (Fig. 4.2), which are rich in organic matter, indicating deposition in dysoxic or anoxic conditions. Calcareous intercalations are also present through the section (Fig. 4.2). In particular, a 1.5-m-thick calcarenitic bank at ~35 m from the base of the measured section has been used as a lithostratigraphic marker to correlate the Pignola-Abriola road section (Fig. 4.2, left panel) to the railway tunnel subsection (Fig. 4.2, right panel).

**Biostratigraphy**

The fossil content of the Pignola-Abriola section consists mainly of conodonts
Figure 4.2: The Pignola-Abriola sections. From left to the right: conodont and radiolarian biostratigraphy (see Plate 4.1 for key species), lithostratigraphy, virtual geomagnetic pole (VGP) latitudes calculated from characteristic remanent magnetization (ChRM) component directions, and derived magnetostratigraphy and chemostratigraphy ($\delta^{13}$C$_{org}$) of the Pignola-Abriola section. To the right is lithostratigraphy and VGP latitudes of the auxiliary subsection B. Black is normal polarity, and white is reverse polarity. The levels containing the first appearance datum (FAD) of conodont Misikella posthernsteini sensu stricto and the marked decrease in the $\delta^{13}$C$_{org}$ to $-30\%$ used to define the Norian-Rhaetian boundary are highlighted by dashed horizontal lines.
and pyritized radiolarians. Here, I present an updated conodont and radiolarian biostratigraphy (Fig. 4.2) after recent biostratigraphic data published by Rigo et al. (2005), Bazzucchi et al. (2005), and Giordano et al. (2010).

Conodonts are well distributed along the entire section (representative specimens are shown in Plate 4.1) and are characterized by a conodont alteration index (CAI) of 1.5 (Epstein et al., 1977; Bazzucchi et al., 2005; Rigo et al., 2005). The following main events have been recognized (Fig. 4.2):

1. the first occurrence (FO) of Mockina bidentata at meter 7;
2. the FO of Misikella hernsteini at meter 21.5, associated with the FO of Parvigondolella andrusovi;
3. the FO of the Misikella hernsteini/posthernsteini morphocline at meter 33.5;
4. the FO of Misikella buseri at meter 32;
5. the FAD of Misikella posthernsteini at meter 45 in sample PIG24, in association with Misikella koessenensis; and
6. the FO of Misikella ultima at meter 54.

The radiolarian associations are well preserved and conform to the biozonation proposed by Carter (1993):

1. Sample PR14 at meter 25 yielded a radiolarian assemblage referable to the Betraccium deweveri zone (Carter, 1993) and consisting of Betraccium deweveri Pessagno and Blome, Praemesosaturnalis gracilis Kozur and Mostler, Tetraporobrachia sp. aff. T. composita Carter, Ayrtonius elizabethae Sugiyama, Citriduma sp. A sensu Carter (1993), Globolaxtorum sp. cf. G. hullea Yeh and Cheng, Lysemela sp. cf. L. olbia Sugiyama, Livarella valida Yoshida and Livarella sp. sensu Carter (1993) (Giordano et al., 2010); a similar assemblage was found also in sample PR15 at meter 23.5,
and sample PR13 at meter 27. The presence of *Globolaxtorum* sp. cf. *G. hullae* Yeh and Cheng in this assemblage is atypical, because the genus *Globolaxtorum* is usually referred only to the *Proparvicingula moniliformis* and *Globolaxtorum tozeri* zones (O’Dogherty et al., 2009).

2. Sample PA25 at meter 41 yielded a radiolarian assemblage referable to the *Proparvicingula moniliformis* zone assemblage 1 (U.A. 2–5 in Carter, 1993)
for the presence of *Fontinella primitiva* Carter, *Praemesosaturnalis* sp. cf. *P. sandspitensis* Blome, *Globolaxtorum* sp. cf. *G. hullae* Yeh and Cheng, and *Livarella densiporata* Kozur and Mostler (Bazzucchi et al., 2005; Giordano et al., 2010).

The Norian-Rhaetian boundary is conventionally placed in stratigraphic levels where the FAD of *Misikella posthernsteini* is documented (Krystyn, 2010), which is a phylogenetic descendent of *M. hernsteini* (e.g., Mostler et al., 1978; Kozur and Mock, 1991; Giordano et al., 2010). The transition from drop-shaped to heart-shaped basal cavity along with a reduction of the number of blade denticles characterize the evolution of the *M. hernsteini/posthernsteini* morphcline (Giordano et al., 2010). Specimens characterized by an evident furrow on the backside of the cusp and the associated inflection of the posterior margin of the basal cavity are here considered *Misikella posthernsteini* sensu stricto, as suggested by Giordano et al. (2010). At Pignola-Abriola, the presence of the *Misikella hernsteini/posthernsteini* morphcline, as well as the presence of the FAD of *M. posthernsteini* sensu stricto (m 45, sample PIG24) provide a reliable (and continuous) biostratigraphic signal. Furthermore, in the Pignola-Abriola section, the conodont *Misikella posthernsteini* sensu stricto appears 4 m above the base of radiolarian *Proparvicingula moniliformis* zone assemblage 1 (Fig. 4.2), which is commonly adopted to define the early Rhaetian (e.g., Carter, 1993; Bertinelli et al., 2005; Giordano et al., 2010).

**Geochemistry**

In total, 41 samples, mostly black to brown shales, from the upper portion of the Pignola-Abriola section (from meter 30 to the top of the section) were analyzed for $\delta^{13}C_{org}$ (data in Appendix A.1). The rock samples were pulverized and acid-washed with 10% HCl in a 70°C water bath for 3 h, and the process was repeated at least three times to thoroughly remove pyrite and carbonates. The samples were subsequently neutralized with high-purity water, dried at 30°C overnight, and then wrapped in tin capsules and analyzed for their isotopic composition. The analyses were carried out using a GVI Isoprime continuous flow-isotope ratio mass spectrometer (CF-IRMS) at Rutgers University, adding multiple blank capsules and isotope standards for each batch of isotopic analyses (NBS 22 = −30.03‰; Coplen et al., 2006) plus a matrix matched in-house standard. Standard deviations for $\delta^{13}C_{org}$ standards during the period of analysis were better than $\sigma = 0.2\%$.

The $\delta^{13}C_{org}$ values of the Pignola-Abriola section are between −29.95‰ and −23.70‰ (Fig. 4.2). After a moderate increase in $\delta^{13}C_{org}$ (from −27.5‰ to −24‰ from meter
30 to 36), a large decrease to $\sim30\%$ was recorded for meter 36 to meter 44.5, immediately followed by a rapid return to higher values ($\sim25\%$, $\sim20$ cm above). A subsequent decrease of $\sim2\%$ is recorded at meter 53.5 (close to the level containing the FO of *Misikella ultima*; Fig. 4.2). Notably, the low $\delta^{13}C_{\text{org}}$ of $\sim30\%$ at meter 44.5 is just below the level containing the FAD of *Misikella posthernsteini* sensu stricto, and within the base of the *Proparvicingula moniliformis* zone (Fig. 4.2).

**Paleomagnetism**

In total, 220 oriented core samples were collected from the Pignola-Abriola section and analyzed at the Alpine Laboratory of Paleomagnetism (Peveragno, Italy). Rock magnetic properties were studied on a representative set of samples by means of thermal decay of a three-component isothermal remanent magnetization (IRM)

![Figure 4.3: Thermal demagnetization of a three-component isothermal remanent magnetization (IRM) (A) and IRM acquisition curves (B) for representative samples from Pignola-Abriola showing the presence of a variable mixture of hematite and magnetite. See text for discussion.](image)

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impacted at fields of 2.5 T, 0.4 T, and 0.12 T (Lowrie, 1990) and IRM acquisition curves. The lower part of the section (samples P1.34, P3.10, P3.43; Fig. 4.3A) is characterized by a high-coercivity mineral with maximum unblocking temperatures (TB) of 650–675 °C, attributed to hematite, coexisting with a lower-coercivity mineral with TB of 525–575°C, interpreted as magnetite; an inflection at ~350°C in the 0.4 T curve observed in sample P1.34 suggests the presence of iron sulfides. Samples from the upper part of the section (GNM497 at 33 m; GNM48 at 43.5 m; GNM119 at 57 m) appear dominated by the high-coercivity hematite phase (Fig. 4.3A). IRM curves of these samples show no tendency to saturate even at applied fields of 2.5 T (Fig. 4.3B). The cumulative log-Gaussian (CLG) analysis (Kruiver et al., 2001) reveals the presence in these samples of two magnetic phases with contrasting coercivities: a high-coercivity phase with coercivity of remanence ($B_{1/2}$) = 1.6–2 T, which accounts for ~60%–85% of the IRM, and a subordinate low-coercivity phase with $B_{1/2} = 0.1$ T, which accounts for the reminder of the IRM (Fig. 4.4). The presence of higher amounts of (detrital) hematite in the upper part of the section may correlate with the increase in terrigenous input (shales and marls) observed in the upper part of the section (Fig. 4.2).

The natural remanent magnetization (NRM) of samples, measured on a 2G Enterprises DC- SQUID cryogenic magnetometer, is on average 0.08 mA/m. All

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**Figure 4.4:** Cumulative log-Gaussian (CLG) IRM curves for samples GNM 497, 48, 119 (see text for details). The saturation magnetization (SIRM), the coercive force ($B_{1/2}$) and the dispersion parameter (DP) are listed for each magnetic component.
samples were thermally demagnetized in steps of 50 °C or 25 °C up to a maximum of 675 °C, and the component structure of the NRM was plotted on vector end-point demagnetization diagrams (Fig. 4.5; Zijderveld, 1967). After removal of spurious magnetizations between room temperature and ~100–300°C, a characteristic remanent magnetization (ChRM) was isolated up to 450–550°C (maximum of 625 °C) in ~55% of the samples (N = 121; ChRM data in Appendix A.1) and found to be broadly oriented either N and down or S and up in tilt-corrected coordinates (Fig. 4.6). These ChRM component directions are distributed in tilt-corrected

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<th>TABLE 4.1: PALEOMAGNETIC DIRECTIONS AND POLE FROM THE PIGNOLA-ABRIOLA SECTION</th>
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<tr>
<td><strong>Mean directions from the Pignola-Abriola section</strong></td>
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<tr>
<td>IN SITU</td>
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<tr>
<td><strong>Comp.</strong></td>
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<tr>
<td>ChRM</td>
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<tr>
<td><strong>Paleomagnetic pole, paleolatitude and rotation from Tilt Corrected Filtered ‘Ch’ directions, corrected for inclination flattening</strong></td>
</tr>
<tr>
<td>Lat.</td>
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<tr>
<td>max.</td>
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Note:  
Comp.: paleomagnetic component  
N: number of samples  
k, α50°S: Fisher statistics parameters  
Dec.: mean declination  
Inc.: mean inclination  
Inc. corr.: mean inclination corrected for inclination flattening  
f: flattening factor  
Lat.: Latitude  
Long.: Longitude  
A95: circular confidence limit  
Rotation: tectonic rotation of the site (relative to the 201 Ma Adria-Africa reference paleopole of Muttoni et al., 2013)
coordinates around an overall mean of Dec = 15.9°E, Inc = 32.5° (k = 8.4, α95 = 4.7°, N = 121; Table 4.1). No fold test could be performed because of the homoclinal bedding tilt of the section, whereas the reversal test (McFadden and McElhinny, 1990) was positive. Based on these results and the consistent magnetostratigraphic correlations with sections from the literature, as discussed later herein, we regard the ChRM component as primary in origin. I checked the ChRM component directions for sedimentary inclination shallowing due to sedimentary and/or compaction processes. The elongation/inclination (E/I) statistical method of Tauxe and Kent (2004) was applied to the ChRM directions, obtaining a flattening factor of f = 0.6 and a corrected mean inclination of 47.7° (min = 39.0°, max = 53.7°), corresponding to a paleolatitude for Pignola-Abriola of ~28.8°N (Table 4.1). A paleomagnetic pole was calculated for Pignola-Abriola using the tilt-corrected mean ChRM direction corrected for inclination shallowing (lat. = 72.5°N, long. = 143.0°E; Table 4.1) and compared to the 201 Ma Adria-Africa paleopole of Muttoni et al. (2013) located at lat. = 69.3°N, long. = 243.8°E. The Pignola-Abriola paleopole is displaced by ~32.8° clockwise relative to the reference Adria-Africa paleopole, probably as a result of vertical-axis tectonic rotation of the sampling area during Apennine tectonics.

A virtual geomagnetic pole (VGP) was calculated for each ChRM component direction in tilt-corrected coordinates. The latitude of the sample VGP relative to the north pole of the paleomagnetic axis was used for interpreting the magnetic polarity stratigraphy, where VGP latitudes approaching +90° or −90° are attributed to normal or reverse polarity, respectively. An overall sequence of five polarity magnetozones, labeled from magnetozone MPA1 to MPA5, was established starting at the base of the section (Fig. 4.2). Each magnetozone was subdivided into a lower, predominantly normal and an upper, predominantly reverse portion, in which submagnetozones can be embedded. No obvious relation was observed between

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<th>CHARACTERISTIC COMPONENT (ChRM)</th>
<th>GEOGRAPHIC</th>
<th>TILT CORRECTED</th>
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Figure 4.6: Equal-area projections for characteristic remanent magnetization (ChRM) component directions isolated at Pignola-Abriola for in situ (geographic) and tilt-corrected coordinates (see Table 4.1 for Fisher statistics parameters).
magnetic polarity stratigraphy and the magnetic mineralogy of the samples. The FAD of *Misikella posthernsteini* sensu stricto falls within magnetozone MPA5r at ~45 m, while the new proposed Norian-Rhaetian boundary coincident with the $\delta^{13}$C$_{org}$ negative spike occurs inside the same magnetozone at ~44.5 m (Fig. 4.2).

**Discussion**

**Correlations with Tethyan Sections from the Literature**

The magnetostratigraphy of the Pignola-Abriola section is comparable with that of the Steinbergkogel section (Hüsing et al., 2011), which at present is the only global boundary stratotype section and point (GSSP) candidate for the base of the Rhaetian Stage (Krystyn et al., 2007a, 2007b), assuming that the occurrence of conodont *Misikella posthernsteini* at Steinbergkogel (plate 1 in Krystyn et al., 2007a) is equivalent to the FO of *Misikella hernsteini/posthernsteini* transitional forms at Pignola-Abriola (sensu Giordano et al., 2010). Hence, the main reversal portion of the Steinbergkogel magnetostratigraphy from magnetozone ST1/B– to magnetozone ST1/H– at Steinbergkogel STKA section (equivalent to ST2/B– to ST2/H– at Steinbergkogel STKB+C section), has been correlated to magnetozones MPA3r to MPA5r of the Pignola-Abriola section (Fig. 4.7). Also, part of the magnetostratigraphy of the Oyuklu section (Gallet et al., 2007), from magnetozone OyB– to OyD–, is comparable with MPA4r to MPA5r of the Pignola-Abriola section, and with ST/D– to ST/H– of the Steinbergkogel section (Fig. 4.7). Furthermore, the lower portion of the Pignola-Abriola section is magnetostratigraphically correlated with the upper part of the Pizzo Mondello section (Muttoni et al., 2004). Using the updated biostratigraphic calibration of the Pizzo Mondello magnetostratigraphy (Mazza et al., 2012), magnetozones MPA1n to MPA3n at Pignola-Abriola have been correlated to magnetozones PM-8n to PM-12n at Pizzo Mondello (Fig. 4.7). Moreover, data from Pignola-Abriola have been compared with the magnetobiostratigraphy of the Brumano and Italcementi Quarry sections (Lombardian Basin, southern Alps, Italy), which encompasses a portion of the Rhaetian (with specimens attributed to *Misikella*) up to the Triassic-Jurassic boundary as defined by pollens (Muttoni et al., 2010, 2014). Awaiting for a formal redefinition of the *Misikella* specimens in the Brumano section following the new definition of *Misikella posthernsteini* sensu stricto adopted in this study (after Giordano et al., 2010), we stress that all *Misikella* specimens at Brumano occur below the recovered magnetostratigraphy (Muttoni et al., 2010, 2014), and thus the sequence of Brumano-Italcementi Quarry magnetozones from BIT1n to
Figure 4.7: The Norian-Rhaetian magnetostratigraphy, biostratigraphy, and chemostratigraphy of the Pignola-Abriola section correlated to data from marine sections from the literature, such as Steinbergkogel (Hising et al., 2011), which is the current global stratotype section and point (GSSP) candidate for the Rhaetian Stage (Krystyn et al., 2007a, 2007b), Oykulu (Gallet et al., 2007), Brumano-Italcementi Quarry (Muttoni et al., 2010, 2014), and Pizzo Mondello (Muttoni et al., 2004). In this work, specimens originally attributed to Misikella posthernsteini at Steinbergkogel (Krystyn et al., 2007a, 2007b) are here considered M. posthernsteini sensu lato (s.l.) and attributed to the M. hernsteini/posthernsteini “transitional forms” (sensu Giardano et al., 2010). Key biostratigraphic events at Pizzo Mondello are after Mazza et al. (2012). The Pignola-Abriola section is correlated to the Newark astrochronological polarity time scale (APTS, left column) using preferred correlation option 7.1. The lower-right panel shows the derived age model of sedimentation for Pignola-Abriola with an increase in sedimentation rate in the upper part of the section where the terrigenous input is higher. The Norian-Rhaetian boundary, placed at a level coincident with a rapid decrease in δ13C, to ~30‰, which virtually coincides with the level containing the first appearance datum (FAD) of conodont Misikella posthernsteini sensu stricto within the Proparvicingula moniliformis radiolarian zone, is traced within Newark magnetozone E20r at 205.7 Ma.
BIT5n is regarded as largely younger than the Pignola-Abriola magnetostratigraphy (Fig. 4.7).

**Correlation with the Newark APTS**

The correlation between the Pignola-Abriola section and the Newark APTS was performed using the statistical approach proposed in Muttoni et al. (2004). Assuming that thickness is a linear proxy of time, the duration of Newark magnetozones was compared with the thickness of Pignola-Abriola magnetozones (Fig. 4.8). The Pignola-Abriola polarity reversal sequence in linear depth coordinates was placed alongside the top of the Newark APTS (at magnetozone E23r) in linear age coordinates. A linear correlation coefficient (R) relating the thickness of each of the N = 22 complete Pignola-Abriola magnetozones to the duration of the correlative Newark magnetozones was calculated, from which a t-value was derived, where $t = R*\text{sqrt}([N − 2]/[1 − R^2])$, R is the linear correlation coefficient, and N is the number of matching magnetozones in the moving window, i.e. 22. The Pignola-Abriola sequence was then slid by two polarity zones along the Newark APTS (in order to maintain internal polarity consistency in correlation), R and t were recalculated, and the exercise was repeated until all 19 possibilities were explored (Fig. 4.8; statistical procedure with correlation options and analysis of t-values is reported in Appendix A.1).

For N = 22 (the number of matching reversals in moving window), each correlation has 20 degrees of freedom. A Student t-test shows that only correlation coefficients with a t-value larger than 1.725 are significant at the 95% level. According to the Student t-test, only correlation option 19 is reliable at more than 95% confidence level (Fig. 4.8), though it can be excluded on stratigraphic grounds. Precisely, option 19 places the Norian-Rhaetian boundary of the Pignola-Abriola section in the Carnian-Norian portion of the Newark sequence, as deduced from correlations of the Pizzo Mondello and Silická Brezová sections to the Newark APTS (Muttoni et al., 2004; Channell et al., 2003). For this reason, I decided to contemplate correlation options characterized by lower values of t (around 1). As a consequence, options 16 and 7 were considered as acceptable (Fig. 4.8). Option 16 is affected by the same problem as option 19 insofar as it places the Norian-Rhaetian boundary within the Norian Stage as implied by the Pizzo Mondello and Silická Brezová to Newark correlations discussed earlier. In addition, option 16 also implies sudden (and unexplained) variations in sediment accumulation rates of the Pignola-Abriola section.
Figure 4.8: Pignola-Abriola geomagnetic reversal sequence in linear depth coordinates was slid aside the Newark astrochronological polarity time scale (APTS) in linear age coordinates maintaining the internal polarity coherency, and a t value was calculated for each of the 19 possible correlation options. Positive (negative) t-values refer to positive (negative) slopes of the linear function relating Pignola-Abriola magnetozone thickness to Newark magnetozone duration. Statistically significant options 16 and 19 were rejected for incoherence with the available stratigraphic data, whereas a modified version of correlation option 7, termed option 7.1, is considered the best solution that is in agreement with (or does not violate) the available stratigraphic data. See text for discussion.
Correlation option 7 results are more coherent with the available magneto-biostratigraphic correlations of Tethyan sections to the Newark APTS and will be investigated in detail. Option 7 links Pignola-Abriola magnetozone MPA1n with Newark E13n.1n at the base, and magnetozone MPA11r with E20r at the top (Fig. 4.8). However, this correlation implies sudden variations in sediment accumulation rates in the middle of the Pignola-Abriola section. Moreover, the lower part of Pignola-Abriola is considered Sevatian (late Norian) in age, but according to correlation 7, it should correspond to Newark magnetozones considered close to the Carnian-Norian boundary (Fig. 4.8; see also Muttoni et al., 2004).

In conclusion, no statistical correlation matches perfectly, and some “adjustments” are necessary. An alternative version of statistical correlation option 7, termed option 7.1 (black bar in Fig. 4.8), solves the problems outlined for option 7, increases statistical significance, and is coherent with the Pizzo Mondello and Pignola-Abriola magneto-biostratigraphies and correlations to the Newark APTS. Preferred option 7.1 is similar to statistical option 7 in range (MPA1n corresponding to E13n.1n, and MPA11r corresponding to E20r.2r) but differs from statistical option 7 in linking MPA1n with E13n.2n, MPA1r (0.1r, 0.1n, 0.2r) with E13r, MPA2n (0.1n, 0.1r, 0.2n) with E14n, and MPA11r with E20r, whereas magnetozones from MPA2r to MPA11n have been correlated with Newark magnetozones from E14r to E20n. Preferred correlation option 7.1 implies a correlation of the Steinbergkogel section to the Newark APTS from magnetozone E17n to E21n (Fig. 4.7) that is substantially equivalent to the correlation originally proposed by Hüsinger et al. (2011). The correlation of the largely younger Brumano-Italcementi Quarry sections to the Newark APTS is the same of Muttoni et al. (2010, 2014), pending a formal redefinition of the Misikella specimens at Brumano (see also earlier discussion).

Using preferred correlation option 7.1, an age model for the Pignola-Abriola section can be derived. The age model shows a change in sedimentation rate from the lower to the upper part of the section (Fig. 4.7). From the base to meter 24.5, the mean sedimentation rate is of ~2.6 m/My, while from meter 24.5 to 40, the mean sedimentation rate increases to ~5.6 m/My. From meter 40 to the section top, the sedimentation rate increases further to ~9.8 m/My. This is coherent with the lithostratigraphy of the section, suggesting a general increase of terrigenous input in the upper part of the section. According to the proposed age model, the Norian-Rhaetian boundary defined by the level containing the FAD of *M. posthernsteini* sensu stricto at meter 45 should correspond to an estimated age of ca. 205.7 Ma.
(Fig. 4.7), which is substantially equivalent to the age of the prominent negative $\delta^{13}C_{org}$ excursion to $\sim$30‰ observed at meter 44.5 (Fig. 4.7).

**GSSP Proposal for the Base of the Rhaetian Stage**

Based on my magneto-bio-chemostratigraphic study of the Pignola-Abriola section, coupled with the recognition of the taxonomic complexities concerning conodont *Misikella posthernsteini*, the current candidate species for the definition of the base of the Rhaetian Stage, I suggest an alternative option for the definition of the Norian-Rhaetian boundary. I favor placing the boundary at the prominent negative $\delta^{13}C_{org}$ spike observed in the Pignola-Abriola section at meter 44.5 (immediately below the level containing the FAD of *M. posthernsteini* sensu stricto and within the base of the radiolarian *Proparvicingula moniliformis* Zone). A similar $\delta^{13}C_{org}$ perturbation around the Norian-Rhaetian boundary was documented in Canada by Ward et al. (2001, 2004) and Whiteside and Ward (2011), coinciding with the disappearance of large *Monotis* (Ward et al., 2004), a typical proxy for the Norian-Rhaetian boundary (McRoberts et al., 2008). The stratigraphic level in the Pignola-Abriola section containing the $\sim$30‰ spike has been magnetostratigraphically correlated to Newark magnetozone E20r.2r at ca. 205.7 Ma. This age was obtained from the Newark astrochronology, calibrated with the new numerical age of 201.5 Ma from the base of the Orange Mountain Basalts in the Newark Supergroup (Blackburn et al., 2013). Assuming an age of ca. 201.3 Ma for the Triassic-Jurassic boundary (Guex et al., 2012), which is broadly consistent with previous estimates (Schoene et al., 2010), and a proposed age of ca. 205.7 Ma for the Norian-Rhaetian boundary, the Rhaetian Stage would have a duration of $\sim$4.4 My (Fig. 4.7). Using a Carnian-Norian boundary at ca. 227 Ma (Muttoni et al., 2004), the Norian would be the longest stage of the Phanerozoic with a duration of $\sim$21.3 My (but see Lucas et al., 2012). Using an approximated Ladinian-Carnian age of 238 Ma, derived from an uppermost Ladinian radiometric age of 237.77 ± 0.14 Ma (Mietto et al., 2012), the Carnian would have lasted almost 10 My According to these figures, the Late Triassic may have lasted $\sim$36 My.

**GSSP proposal for the Rhaetian Stage: an update (after Rigo et al., 2015)**

The Pignola-Abriola section have been proposed as a GSSP (Global boundary Stratotype Section and Point) candidate for the Rhaetian Stage (Rigo et al., 2015). The other major candidate is the section of Steinbergkogel in Austria (Krystyn et al., 2007a, 2007b). The Steinbergkogel section is a basinal sequence of nodular
limestone, subdivided in three outcrops (Krystyn et al., 2007b): STK-A, ~4 meters thick; STK-B+C, ~9 meters thick; ST-4, ~23 meters thick. STK-A overlaps STK-B+C and the upper part of ST-4. The Norian/Rhaetian boundary is placed with the FAD of conodont *Misikella posthernsteini* at meter 2.2, in level 111A, of STK-A. In the Pignola-Abriola section the FAD of *M. posthernsteini* approximates the Rhaetian base, that have been preferentially placed with the negative $\delta^{13}C_{org}$ peak of ~30% at meter 44.5 (Rigo et al., 2015). Pignola-Abriola is more expanded than Steinbergkogel, covering the same time interval in ~60 meters instead of ~30. Moreover, the magnetostratigraphy is detailed as well as organic chemostratigraphy. The integrated stratigraphy of Pignola-Abriola (litho-bio-chemo-magnetostratigraphy) is time constrained by the correlation with the Newark APTS, confirmed by radiometric age around the Norian/Rhaetian boundary from Peru (Wotzlaw et al., 2014). In addition to the ages of the FAD of *M. posthernsteini* and the negative $\delta^{13}C_{org}$ peak, already placed at ~205.7 Ma, Rigo et al. (2015) provided ages for other events recorded in Pignola-Abriola. The Alaunian/Sevatian boundary (FO of conodont *Mockina bidentata*) is dated at ~216.2 Ma, the Sevatian1/Sevatian2 boundary (FO of *Misikella hernsteini*) is at ~210.8 Ma. The base of the *Proparvicingula moniliformis* radiolarian Zone, considered one of the proxy of the Rhaetian (e.g. Giordano et al., 2010), is placed at ~206.2 Ma, while the FO of conodont *Misikella ultima* is dated ~204.7 Ma. Moreover, the FO of the *Misikella hernsteini/posthernsteini* transitional morphotype, associated to the older *Misikella posthernsteini* specimens *sensu* Krystyn et al. (2007a) (Maron et al., 2015; Rigo et al., 2015), is dated ~207.6 Ma. The Pignola-Abriola section satisfies all the requirements to be proposed as GGSSP candidate for the Rhaetian: the outcrop is well exposed and easily accessible; the section is minimal structured deformed and is continuous; is fossiliferous (conodonts and radiolarians); has a detailed magnetostratigraphy and $\delta^{13}C_{org}$ chemostratigraphy; the Norian/Rhaetian is well defined by biocostratigraphy and time constrained after the magnetostratigraphic correlation with the Newark APTS and other marine sections.

