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THE VALANGINIAN WEISSERT OCEANIC ANOXIC EVENT RECORDED IN CENTRAL-EASTERN SARDINIA (ITALY)

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Abstract . Investigations on the S'Ozzastru section from the northern part of the Mt Albo area (central-eastern Sardinia, Italy) for integrated litho- bio- and chemostratigraphy allowed the identification of the Valanginian Weissert Oceanic Anoxic Event (OAE), testified by a positive carbon isotope excursion (CIE). The section, which represents the deepest-water succession of the Valanginian in Sardinia, is composed of the Schiriddè Limestone followed by the Siniscola Marl, both proposed as new lithostratigraphic units. The presence among the ammonites of *Busnardoites campylotoxus* allows the attribution of the Schiriddè Limestone to the upper Lower Valanginian Inostranzewi Zone of Reboulet et al. 2014. Further characterization of this unit was not possible since it is barren/almost barren of nannofossils. The Siniscola Marl can be ascribed to the lower Upper Valanginian on the basis of the ammonite fauna indicating the Verrucosum Zone, and of the nannofossil content suggesting the Zone NK3. The carbon isotope record in the Siniscola Marl is characterized by a positive excursion with values up to 2.98 ‰. In the nannofossil assemblages, nannoconids are not particularly abundant and are found, among others, together with *C. oblongata, D. lehmanii*, and pentaliths. The scarcity of nannoconids is regarded as a biostratigraphic support for the identification of the Weissert OAE, as it possibly reflects the "nannoconid decline" interval which characterizes this event. The end of the Weissert OAE CIE is not recorded probably because of suppression of the upper part of the succession for tectonic causes.

Introduction

The Valanginian is characterized by a globally recorded positive C isotope excursion (CIE) which reflects an episode of carbon cycle perturbation associated with profound variations in climatic and palaeoceanographic conditions. This positive CIE was first detected in successions from the Tethys Ocean (Weissert 1989) and further investigations demonstrated its global significance in different locations (Lini et al. 1992; Channell et al. 1993; Hennig et al. 1999; Erba et al. 2004; Weissert & Erba 2004; Sprovieri et al. 2006; Duchamp-Alphonse et al. 2007; Greselle et al. 2011; Kujau et al. 2013; Charbonnier et al. 2013; Morales et al. 2015; Meissner et al. 2015) including terrestrial records (Gröcke et al. 2005). Palaeoceanographic studies indicated that the CIE coincided with a widespread eutrophication of marine ecosystems associated with platform drowning (Schlager 1981; Föllmi et al. 1994; Graziano 1999; Gradinaru et al. 2016), pelagic carbonate crisis (e.g. Lini et al. 1992; Bersezio et al. 2002; Reboulet et al. 2003; Erba and Tremolada 2004; Erba et al. 2004), and increased evolutionary rates in planktonic communities (Erba et al. 2004; Erba 2006, and reference therein). Erba et al. (2004) formalized the Valanginian event as the "Weissert Oceanic Anoxic Event" (OAE), defined on the basis of the $\delta^{13}C$ curve being comprised between the onset of the positive CIE and the highest values of the isotopic excursion. Stratigraphic studies allowed to calibrate the Valanginian CIE with magnetostratigraphy and biostratigraphy (e.g.