Comparison with Previous Time Scales

I compared my solution with alternative proposals from the literature. Krystyn et al. (2002) used Carnian-Norian data from several Tethyan sections (Kavaalani, Kavur Tepe, Pizzo Mondello lower part, Bolüecktasi Tepe, and Scheiblkogel; see references in Krystyn et al., 2002) to construct a Tethyan composite magneto-biostratigraphic sequence that was correlated to Newark magnetozones E3–E22
and used it to infer a duration of the Rhaetian of only ~2 My Later, Gallet et al. (2007) correlated data from Ouyuklu, Pizzo Mondello (upper part), and the Tethyan composite sequence of Gallet et al. (2003) to the Newark APTS, suggesting that part of the Rhaetian is missing in the Newark sequence, and supporting the ~2 My duration of the Rhaetian as proposed by Krystyn et al. (2002). Muttoni et al. (2010) illustrated that middle Norian (Alaunian) magnetozones in the composite magneto-biostratigraphic sequence of Krystyn et al. (2002) may encompass Newark magnetozones ~E13–E15 rather than ~E13–E17, so that the overlying Sevatian magnetozones may correlate to Newark levels at and immediately above E15 rather than at and above E17 as proposed by Krystyn et al. (2002), thus supporting the existence of a longer (>2 My) Rhaetian.

Coming to more recent times, the long-Tuvalian option of the Geological Time Scale 2012 (Ogg, 2012), which is essentially based on data from Lucas et al. (2012), is characterized by a Carnian-Norian boundary placed at 221 Ma, a Norian-Rhaetian boundary at 205.4 Ma, and a large hiatus in the Newark Supergroup based on inferences from conchostracan biostratigraphy (Lucas et al., 2012, and references therein). According to this option, the preserved portion of the Rhaetian in the Newark Supergroup should have a duration of only ~0.2 My (Lucas et al., 2012). A duration of ~8 My for the Rhaetian, as proposed using marine-Newark magnetostratigraphic correlations by several authors (Channell et al., 2003; Muttoni et al., 2004, 2010; Hüsing et al., 2011), was rejected by Lucas et al. (2012) based on the inference that inserting 7.8 My of missing Rhaetian in the claimed Rhaetian gap of the Newark Supergroup (7.8 My = 8 My of total duration of Rhaetian – 0.2 My of preserved Rhaetian in the Newark) would produce an age for the base of the Newark Supergroup of 240.5 Ma; as Lucas et al. (2012) considered the base of the Newark Supergroup to coincide with the base of the Carnian (based on continental [palynomorphs, conchostracans, tetrapods] biostratigraphy), an age of 240.5 Ma is regarded as inappropriate because it would place the base of the Newark Supergroup close to the age of the Anisian-Ladinian boundary (Mundil et al., 2010). Therefore, a duration of ~8 My for the Rhaetian is considered unacceptable according to Lucas et al. (2012), who instead adopted a duration of ~4 My from Ogg (2004). Under the assumption of a 4 My duration for the Rhaetian and only 0.2 My of Rhaetian time preserved in the Newark Supergroup, Lucas et al. (2012) (and Ogg, 2012 in his long-Tuvalian option) estimated an age of 221.5 Ma for the Carnian-Norian boundary, based on continental biostratigraphy, by counting ~405 ky McLaughlin cycles of Newark astrochronology.
In my opinion, the Rhaetian gap of Lucas et al. (2012) at the basis of the long-Tuvalian option (Ogg, 2012) is flawed by lack of convincing correlations between terrestrial groups and marine-based stage boundaries. For example, conchostracans from the Weser Formation of the Germanic Basin are assigned an early Tuvalian age (late Carnian) because the Weser Formation is considered correlative with the Dolomie de Beaumont of France, which contains marine bivalves considered to be of such age (Lucas et al., 2012). As a further example, the conchostracan fauna from the Coburg Sandstein of the Germanic Basin is considered late Carnian, seemingly because it lies immediately below the beginning of a sporomorph association considered to be late Tuvalian. In general, I find difficult to decipher in Lucas et al. (2012) where and in which stratigraphic context a given continental association was found in direct association with stage-defining marine fossils. The long-Rhaetian option of the Geological Time Scale 2012 (Ogg, 2012) is essentially based on magnetostratigraphic correlations between marine sections bearing stage-defining fossils and the Newark APTS assumed to be continuous in the Rhaetian (Channell et al., 2003; Muttoni et al., 2004, 2010; Hüsing et al., 2011), and it shows a Carnian-Norian boundary at ca. 228.4 Ma and a Norian-Rhaetian boundary at ca. 209.5 Ma. My new time scale for the Late Triassic could be considered an “update” of the long-Rhaetian option of (Ogg, 2012), with a Norian-Rhaetian boundary at 205.7 Ma based on data from this study and a Carnian-Norian boundary at ca. 227 Ma based on correlation of the Pizzo Mondello section with the Newark APTS (both numerical estimates obtained by rescaling the Newark APTS using an age of ca. 201.5 Ma for the base of the Orange Mountain Basalts in the Newark Supergroup; Blackburn et al., 2013). Moreover, my age of 205.7 Ma for the Norian-Rhaetian boundary is coherent with recent U/Pb ages of Wotzlaw et al. (2014) that constrain the Rhaetian base between 205.70 ± 0.15 Ma and 205.3 ± 0.14 Ma.

4.1.2 Mount Messapion section
The Mount Messapion is located near the city of Chalkida (Euboea), in the eastern part of Central Greece (Fig. 4.9). The section is situated along the street leading to the mountaintop, between two hairpin turns respectively at 38°27′52.92″N – 23°28′20.52″E and 38°27′40.2″N – 23°29′46.2″E (Fig. 4.9). The section was already sampled by Pasquale Tiano and Alberto Incoronato (University of Napoli “Federico II”), and I attempted a paleomagnetic analysis on the samples to cover the Rhaetian-Hettangian period as a continuation of the Rhaetian magnetostratigraphy
of Pignola-Abriola section.

**Geological Setting**

The section of the Mount Messapion is a ~710 m long carbonatic sequence covering the Norian to Hettangian interval, including the Triassic/Jurassic boundary (TJB) (see Fig. 3.9 in Chapter 3.4.1). The area investigated for magnetostratigraphy is only 23m long, covering the Rhaetian/Hettangian period. The studied interval is part of the lithological unit C (Romano et al., 2008), made mainly of white limestone organized in peritidal cycles (Fig. 4.10). Below the T/J boundary, lithology is represented by supratidal microbialites alternated to subtidal bivalves rich wackestone (subunit C1). Above the T/J boundary, subunit C2 reveals the same peritidal cycles seen in subunit C1, but without Megalodontids. The upper part of C2 shows a passage to oolithic bioclastic grainstone/packstone, with abundant benthic foraminifera and green algae. The TJB is placed with the Last Occurrence of Megalodontids bivalves (Romano et al., 2008), at the limit between subunits C1 and C2 (Fig. 4.10).

**Biostratigraphy**

Romano et al (2008) distinguished three main associations based on the fossil content:

- **Rhaetian association (Rh):** is in the upper portion of C1 and is characterized
by the extremely abundance of Megalodontids, frequently in life position. Also microfossils are abundant, with small bivalves and gastropods, echinoid spines, foraminifers (e.g. *Triasina hantkeni*) and algae (e.g. *Dasycladacea* *Grifhopperella curvata*).

- **Lower Hettangian association (H1):** is in the lower portion of C2 (130 m) and show the disappearance of Megalodontids, *T. hantkeni* and *G. curvata* (Fig. 4.10). Typical lower Hettangian association of foraminifera, cyanobacteria, bivalves and gastropods was found in these strata.
- **Upper Hettangian association (H2):** is in the middle-upper portion of C2 (100 m) and is characterized by the appearance of typical lower Jurassic Dasycladaceae. Other fossils are Jurassic palynomorphs, echinoids, corals (rare), and bivalves.

The age is constrained mainly by the distribution of foraminifera *Triasina hantkeni*, that appears into the *Paracochloceras suessi* ammonoid Zone (Gazdzicki, 1983), usually referred to Rhaetian (e.g. Moix et al., 2007; Rigo et al., 2015).
The vertical distribution of *T. hantkeni* is reported to reach and not overtake the end of *Choristoceras marshi* ammonoid Zone (De Castro, 1991), which upper limit is usually referred to the end-Triassic. The Last Occurrence of *T. hantkeni*, *Grifhoporella curvata* and the megalodontids (Rh association) is linked to the end-Triassic extinction event (Romano et al., 2008). The fossils of H1 association are typical of Late Triassic and Early Jurassic, representing a post-extinction survival phase (Barattolo and Romano, 2005; Romano et al., 2008). The recovery of dasycladacean algae in H2 association suggests a late Hettangian age (Barattolo and Romano, 2005) for the upper C2, with a probably extension to the Sinemurian (Romano et al., 2008).

**Paleomagnetism**

**Methods**

A total of 34 samples from the Mount Messapion section have been analyzed for magnetostratigraphy at the “Alpine Laboratory of Paleomagnetism” in Peveragno (Italy). To isolate the characteristic component of the natural remanent magnetism (ChRM), the samples have been thermally demagnetized with an ASC TD48 furnace, and then measured with a 2G Enterprises DC-SQUID cryogenic magnetometer. Samples have been demagnetized by steps of 50°C from 100°C to 350°C, then 25°C until 675°C. Directions of magnetization for each step of demagnetization have been plotted on an end-point vector graph (Zijderveld, 1967), one for each sample. In case of samples with magnetization components represented by less than three end-points in sequence (equals to three subsequent demagnetization steps), these samples have been rejected. The low-field magnetic susceptibility (κ) of every sample was measured using an AGICO Kappabridge KLY-3 susceptibility meter. Considering the uniform lithology, only one sample (MES68.1) was treated for backfield IRM using an ASC IM-10-30 impulse magnetizer with a saturation field of 2.5 T.

![Figure 4.11: IRM back field curve of sample MES68.1. Blue line is the smoothed path of remagnetization. Coercivity is around 100 mT.](image)
Magnetic properties

Magnetic susceptibility ($\kappa$) along the Mt Messapion section is mostly negative (~$-10.2 \times 10^{-6}$ SI), sometimes with peaks from $4.3 \times 10^{-6}$ SI to $8.8 \times 10^{-6}$ SI (Fig. 4.10). Negative susceptibility normally indicates a diamagnetic material, suggesting that any kind of magnetic mineral is present in the great part of the section. The hypothesis of an extremely low content of magnetic minerals inside the samples is more probable, with a bulk susceptibility too weak to be detected correctly by the instrument. Intensity of NRM at room temperature is weak, around $3 \times 10^{-5}$ A/m, with few peaks of intensity reaching a maximum of $3 \times 10^{-4}$ A/m.

Sample MES68.1, showing negative susceptibility, has been analyzed for backfield IRM (Fig. 4.11). Total coercivity is low, around 60-100 mT, in line with most of the non-pure magnetic phases, while saturation is reached around $5 \times 10^{-3}$ A/m after an applied field of 500 mT, which is quite low.

The cumulative log-gaussian (CLG) analysis (Kruiver et al., 2001) indicates four phases of magnetic minerals (Fig. 4.12), respectively with a coercivity of:

1. 100 mT at 50% of contribution
2. 80 mT at 35% o.c.
3. 10 mT at 10% o.c.
4. 1600 mT at 5% o.c.

The behavior of the acquisition path is complex near the saturation point, and is difficult to adapt the model to the data.

The kind of magnetic minerals is difficult to determine, but the coercivity is not so low to justify the extremely weak NRM signal and the scattered magnetic vectors. These issues are more probably related to the scarcity of ferromagnetic minerals in the Mt. Messapion rocks.
Magnetostratigraphy

Direction obtained from NRM analysis are quite scattered, probably due to the weak magnetization acquired by the rocks. Only 27 on 34 samples revealed a characteristic component (ChRM), in a large range of temperature (ChRM data in Appendix A.2). In Zijderveld diagrams, directions are both north-down and south-up in in situ coordinates (Fig. 4.13), although some samples revealed north-up or south-down directions not coherent with the area of deposition (northern hemisphere). In facts, filtering the component data for spurious directions, 17 of 27 data remains in in situ coordinates and 9 of 27 in tilt-corrected coordinates (Fig. 4.10). Equal-area projections show extremely scattered data both in in situ and tilt-corrected coordinates (Fig. 4.14). The scarcity of recovered data made difficult to obtain a reliable mean direction, which is Dec:12.2° Inc:-29.2° (k=2.8 a_p=20.8° N=27) in tilt-corrected coordinates. Fold test cannot be performed because the section in substantially homoclinal, but the reversals test (McFadden and McElhinny, 1990) has been applied, resulting indeterminated. Attempting to

Figure 4.13: Vector end-point demagnetization diagrams for representative samples from Mount Messapion section. Closed circles are projections onto the horizontal plane, and open circles are projections onto the vertical plane for in situ (geographic) coordinates.

Figure 4.14: Equal-area projections for characteristic remanent magnetization (ChRM) component directions isolated at Mount Messapion for in situ (geographic) and tilt-corrected coordinates (see text for mean directions).
define the magnetostratigraphy with the few tilt-corrected ChRM data (filtered and non-filtered), Virtual Geomagnetic Poles (VGP) was calculated for each direction. From non-filtered data, five magnetozones have been identified and named MM, where one-point data have been indicated as partial inversions (Fig. 4.10). Using filtered ChRM data, only one large normal magnetozone have been identified (named MMf), with three partial reverse magnetozones and an uncertain polarity zone in the lower part of the section (Fig. 4.10).

Discussion
As seen before, paleomagnetic data from the Mt. Messapion reveal unusual directions of the characteristic component (Fig. 4.13). These directions are quite similar to the directions of the geomagnetic field lines in the southern hemisphere, with upward inclination for normal polarity and downward inclination for reverse polarity. Considering that the Greece in the Mesozoic is assumed as belonging to the northern hemisphere, these directions are not easy to explain. These samples have been sampled by P. Tiano and A. Incoronato (University of Napoli “Federico II”) using a standard criteria for core orientation (as communicated before the analyses), which is clearly indicated by an arrow in every samples. The possibility of an error in orienting the samples correctly on the field has been considered, but this error should be present in all samples (collected by the same operator, with the same instruments), whereas some of them are oriented correctly. As seen, the filter applied to the ChRM directions eliminates these anomalous data, but only few components remain (9 on 27; Fig. 4.10). As obvious, these data-points are not enough to describe a pattern of polarity inversion that could be used for correlations. Without any assurance that the anomalous samples have been rotated during sampling, no changes in orientation data can be applied to obtain directions coherent with the hemisphere in which the sediments deposited. These issues affecting the Mt. Messapion paleomagnetism convinced me to renounce to continue the analyses on this section.

4.1.3 Leg 122: Site 761 (Hole C)
Site 761 of the Leg 122 is located in the Wombat Plateau (16°44’21.78”S; 115°29’12.3”E) (Fig. 4.15), subdivided in three Holes (761A, 761B and 761C). The paleomagnetic analyses provided in this work come from Hole 761C, in particular from the Rhaetian and the uppermost Norian. The core recovery for this Site is poor in the upper Rhaetian, while approaching the Norian/Rhaetian boundary the cores
are more complete (Fig. 4.16).

Geological Setting

Lithostratigraphy

Hole 761C is represented by claystones in the Norian, passing gradually to limestones in the Rhaetian (Fig. 4.16). The top of the Triassic sediments depicts a major unconformity with the Cretaceous strata above.

The detailed lithology here presented (Leg 122 Initial Reports; Haq et al., 1990) is considered from the top to the base of the Late Triassic sediments, following the convention used by the ODP reports.

The first 78.8 m of Rhaetian are poorly recovered white limestones, with facies varying from grainstone to wackestone-packstone and mudstone, indicating littoral, subtidal or intertidal environment. Dissolution is common in cemented sediments, except for micrites that are affected instead by dolomitization and neomorphic replacement of calcite.

Below these white limestones, are present 61 m of carbonate mudstone, wackestone and packstone-grainstone, interbedded with calcareous to silty laminated claystones. The last 23.1 m of the Rhaetian strata contains mainly calcareous and silty claystones with subordinate mudstone, wackestone-packstone and grainstone. Rare levels of
quartzose sandstones are present. The limit with the underlying Norian sediments is considered unconformable, with a hiatus of unknown duration marked by levels rich in crinoidal fragments.

The Norian is represented by 14.3 m of black and dark greenish laminated clayey siltstone, sometimes with pyrite nodules and coal levels.

**Paleoenvironments**

The sediments of the Norian suggest a deposition in a shallow-water coastal environment, dominated by clastic deposition. The abruptly overlain of open-marine strata of the lowermost Rhaetian suggest an intense deepening of the depositional
environment at the Norian/Rhaetian boundary, with an unconformable crinoid-rich pelagic level. The nature of this unconformity is unknown. The presence of quartz-rich sandstone, apparently from erosion of the Norian sediments, suggest that a portion of the Wombat Plateau still remain in high-relief and that the deepening in the Rhaetian is probably related to a rifting phase. The onset of carbonatic coastal facies in the upper Rhaetian suggests a restored shallow-water environment at this Site. At the top of the Rhaetian limestone an unconformity is present, passing immediately to Early Cretaceous (Barremian-Valanginian) sediments. The cavities in uppermost Rhaetian limestone, filled with calcite cements, suggest different options to explain this unconformity. Possibilities include the uplift of the plateau until emersion and subsequent paleokarst, or dissolution after submersion and submarine cementation. The absence of Jurassic sedimentation in Site 761 could have been produced by uplift and subaerial erosion of Jurassic sediments, emersion of the plateau in the latest Triassic/earliest Jurassic followed by erosion or non-deposition, or subsidence and submarine condensation or non-deposition during the Jurassic.

Biostratigraphy
In Hole 761C, relevant fossils for biostratigraphy are calcareous nannofossils, foraminifera, palynomorphs, dinoflagellates, and ostracods (Haq et al., 1990; Brenner et al., 1992) (Fig. 4.16). Depth is indicated in meters below sea floor (mbsf), thickness in meters.

Calcareous nannofossils
Nanoplankton of Site 761C, belonging to the Prinsiosphaera triassica Zone, shows a predominancy of P. triassica in association with Crucirhabdus primulas, Thoracosphaera geometrica and Thoracosphaera wombatensis (Bralower et al., 1992). These species extend from the meter 264 to the base of the recovered 761C. In the uppermost part of the sequence (from meter 375 to 264) Eoconusphaera zlambachensis, Crucirhabdus minitus and Archaeozygodiscus koessenensis have been found. These three species belong to the Eoconusphaera zlambachensis Subzone, exclusively Rhaetian in age (Bralower et al., 1992).

Foraminifera
In the lower part of the 761C (from the base to meter 410), genus Triasina (rare) and Aulotortus are present, where the occurrence of T. oberhauseri indicates a Norian age for these strata (Zaninetti et al., 1992). Aulotortus disappear in the middle part
of 761C (from meter 410 to 344) and diagnostic foraminifers (*Duotaxis birmanica*, genera *Ophthalmidium* and *Tetrataxis*) indicate a late Norian/early Rhaetian age (Zaninetti et al., 1992). In the upper part (from meter 344 to 264), genus *Aulotortus* reappears and *Triasina* species become abundant. The Rhaetian age is confirmed by the presence of *Triasina hantkeni* (Zaninetti et al., 1992).

**Dinoflagellates**

The presence of *Rhaetogonyaulax rhaetica* from meter 422 to 264 suggests a Rhaetian age for this portion (Brenner, 1992). Fossils typical of the *Heibergella balmei* Zone have been found from meter 422 to the base of the Hole (Brenner et al., 1992).

**Palynomorphs**

Most of the palynomorphs in Site 761C are poorly preserved and fragmented. For this Site the zonation is based mainly on the association with the *Rhaetogonyaulax rhaetica* dinoflagellate Zone. From ~340 to ~375 m and from ~400 to ~415 m palynomorphs typical of the *Ashmoripollis reducta* Zone (coeval to the *R. rhaetica* dinoflagellate Zone) (Brenner, 1992). The absence of *Corollina/Classopolis*-type palynomorphs (abundant at the top of the *A. reducta* Zone; e.g. Cirilli, 2010) suggests an early to middle Rhaetian age for this interval. From ~422 to ~430 m palynomorphs are typical of the *Minutosaccatus crenulatus* Zone (Norian) (Brenner, 1992). The absence of *R. rhaetica* dinoflagellates confirms the Norian age of this interval.

**Ostracods**

Samples from ~412 m contain ostracods *Ogmoconcha owthorpensis* (middle Rhaetian) and *Cytherella acuta* (upper Norian to middle Rhaetian) (Dépêche and Crasquin-Soleau, 1992). *Ogmoconcha bristolensis* (middle Rhaetian) was collected from meter ~370 and *Hiatobairdia subsymmetrica* (Rhaetian) from meter ~339 (Dépêche and Crasquin-Soleau, 1992).

**Paleomagnetism**

**Methods**

A total of 84 non-oriented ~10cc minicores and cubes from the Hole 761C have been analysed for paleomagnetism and rock magnetism. Samples have been provided by the Kochi Core Center (Kochi University, Japan) and analyzed at the “Fort
Hoofddijk” Paleomagnetic Laboratory (Utrecht University, Netherlands). Samples have been demagnetized progressively by application of an alternate gradient field (using a 2G Enterprises single-axis AF demagnetizer) and measured using a 2G Enterprises RF-SQUID magnetometer. Samples have been demagnetized by steps of 5 mT until 50 mT, then steps of 10 mT until 100 mT. Single sample NRM directions for each step of demagnetization have been plotted on an end-point vector graph (Zijderveld, 1967), and only the magnetization components made of at least three subsequent end-points have been considered. Cores from Hole 761C are not geographically oriented, and the samples are oriented only respect of the cores. So only the magnetic inclination have been considered to determine the direction of magnetization. The low-field magnetic susceptibility (κ) was measured with an AGICO Kappabridge MK1-A instrument on 86 samples. Rock magnetism experiments have been performed on selected samples to support the paleomagnetic interpretations. Thermomagnetic runs were performed on 3 samples (wba369001, wba3691401, wba3693801) using a modified horizontal translation Curie balance, measuring in air. Powdered samples (70-80 mg) was measured increasing temperature in several cycles, up to 580°C. Field cycles was between 100 and 300 mT, with heating-cooling rates of 10°C/min. Hysteresis cycles, IRM acquisition and backfield IRM have been performed using an alternate field gradient magnetometer (Princeton Measurement Corp. AGM 2900) on 23 samples fragments of about 50 mg (maximum field 500 mT, steps of 10 mT; exception for wba3669701 IRM: max. field 1200 mT, steps of 25 mT).

Magnetic properties
Susceptibility (κ) shows a general decrease from lowermost Rhaetian/uppermost Norian (~110×10⁻⁶ SI) to uppermost Rhaetian (~20×10⁻⁶ SI) strata (Fig. 4.16). The progressive decrease of κ could be related to the reduction of siliciclastic material in Hole 761C after the Norian/Rhaetian boundary (see “Paleoenvironments” paragraph, chapter 4.1.3, for details in environmental changing). Thermomagnetic curves of samples wba3690001, 3691401 and 3693801 indicate a weak magnetization, where wba3693801 shows the most intense magnetization (max. ~0.063 Am²/kg at 400°C) (Fig. 4.17). The curves show a huge increase in magnetization between 400 and 580°C (Fig. 4.17), probably due to the oxidation of a variable quantity of pyrite (FeS₂), in magnetite (Fe³⁺Fe²⁺O₄). Below 400°C there are no evidences of magnetic iron-sulfides (e.g. pyrrhotite – FeS) since the heating-cooling steps are totally reversible until this temperature and Curie temperature
for magnetic Fe-sulfides (~320°C) has not been reached (Fig. 4.17). The original magnetization is carried probably by magnetite but is impossible to differentiate the neo-formed magnetite to the original magnetite.

IRM acquisition curves show variable levels of saturation, from ~300 mT to ~400 mT, coherent with the presence of magnetite; few samples saturate at field intensities higher than 500 mT (Fig. 4.18A). Samples subjected to backfield IRM are characterized by coercivity fields from 40 to 60 mT, except for two samples (wba3669701 and wba3670101) that reaches a coercivity of 200-300 mT (Fig. 4.18B). Lower coercivity fields could be associated to magnetite, while higher fields are probably related to magnetic iron-sulfides as pyrrhotite. Hysteresis cycles are pot-bellied shaped, suggesting a mixture of both single-domain (SD) and superparamagnetic (SP) magnetite (Tauxe et al., 1996) (Fig. 4.18C). Only one sample (wba3670101) has a wasp-waisted shaped hysteresis cycle, indicating a mixture SD/SP magnetite as pot-bellied, but with coarser SP grains (Tauxe et al., 1996) (Fig. 4.18C).
Magnetostratigraphy

Mean intensity of initial NRM is ~0.17 mA/m until meter 420.7, where the intensity increase to ~1.22 mA/m. The higher intensity is located in the Norian part of the hole, just below the Norian/Rhaetian boundary, where siliciclastic sediments are dominant and the magnetic susceptibility is elevated. Vector end-point demagnetization plots (Zijderveld, 1967) reveal a stable magnetic record, showing characteristic magnetization (ChRM) in 60 samples, in a variable range from min. 5 mT to max. 80 mT (Fig. 4.19; ChRM data in Appendix A.3). Equal area-projection for inclination-only data shows a substantial variability of inclinations, with a mean
Figure 4.19: Vector end-point demagnetization diagrams for representative samples from Hole 761C. Closed circles are projections onto the horizontal plane, and open circles are projections onto the vertical plane for cores coordinates.

inclination value of $44.4^\circ \pm 7.2^\circ$ (k=7.4, N=60; McFadden and Reid, 1982) (Fig. 4.20). The comparison with the Geomagnetic Axial Dipole inclination value at the latitude of the site ($\text{Inc}_{GAD} = -31^\circ$) suggests that the paleomagnetic data of Hole 761C are not affected by contamination of VRM. The samples were not oriented geographically (only respect to the vertical axis of the cores), so the only way to determine the paleomagnetic polarity is to consider the magnetic inclination. During the Late Triassic, the Wombat Plateau was situated in the southern hemisphere, so I have to consider negative inclination as representative of normal polarity periods and positive inclination as reverse polarity periods. The stratigraphic sequence of ChRM inclinations provided 20 magnetic polarity reversals defining 5
magnetozones named WMA1n to WMA3n (Fig. 4.16). Single data-points have been considered as partial reversals.

Discussion
The magnetostratigraphy of the Hole 761C is affected by many zones of unknown polarity, mainly due to the scarce recovery in those intervals. The biostratigraphy indicates that the strata from meter 422 to 264 are Rhaetian in age, and the strata below are upper Norian. The Norian/Rhaetian boundary is placed on an unconformity (Haq et al., 1990), suggesting a lack in uppermost Norian/lowermost Rhaetian sediments. The comparison between the discontinuous 761C magnetostratigraphy and the Norian/Rhaetian Tethyan sections of Pignola-Abriola (Maron et al., 2015; Rigo et al., 2015), Steinbergkogel STK A and STK B+C (Krystyn et al., 2007a, 2007b; Hüsing et al., 2011) and Brumano/Italcementi Quarry (Muttoni et al., 2010) seems to confirm a hiatus between Norian and Rhaetian strata at Hole 761C (Fig. 4.21). In facts, the Norian/Rhaetian boundary is placed within a long reverse magnetozone (see Chapter 4.1.1 and Maron et al., 2015), while in Hole 761C it falls within a long normal magnetozone and is indicated as an unconformable boundary (Fig. 4.16). Just above the unconformity the first Rhaetian strata show a partial reverse polarity zone, probably related to the large reversal containing the Rhaetian base. Magnetozones MPA5n of Pignola-Abriola, ST1/G+ of STKA, ST2/G+ of STK B+C and BIT1n of Brumano are considered coeval to the Norian portion of magnetozone WMA1n (below the unconformity; Fig. 4.21). The partial reverse magnetozone just above the unconformity (WMA1n.1r) could be coeval to the upper part of MPA5r (Pignola-Abriola), BIT1r (Brumano) STI/H- of STK A and ST2/H- of STK B+C, while the remainder WMA1n is BIT2n and ST2/I+ of STK B+C (Fig. 4.21). The other magnetozones of 761C are difficult to correlate because biostratigraphy assigns a generic “Rhaetian Age” to this part of the Hole. Only the absence of palynomorphs Classopollis/Corollina constraints the Rhaetian strata of the Hole 760C to the early/middle Rhaetian. Probably, magnetozone WMA1r is coeval to BIT2r in Brumano, BIT3n could be correlated to WMA2n, BIT3r to WMA2r, and BIT4n to WMA3n (Fig. 4.21). A direct correlation with the Newark APTS is not possible because palynomorphs biostratigraphy is not comparable. An association could be made through the sections of Pignola-Abriola and Brumano/Italcementi Quarry, following correlations proposed respectively in Chapter 4.1.1 (and by Maron et al., 2015) and by Muttoni et al. (2010). Hence, the Newark magnetozone E19n should be coeval to the Norian WMA1n, while the Rhaetian
WMA1n is associated to E21n (the small WMA1n.1r should be the top of E20r) (Fig. 4.21). In this case, the estimated gap at the Norian/Rhaetian boundary in the Hole 761C is about 1.5 My. The remainder magnetozones WMA1r, 2n, 2r and 3n have been associated respectively to E21r, 22n.1n, 22n.1r and 22n.2n (Fig. 4.21). The unconformity at the top of the Rhaetian in Hole 761C obliterated the whole Jurassic and, following the correlation with the Newark APTS here proposed, at least 1.5 My of the upper Rhaetian.

Figure 4.21: The Norian-Rhaetian magnetostratigraphy and biostratigraphy of Hole 761C correlated to data from marine sections from the literature, such as Steinbergkogel (Hüsing et al., 2011), Brumano-Italcementi Quarry (Muttoni et al., 2010, 2014), and Pignola-Abriola (Chapter 4.11; Maron et al., 2015; Rigo et al., 2015). The correlation with these sections seems to confirm the presence of a hiatus at the Norian/Rhaetian boundary in Hole 761C, recorded within a large reverse polarity zone in the other sections. Hole 761C is correlated to the Newark astrochronological polarity time scale (APTS) through the statistical correlation of Pignola-Abriola with the APTS (Chapter 4.11; Maron et al., 2015).

4.2 CARNIAN

The Carnian Stage has a detailed magnetostratigraphy around the Carnian/Norian boundary (e.g. Pizzo Mondello, Muttoni et al., 2004, and Silická Brezová, Channell et al., 2003) and at the Ladinian/Carnian boundary (e.g. Prati di Sutuores/Sutuores Wiesen, Broglio Loriga et al., 1999; Mietto et al., 2012), but in the remaining part the sections with magnetostratigraphy are affected by some stratigraphical issues. For example, the Boluçektası Tepe section (Gallet et al., 1992) has a disconformity near the base and a fault that obliterate the lower Tuvalian.

The following Tethyan sections have been chosen for paleomagnetic investigation
to improve the magnetostratigraphy of the Carnian.

4.2.1 Pignola-2 and Dibona sections (from Maron et al., submitted and in review: Newsletter on Stratigraphy; see Attached publications)

Here I presented new magnetostratigraphic and biostratigraphic data from the Pignola-2 section of the Southern Apennines and the Dibona section of the Dolomites (both in Italy), which have been correlated to the Newark APTS in order to provide an independent control on the astrochronological ages of the older (Carnian) part of the Newark APTS (Kent and Olsen, 1999; Olsen and Kent, 1999). Furthermore, I provide a numerical age estimation of a major event occurring in the Carnian, known as the Carnian Pluvial Event (CPE; Simms and Ruffell, 1989). The CPE is represented by a widespread deposition of siliciclastic materials recognized in most of the Carnian sections around the world (e.g., Ruffell et al., 2015). The CPE is attributed to a climatic shift to more humid conditions (Simms and Ruffell, 1989), triggered by the emplacement of the Large Igneous Province (LIP) of Wrangellia in North America (e.g., Furin et al., 2006; Rigo et al., 2007; Preto et al., 2010; Dal Corso et al., 2012; Xu et al., 2014) and consequent emission of greenhouse gasses

Figure 4.22: The Pignola-2 section (coord.: Lat: 40°32’51.44”N, Long: 15°47’17.43”E) is located in the Southern Apennines, near Potenza (Southern Italy). The outcrop is along the main road connecting Pignola to Abriola, on the southern side of the Mt Crocetta. The Dibona section (coord.: Lat: 46°32’2.30”N, Long: 12°04’21.68”E) is located in the Dolomites, near Cortina d’Ampezzo (Belluno, Northern Italy). The outcrop is on the southern side of the Tofane di Rozes. Pignola-2 and Dibona sections were located in central Tethys during the Late Triassic.
in the atmosphere, with temperature increase (Rigo and Joachimski, 2010; Rigo et al., 2012a; Trotter et al., 2015). The CPE is well expressed by the most siliciclastic intervals in both the Pignola-2 and Dibona sections.

**Geological Setting**

**Pignola-2**

The Pignola-2 section (Lat: 40°32’51.44”N, Long: 15°47’17.43”E) crops out in the Southern Apennines, south of the town of Potenza, along the road connecting the two Pignola and Abriola villages (Fig. 4.22). The section is comprised of a 40 m-thick succession of cherty limestones pertaining to the Calcari con Selce Formation Fm (Scandone, 1967; Miconnet, 1983; Amodeo, 1999; Rigo et al., 2012a), and encompassing the Julian/Tuvalian substage boundary (Carnian; Rigo et al., 2007, 2012). The section includes a ~5 m-thick (from meter 8 to 13) green shale and radiolaritic interval (the “green clay-radiolaritic horizon” of Rigo et al., 2007), representing the first documentation of the CPE in Tethyan basinal successions.

![Figure 4.23: The Pignola-2 section. From left to right: lithostratigraphy, conodonts and palynomorphs biostratigraphy, Virtual Geomagnetic Pole (VGP) latitudes (from ChRM directions), magnetostratigraphy and magnetic susceptibility. In the magnetostratigraphy, black is normal polarity and white is reversed polarity. A total of 6 magnetozones have been identified, with a 5 meters interval of unknown polarity (grey shading) corresponding to the shales of the green clay-radiolaritic horizon (that could not be sampled). The U/Pb radiometric age of 230.91±0.33 Ma (Furin et al. 2006) comes from an ash-bed inside the green clay-radiolaritic horizon. Anomaly in the magnetic susceptibility around the “green horizon” represents the Carnian Pluvial Event in basinal environment (light-grey shaded interval).](image-url)
(Rigo et al., 2007, 2012a). In the upper part of the “green clay-radiolaritic horizon” (hereafter “green horizon”), a tuff level, named “Aglianiaco ash-bed” (meter 12) provided a U/Pb radiometric age of 230.91±0.33 Ma (Furin et al., 2006) (Fig. 4.23). The “green horizon” has been interpreted as resulting from a transient rise of $pCO_2$ levels that triggered the shoaling of the calcite compensation depth (CCD). This inferred CCD shoaling is possibly coupled with increased detrital and nutrient input in the basin as a consequence of the CPE, a warm and humid period that fostered silicate weathering and runoff on land (Rigo et al., 2007, 2012b; Rigo and Joachimski, 2010; Trotter et al., 2015). A distinct rise of $pCO_2$ in the coupled ocean-atmosphere system may have been provided by the emplacement of the Wrangellia LIP (e.g., Furin et al., 2006; Rigo et al., 2007; Preto et al., 2010; Dal Corso et al., 2012; Xu et al., 2014), radiometrically dated with Ar/Ar between ~233 and ~222 Ma, with the most likely age comprised between ~230 and ~225 Ma (Greene et al., 2010). The oldest radiometric ($^{207}$Pb/$^{206}$Pb) age available for Wrangellia comes from gabbros in Yukon, associated to the Wrangellian effusions and dated at 232.2 ±1 Ma (Mortensen and Hulbert, 1992; see also Greene et al., 2010). This age interval includes the age of the CPE at Pignola-2 from the “Aglianiaco ash-bed” (Furin et al., 2006).

**Dibona**

The Dibona section (Lat: 46°32’2.50”N, Long: 12°04’21.68”E) is a ~370 m thick shallow-water sedimentary succession located in the Dolomites (Southern Alps), on the southern side of the Tofana di Rozes Mountain, near the Dibona Hut (Fig. 4.22). The section is characterized by mixed carbonate-siliciclastic deposits of shallow-marine (Heiligkreuz/Santa Croce Formation) to marginal-marine (Travenanzes Formation) environments (e.g. Breda et al., 2009). The ~160 m Heiligkreuz Fm. (De Zanche et al., 1993; Preto and Hinnov, 2003; Neri et al., 2007; Gattolin et al., 2013, 2015) is subdivided into three members, which from the base to the top are subsequently the Borca Mb., ~100 m-thick, consisting of limestones and arenites passing to dolostones; the Dibona Sandstones Mb., ~60 m-thick, consisting of arenites, conglomerates, pelites and limestones; the Lagazuoi Mb., ~30 m-thick, consisting mainly of strongly dolomitized oolitic limestones (Fig. 4.24). The shales and arenites of the Borca and Dibona Sandstones Msbs record the CPE in a coastal environment (Breda et al., 2009; Preto et al., 2010). A major negative $\delta^{13}$C spike linked to the eruption of Wrangellia flood basalts has been observed at the base of the Heiligkreuz Fm., close to Dibona section, confirming the connection of
Figure 4.24: The Dibona section. From left to right: lithostratigraphy of the investigated portion, palynomorph and conodont biostratigraphy, magnetostratigraphy, Virtual Geomagnetic Pole (VGP) latitudes (from ChRM directions), magnetic susceptibility and lithostratigraphy of the whole Dibona section. In the magnetostratigraphy, black is normal polarity and white is reversed polarity. In the lower panel is the Dibona Sandstone Mb (Heiligkreuz Fm) site and in the upper panel is the Travenanzes Fm site. A total of 8 magnetozones have been identified, 5 from the Dibona Sandstones and 3 from the Travenanzes Fm. The large portion of unknown polarity (grey shading) in the Travenanzes portion is due to sparse seemingly robust paleomagnetic data (only three meaningful VGP points). Extension of the Carnian Pluvial Event is illustrated with the grey shaded area in the whole Dibona section lithostratigraphy (on the right).

the clastic input to the CPE climatic event (Dal Corso et al., 2012). Above the Lagazuoi Mb., the ~180 m Travenanzes Fm. (De Zanche et al., 1993; Neri et al., 2007; Breda and Preto, 2011) starts with ~25 m of dark clays and aphanitic...
dolostones passing upwards to multicolored clays with carbonatic and evaporitic intercalations deposited in sabbha-like environments. The top of the Travenanzes Fm. is dominated by dolomitic peritidal cycles of carbonate tidal-flat and shallow lagoon environments, with thin dark clay intercalations. It represents the transition to the overlying Dolomia Principale carbonate platform (Breda and Preto, 2011).

Biostratigraphy

Pignola-2

The Pignola-2 section has a detailed conodont and palynomorph biostratigraphy (Rigo et al., 2007, 2012). According to the conodont biostratigraphy, the Julian/Tuvalian (middle/late Carnian) boundary is placed at the base of the “green horizon”. In fact, below the “green horizon” a typical Julian conodont association composed of *Paragondolella praelindae*, *P. polygnathiformis*, and *Gladigondolella* spp is present. Above the “green horizon”, the section bears Tuvalian conodont species, i.e. *Carnepigondolella nodosa*, *C. carpathica*, *Paragondolella noah*, *P. oertlii*, and *Metapolygnathus praecommunisti* (Fig. 4.23) (Rigo et al. 2012a). Specifically, the Julian/Tuvalian boundary is placed at the level with the last occurrence (LO) of the *Gladigondolella* genus (Rigo et al., 2007). Palynomorphs from Pignola-2 have been grouped in two main assemblages (Rigo et al., 2007): Assemblage A is typical of the Julian/Tuvalian interval, while Assemblage B covers a narrower range in the upper Tuvalian (see Rigo et al. 2007 for additional details) (Fig. 4.23).

Dibona

The Dibona section has a detailed pollen and spore biostratigraphy (Roghi et al., 2010). The typical uppermost Julian-lower Tuvalian association with *Patinasporites densus*, *Aulisporites astigmosus* and *Duplicisporites continuus* (Borca Mb, Dibona Sandstones Mb) and *Equisetosporites chinleanus* (Dibona Sandstones Mb) is found in the Heilkgreuz Fm. It is followed by a Tuvalian association of *Granuloperculatipollis rudis* and *Riccisporites* cf. *R. tuberculatus*, found at the base

Plate 4.2: (following page) Conodonts from samples DIN2 and DIN6. The fauna illustrated in the plate includes Paragondolella polygnathiformis, Paragondolella noah, transitional forms from P. noah to Metapolygnathus praecommunisti, and M. praecommunisti. The specimens of M. praecommunisti are basal, showing the accessorional node behind the cup, the posterior prolongation of the keel and a quite centrally located pit, but no nodes on the anterior platform margins. The occurrence of basal representatives of this species in sample DIN6, together with advanced P. noah, suggests a lower Tuvalian age (Mazza et al. 2011; Mazza et al. 2012a). 1: Paragondolella polygnathiformis (Budurov and Stefanov 1965) (DIN2); 1c: the blade of the specimen got broken; 2: Paragondolella noah (Hayashi 1968) transitional to Metapolygnathus praecommunisti Mazza, Rigo and Nicora 2011 (DIN6); 3: Metapolygnathus cf. praecommunisti (DIN6) Figs b–c: the blade termination got broken; 4: Metapolygnathus cf. praecommunisti (DIN6); 5–7: Metapolygnathus praecommunisti (DIN6). a: view from above; b: lateral view; c: view from below. All the conodonts are at the same scale.
of the overlying Travenanzes Fm (Fig. 4.24). The former association, belonging to the *Granuloperculatipollis rudis* Assemblage of Roghi et al. (2006, 2010), is similar to Assemblage B found in the Pignola-2 section. Moreover, additional
sections coeval to the Dibona section reveal pollens and spores comparable with the 
biostratigraphic record of Pignola-2, e.g. the Cave del Predil section in the Southern 
Alps of Friuli (Roghi, 2004), and the Lunz (Köppen, 1997) and Rubland (Kraus, 
1969) sections in the Northern Calcareous Alps of Austria. Six samples for conodont 
analysis have been collected from the Dibona section immediately below the base of 
the Lagazuoi Mb, in the last 10 meters of the Dibona Sandstones Mb. The conodont 
association consists of *Paragondolella polygnathiformis*, *Paragondolella noah*, 
transitional forms from *P. noah* to *Metapolygnathus praecommunisti*, and early 
representatives of *M. praecommunisti* (Fig. 4.24, Plate 4.2) attributed to the early 
Tuvalian age (Mazza et al. 2010, Mazza et al. 2011). Furthermore, the ammonoid 
*Shastites cf. pilari* has been found below the Lagazuoi Mb, in the nodular limestone 
corresponding to the upper portion of the Dibona Sandstone Mb of the Heiligkreuz 
Fm. (Gianolla et al., 1998b; De Zanche et al., 2000; Gattolin et al., 2015).

**Paleomagnetism**

**Sampling and laboratory methods**

A total of 63 oriented paleomagnetic core samples (~10cc) have been collected 
from the Pignola-2 section, 55 from limestones beds and 8 from the radiolaritic 
intervals within the “green horizon”, with a stratigraphic interval of approximately 
0.5 m (Fig. 4.23). The clayey intervals of the “green horizon” have not been 
sampled because they are both too thin and chipped. From the Dibona section a 
total of 45 cores have been collected from the upper Borca Mb. to the base of the 
Lagazuoi Mb. (Heiligkreuz Fm.), and 36 samples from the Travenanzes Fm. To 
isolate the ChRM, all samples have been thermally demagnetized (with an ASC 
TD48 furnace, residual field < 10 nT) and measured with a 2G Enterprises DC-
SQUID magnetometer (magnetic moment noise level <10-12 Am²) at the Alpine 
Laboratory of Paleomagnetism – ALP (Peveragno, Italy). Samples have been 
demagnetized by steps of 50°C from 100°C to 350°C, then 25°C until 675°C. 
Single sample directions of the magnetization vectors have been plotted on an end-
point vector graph (Zijderveld 1967) for each step of demagnetization (Fig. 4.25). 
Samples showing magnetization components made of less than three end-points 
in sequence (representing three subsequent temperature steps) have been rejected. 
The low-field magnetic susceptibility (\(k\)) was measured with a AGICO Kappabridge 
KLY-3 instrument (sensitivity: \(2 \times 10^4\) SI; at the ALP, Peveragno) for all samples. 
Further, to support the paleomagnetic interpretation, thermomagnetic runs were 
performed on a modified horizontal translation Curie balance (paleomagnetic
laboratory ‘Fort Hoofddijk’, Utrecht University, The Netherlands; noise level 5×10⁻⁹ Am², typical signals at least an order of magnitude higher; Mullender et al., 1993) for a subset of the samples. About 70-80 mg of powdered sample was measured in several cycles to increasingly higher temperature up to 670°C; the field was cycled between 100 and 300 mT, heating and cooling rates were 10°C/minute. Measurements were performed in air. Three samples from the Pignola-2 section were investigated and 6 from the Dibona section (3 from the Heiligkreuz Fm. and 3 from the Travenanzes Fm.). Samples PGM0.30, RAD4 and PGM14.64 from the Pignola-2 section and samples MDS12.4, MDS29.1 and MDS52.3 from the Dibona section have been analyzed using the Curie balance.

Magnetic properties

**Pignola-2**

**A**

![Images of demagnetization diagrams for Pignola-2 samples]

**Figure 4.25**: Vector end-point demagnetization diagrams (Zijderveld 1967) of the Pignola-2 section (panel A) and the Dibona section (panel B). Open circles are projections onto the vertical plane, and closed circle are projections onto the horizontal plane for in situ (geographic) coordinates.

The limestones of the Pignola-2 section reveal a very low κ, usually smaller than 5×10⁻⁶ SI (Fig. 4.23). In the “green horizon” the initial magnetic susceptibility is considerably higher (from ~70×10⁻⁶ to ~110×10⁻⁶ SI) than in the rest of the sampled interval, due to an increase of the terrigenous fraction (Fig. 4.23). Interestingly, magnetic susceptibility is high (from 25×10⁻⁶ to 55×10⁻⁶ SI) also in the carbonatic
strata delimiting the “green horizon” (~1 or 2 meters above and below), indicating a significant siliciclastic component we associate to the CPE.

The Curie balance results are shown in Fig. 4.26 and the stratigraphic position of the sample is in Fig. 4.23. The limestones of the Pignola-2 section (PGM0.30 and 14.64; Fig. 4.26A, 4.26C) are very weak, only slightly above instrumental noise level. Nonetheless there seems to be a marginally convex magnetization vs. temperature behavior between ~100-200 and ~450-500°C, that is reversible on intermittent cooling. The final cooling segment from 600°C back to room temperature, however, does not reveal that behavior. It is difficult to interpret this behavior that may be associated with titanomagnetite (sensu lato) with a varying Ti-content. However, a minute amount of magnetic sulfides cannot be excluded with certainty. The “green horizon” sample RAD4 (Fig. 4.26B) is much stronger (but still weak, it remains a sediment) and shows a Curie point (determined by the two-tangent method, Grommé et al. 1969) of ~350°C that is reversible on cooling after the final heating temperature of 600°C.

Dibona

The Dibona Sandstones Mb show a generally high susceptibility (κ), around ~43×10⁻⁶ SI, whereas in the more terrigenous part (meter ~50 to ~60) κ increases up to ~200×10⁻⁶ SI (Fig. 4.24). The Travenanzes Fm show higher κ values, around 50×10⁻⁶ SI, with a peak of ~270×10⁻⁶ SI around meter 170 (Fig. 4.24). The Dibona sandstones Mb samples (MDS12.4, Fig. 4.26D; MDS29.1, Fig. 4.26E; MDS52.3, Fig. 4.26F; stratigraphic position in Fig. 4.24) from the Dibona section all show a variable portion of non-magnetic pyrite (FeS₂) that is oxidized during the thermomagnetic analysis, first to magnetite and finally to hematite, explaining the occasionally huge increase in magnetization between 400 and 600°C. There are no indications for magnetic sulfides below 400°C since the analysis shows reversible heating and cooling segments in that temperature range and no Curie temperature of ~320°C. Plausibly traces of magnetite represent the original magnetic mineralogy but it is impossible to discriminate between left overs of neo-formed magnetite (most of it oxidizes further to hematite) and original magnetite. The three samples from the Travenanzes Fm (MTV9, Fig. 4.26G; MTV52, Fig. 4.26H; MTV67, Fig. 4.26I; stratigraphic position in Fig. 4.24) are all very weak, demonstrating paramagnetic behavior only. During the heating above 600°C a minute amount of magnetic minerals (presumably fine-grained magnetite) is formed because the final
cooling curves lie slightly above the corresponding heating curves.

**Magnetostratigraphy**

**Pignola-2**

Figure 4.26: Thermomagnetic curves determined with a Curie balance of samples PGM (Pignola-2, Calcare con Selce Fm; panels A, C), RAD (Pignola-2, Green clay-radiolaritic horizon; panel B), MDS (Dibona, Heiligkreuz Fm; panels D, E, F), and MTV (Dibona, Travenanzes Fm; panels G, H, I). The PGM samples of Pignola-2 reveal a mixture of different minerals, including magnetite, in the cherty limestones, whereas in the green horizon (RAD sample) there is an increase in more Ti-rich magnetite. Samples MDS from the Heiligkreuz Fm in the Dibona section show a magnetization increase above 400-450°C, coherent with the reaction of pyrite to magnetite.

The mean intensity of the starting NRM is ~0.02 mA/m in the pelagic carbonates, ~0.06 mA/m in the radiolarites of the “green horizon”, and ~0.2 mA/m in the carbonatic levels just above the “green horizon”. Vector end-point demagnetization diagrams (Fig. 4.25A; Zijderveld, 1967) reveal the presence of spurious (viscous) magnetic components from room temperature to 250-300°C; at higher temperatures the characteristic component remanent magnetization (ChRM) direction is isolated (Fig. 4.25A). The demagnetization trajectory trends toward the origin up to a maximum temperature of 675°C. This behavior is observed in 47 samples (ChRM
data in Appendix A.4). Equal area stereographic projections reveal that the ChRM is bipolar being oriented north-and-down or south-and-up in in situ coordinates, and northwest-and-down or southeast-and-up after correction for bedding tilt (Fig. 4.27). The mean direction in tilt-corrected coordinates, calculated with standard Fisher statistics, is of Dec: 28.4°E; Inc: 39.6° (k = 23.9; 95°; N = 47; Table 4.2). No fold test could be performed because the bedding attitude through the section is essentially the same. The reversals test (McFadden and McElhinny, 1990) is positive, suggesting that the ChRM is the original magnetization acquired during or shortly after deposition. The mean directions in in situ coordinates (Dec: 353.5°E; Inc: 59.7°; k: 24.1; 95°; 4.3°; Table 4.2) are similar to the inclination of the geomagnetic axial dipole (GAD Inc: ~ 59.8°), so we cannot exclude some contamination of the ChRM by VRM for normal polarity components. The latitudes of the Virtual Geomagnetic Poles, derived from the ChRM directions, provided a sequence of 12 magnetic polarity reversals defining 12 magnetozones labeled from MP1n to MP6r (Fig. 4.23). The shales of the “green horizon” were not sampled for magnetostratigraphy (see above).

Figure 4.27: Equal area projections for ChRM (characteristic remanent magnetization) of the Pignola-2 (upper panel) and Dibona (lower panel). Mean directions in the text and in Table 4.2.
**Dibona**

The mean intensity of the samples from the Heiligkreuz Fm. is ~0.04 mA/m and in the Travenanzes Fm. is ~0.05 mA/m. The vector end-point diagrams (Fig. 4.25B; Table 4.2) reveal the ChRM between 200°C and 550°C in 25 of 45 samples of the Heiligkreuz Fm. (named MDS), and between 150°C and 400°C in 18 of 36 samples from the Travenanzes Fm (named MTV) (ChRM data in Appendix A.4). The diagrams show both north-down and south-up directions, sometimes scattered (Fig. 4.25B) (typical MAD: ~11). The equal-area stereographic projection reveal fairly scattered directions (Fig. 4.27), failing the reversals test (McFadden and McElhinny, 1990). The mean directions in *in situ* coordinates (Dec: 350.5°E; Inc: 33.9°; k: 4.2; α95: 11.8°; Table 4.2) differs to the inclination of the geomagnetic axial dipole (GAD Inc: ~62.8°), so the ChRM should not be contaminated by a VRM component. As for the Pignola-2 section, also here fold test cannot be performed due to homoclinality of the succession. The sequence of VGPs of the Dibona section shows nine magnetozones labeled MD, where MD1r, 2r and 4r are rather uncertain (grey intervals, Fig. 4.24), due to the poor preservation of NRM in these intervals (only three robust paleomagnetic directions).