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Lini et al. 1992; Channell et al. 1993, 1995; Hennig et al. 1999; Bersezio et al. 2002; Reboulet et al. 2003; Erba et al. 2004; Duchamp-Alphonse et al. 2007; Bornemann & Mutterlose 2008; Charbonnier et al. 2013), and highlighted coeval changes in fossil assemblage composition and abundance of different marine fossil groups. These biotic changes included the decline in nannoconid abundance, called "nannoconid decline", whose onset predated the positive CIE (Channell et al. 1993; Bersezio et al. 2002; Erba & Tremolada 2004; Erba et al. 2004; Duchamp-Alphonse et al. 2007; Bornemann & Mutterlose 2008; Greselle et al. 2011; Barbarin et al. 2012; Mattioli et al. 2014), and coincided with a peak in pentalith abundance (Bersezio et al. 2002; Erba & Tremolada 2004) and the beginning of increased abundance of *Diazomatolithus* evidenced in Tethyan sections and ODP Site 1149 (Bersezio et al. 2002; Erba & Tremolada 2004; Erba et al. 2004). The "nannoconid decline" was explained as the nannoplankton response to the global nutrification episode that occurred over the onset of the OAE (Bersezio et al. 2002; Erba & Tremolada 2004; Erba et al. 2004; Bornemann & Mutterlose 2008; Mattioli et al. 2014) linked to major igneous/tectonic events of the Paranà-Etendeka igneous province. The volcanic activity was identified as the principal cause of altered oceanic/atmospheric conditions (e.g. Lini et al. 1992; Channell et al. 1993, Weissert et al. 1998; Bersezio et al. 2002; Erba & Tremolada 2004; Martinez et al. 2015; Charbonnier et al. 2016, 2017), possibly responsible for increased $p\mathrm{CO_2}$ in the atmosphere and consequent accelerated hydrological cycling (e.g. Weissert 1989; Lini et al. 1992; Föllmi et al. 1994; Weissert et al. 1998; Erba et al. 2004) during the initial phase of OAE. In addition, higher surface water fertility was also suggested to have been induced by the input of biolimiting metals due to increased spreading rates and hydrothermal activity (Lini et al. 1992; Channell et al. 1993; Weissert et al. 1998; Bersezio et al. 2002; Erba et al. 2004; Bornemann & Mutterlose 2008). Under enhanced primary productivity, the carbon burial process was accelerated (Channell et al. 1993; Weissert et al. 1998), locally favoring the onset of oceanic dysoxia/anoxia with black shales deposition (Erba et al. 2004). Reconstructed climatic conditions over the Weissert OAE indicate general cooler temperature with a cooling registered in correspondence of the final phase of OAE (e.g. Kemper 1987; Mutterlose & Kessels 2000; Mutterlose et al. 2003; Reboulet et al. 2003; Greselle et al. 2011; Barbarin et al. 2012; Mattioli et al. 2014).

The purpose of this study is to characterize through integrated litho-, bio- and chemostratigraphy the Mt Albo succession (central-eastern Sardinia) and detect evidence of the Valanginian Weissert OAE in Sardinia. The Mt Albo area records Valanginian sediments of relatively deep water settings and their stratigraphic characterization may constitute the background for further palaeogeographic and palaeoenvironmental studies of the Lower Cretaceous deposits of the island. We provide a detailed geological and stratigraphic setting which includes new data necessary to frame the Valanginian deposits of Sardinia in a wider palaeogeographic context. The data of which the literature sources are not reported, are to be regarded as new.

Material and methods

In this work we focus on the S'Ozzastru quarry section situated near the village of Siniscola, in proximity of the northern termination of the Mt Albo. Field work included a partial updating of the Mt Albo geological map published by Dieni & Massari in 1971, identification and characterization of the lithostratigraphical units, and sedimentological observations on the measured and sampled sections (Figs 1- 3; Pl. 1). Furthermore, the S'Ozzastru section has been analyzed for integrated litho-, bio-, and chemostratigraphy.

Geological setting and stratigraphy

The study area is located in central-eastern Sardinia, where the sedimentary cover of the Palaeozoic metamorphic substrate is represented by a Middle Jurassic-Lower Cretaceous succession, mostly consisting of shallow-marine dolostones and limestones. The investigation is specifically focused on the Lower Cretaceous deposits of the Mt Albo, a fault-bounded massif elongated in a NE-SW direction (Fig. 1). The Middle Jurassic-Lower Cretaceous succession of this massif shows remarkable facies variations, indicative of a north-easterly trend from inner and protected environments to more open and deeper sea. In the northern part (Pl. 1) quartz-arenitic brown dolostone evolving to ooidal dolostone of the Dorgali Formation (unit established by Amadesi et al. 1960) of Middle-Late Bathonian age are followed by fine-grained, well-bedded middle- to outer-ramp limestones and cherty limestones of