**Discussion**

**Correlations between Tethyan sections**

The Pignola-2 section is correlated with other coeval Tethyan sections from the literature containing conodonts to obtain a complete magneto-biostratigraphic record for the Carnian Stage (Fig. 4.28). The upper part of the Pignola-2 magnetostratigraphy (magnetozones MP2n to MP3r) is considered correlative to the basal portion of the Silická Brezová section (up to SB2r) (Channel et al., 2003) and the Pizzo Mondello section (up to PM2r) (Muttoni et al., 2004), whereas the entire Pignola-2 section is comparable with the Guri Zi section in Albania (up to GZ5r) (Muttoni et al., 2005, 2014) (Fig. 4.28). The conodont biostratigraphy of

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**Table 4.2 Paleomagnetic Directions from Pignola-2 and Dibona Sections**

<table>
<thead>
<tr>
<th>Site</th>
<th>Comp.</th>
<th>N</th>
<th>k</th>
<th>α95</th>
<th>Dec.</th>
<th>Inc.</th>
<th>k</th>
<th>α95</th>
<th>Dec.</th>
<th>Inc.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pignola-2</td>
<td>ChRM</td>
<td>47</td>
<td>24.1</td>
<td>4.3°</td>
<td>353.5°E</td>
<td>59.7°</td>
<td>23.9</td>
<td>4.3°</td>
<td>28.4°E</td>
<td>39.6°</td>
</tr>
<tr>
<td>Dibona</td>
<td>ChRM</td>
<td>46</td>
<td>4.2</td>
<td>11.8°</td>
<td>350.5°E</td>
<td>33.9°</td>
<td>4.2</td>
<td>11.8°</td>
<td>10.2°E</td>
<td>41.4°</td>
</tr>
</tbody>
</table>

**Legend**

Comp.: paleomagnetic component  
N: number of samples  
k, α95: Fisher statistics parameters  
Dec.: mean declination  
Inc.: mean inclination

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Zijderveld, 1967) reveal the ChRM between 200°C and 550°C in 25 of 45 samples of the Heiligkreuz Fm. (named MDS), and between 150°C and 400°C in 18 of 36 samples from the Travenanzes Fm (named MTV) (ChRM data in Appendix A.4). The diagrams show both north-down and south-up directions, sometimes scattered (Fig. 4.25B) (typical MAD: ~11). The equal-area stereographic projection reveal fairly scattered directions (Fig. 4.27), failing the reversals test (McFadden and McElhinny, 1990). The mean directions in *in situ* coordinates (Dec: 350.5°E; Inc: 33.9°; k: 4.2; α95: 11.8°; Table 4.2) differs to the inclination of the geomagnetic axial dipole (GAD Inc: ~62.8°), so the ChRM should not be contaminated by a VRM component. As for the Pignola-2 section, also here fold test cannot be performed due to homoclinality of the succession. The sequence of VGPs of the Dibona section shows nine magnetozones labeled MD, where MD1r, 2r and 4r are rather uncertain (grey intervals, Fig. 4.24), due to the poor preservation of NRM in these intervals (only three robust paleomagnetic directions).

**Discussion**

**Correlations between Tethyan sections**

The Pignola-2 section is correlated with other coeval Tethyan sections from the literature containing conodonts to obtain a complete magneto-biostratigraphic record for the Carnian Stage (Fig. 4.28). The upper part of the Pignola-2 magnetostratigraphy (magnetozones MP2n to MP3r) is considered correlative to the basal portion of the Silická Brezová section (up to SB2r) (Channel et al., 2003) and the Pizzo Mondello section (up to PM2r) (Muttoni et al., 2004), whereas the entire Pignola-2 section is comparable with the Guri Zi section in Albania (up to GZ5r) (Muttoni et al., 2005, 2014) (Fig. 4.28). The conodont biostratigraphy of
Silická Brezová has been updated in this study by reclassifying the taxa illustrated in Figs. A1-A3 in Channell et al. (2003), using the new taxonomic criteria illustrated in Mazza et al. (2010, 2011, 2012a,b) (Fig. 4.28).

The magnetostratigraphy of the Dibona section straddling the Dibona Sandstones Mb. of the Heiligkreuz Fm. should be partially coeval with the magnetostratigraphy across the “green horizon” in the Pignola-2 section, as suggested by the first occurrence of conodont *Metapolygnathus praecommunisti* (Figs. 4.24, 4.28).

Consequently, magnetozones MD1n-1r-2n-2r-3n at Dibona have been correlated to Pignola-2 magnetozones MP4n-4r-5n-5r-6n, respectively. Based on the first occurrence of *M. praecommunisti*, magnetozone MD3n is considered coeval to magnetozone SB2n at Silická Brezová and PM2n at Pizzo Mondello (Fig. 4.28).

The correlation between Pignola-2 and Dibona sections implies that the onset of the CPE at Dibona should fall in the lower part of the Heiligkreuz Fm. (basal Borca Mb, as suggested by Dal Corso et al., 2012) and its acme is reasonably represented by the terrigenous-rich levels of the Dibona Sandstone Mb (Fig. 4.24). The absence of strong biostratigraphic constraints does not allow a solid magnetostratigraphic
correlation between the upper Dibona section and other Tethyan sections.

**Correlation with the Newark APTS**
The Pignola-2 magnetostratigraphy has been compared with the Newark APTS (Kent and Olsen, 1999; see also Olsen et al., 2015) using the statistical approach described in Muttoni et al. (2004) and Maron et al. (2015; see also Chapter 4.1.1). The radiometric age of 230.91±0.33 Ma from the Aglianico ash-bed (Furin et al., 2006), comprised within the Pignola-2 magnetozone MP4r, has been taken into account for the correlation. The Dibona section was not considered for statistical correlation with the Newark APTS because of the unreliability of its magnetostratigraphy, due to the variable sedimentation rate typical of shallow-water environments.

We compared the thickness of the Pignola-2 magnetozones with the duration of the magnetozones in the Newark APTS, testing the magnetostratigraphy of Pignola-2 along the APTS and obtaining 24 possible correlation options (Fig. 4.29). The interval of unknown polarity within the Pignola-2 “green horizon” is tentatively interpreted as dominated by normal polarity.

Each correlation is analyzed using linear regression, obtaining 24 t-test values; the higher the t-value, the more reliable the correlation. Only options 1, 2 and 24 pass the 95% confidence level threshold (Fig. 4.29) (statistical procedure with correlation options and analysis of t-values is reported in Appendix A.4). Option 24 is not considered because it is inconsistent with the U/Pb radiometric age of 230.91±0.33 Ma from the “green horizon” (Furin et al., 2006). Option 1 and 2 are consistent with this age. The main features of Option 1 and 2 are as follows:

**Option 1:**
- High t-value (~2.7).
- The radiometric age of Pignola-2 fits more closely (0.6 M.y. older) with the age provided by the Newark APTS for the equivalent stratigraphic level (Fig. 4.30).
- Fits with the correlation of Pizzo Mondello and the APTS. Specifically, magnetozones MP4r and MP5n of Pignola-2 are correlated respectively to E5r and E6n in Newark, as well as to the PM1r and PM2n in Pizzo Mondello. PMr1 and PM2n were correlated to the same Newark magnetozones by Muttoni et al. (2004) (Fig. 4.28).

**Option 2:**
- High t-value (~3.1)
- The correlation with the Newark APTS leads to a 0.9 M.y. discrepancy
between the radiometric age of Pignola-2 and the age of the APTS (Fig. 4.30).

**Figure 4.29:** Sequence of 24 correlation options between the Pignola-2 section and the Newark APTS. Dark grey bars indicate the correlations that are reliable at the 90%, black bars are the correlations reliable at the 95%. Only three options are reliable at 95%: Options 1, 2 and 24. Option 24 was rejected because being inconsistent with the age of Pignola-2, while Options 1 and 2 both covers an interval consistent with the radiometric age of 230.91 Ma. In particular, preferred Option 1 is perfectly coherent with the time constraint in Pignola-2 and with the previous correlation between Pizzo Mondello section and the Newark APTS (Muttoni et al., 2004), performed using the same statistical method.

- This option does not fit with the previous correlation between Pizzo Mondello and the Newark APTS (Muttoni et al., 2004).

Option 1 implies only a minor discrepancy between the Pignola-2 U/Pb age and the Newark astrochronology, considering that in the lower Stockton Fm astrochronology is extrapolated from the upper Stockton and Lockatong Fms, where the 404 kyr
McLaughlin cycles are better expressed (Kent and Olsen, 1999; Olsen and Kent, 1999). Moreover, Option 1 is coherent with previous correlations from the literature (Pizzo Mondello; Muttoni et al., 2004) and is preferred over Option 2.

We derived an age model from Option 1 that reveals a complex pattern of sedimentation rate along the Pignola-2 section (Fig. 4.31). In the cherty limestones, the sedimentation rate is mostly constant, except for a decrease just below and above the “green horizon”. In the “green horizon” the sedimentation rate increases, probably due to an enhanced runoff of siliciclastic sediments from the continent caused by increased rainfall and weathering, in consequence of the intensified humid conditions at the CPE.

The age model derived from Option 1 (Fig. 4.31) suggests an age of ~230.7 Ma for the Julian/Tuvalian boundary, approximated in Pignola-2 by the LO of conodont *Gladigondolella* spp.. Assuming a Carnian/Norian boundary at ~227 Ma (Muttoni et al., 2004) and a Ladinian/Carnian boundary at ~237 Ma (Mietto et al., 2012), the Julian should be ~6.3 My-long and the Tuvalian ~3.7 My-long. Assuming the
magnetic susceptibility anomaly in Pignola-2 (covering the “green horizon” and
the closest limestone beds) as expression of the CPE, its duration was about 1 My.

**4.2.2 Leg 122: Site 759 (Hole B) and Site 760 (Hole B)**

Sites 759 and 760 of the Leg 122 are located in the Wombat Plateau (16°44’21.78”S;
115°29’12.3”E) (Fig. 4.32). Sites are subdivided in Holes: 759A-759B; 760A-760B. The
paleomagnetic analyses provided here come from Holes 759B and 760B, in
particular from the Late Triassic portion. The core recovery in the Norian portion of
Hole 759B is poor, while the Carnian portion is more available (Fig. 4.33). In Hole
760B the recovery is good for the entire Hole, and is greater in the Carnian portion
(Fig. 4.34).

**Geological Setting of the sites**

Lithologies of Holes 759B and 760B show a transition from the siliciclastic
sedimentation in the Carnian/lowermost Norian to an increased presence of
carbonates in the middle Norian (Figs. 4.33, 4.34). In Hole 760B the siliciclastic
sediments are more abundant and coarser than in Hole 759B. The detailed lithology
here presented (Leg 122 Initial Reports; Haq et al., 1990) is considered from the
top to the base of the Late Triassic sediments, following the convention used by
the ODP reports (depth is indicated in meters below sea floor – mbsf, thickness in
meters).

Figure 4.32: Sites 759-760 are located in the Wombat Plateau (16°44’21.78”S; 115°29’12.3”E),
north-western Australia. The area of the Wombat Basin was located in southern Tethys during Late
Triassic.
**Hole 759B**

The first 95.4 m of sediments (Norian) are fossiliferous limestone (mudstone to wackestone, sometimes packstone) and dolomite, with interbedded silty claystone (Fig. 4.33). Minor lithologies associated with claystone are siltstone and fine-grained sandstone. In carbonates, oolitic/oncolitic grainstones are present, as well as calcarenites with variable amounts (1% to 20%) of siliciclastic component. Different degrees of dolomitization affect this interval, but they do not obliterate the

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**Figure 4.33:** Hole 759B. From left to the right: lithostratigraphy, core recovery, main biostratigraphy (see text for discussion), magnetostratigraphy, virtual geomagnetic pole (VGP) latitudes calculated from characteristic remanent magnetization (ChRM) component directions, and magnetic susceptibility. Carnian/Norian boundary is placed with the beginning of Minutosaccatus crenulatus palynomorph Zone.
sedimentary features for most of the interval. The interbedded claystone become more abundant moving to the lowermost Norian. The boundary between the Late Triassic portion of the Hole and the overlying Cenozoic sediments is not recovered, so the type of stratigraphic contact is unknown.

The following 69 m (Carnian/early Norian) of sediments are dominated by parallel-laminated black silty claystone and clayey siltstone with disseminated pyrite nodules (Fig. 4.33). These lithologies are interbedded sometimes with coarse-grained levels of mixed siliciclastics and carbonates. Coarse-grained levels include a quartz-rich sandstones interval from ~140 to ~150 mbsf (sometimes fine-grained) and carbonate wackestone, packstone and grainstone.

The last 103 m (Carnian) are represented by alternation of silty claystone and clayey siltstone, with silty claystone becoming dominant in the lower part of the interval (Fig. 4.33). Minor lithologies are claystone (sideritic in part), sandy siltstone, coarse-grained quartz sandstone and pyrite nodules (decreasing downward).

Paleoenvironments

In the Carnian, the Wombat Plateau was characterized by a distal deltaic claystone sequence. The presence of pyrite/siderite, the absence of shelly faunas, the black color and the well-preserved lamination indicate restricted marine depositional conditions. The upward increase of carbonatic levels interbedded with claystone and sandstone indicates shallowing. The general coarsening upward trend in this interval is interpreted as marine regression and progradation of the delta. The presence of scattered pyrite, abundant carbonaceous matter in siltier levels and the lack of many faunas suggest oxygen-depleted, reducing conditions. The presence of quartz sandstone at ~140-150 mbsf marks the period of maximum marine regression. In the Norian sediments, the dominant carbonate-rich sediments suggest a marginal-marine, moderately low-energy environment. The presence of interbedded claystone and siltstone indicate carbonate banks formed in a coastal area with fluvial discharge. The acyclic repetition of carbonate and terrigenous claystone is related probably by change of terrigenous depocenter, consequent to the movement of distributary channels. These shifts could have driven the onset and the demise of the carbonate sedimentation in the Norian. The Norian sequence is truncated by a major unconformity beneath the Cenozoic pelagic sediments. The Jurassic and Cretaceous deposits have been eroded because of uplift, as well as the upper part of Late Triassic (upper Norian/Rhaetian). Reworked sands made of quartz, manganese fragments and early Miocene foraminifera (although they are
probably due to borehole contamination) overlie the erosive surface.

Hole 760B

Lithological information of Hole 760B starts from Core 6R (283 mbsf; Fig. 4.34), whereas the previous Cores were drilled only to reach the stratigraphic level corresponding to the end of the Hole 760A.

The first lithological unit of Hole 760B (179 m thick, Carnian-Norian) is characterized by interbedded carbonates and siliciclastic sedimentary rocks (Fig. 4.34). The main lithologies are dark gray clayey siltstone, black silty claystone, gray grainstone,
wackestone, mudstone, and quartz-rich silty sandstone. The carbonate rocks are abundant from 283 to 321 mbsf, 360 to 396 mbsf, and 416 to 464 mbsf (Fig. 4.34). Many of the carbonate rocks contains a high percentage of secondary dolomite. Then, the second unit (42 m thick, Carnian) is dominated by black silty claystone, dark gray clayey siltstone, and dark greenish gray silty sandstone (Fig. 4.34). Coarser grained rocks are localized in the lower part of the unit, indicating a fining-upward trend. Sideritic levels are present, increasing with depth.

*Paleoenvironments*

Carnian sediments (lower unit) could have been deposited either in a distal prodelta environment, or in a protected shallow-water setting (i.e. tidal flat, estuarine bay). Chemically reducing conditions are locally present, with deposition of siderite and sulfurs. The upper unit, around the Carnian/Norian boundary, shows lithologies and features typical of a marginal-marine environment, near the interfingering between siliciclastic and shallow-water carbonate depositional systems. Siliciclastic sediments were mostly deposited by currents and bioturbated. Carbonates contain peloidal and bioclastic components typical of a shallow to marginal-marine environment. Energy conditions were generally higher than in the lower unit.

*Biostratigraphy*

**Hole 759B**

In Hole 759B the more relevant fossils for biostratigraphy are calcareous nannofossils, dinoflagellates and palynomorphs (Haq et al., 1990; Brenner et al., 1992) (Fig. 4.33).

**Calcareous nannofossils**

An assemblage containing calcareous nannofossils *Tetralithus cassianus*, *Prinsiosphaera triassica* and *Hayococcus floralis* was found from ~230 mbsf to the bottom of the Hole. *T. cassianus* is a typical Carnian species, *H. floralis* is exclusively Norian and *P. triassica* is Norian/Rhaetian (see Tethyan palynology from Germany and Austria of Jafar, 1983) and they generally indicate a middle Late Triassic period. In some specimens of calcareous nannofossils Ca-carbonate have been replaced by siderite (FeCO₃).
Dinoflagellates
From ~41 to 45 mbsf, the dinoflagellates cysts *Heibergella balmei* and *Suessa listeri* indicate the *Heibergella balmei* Zone, which is late Norian in age (Brenner, 1992).

Palynomorphs
Palynomorphs assemblage belonging to the *Samaropollenites speciosus* Zone (Carnian) has been found in the lower part of the Hole, from ~175 mbsf to the bottom (Brenner, 1992; Brenner et al., 1992). The interval between ~41 mbsf and ~45 mbsf is in the *Minutosaccatus crenulatus* Zone (middle to late Norian), as confirmed by the coeval *Heibergella balmei* dinoflagellates Zone (Brenner, 1992). The interval between ~45 mbsf and ~175 mbsf provides a transitional assemblage between *S. speciosus* and *M. crenulatus* Zones. The presence of *Camerosporites secatus* and *Camerosporites pseudoverrucatus* suggests a Carnian age for the transition interval, whereas the permanent low abundance of *S. speciosus* indicates a Norian age (Brenner, 1992). Therefore, the *Samaropollenites speciosus/Minutosaccatus crenulatus* transition is considered late Carnian to early Norian in age.

Hole 760B
Main relevant fossils in Hole 760B are calcareous nannofossils, palynomorphs, dinoflagellates, and ostracods (Fig. 4.34).

Calcareous nannofossils
Main species are *Hayococcus floralis* and *Tetralithus* spp., found between 465 and 505 mbsf, in which calcite is often replaced by siderite. *H. floralis* indicates a Carnian age, whereas genus *Tetralithus* is usually identified as a Norian genus.

Palynomorphs
An assemblage typical of the *Minutosaccatus crenulatus* Zone has been found from ~350 to ~285 mbsf. The rest of the Hole 760B yielded palynomorphs indicating a transition zone between *Samaropollenites speciosus* and *M. crenulatus* (as seen also in Hole 759B). *Enzonalasporites vigens* is abundant from ~396 to ~400 mbsf, indicating a Carnian age (Brenner, 1992).

Dinoflagellates
First occurrence of dinoflagellates *Suessia listeri* and *Suessia swabiana* (*Suessia listeri* Zone; Helby et al., 1987) occurred at ~350 mbsf (at the base of the
Minutosaccatus crenulatus palynozone). The sporadic abundance of Bartenia communis suggest a middle-late Norian age for this interval (Helby et al., 1987), although the absence of Heibergella balmei suggest an early Norian age (although the presence or absence of H. balmei could be environmentally controlled). From ~350 mbsf to the bottom of the Hole, the absence of Suessia listeria and the presence of Shublikodinium spp. suggest the Shublikodinium wigginsii Zone. Unfortunately, the rare occurrence of Shublikodinium spp. and the absence of Suessia swabiana (typical of the upper Shublikodinium wigginsii Zone) indicate that dinoflagellates of this interval are environmentally controlled and that cannot be used for biostratigraphic investigations.

Ostracods
From ~305 to ~312 mbsf, the presence of ostracods Omoconcha martini and Rhombocythere penarthensis suggests a late Norian to Rhaetian age (Dépêche and Crasquin-Soleau, 1992). The age based on ostracods is in accordance with palynology (presence of middle-late Norian Bartenia communis in the Minutosaccatus crenulatus palynozone).

Paleomagnetism
Methods
A total of 90 non-oriented ~10cc minicores and cubes from the Hole 759B, and 103 from Hole 760B, have been analysed for paleomagnetism and rock magnetism. Samples have been provided by the Kochi Core Center (Kochi University, Japan) and analyzed at the “Fort Hoofddijk” Paleomagnetic Laboratory (Utrecht University, The Netherlands). Samples have been demagnetized progressively by application of an alternate gradient field (using a 2G Enterprises singe-axis AF demagnetizer) and measured using a 2G Enterprises RF-SQUID magnetometer. Samples have been demagnetized by steps of 5 mT up to 50 mT, then steps of 10 mT up to 100 mT. Single sample NRM directions for each step of demagnetization have been plotted on an end-point vector graph (Zijderveld, 1967), and only the magnetization components made of at least three subsequent end-points have been considered. Cores from Holes 759B and 760B are not oriented, and the samples are oriented only respect to the cores. So only the magnetic inclinations have been considered to determine the directions of magnetization. The low-field magnetic susceptibility (κ) was measured with an AGICO Kappabridge MFK1-A instrument on 87 samples from Hole 759B and on 84 samples from Hole 760B. Rock magnetism experiments have
been performed on selected samples to support the paleomagnetic interpretations. Thermomagnetic runs were performed on 8 samples from Hole 759B and 8 samples from Hole 760B, using a modified horizontal translation Curie balance (noise level 5x10^{-9} \text{ Am}^2; \text{Mullender et al., 1993}, measuring in air. Powdered samples (70-80 mg) was measured increasing temperature in several cycles, up to 580\textdegree C. Field cycles was between 100 and 300 mT, with heating-cooling rates of 10\textdegree C/min. Hysteresis cycles, IRM acquisition and backfield IRM have been performed on 23 samples from Hole 759B and on 19 samples from Hole 760B, using an alternate field gradient magnetometer (Princeton Measurement Corp. AGM 2900) on rock fragments of about 50 mg (maximum field 500 mT, steps of 10 mT).

**Magnetic properties**

*Hole 759B*

Susceptibility (\(\kappa\)) is almost constant (~125\times10^{-6} \text{ SI}) up to 110 mbsf, where it decreases abruptly (~46\times10^{-6} \text{ SI}) (Fig. 4.33). The sudden decrease of \(\kappa\) could be related to a strong reduction of siliciclastic material in carbonate levels (usually contaminated by terrigenous from fluvial discharge), probably due to one of the frequent movement of distributary channels (for details in environmental changing see “Paleoenvironments” paragraph, Chapter 4.2.3). Thermomagnetic curves of samples wbb3709801, 3711801, 3713201, 3714901, 3716501, 3720701, 3722201 and 3722901 indicate a weak magnetization (frequently below 0.02 Am^2/kg), in particular for wbb3713201 that does not reach 0.005 Am^2/kg (Fig. 4.35). The curves of samples wbb3709801 (~165 mbsf), 3720701, 3722201 and 3722901 (from ~262.1 to ~299 mbsf) show an increase in magnetization between 400 and 580\textdegree C (followed by a gentle increase during cooling, until room temperature is reached) (Fig. 4.35), probably due to the oxidation of a variable quantity of pyrite (FeS_2) and siderite (FeCO_3) in magnetite (Fe^{3+}_2\text{Fe}^{2+}_3\text{O}_4). Probably pyrite is the dominating magnetite-forming mineral in wbb3709801, while the stratigraphic interval containing the other three samples is rich in siderite (but also pyrite is present). Below 400\textdegree C there are no evidences of magnetic iron-sulfides (e.g. pyrrhotite – FeS), since the heating-cooling steps are totally reversible up to this temperature, and the Curie temperature for magnetic Fe-sulfides (~320\textdegree C) has not been reached (Fig. 4.35). The original magnetization is carried probably by magnetite but is impossible to differentiate the neo-formed magnetite to the original magnetite. Samples wbb3711801, 3713201, 3714901 and 3716501 (from ~186.3 to ~246.6 mbsf) show a similar behavior, but the increase
is very (sometimes extremely) slight (Fig. 4.35). This difference with the other samples is probably due to a smaller amount of magnetite-forming minerals (in this interval pyrite is common).
IRM acquisition curves show variable levels of saturation, from ~300 mT to ~500 mT, coherent with the presence of magnetite (Fig. 4.36A). Samples subjected to backfield IRM are characterized by coercivity fields from 40 to 60 mT, associated to magnetite (Fig. 4.36B). Hysteresis cycles are pot-bellied shaped (Fig. 4.36C), suggesting a mixture of single-domain (SD) and superparamagnetic (SP) magnetite (Tauxe et al., 1996).

**Hole 760B**

Susceptibility ($\kappa$) is normally around $100 \times 10^{-6}$ SI, increasing to $250 \times 10^{-6}$ SI from ~426 to ~440 mbsf and to $300 \times 10^{-6}$ SI from ~380 to ~400 mbsf (Fig. 4.34). Two
major peaks of 550-600×10^6 SI at ~387 mbsf and at ~316 mbsf are present, as well as a minor peak of 400×10^6 SI at ~350 mbsf (Fig. 4.34). The higher values of κ seem localized in the more carbonatic levels, probably related to the presence of

Figure 4.37: Thermomagnetic curves determined with a Curie balance of samples wbc3724701, 3727301, 3730401, 3732301, 3745501, 3745601, 3747101 and 3749401. See text for discussion.
siderite.

Thermomagnetic curves of samples wbc3724701, 3727301, 3730401, 3732301, 3745501, 3745601, 3747101 and 3749401 indicate a weak magnetization, normally below 0.02-0.03 Am²/kg (only sample wbc3732301 reach 0.05 Am²/kg) (Fig. 4.37). The curves of samples wbc3727301 (~323 mbsf), 3732301 (~392 mbsf), 3745501, 3745601 (from ~440 to ~441 mbsf) and 3747101 (~465 mbsf) show an increase in magnetization between 400 and 580°C, and then a smooth increase in cooling curve until room temperature (Fig. 4.37). This behavior is probably due to the oxidation of pyrite (or siderite, in the lower part of the Hole) in magnetite. Magnetic iron sulfides (e.g. pyrrhotite) seems not to be present since the heating-cooling steps are

Figure 4.38: IRM acquisition curve (A), IRM backfield (B), and hysteresis cycles (C) of wba samples from Hole 760B (see discussion in text). Purple line is the data-fitting curve.
reversible until 400°C and Curie temperature for magnetic Fe-sulfides (~320°C) has not been reached (Fig. 4.37). A small amount of primary magnetite is present, but it cannot be differentiated from the neo-formed magnetite. The behavior of samples wbc3724701 (~285 mbsf), 3730401 (~363 mbsf), and 3749401 (~492 mbsf) is similar, but the increasing in magnetization after 400°C is less intense (Fig. 4.37). These small differences are probably related to the different amount of pyrite (or siderite, probably in sample wbc3747101) in the rocks, which seems not to be controlled by lithology. More likely, the various pyrite content is related to temporary reducing conditions.

Sample subjected to IRM acquisition show variability in saturation field, generally from ~300 mT to ~500 mT (Fig. 4.38A), coherent with the presence of magnetite. Few samples seems not to reach saturation above 500 mT. Backfield IRM curves revealed coercivity fields from 40 to 80 mT, ascribed to magnetite (Fig. 4.38B). The shape of hysteresis cycles is pot-bellied (Fig. 4.38C), typical of single-domain (SD) and superparamagnetic (SP) magnetite mixture (Tauxe et al., 1996).

**Magnetostratigraphy**

**Hole 759B**

Mean intensity of initial NRM is ~0.58 mA/m, lower in the upper part (~0.07

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Figure 4.39: Vector end-point demagnetization diagrams for representative samples from Hole 759B. Closed circles are projections onto the horizontal plane, and open circles are projections onto the vertical plane for cores coordinates.
mA/m; 52 to 90 mbsf) and higher in the middle (~1.1 mA/m; 186 to 243 mbsf). The lower intensities are located around the Norian carbonate, where the siliciclastic input decreased. Characteristic magnetization (ChRM) have been identified in 71 samples, mainly above 15-20 mT and maximum until 100 mT (ChRM data in Appendix A.3), resulting stable in vector end-point demagnetization plots (Zijderveld, 1967) (Fig. 4.39). Equal-area projection for inclination-only data show a substantial variability of inclinations, mainly localized around 30°-50° (Fig. 4.40). This is confirmed by the calculation of the mean inclination: 42.2°±5.4° (k=10.4, N=71; McFadden and Reid, 1982). The mean inclination of 759B differs from the Geomagnetic Axial Dipole (GAD) inclination value at the latitude of the site (Inc_{GAD} = -31°), suggesting
that the paleomagnetic data of the Hole are not affected by contamination of VRM. Samples have no geographic orientation, so the paleomagnetic polarity must be derived from the magnetic inclination. In Late Triassic, the Wombat Plateau was in the southern hemisphere, hence positive inclination means normal polarity and negative inclination means reverse polarity. The sequence of ChRM inclinations define 17 polarity inversions, grouped in 5 magnetozones named WMB1n to WMB3n, where single data-points have been considered as partial reversals (Fig. 4.33).

**Hole 760B**

The mean intensity of initial NRM is ~0.67 mA/m, and is higher between 380 and 420 mbsf (~1.21 mA/m). A characteristic NRM magnetization (ChRM) has been recognized in 67 samples generally between 10 and 60 mT (maximum until 100 mT; ChRM data in Appendix A.3), as shown in vector end-point demagnetization diagrams (Zijderveld, 1967) (Fig. 4.41). Paleomagnetic inclinations distribution in equal-area projection show a certain variability (Fig. 4.42), with a mean inclination of 31.5°±5.6° (k=10.5, N=67; McFadden and Reid, 1982). Mean inclination of Hole 760B is dramatically similar to the GAD inclination at the latitude of the Site (Inc_{GAD} = -31°), which means that there is a probable contamination of VRM. As for Hole 759B, only the magnetic inclination must be considered to define the polarity inversions, following the criteria for the southern hemisphere. The stratigraphic sequence of ChRM inclinations shows 16 normal/reverse polarity shifts, defining two long magnetozones named WMC1n and 1r; 8 single data points define likewise partial inversions (Fig. 4.34).

**Discussion**

The magnetostratigraphy of Holes 759B and 760B have been compared with other
Figure 4.43: The Carnian-Norian magnetostratigraphy and biostratigraphy of Holes 759B and 760B, correlated to data from marine sections from the literature, such as Pignola-2, D'borea (Chapter 4.2.1), Silická Brezová (Channell et al., 2003), Pizzo Mondello (Muttoni et al., 2004), and Guri Zì (Muttoni et al., 2005, 2014). Holes 759B and 760B are correlated to the Newark astrochronological polarity time scale (APTS) through the statistical correlation of Pignola-2 with the APTS (Chapter 4.2.1).
Carnian/Norian sections, as Pignola-2, Dibona, Silická Brezová (Channell et al., 2003), Pizzo Mondello (Muttoni et al., 2004) and Guri Zi (Muttoni et al., 2005, 2014) (Fig. 4.43). The palynomorph biozonation places the magnetozones WMB1n to 3n in the Carnian (see paragraph “Palynomorphs”, Chapter 4.2.3; Fig. 4.33), whereas the unknown polarity zone covers mainly the transitional Carnian/Norian interval (Samaropollenites speciosus/Minutosaccatus crenulatus transitional Zone). The partial reverse magnetozone between Cores 759B-7R and 759B-8R is considered Norian in age, following the palynomorphs and dinoflagellates biozonation (Fig. 4.33). In Hole 760B, palynomorphs and dinoflagellates diagnostic for Norian age have been found in WMC1r, while the large normal magnetozone below (WMC1n) is considered Carnian to Norian (Fig. 4.34). Using these calibrations, is reasonable to think that partial reversals WMB3n.1r and WMB3n.2r are coeval respectively to WMC1n.1r and WMC1n.2r (Fig. 4.43). Moreover, the palynomorph Samaropollenites speciosus is found also in Pignola-2 section, in correspondence of the transition to magnetozones MP4n and MP4r (Fig. 4.43). Magnetozone WMB1n should correspond to MD1n (Dibona), MP4n (Pignola-2) and GZ3n (Guri Zi); WMB1r to MD1r, MP4r and GZ3r; WMB2n to MD2n, MP5n, GZ4n, SB-1n (Silická Brezová) and PM1n (Pizzo Mondello); WMB2r to MD2r, MP5r, GZ4r, SB-1r and PM1r; WMB3n and WMC1n to MD3n, MP6n, GZ5n, SB-2n and PM2n (Fig. 4.43). Magnetozone WMC1r is considered middle-late Norian for the abundance of dinoflagellate Bartenia communis (coherent with the founding of upper Norian ostracod Ogmocconcha martini), so is reasonable to think that the base of this magnetozone should be coeval to the base of SB-4r in Silická Brezová, of PM5r in Pizzo Mondello, and of GZ6r in Guri Zi (Fig. 4.43). Partial reversals within WBC1r could be part of the normal magnetozones included in the mainly reverse interval SB-4r/SB8r in Silická Brezová (coeval to PM5r/PM8r interval in Pizzo Mondello) (Fig. 4.43).

Following the correlation between the Carnian/Norian sections here considered and the Newark APTS (see paragraph “Correlation with the Newark APTS” in Chapter 4.2.1), the magnetostratigraphy of Holes 759B and 760B should cover the interval E5n to E8r (~6 My), although it could be more extended, considering that probably the partial reversal within WBC1r could be the normal magnetozones E9n and above (Fig. 4.43).
Chapter V
A GEOMAGNETIC POLARITY TIME SCALE FOR
THE LATE TRIASSIC

5.1 BRIEF HISTORY OF THE GPTS
A Geomagnetic Polarity Time Scale (GPTS) assigns a duration to each magnetozone in a sequence of geomagnetic polarity reversals. Using a GPTS is possible to assign an age to the events (biologic, climatic, geodynamic, etc.) calibrated to the magnetostratigraphy of the same period.

The first attempts to construct a GPTS for the Triassic started in 1960s-1970s, but they were fragmentary and poor-detailed. Khramov (1963) proposed the first GPTS for Early-Middle Triassic, based on the Moscow Basin and correlated with other studies from literature. Later, other authors (McElhinny and Burek, 1971; Pergament et al., 1971; Pechersky and Khramov, 1973; Molostovsky et al., 1976) proposed their version of GPTS for the entire Triassic. Contrary to the Late Jurassic to Pleistocene magnetostratigraphic record, mainly derived from linear magnetic anomalies of the sea floor, Triassic magnetostratigraphy was derived mainly from sedimentary successions in which the magnetization was very low-preserved. After the introduction of the more sensitive SQUID magnetometer in the late 1980s, paleomagnetic analyses provided more detailed magnetostratigraphic investigations. The integrations of detailed magnetostratigraphic studies with detailed biostratigraphies, mostly based on ammonoids, conodonts, palynomorphs and radiolarians, paved the way to the construction of a precise GPTS. In 2010, Hounslow and Muttoni proposed a GPTS for the Triassic, based on marine sections with biostratigraphically calibrated magnetostratigraphy, focusing on the Stage boundaries that are fixed points in the geochronology. To accommodate the differences in sedimentation rates, the magnetostratigraphy of the considered sections has been stretched (or shrunked), creating a Triassic composite magnetostratigraphy in a pseudo-height scale, which permits a graphic correlation. Marine sections were preferred for their greater chances to confirm correlations using different constraints (biostratigraphic, chemostratigraphic, etc.). To reshape the Triassic composite magnetostratigraphy in a GPTS, the metric length scale have been converted in a time scale (Hounslow and Muttoni, 2010) using linear interpolation.
or radiometric ages from Induan (first Stage of the Triassic) to Carnian. The Newark APTS ages (Olsen and Kent, 1999) has been applied to calibrate the Rhaetian, using the Triassic/Jurassic boundary age of ~201.6 Ma (Schaltegger et al., 2008) as a tie-point. Problems remain for the Norian Stage, which is poorly constrained by radiometric ages and the correlation with the Newark APTS was not certain.

5.2 A NEW PROPOSAL OF GPTS

The correlations between the Carnian/Norian sections proposed in Chapter 4.2.1 (Pignola-2 and Dibona, Fig. 4.28), in association with the correlations between the Norian/Rhaetian sections in Chapter 4.1.1 (Pignola-Abriola, Fig. 4.7), allow to assemble a composite magnetostratigraphy of Tethyan sections valid for the Late Triassic. The magnetostratigraphy of Leg 122 (Sites 759-760, Chapter 4.2.2; Site 761, Chapter 4.1.3) is not included in the Late Triassic composite, because of the extended stratigraphic discontinuities and poor bio- and chemostratigraphic constraints. Similarly, magnetostratigraphy of Mt. Messapion (Chapter 4.1.2) is excluded for its scarce reliability. This long-time extended composite magnetostratigraphy includes many reversals (Fig. 5.1; high-definition version in Appendix B), defining a unique pattern that allows a more precise correlation with the Newark APTS. The chance to build a Geomagnetic Polarity Time Scale for the Late Triassic, connecting our composite polarity scale to the Newark APTS, is given by radiometric constraints. In facts, the Carnian magnetostratigraphy of the Pignola 2 section is constrained at 230.91±0.33 Ma (Furin et al., 2006) and the age of the Rhaetian at Pignola-Abriola is confirmed by the U/Pb ages around the Norian/Rhaetian boundary in Peru (Wotzlaw et al., 2014). The Pignola-2 section is correlated confidently with the Pizzo Mondello section, and the correlation with the Newark APTS of which (Muttoni et al., 2004; Hounslow and Muttoni, 2010) was confirmed by the correlation of Norian/Rhaetian sections proposed in Chapter 4.1.1. Using the radiometric age of Pignola-2 (Furin et al., 2006) was possible to constrain the magnetostratigraphy of the Carnian in the lower APTS, strictly correlating magnetozone MP2r of Pignola-2 section with E5n.1r of the APTS (Fig. 5.1) using statistics (Fig. 4.29), while for the upper APTS the Rhaetian was constrained by the correlation between MPA5r in Pignola-Abriola and E20r of the APTS. Therefore, the Tethyan magnetostratigraphy of the Late Triassic is both anchored at the bottom (at 230.91 Ma) and the top of the APTS (at 205.7 Ma). The magnetostratigraphy of the APTS older than 230.9 Ma is associated to a composite of Dibona, Pignola-2, Guri Zi (Muttoni et al., 2005, 2014) and Bolücektasi Tepe
been derived considering a basin with constant sedimentation rate and a radiometric
for the lowermost Carnian time interval. The duration of these magnetozones have

(Gallet et al., 1992) sections until ~233 Ma, as explained in Chapter 4.1.1. The
Late Triassic composite magnetostratigraphy here proposed defines a pattern of
geomagnetic polarity reversals coherent with the Newark magnetostratigraphy,
confirming the correlations between the considered Tethyan marine sections and
the Newark APTS as proposed in the previous Chapters (4.1.1 and 4.2.1).

Since the Newark APTS does not record sediments younger than lower Tuvalian
(upper Carnian), the magnetostratigraphy of the Mayerling (Gallet et al., 1994)
and the Prati di Stuores/Stuores Wiesen (Mietto et al., 2012) have been integrated
for the lowermost Carnian time interval. The duration of these magnetozones have
been derived considering a basin with constant sedimentation rate and a radiometric
The duration of the magnetozones of Bolüçktası Tepe was derived extending the duration/thickness ratio of magnetozone BT5n, tied to E2n of the APTS, to the other magnetozones, up to ca. 234.5 Ma. The duration of the magnetozones of Mayerling and Prati di Stuores has been calculated comparing the duration and thickness of the whole interval between the beginning of magnetozone MY5n.2n/S2n.2n and the radiometric age of 237.77±0.14 Ma, and then assigning duration to each magnetozone proportionally to its thickness.

Furthermore, magnetostratigraphic correlation between the Tethyan GPTS and the continental section of Chinle Group have been attempted using the radiometric ages of the Petrified Forest (Ramezani et al., 2011) (Fig. 5.2). The magnetostratigraphy of the Petrified Forest (Steiner and Lucas, 2000) is partially inconsistent with the GPTS and the APTS, as well as with coeval sections of the Chinle Group, in particular in the Sonsela Sandstone Mb (Zeigler and Geissmann, 2012; Lucas and Spielmann, 2013). In fact, the magnetostratigraphy of the Poleo Fm (Zeigler et al., 2005) shows a large interval of reversal polarity (~30 m) that is considered time equivalent with the less extended Sonsela Sandstone Mb (~5 m; Steiner and Lucas, 2000). Lucas and Spielmann (2013) propose a composite magnetostratigraphy of the Chinle Group, including the Petrified Forest and relative radiometric ages, which appear more coherent with the APTS and the GPTS. After these considerations,
the GPTS of the Tethys is confirmed also by sections of other realms, as the ones belonging to the Chinle Group.

The potential of the Late Triassic GPTS is to assign an age or a duration to each event calibrated with the magnetostratigraphy occurring in this time interval. Virtually, the events of every realm could be calibrated with the GPTS, using markers of global extension (i.e. fossil markers as radiolarians, geochemical curves or magnetostratigraphy itself). There are many advantages of the GPTS against the APTS, for instance the full Late Triassic coverage (from Ladinian/Carnian to Rhaetian/Hettangian) and the clear definition of the Stage boundaries. In facts, the GPTS includes the Ladinian/Carnian GSPP of Prati di Stuores/Stuores Wiesen (Mietto et al., 2012) that is the only ratified Stage boundary in the Late Triassic. Regarding the Carnian/Norian boundary, the Stage is placed with the FO of the conodont *Carnepipgondolella gulloae* (Pizzo Mondello section), proposed as a primary biomarker for the Norian base (Mazza et al., 2012a). The Norian/Rhaetian boundary is proposed with a negative $\delta^{13}C_{org}$ spike (Maron et al., 2015; Rigo et al., 2015; Chapter 4.1.1) and approximated by the FAD of conodont *Misikella posthernsteini* (Pignola-Abriola section), proposed as one of the primary biomarkers for the Rhaetian Stage (i.e. Krystyn, 2010; Ogg, 2012). The Triassic/Jurassic

<table>
<thead>
<tr>
<th>Event</th>
<th>Stage</th>
<th>Substage</th>
<th>Section</th>
<th>Age</th>
</tr>
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<tr>
<td>LO of <em>Gladigondolella</em> spp.</td>
<td>Carnian</td>
<td>Julian</td>
<td>Pignola-2</td>
<td>~230.8 Ma</td>
</tr>
<tr>
<td>Carnian Pluvial Event</td>
<td>Carnian</td>
<td>Julian to Tuvalian</td>
<td>Pignola-2 Dibona</td>
<td>~229.7 Ma to ~230.7 Ma</td>
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<td>FO of <em>Metapolygnathus communisti</em></td>
<td>Norian</td>
<td>Lacin</td>
<td>Pizzo Mondello</td>
<td>~227.3 Ma</td>
</tr>
<tr>
<td>FO of <em>Epigondolella rigoi</em></td>
<td>Norian</td>
<td>Lacin</td>
<td>Pizzo Mondello Silická Brezová</td>
<td>~227 Ma to ~226.6 Ma</td>
</tr>
<tr>
<td>FO of <em>Carnepipgondolella gulloae</em></td>
<td>Norian</td>
<td>Lacin</td>
<td>Pizzo Mondello</td>
<td>~226.6 Ma</td>
</tr>
<tr>
<td>FO of <em>Mockina tozeri</em></td>
<td>Norian</td>
<td>Alaunian</td>
<td>Pizzo Mondello</td>
<td>~221.5 Ma</td>
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<td>FO of <em>Mockina bidentata</em></td>
<td>Norian</td>
<td>Sevatin</td>
<td>Pignola-Abriola</td>
<td>~216.2 Ma</td>
</tr>
<tr>
<td>FO of <em>Misikella hernsteini</em></td>
<td>Norian</td>
<td>Sevatin</td>
<td>Pignola-Abriola</td>
<td>~210.8 Ma</td>
</tr>
<tr>
<td><em>Proparvicingula moniliformis</em> Zone</td>
<td>Norian</td>
<td>Sevatin</td>
<td>Pignola-Abriola</td>
<td>~206.2 Ma</td>
</tr>
<tr>
<td>FO of <em>Misikella posthernsteini</em></td>
<td>Rhaetian</td>
<td>--------</td>
<td>Pignola-Abriola</td>
<td>~205.7 Ma</td>
</tr>
<tr>
<td>Negative $\delta^{13}C_{org}$ spike</td>
<td>Rhaetian</td>
<td>--------</td>
<td>Pignola-Abriola</td>
<td>~205.7 Ma</td>
</tr>
<tr>
<td>FO of <em>Misikella ultima</em></td>
<td>Rhaetian</td>
<td>--------</td>
<td>Pignola-Abriola</td>
<td>~204.7 Ma</td>
</tr>
</tbody>
</table>

Table 5.1: Ages of the main Late Triassic events. Here are reported the events defining the Stage and substage boundaries, as well as the Carnian Pluvial Event, the largest climatic episode occurred in the Late Triassic before the End Triassic Extinction event.
boundary (TJB) is placed in the GPTS using the palynomorphs biostratigraphy of Brumano/Italcementi Quarry section (Muttoni et al., 2010), obtaining an age of about 201.3-201.4 Ma that is coherent with the radiometric age of TJB relative to the FO of ammonoid *Psiloceras spelae* (201.36±0.17 Ma; Wotzlau et al., 2014) (Fig. 5.2). The age of substages is approximated by conodonts biostratigraphy in the considered sections. The Julian/Tuvalian boundary (Carnian) is placed with the LO of *Gladigondolella* spp. in the Pignola-2 section (Rigo et al., 2007), just below the “green clay-radiolaritic horizon”, where the typical Tuvalian conodonts was found only above this horizon. The base of the Alaunian (Norian) can be placed with the FO of *Mockina tozeri* in Pizzo Mondello (Mazza et al., 2012a), and the Sevatian is normally defined by the FO of *Mockina bidentata* in Pignola-Abriola (Maron et al., 2015; Rigo et al., 2015). The ages of the events here reported are presented in Table 5.1.
Chapter VI
CONCLUSIONS

6.1 RHAETIAN

6.1.1 Pignola-Abriola (from Maron et al., 2015: GSA Bulletin, v. 127, p. 962-974; see Attached publications)
Paleomagnetic data obtained from the Pignola-Abriola section provided a sequence of 22 polarity reversals grouped in 10 magnetozones. The correlation between the Pignola-Abriola section and additional Norian and Rhaetian Tethyan marine sections from the literature (Steinbergkogel, Oyuklu, Brumano, Italcementi Quarry, and Pizzo Mondello) reveals significant internal consistency.
To provide numerical age constraints on the Pignola-Abriola section, I applied a statistical correlation to the Newark APTS, which is assumed to be continuous in its younger part (contra Lucas et al., 2012). Three out of a total of 19 explored correlation options produced statistically reliable results, and after a thorough analysis, one option (7.1) is considered as the most reliable. According to this option, the Pignola-Abriola section correlates to Newark magnetozones E13n to E20r. I place the Norian-Rhaetian boundary at Pignola-Abriola at a level coincident with a rapid decrease in $\delta^{13}C_{\text{org}}$ to $\sim$30%, which virtually coincides with the level containing the FAD of conodont Misikella postherneudi sensu stricto within the Proparvicingula moniliformis radiolarian zone. This level is traced within Newark magnetozone E20r at 205.7 Ma.
Assuming an age of ca. 201.3 Ma for the Triassic-Jurassic boundary (Schoene et al., 2010; Guex et al., 2012; Wotzlaw et al., 2014), our study shows that the Rhaetian is $\sim$4.4 m.y. long. Assuming a Carnian-Norian boundary age of ca. 227 Ma (Muttoni et al., 2004, 2014, and references therein), our study shows that the Norian is $\sim$21.3 m.y. long.

6.1.2 Mount Messapion
Paleomagnetic analysis of the Mt. Messapion sample provided only a very small number of reliable data. The resulting magnetostratigraphy consisting only of few data points, which made impossible to define clearly the magnetozones. The possibility of a bad-preserved signal was suggested by the extremely low
susceptibility, mostly negative, although the demagnetization path suggest that some kind of magnetization is retained by rock. Unfortunately, in the great part of the samples the directions are unusual, mirroring the typical directions of the southern hemisphere (whereas the site was in the northern hemisphere in the Late Triassic). Filtering the data for these strange directions, the final dataset resulted too small to attempt a magnetostratigraphy of this section.

6.1.3 Leg 122 – Hole 761C
The paleomagnetic analysis on the Hole 761C of the Wombat Plateau provided a sequence of 5 magnetozones, showing discontinuity because of the scarce recovery in some of the Cores. The magnetostratigraphy is constrained in the uppermost Norian to middle Rhaetian by calcareous nanofossils, ostracods, foraminifera, dinoflagellates and palynomorphs biostratigraphy. The correlation with the Norian/Rhaetian sections of Pignola-Abriola, Brumano/Italcementi Quarry and Steinbergkogel seems to confirm the partial magnetostratigraphy of Hole 761C, as well as a lack of sediments around the Norian/Rhaetian boundary. By a comparison with the Newark APTS the hiatus should be ~1.5 My. The Rhaetian strata in Hole 761C are in unconformity with overlying Lower Cretaceous (Berriasian/Valanginian) strata. The whole Jurassic and the upper Rhaetian is missed, probably by erosion or condensation/non-deposition (at least 1.5 My of missing Rhaetian, by comparison with the Newark APTS). Hence, at least ~57.5 My of Late Triassic/Jurassic sediments are missing in the Wombat Plateau.

6.2 CARNIAN

6.2.1 Pignola-2 and Dibona (from Maron et al., submitted and in review, Newsletter on Stratigraphy; see Attached publications)
The paleomagnetic analyses of the Carnian sections of Pignola-2 (Southern Apennines, Italy) and Dibona (Dolomites, Italy) provided respectively a sequence of 12 and 8 magnetozones.
The correlation with other Tethyan sections of the same time interval (Pizzo Mondello, Silická Brezová, Guri Zi) reveals a virtually continuous magnetostratigraphic record for the Carnian, constrained by a radiometric age of 230.91±0.33 Ma.
Using a statistical approach, I performed a correlation between the Pignola-2 section and the Newark APTS that led to three statistically relevant options. Only two of them (Options 1 and 2) appear to be broadly consistent with both the radiometric age
of the Pignola-2 ash-layer and their correlative ages in the Newark APTS. Although Option 2 was statistically slightly more robust, Option 1 is provisionally preferred as it shows the highest matching between radiometric and astrochronologic age estimates of the Pignola-2 ash-layer and does not violate the correlation between the Newark APTS and the Pizzo Mondello section as proposed by Muttoni et al (2004) using the same statistical method adopted in this study. Ages of the main events of the Pignola-2 and Dibona sections were calculated using a model derived from Option 1. The level containing the Julian/Tuvalian boundary defined by conodonts is now calibrated at ~230.7 Ma, and the levels attributed to the Carnian Pluvial Event should have deposited between ~229.7 and 230.7 Ma.

6.2.2 Leg 122 – Holes 759B and 760B
Paleomagnetic analysis of Holes 759B and 760B (Wombat Plateau) provided respectively a sequence of 5 and 2 magnetozones. Due to poor recovery and bad-preserved ChRM, magnetostratigraphy of the upper part of Hole 759B and middle part of Hole 760B is considered as unknown polarity. Calcareous nanofossils, palynomorphs, dinoflagellates and ostracods biostratigraphy constraint Holes 759B/760B into the Carnian to middle Norian. Although some parts of Holes 759B/760B magnetostratigraphy are missing or have a low resolution, it seems confirmed by the correlation with the Carnian/Norian sections of Pignola-2, Dibona, Silická Brezová, Pizzo Mondello and Guri Zi. The comparison with the Newark APTS, obtained through the correlations between the APTS and Pignola-2, Dibona, Pizzo Mondello and Silická Brezová sections (see Chapter 4.2.1 and references within), suggest that the stratigraphic record of Holes 759B/760B covers ~6 My, straddling the Carnian/Norian boundary.