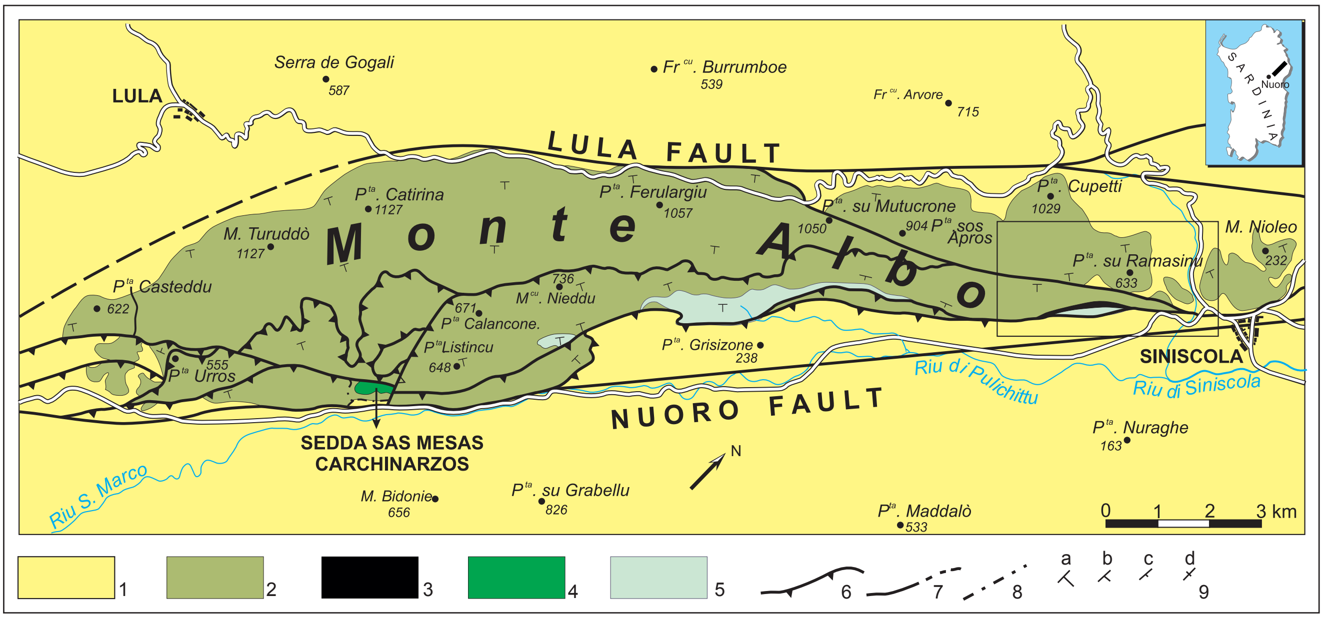


Fig. 1 - Geological sketch map of the Mt Albo massif. Legend: 1) Palaeozoic basement. 2) Bajocian(?) to Berriasian siliciclastic to carbonate succession (Genna Selole Fm., Dorgali Fm. and Monte Bardia Limestone). 3) Valanginian limestone and marl (Schiriddè Limestone and Siniscola Marl). 4) Valanginian to Lower Hauterivian limestone emplaced by palaeo-slide. 5) Middle Eocene Cuccuru 'e Flores Conglomerate. 6) Overthrust. 7) Fault. 8) Limit of palaeo-slide (Sedda sas Mesas – Carchinarzos). 9) Attitude and dip of beds (a= 20°- 40°; b= 41°- 60°; c= 61°- 80°; d= 81°- 90°). Quaternary deposits omitted. Inset shows the area of the geological map of Pl. 1 (From Dieni & Massari 1987, fig. 42, modified).

latest Bathonian - earliest Tithonian age (S'Adde Limestone, formational name proposed by Dieni & Massari 1985, and formalised by Casellato et al. 2012), with hardgrounds in the lowermost part and cherty limestones in the upper part, containing belemnites and ammonites particularly of the Upper Kimmeridgian Beckeri Zone (Dieni et al. 1966). This interval is followed by a thick carbonate platform complex of Early Tithonian to Late Berriasian age (Monte Bardia Limestone, formation established by Amadesi et al. 1960), consisting of thick-bedded to massive white limestones, mostly made up of biosparites and algal bindstones, and in places with patch reefs in the lower part (Dieni & Massari 1985, 1991), with corals, nerineids, *Clypeina jurassica* Favre in Joukovsky & Favre, 1927, *Campbelliella striata* (Carozzi, 1954), *Neoteutloporella socialis* (Praturlon, 1963), *Mohlerina basiliensis* (Mohler, 1938), abundant *Crescentiella morronensis* (Crescenti, 1969), *Koskinobullina socialis* Cherchi & Schroeder, 1979, and "*Bacinella-Lithocodium*" boundstones.