6.3 LATE TRIASSIC GPTS
A composite magnetostratigraphy have been obtained combining the Late Triassic Tethyan marine sections of Pignola-Abriola (Norian/Rhaetian), Brumano/Italcementi Quarry (Rhaetian/Hettangian), Pizzo Mondello (Carnian/Norian), Silická Brezová (Carnian/Norian), Pignola-2 (Carnian), Dibona (Carnian), Guri Zi (Carnian), Bolücektasi Tepe (Carnian), Mayerling (Ladinian/Carnian), and Prati di Stuores/Stuores Wiesen (Ladinian/Carnian). Through the statistical correlation between the Newark APTS of Pizzo Mondello (Muttoni et al., 2004), Pignola-Abriola (Chapter 4.1.1; Maron et al., 2015), and Pignola-2/Dibona (Chapter 4.2.1),
the Tethyan magnetostratigraphic composite has been calibrated with the APTS, also using the radiometric age of 230.91±0.33 Ma from Pignola-2 (Aglianico ash-bed) as a tie point. The part of GPTS not covered by the Newark APTS, which is the portion of late Ladinian/early Carnian represented by lower Bolücektasi Tepe, Mayerling and Prati di Stuores/Stuores Wiesen magnetostratigraphy, is time-calibrated using the radiometric age of 237.77±0.14 Ma for the lattermost Ladinian (Mietto et al., 2012) as a tie-point. The time-calibration of the magnetozones in this interval is calculated assuming a mostly constant sedimentation rate within sections, using the thickness of the magnetozones as a proxy of their duration. The resulting sequence of magnetochrons of the GPTS have been named LTM 1 to LTM 25. Using the GPTS is possible to assign an age to the events occurring in the Late Triassic that are calibrated with magnetostratigraphy. Ages of key events have been reported as obtained from the GPTS.
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APPENDIX

A

ChRM DATA AND STATISTICS
## APPENDIX A.1
### PIGNOLA-ABRIOLA SECTION

### ORGANIC CARBON

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<thead>
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**Legend**
- **Comp.** Type of component
- **N** Number of demagnetization steps used to define the magnetic component
- **A/F** Magnetic component anchored to origin (A) or free from origin (F)
- **MAD** Mean Angular Deviation of the magnetic component
- **%VAR** Percentage of Variance
- **Inc** Inclination of the magnetic component
- **Dec** Declination of the magnetic component
- **Trt. min-max (°C)** Treatment min-max (°C)
# Appendix A.2

## Mount Messapion Section

### Characteristic Components of Magnetization (ChRM)

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### Legend

- **Comp.**: Type of component
- **N**: Number of demagnetization steps used to define the magnetic component
- **A/F**: Magnetic component anchored to origin (A) or free from origin (F)
- **MAD**: Mean Angular Deviation of the magnetic component
- **%VAR**: Percentage of Variance
- **Dec**: Declination of the magnetic component
- **Inc**: Inclination of the magnetic component
### FILTERED CHARACTERISTIC COMPONENTS OF MAGNETIZATION (ChRM)

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#### LEGEND

- **MAD**: Mean Angular Deviation of the magnetic component
- **Det**: Declination of the magnetic component
- **Inc**: Inclination of the magnetic component
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**PIGNOLA-2 Characteristic components (ChRM)**

**DIBONA Characteristic component (ChRM)**

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### LEGEND

- **Comp.** Type of component
- **N.** Number of demagnetization steps used to define the magnetic component
- **A/F.** Magnetic component anchored to origin (A) or free from origin (F)
- **MAD.** Mean Angular Deviation of the magnetic component
- **%VAR.** Percentage of Variance
- **Inc.** Inclination of the magnetic component
- **Dec.** Declination of the magnetic component

### STATISTIC CORRELATION OPTIONS

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- **NEWARK**
- **Mz.** Zone thickness

### STATISTIC CORRELATION OPTIONS

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### Graph

The graph illustrates the t-value distribution for various options, with confidence limits marked at 90% and 95% levels. Each point on the graph corresponds to a specific option, with its t-value and confidence limits indicated. The t-value scale ranges from -5.000000 to 5.000000, and the confidence levels indicate the range within which the true t-value is expected to lie with 90% and 95% confidence, respectively.
APPENDIX B

GEOMAGNETIC POLARITY TIME SCALE
The Geomagnetic Polarity Time Scale (GPTS) here proposed (on the right) is based on the magnetostratigraphic correlations between Tethyan marine sections and the Newark APTS (Olsen and Kent, 1999; Olsen et al., 2015). The marine sections considered are Brumano/Ialcumenti Quarry (Muttoni et al., 2010), Pignola–Abriola (Chapter 4.1.1; Maron et al., 2015; Rigo et al., 2015), Steinbergkogel (Hising et al., 2011), Pizzo Mondello (Muttoni et al., 2004), Sílvia–Brezavá (Channell et al., 2005), Pignola-2 (Chapter 4.2.1), Dibona (Chapter 4.2.1), Guri Zi (Muttoni et al., 2005, 2014), Bolívaríkasi Tepe (Gallet et al., 1992, 1994), Prati di Stuores/Stuores Wiesen (Mietto et al., 2012), and Mayerling (Gallet et al., 1994).

A composite magnetostratigraphy based on these Tethyan marine sections has been compared to the Newark APTS, following the statistical correlation between the APTS and Pizzo Mondello (Muttoni et al., 2004), Pignola–Abriola (Chapter 4.1.1; Maron et al., 2015), and Pignola-2 (Chapter 4.2.1). The correlation between the APTS and Brumano/Ialcumenti Quarry is based on the proposal of Muttoni et al. (2004) and is confirmed by the correlation with the Pignola–Abriola section (Chapter 4.1.1; Maron et al., 2015). The Newark APTS does not reach the Ladinian/Carnian boundary, so it is not possible to date directly the magnetostratigraphy of Prati di Stuores and Mayerling (both deposited in a basin, where the accumulation rates normally show slight fluctuations) and as a tie-point the U/Pb age of 237.73±0.052 Ma at the Ladinian/Carnian boundary (Mietto et al., 2012). The colored intervals besides the GPTS represent the contribution of the considered sections in the Late Triassic composite magnetostratigraphy (matching the colored names above each section), containing the names of the considered magnetozones. For a detailed description of the GPTS see Chapter 3.2.

Magnetostratigraphy, biostratigraphy, and chemostratigraphy of the Pignola-Abriola section: New constraints for the Norian-Rhaetian boundary

Matteo Maron, Manuel Rigo, Angela Bertinelli, Miriam E. Katz, Linda Godfrey, Mariachiara Zaffani and Giovanni Muttoni

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doi: 10.1130/B31106.1

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Notes
M magnetostratigraphy, biostratigraphy, and chemosтратigraphy of the Pignola-Abriola section: New constraints for the Norian-Rhaetian boundary

Matteo Maron1,‡, Manuel Rigo1, Angela Bertinelli2, Miriam E. Katz1, Linda Godfrey4, Mariachiara Zaffani1, and Giovanni Muttoni5

1Department of Geosciences, University of Padova, via G. Gradenigo 6, 35131 Padova, Italy
2Department of Physics and Geology, University of Perugia, via G. Pascoli, 06123 Perugia, Italy
3Department of Earth and Environmental Sciences, Rensselaer Polytechnic Institute, 110 8th Street, Troy, New York 12180, USA
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5Department of Earth Sciences, University of Milan, via L. Mangiagalli 34, 20133 Milan, Italy

ABSTRACT

A detailed magnetostratigraphic investigation of the Pignola-Abriola section of Norian to Rhaetian age permits identification of 22 magnetic polarity reversals grouped in 10 magnetozones. We correlate the magnetostratigraphy of the Pignola-Abriola section with the Newark astrochronological polarity time scale (APTS). In total, 19 correlation options were tested, and only one (option 7) yielded a statistically significant correlation that was consistent with the available information on the stratigraphic age of the Newark APTS. After some adjustments to minimize errors with sedimentation rates, a final correlation (option 7.1) was used to generate an age model of sedimentation for the Pignola-Abriola section. The Pignola-Abriola section has been correlated with Rhaetian sections from the literature, notably the current global boundary stratotype section and point candidate for the base of the Rhaetian at Steinbergkogel, Austria, where the Norian-Rhaetian boundary is proposed to be placed at a stratigraphic level containing the first appearance datum (FAD) of conodont Missikella posthersteinii, traced on the Newark APTS to ca. 209–210 Ma. Issues regarding the taxonomy of M. posthersteinii, a species characterized by transitional forms with its ancestor Missikella hernsteini, lead us to propose the alternative option of placing the Norian-Rhaetian boundary at a prominent negative δ13Corg spike observed in the Pignola-Abriola section at meter 44.5, 50 cm below the level containing the FAD of M. posthersteinii sensu stricto and close to the base of radiolarian Proparvicingula moniliformis zone. This level has been magnetostratigraphically correlated to Newark magnetozone E19r.2r at ca. 205.7 Ma. Assuming an age of ca. 201.3 Ma for the Triassic-Jurassic boundary, the Rhaetian Stage would have a duration of ~4.4 m.y.

INTRODUCTION

Current generations of time scales for the Triassic System (e.g., Ogg, 2012) are based on magnetostratigraphic correlations between marine sections and the continental Newark astrochronological polarity time scale (APTS). For marine sections, magnetostratigraphy is tied to stage boundaries that are defined biostratigraphically (e.g., Channell et al., 2003; Muttoni et al., 2004; Gallet et al., 2007), whereas the magnetostratigraphy of the Newark APTS (Kent et al., 1995; Kent and Olsen, 1999; Olsen and Kent, 1999; Olsen et al., 2011) is constrained by terrestrial biostratigraphy such as sporomorphs, tetrapod footprints, and conchostracans (e.g., Cornet, 1977, 1993; Olsen and Sues, 1986; Fowell et al., 1994; Lucas and Huber, 1993; Kozur and Weems, 2005, 2010; Lucas et al., 2012). The base of the Rhaetian Stage (Norian-Rhaetian boundary), pending formal designation by the International Commission on Stratigraphy, is currently placed at a stratigraphic level of the Steinbergkogel section (Austria) containing the first appearance datum (FAD) of the conodont Missikella posthersteinii (Krystyn, 2010). This level was magnetostratigraphically correlated to magnetozone E16r of the Newark APTS (Häsing et al., 2011), in substantial agreement with previous inferences based on magnetostratigraphic correlations between the Pizzo Mondello (Italy) and Silická Brezová (Slovakia) marine sections, which are Carnian–Norian in age, and the Newark APTS (Muttoni et al., 2004; Channell et al., 2003). The base of Newark magnetozone E16r is currently dated at ca. 210.3 Ma as the result of rescaling the Newark APTS (Kent and Olsen, 1999) from the base of the Orange Mountain Basalts of the Central Atlantic magmatic province (CAMP), recently dated at ca. 201.5 Ma (Blackburn et al., 2013). This would imply a duration for the Rhaetian Stage of ~9 m.y., in broad agreement with the long-Rhaetian option of Muttoni et al. (2010; see also Muttoni et al., 2004) and in contrast to the short-Rhaetian option (~2 m.y.) of Gallet et al. (2007; see discussions in Muttoni et al., 2010; Hüsing et al., 2011). The short-Rhaetian option was recently revived by Callegaro et al. (2012), who associated negative δ87Sr/86Sr and 187Os/188Os (Cohen and Coe, 2007; Kuroda et al., 2010) excursions observed starting at the base of the Rhaetian (defined by the occurrence of M. posthersteinii) with the emplacement of the CAMP which was considered to have started as early as ca. 202–203 Ma on the basis of 40Ar/39Ar dates (Marzoli et al., 2011; Callegaro et al., 2012), albeit recent U-Pb dates suggest rapid emplacement around ca. 201.56 Ma, coincident with the end-Triassic extinction event (Blackburn et al., 2013).

The debate over the duration of the Rhaetian (and Norian) is not yet settled, with two options proposed in the Geological Time Scale 2012 (Ogg, 2012): The long-Tuvalian option places the Norian-Rhaetian boundary at 205.4 Ma and the Carnian-Norian boundary at 221 Ma, whereas the long-Rhaetian option has a Norian-Rhaetian boundary at 209.5 Ma and a Carnian-Norian boundary at 228.4 Ma. These alternative options arise from different approaches to time scale construction. The long-Tuvalian option is grounded in biostratigraphic correlations of terrestrial groups (conchostracans, pollens, tetrapods) between the

GSA Bulletin; July/August 2015; v. 127; no. 7/8; p. 962–974; doi: 10.1130/B31106.1; 7 figures; 2 tables; Data Repository item 2015069; published online 3 February 2015.

E-mail: matteo.maron.1@studenti.unipd.it.
continental sequences of the Germanic Basin and the Newark Supergroup–based APTS (Lucas et al., 2012). The long-Rhaetian option is based on magnetostratigraphic correlations between marine sections bearing stage-defining fossils (conodonts, ammonoids) and the Newark APTS (Channell et al., 2003; Muttoni et al., 2004, 2010; Hüsing et al., 2011). In our opinion, the terrestrial correlation approach as the basis for the long-Tuvalian option (Lucas et al., 2012) is flawed by inherent difficulties in correlating terrestrial associations of, e.g., freshwater clam shrimps (conchostracans) to marine-based stage boundaries. The base of the Rhaetian was assigned directly in the Newark Supergroup using sporomorphs (base of the Upper Balls Bluff–Upper Passaic palynofloral zone of Cornet 1977, 1993), although the typical Rhaetian taxa of Europe and Greenland have not been found in the Newark Supergroup (Cornet, 1977).

In this paper, we contribute to the definition of the Norian-Rhaetian boundary by presenting new biostratigraphic, magnetostratigraphic, and chemostratigraphic data from the Pignola-Abriola section of Italy. This section records the FAD of *M. posthernsteini*, occurring in the lower *Proparvicingula moniliformis* radiolarian zone (Giordano et al., 2010). We date these events by means of magnetostratigraphic correlation with the Newark APTS, while addressing in detail the taxonomic complexities vexing the use of the conodont *M. posthernsteini* as proxy for the Norian-Rhaetian boundary level. We also illustrate the occurrence of a prominent negative $\delta^{13}C_{\text{org}}$ excursion at meter level 44.5, -0.5 m below the FAD of *M. posthernsteini* (within the base of the *P. moniliformis* zone), which serves as a useful geochemical proxy for the Norian-Rhaetian boundary level. The levels containing the negative $\delta^{13}C_{\text{org}}$ excursion and the FAD of *M. posthernsteini* are traced to Newark magnetozone E20r.2r at ca. 205.7 Ma, providing a younger age for the Norian-Rhaetian boundary relative to Hüsing et al. (2011), and very similar to the Norian-Rhaetian boundary in the long-Tuvalian option described in Ogg (2012).

**GEOLOGICAL SETTING**

The Pignola-Abriola section crops out on the western side of Mount Crocetta, along the road SP5 connecting the village of Pignola to Abriola (Potenza, southern Italy; Fig. 1, section A, coordinates: 40°33′23.50″N, 15°47′1.71″E). The road section is ~58 m thick (Fig. 2, left panel) and is complemented by an ancillary 7-m-thick
subsection (Fig. 2, right panel) outcropping close to a unused railway tunnel located ~10 m below the SP5 road level (Fig. 1, section B, coordinates: 40°33′24.74″N, 15°46′59.59″E). The stratigraphic sequence is composed of the Calcari con Selce (i.e., Cherty Limestone) Formation, which was deposited in the Lagonegro Basin, a branch of the western Tethys Ocean characterized by pelagic sedimentation since the Permian (Finetti, 1982, 2005; Catalano et al., 2001; Ciarapica and Passeri, 2002, 2005; Argnani, 2005; Rigo et al., 2012). The Calcari con Selce Formation consists of thinly bedded cherty hemipelagic to pelagic limestones (mudstones, wackestones, and rare packstones), interbedded with shales and marls, with common radiolarians, conodonts, and sporadic bivalves. The lower part of the section is dominated by cherty limestones, often dolomitized, intercalated with very thin marls or clayey levels (Fig. 2). The upper portion is instead dominated by an alternation of silicified limestones and black to brown or greenish, thinly laminated shales (Fig. 2), which are rich in organic matter, indicating deposition in dysoxic or anoxic conditions. Calcarenitic intercalations are also present throughout the section (Fig. 2). In particular, a 1.5-m-thick calcarenitic bank at ~35 m from the base of the measured section has been used as a lithostratigraphic marker to correlate the Pignola-Abriola road section (Fig. 2, left panel) to the railway tunnel subsection (Fig. 2, right panel).

BIOSTRATIGRAPHY

The fossil content of the Pignola-Abriola section consists mainly of conodonts and pyritized radiolarians. Here, we present an updated conodont and radiolarian biostratigraphy (Fig. 2) after recent biostratigraphic data published by Rigo et al. (2005), Bazzucchi et al. (2005), and Giordano et al. (2010).

Conodonts are well distributed along the entire section (representative specimens are listed in Table 1) and are characterized by a conodont alteration index (CAI) of 1.5 (Epstein et al., 1977; Bazzucchi et al., 2005; Rigo et al., 2005). The following main events have been recognized (Fig. 2):

1. (1) the first occurrence (FO) of Mockina bidentata at meter 7;
2. (2) the FO of Misikella hermsteini at meter 21.5, associated with the FO of Parvigondolella andrusovii;
3. (3) the FO of the Misikella hermsteini/post-hermsteini morphoclone at meter 33.5;
4. (4) the FO of Misikella buseri at meter 32;
5. (5) the FAD of Misikella posthermsteini at meter 45 in sample PIG24, in association with Misikella koessenensis; and

Figure 2. The Pignola-Abriola sections. From left to the right: conodont and radiolarian biostratigraphy (see Table 1 for key species), lithostratigraphy, virtual geomagnetic pole (VGP) latitudes calculated from characteristic remanent magnetization (ChRM) component directions, and derived magnetostratigraphy and chemostatigraphy ($\delta^{13}$Corg) of the Pignola-Abriola section. To the right is lithostratigraphy and VGP latitudes of the auxiliary subsection B. Black is normal polarity, and white is reverse polarity. The levels containing the first appearance datum (FAD) of conodont Misikella posthermsteini sensu stricto and the marked decrease in the $\delta^{13}$Corg to ~30‰ used to define the Norian-Rhaetian boundary are highlighted by dashed horizontal lines.
(6) the FO of Misikella ultima at meter 54. The radiolarian associations are well preserved and conform to the biozonation proposed by Carter (1993):

(1) Sample PR14 at meter 25 yielded a radiolarian assemblage referable to the Betraccium deweveri zone (Carter, 1993) and consisting of Betraccium deweveri Pessagno and Blome, Praemerosatunaris gracilis Kozur and Mostler, Tetraporobrachia sp. aff. T. composita Carter, Aytontius elizabethae Sugiyama, Citriduma sp. A sensu Carter (1993), Globolaxtorum sp. cf. G. hullae Yeh and Cheng, Lysemela sp. cf. L. olbia Sugiyama, Livarella valida Yoshida and Livarella sp. sensu Carter (1993) (Giordano et al., 2010); a similar assemblage was found also in sample PR15 at meter 23.5, and sample PR13 at meter 27. The presence of Globolaxtorum sp. cf. G. hullae Yeh and Cheng in this assemblage is atypical, because the genus Globolaxtorum is usually referred only to the Proparvicingula moniliformis and Globolaxtorum tozeri zones (O’Dogherty et al., 2009).

(2) Sample PA25 at meter 41 yielded a radiolarian assemblage referable to the Proparvicingula moniliformis zone assemblage 1 (U.A. 2–5 in Carter, 1993) for the presence of Fontinella primitiva Carter, Praemerosatunaris sp. cf. P. sandspitsensis Blome, Globolaxtorum sp. cf. G. hullae Yeh and Cheng, and Livarella densisiporata Kozur and Mostler (Bazzucchi et al., 2005; Giordano et al., 2010).

The Norian-Rhaetian boundary is conventionally placed in stratigraphic levels where the FAD of Misikella posternsteinii is documented (Krystyn, 2010), which is a phylogenetic descendent of M. hersteinii (e.g., Mostler et al., 1978; Kozur and Mock, 1991; Giordano et al., 2010). The transition from drop-shaped to heart-shaped basal cavity along with a reduction of the number of blade denticles characterize the evolution of the M. hersteinii posternsteinii morpholine (Giordano et al., 2010). Specimens characterized by an evident furrow on the backside of the cup and the associated inflection of the posterior margin of the basal cavity are here considered Misikella posternsteinii sensu stricto, as suggested by Giordano et al. (2010). At Pignola-Abriola, the presence of the Misikella hersteinii posternsteinii morpholine, as well as the presence of the FAD of M. posternsteinii sensu stricto (m 45, sample PIG24) provide a reliable (and continuous) biostratigraphic signal. Furthermore, in the Pignola-Abriola section, the conodont Misikella posternsteinii sensu stricto appears 4 m above the base of radiolarian Proparvicingula moniliformis zone assemblage 1 (Fig. 2), which is commonly adopted to define the early Rhaetian (e.g., Carter, 1993; Bertinelli et al., 2005; Giordano et al., 2010).

Table 1 (on following page). Scanning electron microscope (SEM) micrographs of Upper Norian and Rhaetian radiolarians and conodonts from the Calcarci con Selzione, Pignola-Abriola section. Radiolarians: samples PR13, PR14, and PR15 are referred to the Betraccium deweveri zone; sample PA25 is referred to the Proparvicingula moniliformis zone, assemblage 1. Scale bar = 100 μm for 1–2, 7, 9, 11–14; 112.5 μm for 3–5, 8; 150 μm for 6, 10 (after Bazzucchi at al., 2005; Giordano et al., 2010, modified). 1–2—Betraccium deweveri Pessagno and Blome, sample PR14. 3—Praemerosatunaris gracilis (Kozur and Mostler), sample PR14. 4—Tetraporobrachia sp. aff. T. composita Carter, sample PR14. 5—Aytontius elizabethae Sugiyama, sample PR15. 6—Citriduma sp. A, sensu Carter (1993), sample PR13. 7—Globolaxtorum sp. cf. G. hullae (Yeh and Cheng), sample PR14. 8—Lysemela sp. cf. L. olbia Sugiyama, sample PR15. 9—Livarella valida Yoshida, sample PR15. 10—Livarella sp., sensu Carter (1993), sample PR14. 11—Fontinella primitiva Carter, section sample PA 25. 12—Praemerosatunaris sp. cf. P. sandspitsensis (Blome), sample PA25. 13—Globolaxtorum hullae (Yeh and Cheng), sample PA25. 14—Livarella densisiporata Kozur and Mostler, sample PA25. Conodonts: Scale bar = 75 μm (after Bazzucchi et al., 2005; Giordano et al., 2010, modified): 15 (a, b, c)—Mockina zupfii (Kozur), sample PIG 0. 16 (a, b, c)—Mockina slovakensis (Kozur), sample PIG 0. 17 (a, b, c)—Misikella hersteinii (Mostler), sample PIG 16. 18 (a, b)—Misikella posternsteinii Kozur and Mock, sample PIG 24. 19 (a, b)—Misikella kovacevich Orchard, sample PIG 40. 20 (a, b, c)—Misikella ultima and Mock, sample PIG 40.

GEOCHEMISTRY

In total, 41 samples, mostly black to brown shales, from the upper portion of the Pignola-Abriola section (from meter 30 to the top of the section) were analyzed for δ13Corg (worksheet 1 in GSA Data Repository1). The rock samples were pulverized and acid-washed with 10% HCl in a 70 °C water bath for 3 h, and the process was repeated at least three times to thoroughly remove pyrite and carbonates. The samples were subsequently neutralized with high-purity water, dried at 30 °C overnight, and then wrapped in tin capsules and analyzed for their isotopic composition. The analyses were carried out using a GVI Isoprobe continuous flow/isotope ratio mass spectrometer (CF-IRMS) at Rutgers University, adding multiple blank capsules and isotope standards for each batch of isotopic analyses (NBS 22 = −30.03‰; Coplen et al., 2006) plus a matrix matched in-house standard. Standard deviations for δ13Corg standards during the period of analysis were better than σ = 0.2‰.

The δ13Corg values of the Pignola-Abriola section are between −29.95‰ and −23.70‰ (Fig. 2). After a moderate increase in δ13Corg (from −27.5‰ to −24‰ from meter 30 to 36), a large decrease to −30‰ was recorded for meter 36 to meter 44.5, immediately followed by a rapid return to higher values (~−25‰, ~20 cm above). A subsequent decrease of ~2‰ is recorded at meter 53.5 (close to the level containing the FO of Misikella ultima; Fig. 2). Notably, the low δ13Corg of ~−30‰ at meter 44.5 is just below the level containing the FAD of Misikella posternsteinii sensu stricto, and within the base of the Proparvicingula moniliformis zone (Fig. 2).

PALEOMAGNETISM

In total, 220 oriented core samples were collected from the Pignola-Abriola section and analyzed at the Alpine Laboratory of Paleomagnetism (Peveragno, Italy). Rock magnetic properties were studied on a representative set of samples by means of thermal decay of a three-component isothermal remanent magnetization (IRM) imparted at fields of 2.5 T, 0.4 T, and 0.12 T (Lowrie, 1990) and IRM acquisition curves. The lower part of the section (samples P1.34, P3.10, P3.43; Fig. 3A) is characterized by a high-coercivity mineral with maximum unblocking temperatures (Tob) of 650–675 °C, attributed to hematite, coexisting with a lower-coercivity mineral with Tob of 525–575 °C, interpreted as magnetite; an inflection at ~350 °C in the 0.4 T curve observed in sample P1.34 suggests the presence of iron sulfides. Samples from the upper part of the section (GMM497 at 33 m; GMM48 at 43.5 m; GMM119 at 57 m) appear dominated by the high-coercivity hematite phase (Fig. 3A). IRM curves of these samples show no tendency to saturate even at applied fields of 2.5 T (Fig. 3B). The cumulative log-Gaussian (CLG) analysis (Kruiver et al., 2001) reveals the presence in these samples of two magnetic phases with contrasting coercivities: a high-coercivity phase with coercivity of remanence (Brc) ~1.6–2 T, which accounts...
for ~60%–85% of the IRM, and a subordinate low-coercivity phase with $B_{k2} = 0.1$ T, which accounts for the remainder of the IRM (Fig. 1; GSA Data Repository [see footnote 1]). The presence of higher amounts of (detrital) hematite in the upper part of the section may correlate with the increase in terrigenous input (shales and marls) observed in the upper part of the section (Fig. 2).

The natural remanent magnetization (NRM) of samples, measured on a 2G Enterprises DC-SQUID cryogenic magnetometer, is on average 0.08 mA/m. All samples were thermally demagnetized in steps of 50 °C or 25 °C up to a maximum of 675 °C, and the component structure of the NRM was plotted on vector end-point demagnetization diagrams (Fig. 4; Zijderveld, 1967). After removal of spurious magnetizations between room temperature and ~100–300 °C, a characteristic remanent magnetization (ChRM) was isolated up to 450–550 °C (maximum of 625 °C) in ~55% of the samples ($N = 121$; worksheet 3, GSA Data Repository [see footnote 1]) and found to be broadly oriented either N and down or S and up in tilt-corrected coordinates (Fig. 5). These ChRM component directions are distributed in tilt-corrected coordinates around an overall sequence of five polarity magnetozones, to normal or reverse polarity, respectively. An overall sequence of five polarity magnetozones, labeled from magnetozone MPA1 to MPA5, was established starting at the base of the section (Fig. 2). Each magnetozone was subdivided into a lower, predominantly normal and an upper, predominantly reverse portion, in which submagnetozones can be embedded. No obvious relation was observed between magnetic polarity stratigraphy and the magnetic mineralogy of the samples. The FAD of Misikella posthersteinii sensu stricto falls within magnetozone MPA5 at ~45 m, while the new proposed Norian-Rhaetian boundary coincident with the δ13Corg negative spike occurs inside the same magnetozone at ~44.5 m (Fig. 2).