Contrasting with the northern part of the Mt Albo massif, the southern part is characterized by a much thicker succession of dolostones of the Dorgali Formation, locally underlain by fluvio-deltaic quartz-conglomerates and sandstones (Genna Selole Formation, Bajocian(?)-Lower Bathonian; Dieni & Massari in Dieni et al. 1983), and in places including limestone lenses particularly rich in infaunal bivalves, usually in life position [among which *Pholadomya lirata* (J. Sowerby, 1818), *Ph. ovalis* (J. Sowerby, 1819), *Homomya gibbosa* (J. Sowerby, 1813), *Ceratomya concentrica* (J. de C. Sowerby, 1825)], and with brachiopods (including *Burmirhynchia turgida* Buckman, 1917) of Middle-Late Bathonian age. This succession is followed upwards in apparent continuity by the Tithonian Monte Bardia Limestone with *Clypeina jurassica* and *Campbelliella striata*, represented here by back-ramp lagoonal and protected-flat limestones with stromatolite horizons.

Tithonian-Berriasian deposits were formed on a carbonate platform with a spectrum of variable environments. In the northern Mt Albo massif, platform-margin limestones predominate in the upper Monte Bardia Limestone, where they mostly consist of bioclastic sediments including various microencrusters such as "*Bacinella-Lithocodium*" boundstone and *Crescentiella morronensis*.

In the southern Mt Albo massif, the upper Monte Bardia Limestone includes in the Lower Berriasian a characteristic interval which is well documented in central-eastern Sardinia [in the Orosei area (Mt Tuttavista) and extensively in the Oliena-Orgosolo-Urzulei "Supramonte" (e.g. Lanaittu area near Oliena; see Dieni & Massari 1985, fig. 3)]. It consists of predominant mud-cracked micritic laminites with intercalated layers of storm-related washover deposits and black pebble breccias (the latter resulting from fragmentation of blackened carbonate sediments rich in organic matter mostly formed in freshwater to brackish pools with reducing conditions), corresponding to the well-known "Purbeckian facies" of the literature. This interval contains oligotypic assemblages of charophytes, ostracods and small molluscs (among which planorbid gastropods) suggesting a scenario of restricted environments of very shallow-water and temporarily emergent wide inter- to supratidal flats disseminated with lagoonal to lacustrine ponds. Wholly similar Berriasian deposits indicating a regressive transition from the marine platform deposits of the Upper Tithonian to lagoonal-lacustrine "Purbeckian" marls and limestones with charophytes and ostracods are also known in the Nurra area of NW Sardinia (Maxia & Pecorini 1963; Chabrier & Fourcade 1975; Colin et al. 1985). The facies association of these deposits points to a marked regressive trend which is consistent with a scenario of lowstand conditions, as generally observed in the uppermost Jurassic to Lower Berriasian successions of the central and northern European area (Hallam 1986; Haq et al. 1987; Hallam et al. 1991; Abbink et al. 2001; Dera et al. 2011).

A sedimentary record of the Valanginian in central-eastern Sardinia is only known in the Mt Albo massif, in the Orosei area and in the Lanaittu valley (Oliena).

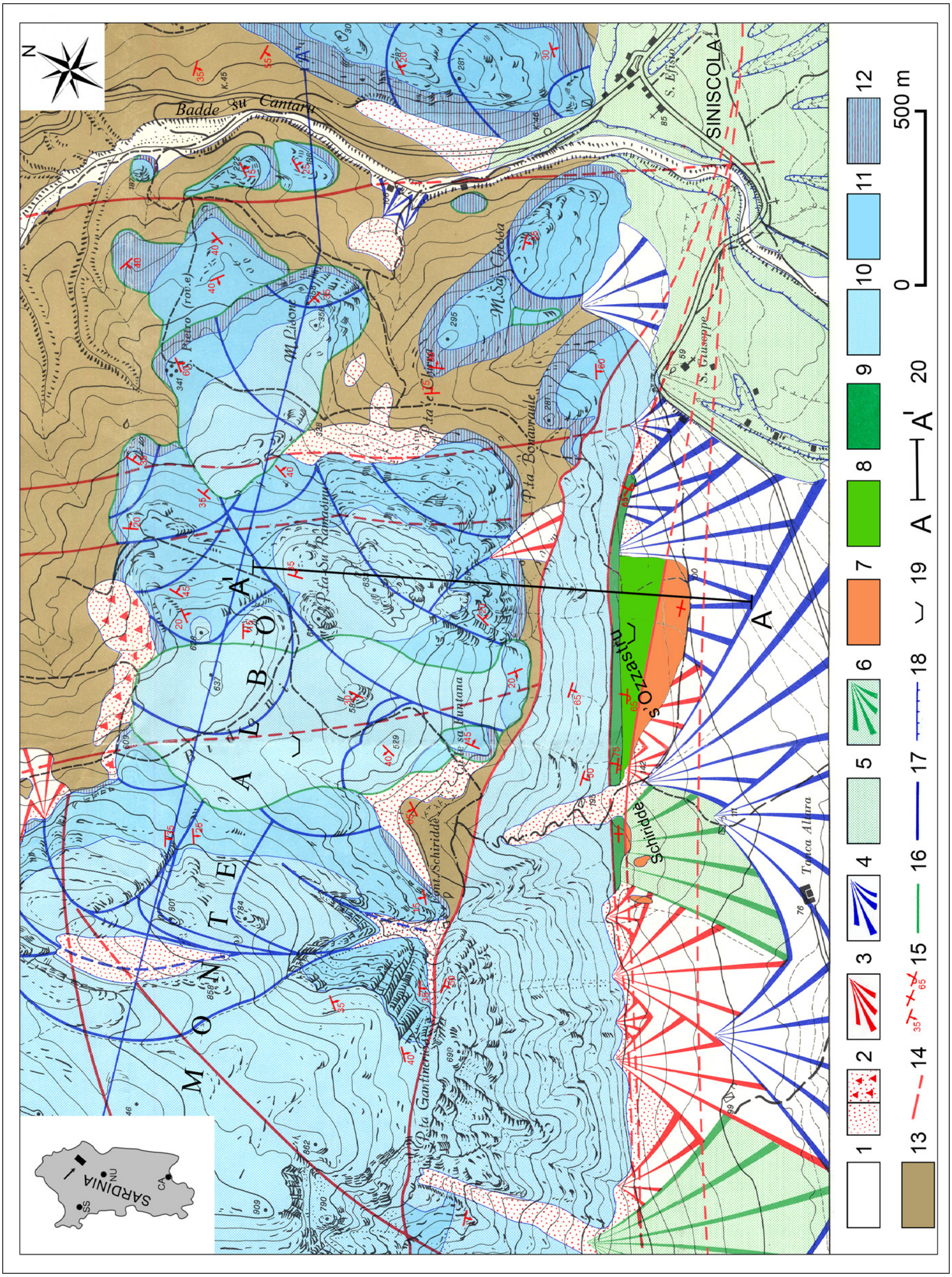
The Valanginian deposits cropping out in proximity of the northern termination of the Mt Albo, which are the focus of this paper, are significantly different in facies from those of adjacent areas (see below) for their deeper-water depositional setting. The deposits appear tectonically squeezed between Upper Berriasian limestones, representing the last terms of the Monte Bardia Limestone, and Middle Eocene rudites and arenites of the Cuccuru 'e Flores Conglomerate (formation described by Dieni & Massari 1966 and formalized by Dieni et al. 2008) (Pl. 1 and Fig. 1). In central-eastern Sardinia the latter are commonly localized at the feet of fault scarps bounding the carbonate massifs and linked to the Cenozoic tectonics.

In general, the Lower Cretaceous deposits are reduced in extension at the foot of the Mt Albo and appear as slices bounded by faults delimiting the massif on its eastern side (Fig. 1). The faults are part of a transpressional system, which is a major, polyphase structural element of regional importance crossing the whole island. The different depositional setting, particularly comparing the marine and almost continuous succession of the northern Mt Albo (see below) to the incomplete sequences of the other areas, laid down in significantly shallower environments, may be interpreted to result from Tertiary transcurrent movements, which brought near to one another originally more spaced areas. This fault system was activated in the Middle Eocene in an inferred foreland setting with respect to the Alpine orogen, and re-activated in the Late Oligocene – Early Miocene initial stages of the Apenninic history, when the Corsica-Sardinia microplate behaved as hinterland of the Apenninic migrating front (Dieni et al. 2008; Massari &

PLATE 1

Geologic map of the Siniscola area, central-eastern Sardinia (from Dieni & Massari 1971, updated in 1986).