**DISCUSSION**

**Correlations with Tethyan Sections from the Literature**

The magnetostratigraphy of the Pignola-Abriola section is comparable with that of the Steinbergkogel section (Hüssing et al., 2011), which at present is the only global boundary stratotype section and point (GSSP) candidate

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Figure 3. Thermal demagnetization of a three-component isothermal remanent magnetization (IRM) (A) and IRM acquisition curves (B) for representative samples from Pignola-Abriola showing the presence of a variable mixture of hematite and magnetite. See text for discussion.
for the base of the Rhaetian Stage (Krystyn et al., 2007a, 2007b), assuming that the occurrence of conodont *Misikella posthernsteini* at Steinbergkogel (plate 1 in Krystyn et al., 2007a) is equivalent to the FO of *Misikella hernsteini* / *posthernsteini* transitional forms at Pignola-Abriola (sensu Giordano et al., 2010). Hence, the main reversal portion of the Steinbergkogel magnetostratigraphy from magnetozone ST1/B– to magnetozone ST1/H– at Steinbergkogel STKA section (equivalent to ST2/B– to ST2/H– at Steinbergkogel STKB+C section), has been correlated to magnetozones MPA3r to MPA5r of the Pignola-Abriola section (Fig. 6). Also, part of the magnetostratigraphy of the Oyuklu section (Gallet et al., 2007), from magnetozone OyB– to OyD–, is comparable with MPA4r to MPA5r of the Pignola-Abriola section, and with ST/D– to ST/H– of the Steinbergkogel section (Fig. 6). Furthermore, the lower portion of the Pignola-Abriola section is magnetostatigraphically correlated with the upper part of the Pizzo Mondello section (Muttoni et al., 2004). Using the updated biostratigraphic calibration of the Pizzo Mondello magnetostratigraphy (Mazza et al., 2012).

**CHARACTERISTIC COMPONENT (ChRM)**

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<th>GEOGRAPHIC</th>
<th>TILT CORRECTED</th>
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Figure 4. Vector end-point demagnetization diagrams for representative samples from Pignola-Abriola. Closed circles are projections onto the horizontal plane, and open circles are projections onto the vertical plane for in situ (geographic) coordinates. Temperatures are expressed in °C. NRM—natural remanent magnetization.

Figure 5. Equal-area projections for characteristic remanent magnetization (ChRM) component directions isolated at Pignola-Abriola for in situ (geographic) and tilt-corrected coordinates (see Table 2 for Fisher statistics parameters). For the ChRM component directions, see also worksheet 3, GSA Data Repository (see text footnote 1).
Paleomagnetic pole, paleolatitude, and rotation from tilt-corrected filtered ‘Ch’ directions, corrected for inclination flattening:

22. The Pignola-Abriola sequence was then slid from which a t-test value was derived, where \( t = R \times \sqrt{(N - 2) / (1 - R^2)} \), \( R \) is the linear correlation coefficient, and \( N \) is the number of matching magnetozones in the moving window, i.e., 22. The Pignola-Abriola sequence was then slid by two polarity zones along the Newark APTS (in order to maintain internal polarity consistency in correlation), \( R \) and \( t \) were recalculated, and the exercise was repeated until all 19 possibilities were explored (Fig. 7; statistical procedure with correlation options and analysis of \( t \)-values is reported in worksheet 2 of GSA Data Repository [see footnote 1]).

For \( N = 22 \) (the number of matching reversals in moving window), each correlation has 20 degrees of freedom. A Student \( t \)-test shows that only correlation coefficients with a \( t \) value larger than 1.725 are significant at the 95% level. According to the Student \( t \)-test, only correlation option 19 is reliable at more than 95% confidence level (Fig. 7), though it can be excluded on stratigraphic grounds. Precisely, option 19 places the Norian-Rhaetian boundary of the Pignola-Abriola section in the Carnian-Norian portion of the Newark sequence, as deduced from correlations of the Pizzo Mondello and Silicà Brezová sections to the Newark APTS (Muttoni et al., 2004). In this work, specimens originally attributed to *M. post hern steini* at Steinbergkogel (Krystyn et al., 2007a, 2007b) are here considered *M. post hern steini* sensu lato (s.l.) and attributed to the *M. hernsteini/post hern steini* “transitional forms” (sensu Giordano et al., 2010). Key biostratigraphic events at Pizzo Mondello are after Mazza et al. (2012). The Pignola-Abriola section is correlated to the Newark APTS and will be investigated in detail. Option 7 links Pignola-Abriola magnetozone MPA1n with Newark E13n.In at the base, and magnetozone MPA11r with E20r at the top (Fig. 7). However, this correlation implies sudden variations in sediment accumulation rates in the middle of the Pignola-Abriola section. Moreover, the lower part of Pignola-Abriola is considered Sevatin (late Norian) in age, but according to correlation 7, it should correspond to Newark magnetozones MPA1n to MPA3n at Pignola-Abriola have been correlated to magnetozones PM-8a to PM-12n at Pizzo Mondello (Fig. 6). Moreover, data from Pignola-Abriola have been compared with the magneto-biostratigraphy of the Brumano and Italcementi Quarry sections (Lombardian Basin, southern Alps, Italy), which encompasses a portion of the Rhaetian (with specimens attributed to *Misikella* up to the Triassic-Jurassic boundary as defined by pollenists (Muttoni et al., 2010, 2014). Awaiting for a formal redefinition of the *Misikella* specimens in the Brumano section following the new definition of *Misikella post hern steini* sensu stricto adopted in this study (after Giordano et al., 2010), we stress that all *Misikella* specimens at Brumano occur below the recovered magnetostratigraphy (Muttoni et al., 2010, 2014), and thus the sequence of Brumano-Italcementi Quarry magnetozones from BIT1n to BIT5n is regarded as largely younger than the Pignola-Abriola magnetostratigraphy (Fig. 6).

**Correlation with the Newark APTS**

The correlation between the Pignola-Abriola section and the Newark APTS was performed using the statistical approach proposed in Muttoni et al. (2004). Assuming that thickness is a linear proxy of time, the duration of Newark magnetozones was compared with the thickness of Pignola-Abriola magnetozones (Fig. 7). The Pignola-Abriola polarity reversal sequence in linear depth coordinates was placed alongside the top of the Newark APTS (at magnetozone E23r) in linear age coordinates. A linear correlation coefficient (\( R \)) relating the thickness of each of the \( N = 22 \) complete Pignola-Abriola magnetozones to the duration of the correlative Newark magnetozones was calculated, from which a \( t \) value was derived, where \( t = R \times \sqrt{(N - 2) / (1 - R^2)} \), \( R \) is the linear correlation coefficient, and \( N \) is the number of matching magnetozones in the moving window, i.e., 22. The Pignola-Abriola sequence was then slid by two polarity zones along the Newark APTS (in order to maintain internal polarity consistency in correlation), \( R \) and \( t \) were recalculated, and the exercise was repeated until all 19 possibilities were explored (Fig. 7; statistical procedure with correlation options and analysis of \( t \)-values is reported in worksheet 2 of GSA Data Repository [see footnote 1]).

For \( N = 22 \) (the number of matching reversals in moving window), each correlation has 20 degrees of freedom. A Student \( t \)-test shows that only correlation coefficients with a \( t \) value larger than 1.725 are significant at the 95% level. According to the Student \( t \)-test, only correlation option 19 is reliable at more than 95% confidence level (Fig. 7), though it can be excluded on stratigraphic grounds. Precisely, option 19 places the Norian-Rhaetian boundary of the Pignola-Abriola section in the Carnian-Norian portion of the Newark sequence, as deduced from correlations of the Pizzo Mondello and Silicà Brezová sections to the Newark APTS (Muttoni et al., 2004; Channell et al., 2003). For this reason, we decided to contemplate correlation options characterized by lower values of \( t \) (around 1). As a consequence, options 16 and 7 were considered as acceptable (Fig. 7). Option 16 is affected by the same problem as option 19 insofar as it places the Norian-Rhaetian boundary within the Norian Stage as implied by the Pizzo Mondello and Silicà Brezová to Newark correlations discussed earlier. In addition, option 16 also implies sudden (and unexplained) variations in sediment accumulation rates of the Pignola-Abriola section.

Correlation option 7 results are more coherent with the available magneto-biostratigraphic correlations of Tethyan sections to the Newark APTS and will be investigated in detail. Option 7 links Pignola-Abriola magnetozone MPA1n with Newark E13n.In at the base, and magnetozone MPA11r with E20r at the top (Fig. 7). However, this correlation implies sudden variations in sediment accumulation rates in the middle of the Pignola-Abriola section. Moreover, the lower part of Pignola-Abriola is considered Sevatin (late Norian) in age, but according to correlation 7, it should correspond to Newark...
magnetozones considered close to the Carnian-Norian boundary (Fig. 7; see also Muttoni et al., 2004).

In conclusion, no statistical correlation matches perfectly, and some “adjustments” are necessary. An alternative version of statistical correlation option 7, termed option 7.1 (black bar in Fig. 7), solves the problems outlined for option 7, increases statistical significance, and is coherent with the Pizzo Mondello and Pignola-Abriola magneto-biostratigraphies and correlations to the Newark APTS. Preferred correlation option 7.1 is similar to statistical option 7 in range (MPA1n corresponding to E13n.1n, and MPA1r (0.1r, 0.1n, 0.2r) with E13r, MPA2n (0.1n, 0.1r, 0.2n) with E14n, and MPA11r with E20r, whereas magnetozones from MPA2r to MPA11n have been correlated with Newark magnetozones from E14r to E20n. Preferred correlation option 7.1 implies a correlation of the Steinbergkogel section to the Newark APTS from magnetozone E17n to E21n (Fig. 6) that is substantially equivalent to the correlation originally proposed by Hüsing et al. (2011). The correlation of the largely younger Brumano-Italcementi Quarry sections to the Newark APTS is the same of Muttoni et al. (2010, 2014), pending a formal redefinition of the Misikella specimens at Brumano (see also earlier discussion).

Using preferred correlation option 7.1, an age model for the Pignola-Abriola section can be derived. The age model shows a change in sedimentation rate from the lower to the upper part of the section (Fig. 6). From the base to meter 24.5, the mean sedimentation rate is of ~2.6 m/m.y., while from meter 24.5 to 40, the mean sedimentation rate increases to ~5.6 m/m.y. From meter 40 to the section top, the sedimentation rate increases further to ~9.8 m/m.y. This is coherent with the lithostratigraphy of the section, suggesting a general increase of terrigenous input in the upper part of the section. According to the proposed age model, the Norian-Rhaetian boundary defined by the level containing the FAD of M. post hern steini sensu stricto at meter 45 should correspond to an estimated age of ca. 205.7 Ma (Fig. 6), which is substantially equivalent to the age of the prominent negative δ13Corg excursion to ~–30‰ observed at meter 44.5 (Fig. 6).

GSSP Proposal for the Base of the Rhaetian Stage

Based on our magneto-bio-chemostratigraphic study of the Pignola-Abriola section, coupled with the recognition of the taxonomic complexities concerning conodont Misikella post hern steini, the current candidate species for...
the definition of the base of the Rhaetian Stage, we suggest an alternative option for the definition of the Norian-Rhaetian boundary. We favor placing the boundary at the prominent negative δ¹³Corg spike observed in the Pignola-Abrilia section at meter 44.5 (immediately below the level containing the FAD of M. posthornsteini sensu stricto and within the base of the radiolarian Proparvicingula moniliformis zone). A similar δ¹³Corg perturbation around the Norian-Rhaetian boundary was documented in Canada by Ward et al. (2001, 2004) and Whiteside and Ward (2011), coinciding with the disappearance of large Monotis (Ward et al., 2004), a typical proxy for the Norian-Rhaetian boundary (McRoberts et al., 2008). The stratigraphic level in the Pignola-Abrilia section containing the −30‰ spike has been magnetostratigraphically correlated to Newark magnetzone E2/2r at ca. 205.7 Ma. This age was obtained from the Newark astrochronology, calibrated with the new numerical age of 201.5 Ma from the base of the Orange Mountain Basalts in the Newark Supergroup (Blackburn et al., 2013). Assuming an age of ca. 201.3 Ma for the Triassic-Jurassic boundary (Gue et al., 2012), which is broadly consistent with previous estimates (Schoene et al., 2010), and a proposed age of ca. 205.7 Ma for the Norian-Rhaetian boundary, the Rhaetian Stage would have a duration of ~4.4 m.y. (Fig. 6). Using a Carnian-Norian boundary at ca. 227 Ma (Muttoni et al., 2004), the Norian would be the longest stage of the Phanerozoic with a duration of ~21.3 m.y. (but see Lucas et al., 2012). Using an approximated Ladinian-Carnian age of 238 Ma, derived from an uppermost Ladinian radiometric age of 237.77 ± 0.14 Ma (Mietto et al., 2012), the Carnian would have lasted almost 10 m.y. According to these figures, the Late Triassic may have lasted ~36 m.y.

Comparison with Previous Time Scales

We compared our solution with alternative proposals from the literature. Krystyn et al. (2002) used Carnian-Norian data from several Tethyan sections (Kavaalani, Kavur Tepe, Pizzo Mondello lower part, Bolücektsü Tepe, and Scheiblkogel; see references in Krystyn et al., 2002) to construct a Tethyan composite magneto-biostратigraphic sequence that was correlated to Newark magnetozones E3–E22 and used it to infer a duration of the Rhaetian of only ~2 m.y. Later, Gallet et al. (2007) correlated data from Oyuklu, Pizzo Mondello (upper part), and the Tethyan composite sequence of Gallet et al. (2003) to the Newark APTS, suggesting that part of the Rhaetian is missing in the Newark sequence, and supporting the ~2 m.y. duration of the Rhaetian as proposed by Krystyn et al. (2002). Muttoni et al. (2010) illustrated that middle Norian (Aluanian) magnetozones in the composite magneto-biostратigraphic sequence of Krystyn et al. (2002) may encompass Newark magnetozones –E13–E15 rather than –E13–E17, so that the overlying Sevitan magnetozones may correlate to Newark levels at and immediately above E15 rather than at and above E17 as proposed by Krystyn et al. (2002), thus supporting the existence of a longer (>2 m.y.) Rhaetian.

Coming to more recent times, the long-Tuvalian option of the Geological Time Scale 2012 (Ogg, 2012), which is essentially based on data from Lucas et al. (2012), is characterized by a Carnian-Norian boundary placed at 221 Ma, a Norian-Rhaetian boundary at 205.4 Ma, and a large hiatus in the Newark Supergroup based on inferences from conchostracan biostratigraphy (Lucas et al., 2012, and references therein). According to this option, the preserved portion of the Rhaetian in the Newark Supergroup should have a duration of only ~0.2 m.y. (Lucas et al., 2012). A duration of ~8 m.y. for the Rhaetian, as proposed using marine-Newark magnetostratigraphic correlations by several authors (Channell et al., 2003; Muttoni et al., 2004, 2010; Hüsing et al., 2011), was rejected by Lucas et al. (2012) based on the inference that inserting 7.8 m.y. of missing Rhaetian in the claimed Rhaetian gap of the Newark Supergroup (7.8 m.y. = 8 m.y. of total duration of Rhaetian – 0.2 m.y. of preserved Rhaetian in the Newark) would produce an age for the base of the Newark Supergroup of 240.5 Ma; as Lucas et al. (2012) considered the base of the Newark Supergroup to coincide with the base of the Carnian (based on continental [palynomorphs, conchostracans, tetrapods] biostratigraphy), an age of 240.5 Ma is regarded as inappropriate because it would place the base of the Newark Supergroup close to the age of the Anisian-Ladinian boundary (Mundil et al., 2010). Therefore, a duration of ~8 m.y. for the Rhaetian is considered unacceptable according to Lucas et al. (2012), who instead adopted a duration of ~4 m.y. from Ogg (2004). Under the assumption of a 4 m.y. duration for the Rhaetian and 0.2 m.y. of Rhaetian time preserved in the Newark Supergroup, Lucas et al. (2012) (and Ogg [2012] in his long-Tuvalian option) estimated an age of 221.5 Ma for the Carnian-Norian boundary, based on continental biostratigraphy, by counting ~405 k.y. McLaughlin cycles of Newark astrochronology.

In our opinion, the Rhaetian gap of Lucas et al. (2012) at the basis of the long-Tuvalian option (Ogg, 2012) is flawed by lack of convincing correlations between terrestrial groups and marine-based stage boundaries. For example, conchostracans from the Weser Formation of the Germanic Basin are assigned an early Tuvalian age (late Carnian) because the Weser Formation is considered correlative with the Dolomie de Beaumont of France, which contains marine bivalves considered to be of such age (Lucas et al., 2012). As a further example, the conchostracan fauna from the Coburg Sandstein of the Germanic Basin is considered late Carnian, seemingly because it lies immediately below the beginning of a sporomorph association considered to be late Tuvalian. In general, we find difficult to decipher in Lucas et al. (2012) where and in which stratigraphic context a given continental association was found in direct association with stage-defining marine fossils.

The long-Rhaetian option of the Geological Time Scale 2012 (Ogg, 2012) is essentially based on magnetostratigraphic correlations between marine sections bearing stage-defining fossils and the Newark APTS assumed to be continuous in the Rhaetian (Channell et al., 2003; Muttoni et al., 2004, 2010; Hüsing et al., 2011), and it shows a Carnian-Norian boundary at ca. 228.4 Ma and a Norian-Rhaetian boundary at ca. 209.5 Ma. Our new time scale for the Late Triassic could be considered an “update” of the long-Rhaetian option of (Ogg, 2012), with a Norian-Rhaetian boundary at 205.7 Ma based on data from this study and a Carnian-Norian boundary at ca. 227 Ma based on correlation of the Pizzo Mondello section with the Newark APTS (both numerical estimates obtained by rescaling the Newark APTS using an age of ca. 201.5 Ma for the base of the Orange Mountain Basalts in the Newark Supergroup; Blackburn et al., 2013). Moreover, our age of 205.7 Ma for the Norian-Rhaetian boundary is coherent with recent U/Pb ages of Wotzlaw et al. (2014) that constrain the Rhaetian base between 205.70 ± 0.15 Ma and 205.3 ± 0.14 Ma.

**CONCLUSIONS**

Paleomagnetic data obtained from the Pignola-Abrilia section provided a sequence of 22 polarity reversals grouped in 10 magnetozones. The correlation between the Pignola-Abrilia section and additional Norian and Rhaetian Tethyan marine sections from the literature (Steinbergkogel, Oyuklu, Brumano, Halcementi Quarry, and Pizzo Mondello) reveals significant internal consistency.

To provide numerical age constraints on the Pignola-Abrilia section, we applied a statistical correlation to the Newark APTS, which is assumed to be continuous in its younger part (contra Lucas et al., 2012). Three out of a total of 19 explored correlation options produced statistically reliable results, and after a thorough analysis, one option (7.1) is considered as the
most reliable. According to this option, the Pignola-Abriola section correlates to Newark magmatozonal E13n to E20r. We place the Norian-Rhaetian boundary at Pignola-Abriola at a level coincident with a rapid decrease in δ13Corg to ~−30‰, which virtually coincides with the FAD of conodont Mistikella posthersteinii sensu stricto within the Proparviviliga multiformis radiolarian zone. This level is traced within Newark magmatozone E20r at 205.7 Ma.

Assuming an age of ca. 201.3 Ma for the Triassic-Jurassic boundary (Schoenew, et al., 2010; Guex et al., 2012), our study shows that the Rhaetian is ~4.4 m.y. long. Assuming a Carnian-Norian boundary of ca. 227 Ma (Muttoni et al., 2004, 2014, and references therein), our study shows that the Norian is ~21.3 m.y. long.

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IMPROVING THE GEOMAGNETIC POLARITY TIME SCALE FOR THE LATE TRIASSIC: NEW MAGNETO-BIOSTRATIGRAPHIC CONSTRAINTS FROM PIGNOLA-2 AND DIBONA MARINE SECTIONS, ITALY

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Abstract
To contribute to the calibration of the Late Triassic time scale, two sections in Italy were investigated for magnetostratigraphy: the Pignola-2 (Southern Apennines) and the Dibona sections (Dolomites). These sections reveal a sequence of biostratigraphically (conodonts and palynomorphs) calibrated magnetic polarity reversals encompassing the Julian/Tuvalian boundary (Carnian). A total of 63 samples have been collected from the Pignola-2 section that helped defining 12 magnetozones. From the Dibona section, 81 samples have been collected, revealing 9 magnetozones. These data are furthermore constrained by a published radiometric U/Pb age of 230.91±0.33 Ma from the Aglianico ash-bed in the Pignola-2 section. Correlations of the Pignola-2 and the Dibona stratigraphic successions with other Carnian sections from the literature were used to further define the magnetostratigraphy around the Julian/Tuvalian boundary in the Tethys realm. A statistical correlation between the Pignola-2 and the Newark Astrochronological Polarity Time Scale (APTS) provides new constraints on the chronology of the Carnian, also suggesting a duration for the Carnian Pluvial Event (CPE) of ~1 Myr, between ~229.7 and 230.7 Ma.

1. Introduction

Substantial progress has been made in the recent years on the chronology of the Late Triassic, particularly on the duration of the Carnian, Norian and Rhaetian stages, by means of magnetostratigraphic correlations of marine fossil-bearing Tethyan sections and the Newark Astrochronological Polarity Time Scale (APTS) (e.g. Kent and Olsen 1999, Olsen et al. 2015). A first step forward came from the correlation between the Pizzo Mondello section (Southern Italy) and the Newark APTS, which helped assigning a numerical age of ~227 Ma to the level hosting the Carnian/Norian boundary based on conodonts (first appearance datum of *Carnepigondolella gulloae*; Mazza et al., 2012a) close to a positive δ¹³C excursion of ~1‰ (Muttoni et al. 2004, Mazza et al. 2010). Recently, a similar
correlation exercise applied to the Pignola-Abriola section of Southern Italy assigned an
age of 205.7 Ma to the level containing the Norian/Rhaetian boundary based on conodonts
(first appearance of *Misikella posthernsteini*), which was found to fall close to a negative
$\delta^{13}C_{org}$ spike of $\sim -30$ ‰ (Maron et al. 2015, Rigo et al. 2015). This latter numerical age
estimate for the N/R boundary, derived from correlation with the Newark APTS, is
compatible with radiometric (U/Pb) age estimates of 205.3±0.14 Ma, 205.4±0.09 Ma, and
205.7±0.15 Ma obtained from a level in the Levanto section of Peru close to the N/R
boundary approximated by the last occurrence of the bivalve *Monotis subcircularis*
(Wotzlaw et al. 2014).

In this paper, we focus on the Carnian interval by providing new magnetostratigraphic and
biostratigraphic data from the Pignola-2 section of the Southern Apennines and the Dibona
section of the Dolomites (both in Italy), which have been correlated to the Newark APTS
in order to provide an independent control on the astrochronological ages of the older
(Carnian) part of the Newark APTS (Kent and Olsen 1999, Olsen and Kent 1999).

Furthermore, we provide a numerical age estimation of a major event occurring in the
Carnian, known as the Carnian Pluvial Event (CPE; Simms and Ruffell 1989). The CPE is
represented by a widespread deposition of siliciclastic materials recognized in most of the
Carnian sections around the world (e.g., Ruffell et al. 2015). The CPE is attributed to a
climatic shift to more humid conditions (Simms and Ruffell 1989), triggered by the
emplacement of the Large Igneous Province (LIP) of Wrangellia in North America (e.g.,
and consequent emission of greenhouse gasses in the atmosphere. The CPE is well
expressed by the most siliciclastic intervals in both the Pignola-2 and Dibona sections.

2. Geological Setting
2.1 Pignola-2

The Pignola-2 section (Lat: 40°32'51.44"N, Long: 15°47'17.43"E) crops out in the Southern Apennines, south of the town of Potenza, along the road connecting the two Pignola and Abriola villages (Fig. 1). The section is comprised of a 40 m-thick succession of cherty limestones pertaining to the Calcari con Selce Formation Fm. (Scandone 1967, Miconnet 1983, Amodeo 1999, Rigo et al. 2012), encompassing the Julian/Tuvalian substage boundary (Carnian; Rigo et al. 2007, 2012). The section includes a ~5 m-thick (from meter 8 to 13) green shale and radiolaritic interval (the “green clay-radiolaritic horizon” of Rigo et al. 2007), representing the first documentation of the CPE in Tethyan basinal successions (Rigo et al. 2007, 2012a). In the upper part of the “green clay-radiolaritic horizon” (hereafter “green horizon”), a tuff level, named “Aglianico ash-bed” (meter 12) provided a U/Pb radiometric age of 230.91±0.33 Ma (Furin et al. 2006) (Fig. 2). The “green horizon” has been interpreted as resulting from a transient rise of $pCO_2$ levels that triggered the shoaling of the calcite compensation depth (CCD). This inferred CCD shoaling is possibly coupled with increased detrital and nutrient input in the basin as a consequence of the CPE, a warm and humid period that fostered silicate weathering and runoff on land (Rigo et al. 2007, 2012b, Rigo and Joachimski 2010, Trotter et al. 2015). A distinct rise of $pCO_2$ in the coupled ocean-atmosphere system may have been provided by the emplacement of the Wrangellia LIP (e.g., Furin et al. 2006, Rigo et al. 2007, Preto et al. 2010, Dal Corso et al. 2012, Xu et al. 2014), radiometrically dated with Ar/Ar between ~233 and ~222 Ma, with the most likely age comprised between ~230 and ~225 Ma (Greene et al. 2010). The oldest radiometric ($^{207}$Pb/$^{206}$Pb) age available for Wrangellia comes from gabbros in Yukon, associated to the Wrangellian effusions and dated at 232.2...
±1 Ma (Mortensen and Hulbert 1992; see also Greene et al. 2010). This age interval includes the age of the CPE at Pignola-2 from the “Aglianico ash-bed” (Furin et al. 2006).

2.2 Dibona

The Dibona section (Lat: 46°32'2.50"N, Long: 12°04'21.68"E) is a ~370 m thick shallow-water sedimentary succession located in the Dolomites (Southern Alps), on the southern side of the Tofana di Rozes Mountain, near the Dibona Hut (Fig. 1). The section is characterized by mixed carbonate-siliciclastic deposits of shallow-marine (Heiligkreuz/Santa Croce Formation) to marginal-marine (Travenanzes Formation) environments (e.g. Breda et al. 2009). The ~160 m Heiligkreuz Fm. (De Zanche et al. 1993, Preto and Hinnov 2003, Neri et al. 2007, Gattolin et al. 2013, 2015) is subdivided into three members, which from the base to the top are subsequently the Borca Mb., ~100 m-thick, consisting of limestones and arenites passing to dolostones; the Dibona Sandstones Mb., ~60 m-thick, consisting of arenites, conglomerates, pelites and limestones; the Lagazuoi Mb., ~30 m-thick, consisting mainly of strongly dolomitized oolitic limestones (Fig. 3). The shales and arenites of the Borca and Dibona Sandstones Mbs record the CPE in a coastal environment (Breda et al. 2009, Preto et al. 2010). A major negative $\delta^{13}$C spike linked to the eruption of Wrangellia flood basalts has been observed at the base of the Heiligkreuz Fm., close to Dibona section, confirming the connection of the clastic input to the CPE climatic event (Dal Corso et al. 2012). Above the Lagazuoi Mb., the ~180 m Travenanzes Fm. (De Zanche et al. 1993, Neri et al. 2007, Breda and Preto 2011) starts with ~25 m of dark clays and aphanitic dolostones passing upwards to multicolored clays with carbonatic and evaporitic intercalations deposited in sabkha-like environments. The top of the Travenanzes Fm. is dominated by dolomitic peritidal cycles of carbonate tidal-
flat and shallow lagoon environments, with thin dark clay intercalations. It represents the transition to the overlying Dolomia Principale carbonate platform (Breda and Preto 2011).