Legend: 1) Recent and present alluvial deposits (*Holocene*). 2) Scree deposits; idem with boulders (*Würm-Holocene*). 3) Detrital cones mostly composed of gelifraction limestone clasts ("éboulis ordonnés") (*Würm-Holocene*). 4) Uncemented alluvial cones mostly consisting of limestone clasts (*Würm-Holocene*). 5) Terraced alluvial deposits composed of clasts of basement rocks (*Riss*?). 6) Tightly cemented and dismembered alluvial cones mostly made up of gelifraction limestone clasts (*Riss*?). 7) Cuccuru 'e Flores Conglomerate: polymict, chaotic and thick-bedded breccia and stratified sandstone (*Middle Eocene*). 8) Siniscola Marl: grey and grey-yellow marl and, in the upper part, alternating marl/marly limestone with ammonites and very abundant radiolarians and sponge spicules (*Late Valanginian*). 9) Schiriddè Limestone: light-brown cherty limestone and limestone with marly interbeds, containing ammonites, radiolarians, sponge spicules and calpionellids (*Early Valanginian*). 10) Monte Bardia Limestone: thick-bedded to massive white limestone in places bioconstructed, with corals, nerineids, etc. (*Early Tithonian - Late Berriasian*). 11) S'Adde Limestone: light-brown to grey fine-grained, well-bedded limestone, condensed with hardgrounds near the base, cherty in the upper part, with ammonites and belemnites in the uppermost part (*latest Bathonian - earliest Tithonian*). 12) Dorgali Formation: well bedded quartz-arenitic and locally ooidal brown dolostone (*Middle-Late Bathonian*). 13) Metamorphic basement (phyllite, micaschist, paragneiss, quartzite, etc.) (*Palaeozoic*). 14) Fault. 15) Attitude of beds and schistosity planes. 16) Limits of slumped masses of Jurassic formations (*Quaternary*). 17) Concave-upward slip surfaces delimiting rotational slumps (*Quaternary*). 18) Edge of fluvial terrace. 19) Quarry. 20) Trace of the geological section of Fig. 2A.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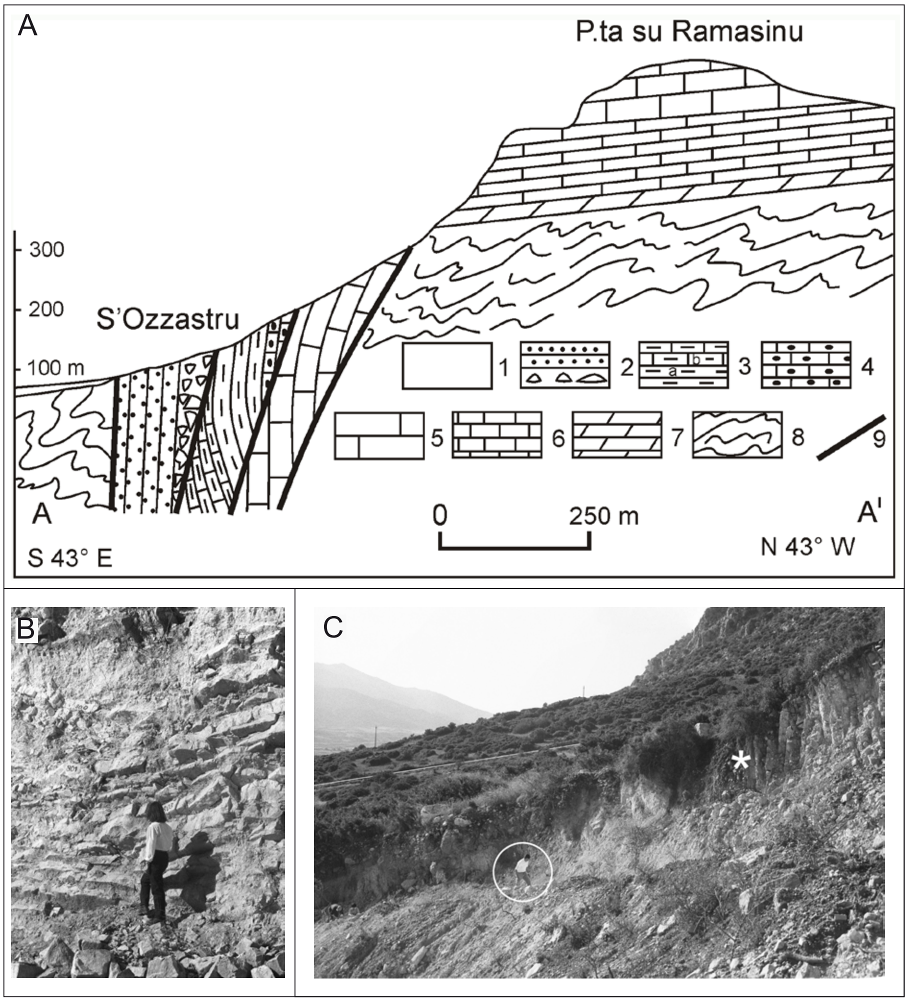

Fig. 2 - A) Geological section across the S'Ozzastru area (trace A--A' in the geological map of Pl. 1). Legend: 1) Recent and present alluvial deposits (*Holocene*). 2) Cuccuru 'e Flores Conglomerate: sandstone and breccia (*Middle Eocene*). 3) Siniscola Marl: marl (a) and marly limestone (b) (*Late Valanginian*). 4) Schiriddè Limestone: cherty limestone (*Early Valanginian*). 5) Monte Bardia Limestone: limestone (*Early Tithonian - Late Berriasian*). 6) S'Adde Limestone: limestone and cherty limestone (*latest Bathonian - earliest Tithonian*). 7) Dorgali Formation: dolostone (*Middle-Late Bathonian*). 8) Metamorphic basement (*Palaeozoic*). 9) Fault.
B) Overturned succession of the Schiriddè Limestone between beds 799 and 804 of the log, consisting of alternating cherty limestone and marl.
C) Transition (asterisk) from subvertical strata of the Schiriddè Limestone (right) to the Siniscola Marl, partly covered by scree deposits.

Dieni 2014). As a result of this polyhistory tectonics, the Mt Albo area was sliced in a number of mutually overlapping slivers involved in a positive flower structure, the last motion being represented by sinistral transpression (Dieni & Massari 1991, fig. 26; Carmignani et al. 1992) (Pl. 1; Figs 1 and 2A).

The Berriasian-Valanginian succession of Mt Albo

In the Mt Albo massif the Lower Cretaceous succession has been analyzed in the S'Ozzastru section (already briefly described by Dieni et al. 1987), located near the village of Siniscola, close to the northern corner of the massif (Pl. 1), and in the Sas Mesas-Carchinarzos section, located in the south-eastern part of the massif. The focus of this paper is specifically on the S'Ozzastru section, whereas the latter section has been taken into account for its sharp differences in the stratigraphy and facies association with respect to the former and was not analyzed for the nannofossil assemblages.

A - S'Ozzastru area

The last thirty metres of the carbonate platform complex of the Monte Bardia Limestone in the small Riu Siccu valley, located about 2.2 Km SW of S'Ozzastru (Pl. 1), are represented by platform-margin white biosparites with *Coscinoconus* ex gr. *alpinus-elongatus* Leupold, 1935, *Pseudocyclammina* ex gr. *lituus* (Yokoyama, 1890), *Protopeneroplis ultragranulata* (Gorbatchick, 1971), *Montsalevia elevata* Zaninetti, Salvini-Bonnard & Charollais, 1987, *Pseudocymopolia jurassica* (Dragastan, 1968), *Crescentiella morronensis*, *Favreina salevensis* (Paréjas, 1948), associated with rare calpionellids, including *Calpionella alpina* Lorenz, 1902, and *C.* cf. *elliptica* Cadisch, 1932 (see also Azéma et al. 1978).