3. Biostratigraphy

The Pignola-2 section has a detailed conodont and palynomorph biostratigraphy (Rigo et al. 2007, 2012). According to the conodont biostratigraphy, the Julian/Tuvalian (middle/late Carnian) boundary is placed at the base of the “green horizon”. In fact, below the “green horizon” a typical Julian conodont association composed of Paragondolella praelindae, P. polygnathiformis, and Gladigondolella spp is present. Above the “green horizon”, the section bears Tuvalian conodont species, i.e. Carnepigondolella nodosa, C. carpathica, Paragondolella noah, P. oertlii, and Metapolygnathus praecommunisit (Fig. 2) (Rigo et al. 2012). Specifically, the Julian/Tuvalian boundary is placed at the level with the last occurrence (LO) of the Gladigondolella genus (Rigo et al. 2007). Palynomorphs from Pignola-2 have been grouped in two main assemblages (Rigo et al. 2007): Assemblage A is typical of the Julian/Tuvalian interval, while Assemblage B covers a narrower range in the upper Tuvalian (see Rigo et al. 2007 for additional details) (Fig. 2).

The Dibona section has a detailed pollen and spore biostratigraphy (Roghi 2004, Roghi et al. 2010). The typical uppermost Julian-lower Tuvalian association with Patinasporites densus, Aulisporites astigmosus and Duplicisporites continuus (Borca Mb, Dibona Sandstone Mb) and Equisetosporites chinleanus (Dibona Sandstones Mb) is found in the Heilgkreuz Fm. It is followed by a Tuvalian association of Granuloperculatipollis rudis and Riccisporites cf. R. tuberculatus, found at the base of the overlying Travenanzes Fm (Fig. 3). The former association, belonging to the Granuloperculatipollis rudis Assemblage of Roghi et al. (2006, 2010), is similar to Assemblage B found in the Pignola-2 section. Moreover, additional sections coeval to the Dibona section reveal pollens and spores
comparable with the biostratigraphic record of Pignola-2, e.g. the Cave del Predil section in the Southern Alps of Friuli (Roghi 2004), and the Lunz (Köppen 1997) and Rubland (Kraus 1969) sections in the Northern Calcareous Alps of Austria. Six samples for conodont analysis have been collected from the Dibona section immediately below the base of the Lagazuoi Mb, in the last 10 meters of the Dibona Sandstones Mb. The conodont association consists of *Paragondolella polygnathiformis*, *Paragondolella noah*, transitional forms from *P. noah* to *Metapolygnathus praecommunisti*, and early representatives of *M. praecommunisti* (Fig. 3, Pl. 1) attributed to the early Tuvalian age (Mazza et al. 2010, Mazza et al. 2011). Furthermore, the ammonoid *Shastites cf. pilari* has been found below the Lagazuoi Mb, in the nodular limestone corresponding to the upper portion of the Dibona Sandstone Mb of the Heiligkreuz Fm. (Gianolla et al. 1998, De Zanche et al. 2000, Gattolin et al. 2015).

4. Paleomagnetism

4.1 Sampling and laboratory methods

A total of 63 oriented paleomagnetic core samples (~10cc) have been collected from the Pignola-2 section, 55 from limestones beds and 8 from the radiolaritic intervals within the “green horizon”, with a stratigraphic interval of approximately 0.5 m (Fig. 2). The clayey intervals of the “green horizon” have not been sampled because they are both too thin and chipped. From the Dibona section a total of 45 cores have been collected from the upper Borca Mb. to the base of the Lagazuoi Mb. (Heiligkreuz Fm.), and 36 samples from the Travenanzes Fm. To isolate the ChRM, all samples have been thermally demagnetized (with an ASC TD48 furnace, residual field < 10 nT) and measured with a 2G Enterprises DC-SQUID magnetometer (magnetic moment noise level <10^{-12} Am^2) at the Alpine Laboratory of Paleomagnetism – ALP (Peveragno, Italy). Samples have been demagnetized
by steps of 50°C from 100°C to 350°C, then 25°C until 675°C. Single sample directions of
the magnetization vectors have been plotted on an end-point vector graph (Zijderveld 1967)
for each step of demagnetization (Fig. 4). Samples showing magnetization components
made of less than three end-points in sequence (representing three subsequent temperature
steps) have been rejected.

The low-field magnetic susceptibility ($\kappa$) was measured with a AGICO Kappabridge KLY-
3 instrument (sensitivity: $2 \times 10^{-8}$ SI; at the ALP, Peveragno) for all samples. Further, to
support the paleomagnetic interpretation, thermomagnetic runs were performed on a
modified horizontal translation Curie balance (paleomagnetic laboratory ‘Fort Hoofddijk’,
Utrecht University, The Netherlands; noise level $5 \times 10^{-9}$ Am², typical signals at least an
order of magnitude higher; Mullender et al. 1993) for a subset of the samples. About 70-80
mg of powdered sample was measured in several cycles to increasingly higher temperature
up to 670°C; the field was cycled between 100 and 300 mT, heating and cooling rates were
10°C/minute. Measurements were performed in air. Three samples from the Pignola-2
section were investigated and 6 from the Dibona section (3 from the Heiligkreuz Fm. and
3 from the Travenanzes Fm.).

Samples PGM0.30, RAD4 and PGM14.64 from the Pignola-2 section and samples
MDS12.4, MDS29.1 and MDS52.3 from the Dibona section have been analyzed using the
Curie balance.

4.2 Magnetic properties

The limestones of the Pignola-2 section reveal a very low $\kappa$, usually smaller than $5 \times 10^{-6}$ SI
(Fig. 2). In the “green horizon” the initial magnetic susceptibility is considerably higher
(from ~$70 \times 10^{-6}$ to ~$110 \times 10^{-6}$ SI) than in the rest of the sampled interval, due to an increase
of the terrigenous fraction (Fig. 3). Interestingly, magnetic susceptibility is high (from
25×10^{-6} to 55×10^{-6} SI) also in the carbonatic strata delimiting the “green horizon” (~1 or 2 meters above and below), indicating a significant siliciclastic component we associate to the CPE.

The Curie balance results are shown in Fig. 5 and the stratigraphic position of the sample is in Fig. 3. The limestones of the Pignola-2 section (PGM0.30 and 14.64; Fig. 5A, 5C) are very weak, only slightly above instrumental noise level. Nonetheless there seems to be a marginally convex magnetization vs. temperature behavior between ~100-200 and ~450-500°C, that is reversible on intermittent cooling. The final cooling segment from 600°C back to room temperature, however, does not reveal that behavior. It is difficult to interpret this behavior that may be associated with titanomagnetite (sensu lato) with a varying Ti-content. However, a minute amount of magnetic sulfides cannot be excluded with certainty. The “green horizon” sample RAD4 (Fig. 5B) is much stronger (but still weak, it remains a sediment) and shows a Curie point (determined by the two-tangent method, Grommé et al. 1969) of ~350°C that is reversible on cooling after the final heating temperature of 600°C.

The Dibona Sandstones Mb samples (MDS12.4, Fig. 5D; MDS29.1, Fig. 5E; MDS52.3, Fig. 5F; stratigraphic position in Fig. 3) from the Dibona section all show a variable portion of non-magnetic pyrite (FeS$_2$) that is oxidized during the thermomagnetic analysis, first to magnetite and finally to hematite, explaining the occasionally huge increase in magnetization between 400 and 600°C. There are no indications for magnetic sulfides below 400°C since the analysis shows reversible heating and cooling segments in that temperature range and no Curie temperature of ~320°C. Plausibly traces of magnetite represent the original magnetic mineralogy but it is impossible to discriminate between left overs of neo-formed magnetite (most of it oxidizes further to hematite) and original magnetite. The three samples from the Travenanzes Fm (MTV9, Fig. 5G; MTV52, Fig. 5H; MTV67, Fig. 5I; stratigraphic position in Fig. 3) are all very weak demonstrating
paramagnetic behavior only. During the heating above 600°C a minute amount of magnetic minerals (presumably fine-grained magnetite) is formed because the final cooling curves lie slightly above the corresponding heating curves.

4.3 Magnetostratigraphy

4.3.1 Pignola-2

The mean intensity of the starting NRM is ~0.02 mA/m in the pelagic carbonates, ~0.06 mA/m in the radiolarites of the “green horizon”, and ~0.2 mA/m in the carbonatic levels just above the “green horizon”. Vector end-point demagnetization diagrams (Fig. 4A; Zijderveld 1967) reveal the presence of spurious (viscous) magnetic components from room temperature to 250-300°C; at higher temperatures the characteristic component remanent magnetization (ChRM) direction is isolated (Fig. 4A). The demagnetization trajectory trends toward the origin up to a maximum temperature of 675°C. This behavior is observed in 47 samples. Equal area stereographic projections reveal that the ChRM is bipolar being oriented north-and-down or south-and-up in in situ coordinates, and northwest-and-down or southeast-and-up after correction for bedding tilt (Fig. 6). The mean direction in tilt-corrected coordinates, calculated with standard Fisher statistics, is of Dec: 28.4°E; Inc: 39.6° (k=23.9; α95=4.3°; N = 47; Table 2). No fold test could be performed because the bedding attitude trough the section is essentially the same. The reversals test (McFadden and McElhinny 1990) is positive, suggesting that the ChRM is the original magnetization acquired during or shortly after deposition. The mean directions in in situ coordinates (Dec: 353.5°E; Inc: 59.7°; k: 24.1; α95: 4.3°) are similar to the inclination of the geomagnetic axial dipole (GAD Inc: ~59.8°), so we cannot exclude some contamination of the ChRM by VRM for normal polarity components. The latitudes of the
Virtual Geomagnetic Poles, derived from the ChRM directions, provided a sequence of 12 magnetic polarity reversals defining 12 magnetozones labeled from MP1n to MP6r (Fig. 2). The shales of the “green horizon” were not sampled for magnetostratigraphy (see above).

4.3.2 Dibona

The mean intensity of the samples from the Heiligkreuz Fm. is ~0.04 mA/m and in the Travenanzes Fm. is ~0.05 mA/m. The vector end-point diagrams (Fig. 4B; Zijderveld 1967) reveal the ChRM between 200°C and 550°C in 25 of 45 samples of the Heiligkreuz Fm. (named MDS), and between 150°C and 400°C in 18 of 36 samples from the Travenanzes Fm (named MTV). The diagrams show both north-down and south-up directions, sometimes scattered (Fig. 4B) (typical MAD: ~11). The equal-area stereographic projection reveals fairly scattered directions (Fig. 6), failing the reversals test (McFadden and McElhinny 1990). As for the Pignola-2 section, also here fold test cannot be performed due to homoclinality of the succession. The sequence of VGPs of the Dibona section shows nine magnetozones labeled MD, where MD1r, 2r and 4r are rather uncertain (grey intervals, Fig. 3), due to the poor preservation of NRM in these intervals (only three robust paleomagnetic directions).

5. Discussion

5.1 Correlations between Tethyan sections

The Pignola-2 section is correlated with other coeval Tethyan sections from the literature containing conodonts to obtain a complete magneto-biostratigraphic record for the Carnian
Stage (Fig. 7). The upper part of the Pignola-2 magnetostratigraphy (magnetozones MP2n to MP3r) is considered correlative to the basal portion of the Silická Brezová section (up to SB2r) (Channel et al. 2003) and the Pizzo Mondello section (up to PM2r) (Muttoni et al. 2004), whereas the entire Pignola-2 section is comparable with the Guri Zi section in Albania (up to GZ5r) (Muttoni et al. 2005, 2014) (Fig. 7). The conodont biostratigraphy of Silická Brezová has been updated in this study by reclassifying the taxa illustrated in Figs. A1-A3 in Channell et al. (2003), using the new taxonomic criteria illustrated in Mazza et al. (2010, 2011, 2012a,b) (Fig. 7).

The magnetostratigraphy of the Dibona section straddling the Dibona Sandstones Mb. of the Heiligkreuz Fm. should be partially coeval with the magnetostratigraphy across the “green horizon” in the Pignola-2 section, as suggested by the first occurrence of conodont *Metapolygnathus praecommunisti* (Figs. 3, 7). Consequently, magnetozones MD1n-1r-2n-2r-3n at Dibona have been correlated to Pignola-2 magnetozones MP4n-4r-5n-5r-6n, respectively. Based on the first occurrence of *M. praecommunisti*, magnetozone MD3n is considered coeval to magnetozone SB2n at Silická Brezová and PM2n at Pizzo Mondello (Fig. 7).

The correlation between Pignola-2 and Dibona sections implies that the onset of the CPE at Dibona should fall in the lower part of the Heiligkreuz Fm. (basal Borca Mb, as suggested by Dal Corso et al., 2012) and its acme is reasonably represented by the terrigenous-rich levels of the Dibona Sandstone Mb (Fig. 3). The absence of strong biostratigraphic constraints does not allow a solid magnetostratigraphic correlation between the upper Dibona section and other Tethyan sections.

5.2 Correlation with the Newark APTS
The Pignola-2 magnetostratigraphy has been compared with the Newark APTS (Kent and Olsen 1999; see also Olsen et al. 2015) and then been exploited also in Olsen et al. (2015) using the statistical approach described in Muttoni et al. (2004) and Maron et al. (2015).

The radiometric age of 230.91±0.33 Ma from the Aglianico ash-bed (Furin et al. 2006), comprised within the Pignola-2 magnetozone MP4r, has been taken into account for the correlation. The Dibona section was not considered for statistical correlation with the Newark APTS because of the unreliability of its magnetostratigraphy, due the variable sedimentation rate typical of shallow-water environments.

We compared the thickness of the Pignola-2 magnetozones with the duration of the magnetozones in the Newark APTS, testing the magnetostratigraphy of Pignola-2 along the APTS and obtaining 24 possible correlation options (Fig. 8). The interval of unknown polarity within the Pignola-2 “green horizon” is tentatively interpreted as dominated by normal polarity.

Each correlation is analyzed using linear regression, obtaining 24 t-test values; the higher the t-value, the more reliable the correlation. Only options 1, 2 and 24 pass the 95% confidence level threshold (Fig. 8). Option 24 is not considered because it is inconsistent with the U/Pb radiometric age of 230.91±0.33 Ma from the “green horizon” (Furin et al. 2006). Option 1 and 2 are consistent with this age. The main features of Option 1 and 2 are as follows:

Option 1:

- High t-value (~2.7).
- The radiometric age of Pignola-2 fits more closely (0.6 M.y. older) with the age provided by the Newark APTS for the equivalent stratigraphic level (Fig. 9).
Fits with the correlation of Pizzo Mondello and the APTS. Specifically, magnetozones MP4r and MP5n of Pignola-2 are correlated respectively to E5r and E6n in Newark, as well as to the PM1r and PM2n in Pizzo Mondello. PMr1 and PM2n were correlated to the same Newark magnetozones by Muttoni et al. (2004) (Fig. 7).

Option 2:
- High t-value (~3.1)
- The correlation with the Newark APTS leads to a 0.9 M.y. discrepancy between the radiometric age of Pignola-2 and the age of the APTS (Fig. 9).
- This option does not fit with the previous correlation between Pizzo Mondello and the Newark APTS (Muttoni et al. 2004).

Option 1 implies only a minor discrepancy between the Pignola-2 U/Pb age and the Newark astrochronology, considering that in the lower Stockton Fm astrochronology is extrapolated from the upper Stockton and Lockatong Fms, where the 404 kyr McLaughlin cycles are better expressed (Kent and Olsen 1999, Olsen and Kent 1999). Moreover, Option 1 is coherent with previous correlations from the literature (Pizzo Mondello; Muttoni et al. 2004) and is preferred over Option 2.

We derived an age model from Option 1 that reveals a complex pattern of sedimentation rate along the Pignola-2 section (Fig. 10). In the cherty limestones, the sedimentation rate is mostly constant, except for a decrease just below and above the “green horizon”. In the “green horizon” the sedimentation rate increases, probably due to an enhanced runoff of siliciclastic sediments from the continent caused by increased rainfall and weathering, in consequence of the intensification of the humid conditions at the CPE.
The age model derived from Option 1 (Fig. 10) suggests an age of ~230.7 Ma for the Julian/Tuvalian boundary, approximated in Pignola-2 by the LO of conodont *Gladigondolella* spp.. Assuming a Carnian/Norian boundary at ~227 Ma (Muttoni et al. 2004) and a Ladinian/Carnian boundary at ~237 Ma (Mietto et al., 2012), the Julian should be ~6.3 My-long and the Tuvalian ~3.7 My-long. Assuming the magnetic susceptibility anomaly in Pignola-2 (covering the “green horizon” and the closest limestone beds) as expression of the CPE, its duration was about 1 My.

6. Conclusions

The paleomagnetic analyses of the Carnian sections of Pignola-2 (Southern Apennines, Italy) and Dibona (Dolomites, Italy) provided respectively a sequence of 12 and 8 magnetozones. The correlation with other Tethyan sections of the same time interval (Pizzo Mondello, Silická Brezová, Guri Zi) reveals a virtually continuous magnetostratigraphic record for the Carnian, constrained by a radiometric age of 230.91±0.33 Ma. Using a statistical approach, we performed a correlation between the Pignola-2 section and the Newark APTS that led to three statistically relevant options. Only two of them (Options 1 and 2) appear to be broadly consistent with both the radiometric age of the Pignola-2 ash-layer and their correlative ages in the Newark APTS. Although Option 2 was statistically slightly more robust, Option 1 is provisionally preferred as it shows the highest matching between radiometric and astrochronologic age estimates of the Pignola-2 ash-layer and does not violate the correlation between the Newark APTS and the Pizzo Mondello section as proposed by Muttoni et al. (2004) using the same statistical method adopted in this study. Ages of the main events of the Pignola-2 and Dibona sections were calculated using a model derived from Option 1. The level containing the Julian/Tuvalian boundary defined
by conodonts is now calibrated at ~230.7 Ma, and the levels attributed to the Carnian Pluvial Event should have deposited between ~229.7 and 230.7 Ma.

7. Acknowledgments
Nereo Preto, Leonardo Solazzi and Jacopo Dal Corso are acknowledged for their help during the fieldwork at the Dibona section. The “Fort Hoofddijk” crew is thanked for the help during the analysis and for the hospitality. Field and laboratory activities were funded by grants ex-60% (60A05-7013/15) to M. Rigo, ex-60% (60A05-0287/142) to A. Breda.

8. References


and correlation to the Late Triassic Newark astrochronological polarity time scale.


9. Figure captions, Table captions

Figure 1: The Pignola-2 section (coord.: Lat: 40°32'51.44"N, Long: 15°47'17.43"E) is located in the Southern Apennines, near Potenza (Southern Italy). The outcrop is along the main road connecting Pignola to Abriola, on the southern side of the Mt Crocetta. The Dibona section (coord.: Lat: 46°32'2.50"N, Long: 12°04'21.68"E) is located in the Dolomites, near Cortina d’Ampezzo (Belluno, Northern Italy). The outcrop is on the southern side of the Tofane di Rozes. Pignola-2 and Dibona sections were located in central Tethys during the Late Triassic.

Figure 2: The Pignola-2 section. From left to right: lithostratigraphy, conodonts and palynomorphs biostratigraphy, Virtual Geomagnetic Pole (VGP) latitudes (from ChRM directions), magnetostratigraphy and magnetic susceptibility. In the
magnetostratigraphy, black is normal polarity and white is reversed polarity. A total of 6 magnetozones have been identified, with a 5 meters interval of unknown polarity (grey shading) corresponding to the shales of the green clay-radiolaritic horizon (that could not be sampled). The U/Pb radiometric age of 230.91±0.33 Ma (Furin et al. 2006) comes from an ash-bed inside the green clay-radiolaritic horizon. Anomaly in the magnetic susceptibility around the “green horizon” represents the Carnian Pluvial Event in basinal environment (light-grey shaded interval).

Figure 3: The Dibona section. From left to right: lithostratigraphy of the investigated portion, palynomorph and conodont biostratigraphy, magnetostratigraphy, Virtual Geomagnetic Pole (VGP) latitudes (from ChRM directions), magnetic susceptibility and lithostratigraphy of the whole Dibona section. In the magnetostratigraphy, black is normal polarity and white is reversed polarity. In the lower panel is the Dibona Sandstone Mb (Heiligkreuz Fm) site and in the upper panel is the Travenanzes Fm site. A total of 8 magnetozones have been identified, 5 from the Dibona Sandstones and 3 from the Travenanzes Fm. The large portion of unknown polarity (grey shading) in the Travenanzes portion is due to sparse seemingly robust paleomagnetic data (only three meaningful VGP points). Extension of the Carnian Pluvial Event is illustrated with the grey shaded area in the whole Dibona section lithostratigraphy (on the right).

Figure 4: Vector end-point demagnetization diagrams (Zijderveld 1967) of the Pignola-2 section (panel A) and the Dibona section (panel B). Open circles are projections onto the vertical plane, and closed circle are projections onto the horizontal plane for in situ (geographic) coordinates.

Figure 5: Thermomagnetic curves determined with a Curie balance of samples PGM (Pignola-2, Calcari con Selce Fm; panels A, C), RAD (Pignola-2, Green clay-
radiolaritic horizon; panel B), MDS (Dibona, Heiligkreuz Fm; panels D, E, F), and MTV (Dibona, Travenanzes Fm; panels G, H, I). The PGM samples of Pignola-2 reveal a mixture of different minerals, including magnetite, in the cherty limestones, whereas in the green horizon (RAD sample) there is an increase in more Ti-rich magnetite. Samples MDS from the Heiligkreuz Fm in the Dibona section show a magnetization increase above 400-450°C, coherent with the reaction of pyrite to magnetite. Figure 6: Equal area projections for ChRM (characteristic remanent magnetization) of the Pignola-2 (upper panel) and Dibona (lower panel).

Figure 7: Correlations between Tethyan sections of Late Triassic and their calibration with the Newark Astrochronological Polarity Time Scale (APTS). The Norian to Rhaetian calibration and correlations are after Maron et al. (2015). All correlations between Tethyan sections are based on integrated conodont bio-magnetostratigraphy. Biostratigraphy of Silická Brezová has been updated after Mazza et al. (2010, 2011, 2012a,b). Pignola-2 provides a chronological tie point with the Newark APTS (U/Pb age of 230.91±0.33 Ma; Furin et al., 2006) and part of the Dibona magnetostratigraphy (MD1n) covers the missing interval in Pignola-2. Correlation between Pignola-2 and the Newark APTS based on the statistical method of Muttoni et al. (2004) and Maron et al. (2015).

Figure 8: Sequence of 24 correlation options between the Pignola-2 section and the Newark APTS. Dark grey bars indicate the correlations that are reliable at the 90%, black bars are the correlations reliable at the 95%. Only three options are reliable at 95%: Options 1, 2 and 24. Option 24 was rejected because being inconsistent with the age of Pignola-2, while Options 1 and 2 both covers an interval consistent with the radiometric age of 230.91 Ma. In particular, preferred Option 1 is perfectly coherent with...
with the time constraint in Pignola-2 and with the previous correlation between
Pizzo Mondello section and the Newark APTS (Muttoni et al. 2004), performed
using the same statistical method.

Figure 9: Comparison between Option 1 and Option 2. Both Options 1 and 2 show a
discrepancy between the radiometric age of 230.91 (Furin et al. 2006) and the
corresponding level in the Newark APTS. In Option 1 is the age discrepancy (0.6
M.y.) is smaller than in Option 2 (0.9 M.y.). A possible cause of this age discrepancy
is the absence of astronomic cycles recorded below E8n, and the age is presumed
assuming sedimentation rates similar to the upper Lockatong Formation (Olsen and
Kent 1999).

Figure 10: Age model based on Option 1. The inclination of the trend lines of the plot
reflects the sedimentation rates, which decrease just after and before the “green
clay-radiolaritic horizon”. The model provides an age of ~230.7 Ma for the
Julian/Tuvalian boundary and a duration of ~1 M.y. for the Carnian Pluvial Event.

Plate 1: Conodonts from samples DIN2 and DIN6. The fauna illustrated in the plate
includes Paragondolella polygnathiformis, Paragondolella noah, transitional
forms from P. noah to Metapolygnathus praecommunisti and M. praecommunisti.
The specimens of M. praecommunisti are basal, showing the accessoriral node
behind the cusp, the posterior prolongation of the keel and a quite centrally located
pit, but no nodes on the anterior platform margins. The occurrence of basal
representatives of this species in sample DIN6, together with advanced P. noah,
suggests a lower Tuvalian age (Mazza et al. 2011; Mazza et al. 2012a). 1:
Paragondolella polygnathiformis (Budurov and Stefanov 1965) (DIN2); 1c: the
blade of the specimen got broken; 2: Paragondolella noah (Hayashi 1968)
transitional to Metapolygnathus praecommunisti Mazza, Rigo and Nicora 2011
3: *Metapolygnathus* cf. *praecommunisti* (DIN6) Figs b-c: the blade termination got broken; 4: *Metapolygnathus* cf. *praecommunisti* (DIN6); 5-7: *Metapolygnathus praecommunisti* (DIN6). a: view from above; b: lateral view; c: view from below. All the conodonts are at the same scale.

Table 1: Mean directions from Pignola-2 and Dibona.
Fig. 3

DIBONA SECTION

HETIKGREUX FM
BORCAMB
DIBONA SANDSTONE MB
LAGAZUOI MB
TRAVENANZES FM

PALYNOMORPHS
Conodonts

Patinasporites densus
Granuloperculatipollis rudis
Riccisporites cf. R. tuberculatus
Equisetosporites chinleanus
Aurisporites astigmosus
Duplicisporites continuus

VIRTUAL GEOMAGNETIC POLE LATITUDE
MAGNETIC SUSCEPTIBILITY

CARNIAN PLUVIAL EVENT
MAGNETOSTRATIGRAPHY

SANDSTONE
CASSIANO FM
SAN

DOLOMIA PRINCIPALE
FIGURE 8

Newark/APT S

Pignola-2

Aglianico ash-bed
230.91±0.33 Ma
(Furin et al. 2006)
## TABLE 1. PALEOMAGNETIC DIRECTIONS FROM PIGNOLA-2 AND DIBONA SECTIONS

<table>
<thead>
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<th>Site</th>
<th>Comp.</th>
<th>N</th>
<th>k</th>
<th>α95</th>
<th>Dec.</th>
<th>Inc.</th>
<th>k</th>
<th>α95</th>
<th>Dec.</th>
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<td>24.1</td>
<td>4.3°</td>
<td>353.5°E</td>
<td>59.7°</td>
<td>23.9</td>
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<td>28.4°E</td>
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<tr>
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<td>ChRM</td>
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<td>4.2</td>
<td>11.8°</td>
<td>350.5°E</td>
<td>33.9°</td>
<td>4.2</td>
<td>11.8°</td>
<td>10.2°E</td>
<td>41.4°</td>
</tr>
</tbody>
</table>

**LEGEND**

- **Comp.**: paleomagnetic component
- **N**: number of samples
- **k, α95**: Fisher statistics parameters
- **Dec.:** mean declination
- **Inc.:** mean inclination