In the S'Ozzastru quarry section (Pl. 1) the above deposits grade upwards into light-brown biomicrites and intramicrites of about 10 m thickness (uppermost Monte Bardia Limestone) involved in the fault bundle affecting the Lower Cretaceous succession in this area and showing therefore intense tectonic shearing. The microfaunal content includes

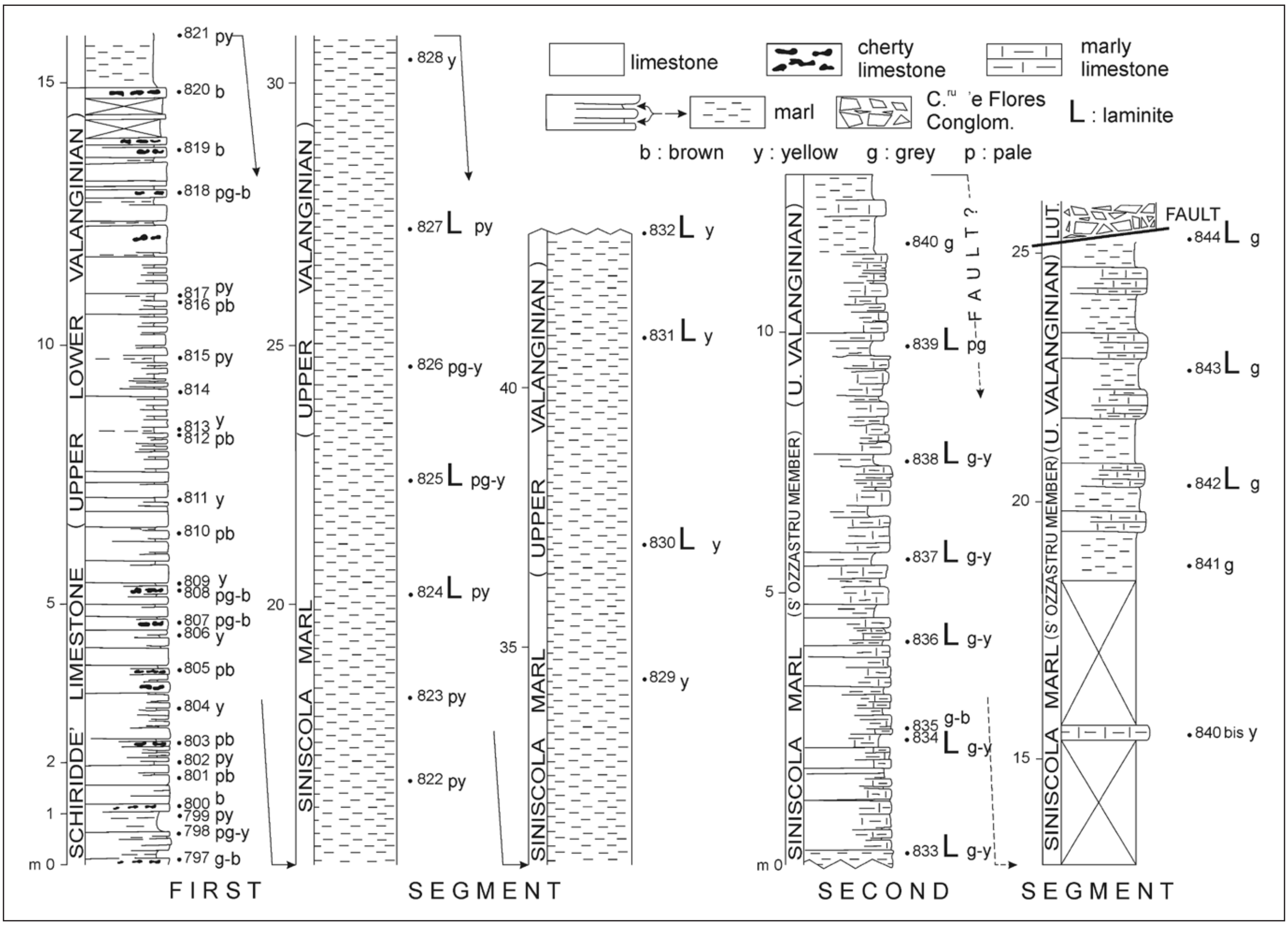

Fig. 3 - Stratigraphic log of the Valanginian succession in the S'Ozzastru quarry. Numbers refer to the studied samples.

Coscinoconus ex gr. *alpinus-elongatus* and *P. ultragranulata*, accompanied by common calpionellids, among which *Calpionellopsis simplex* (Colom, 1939), *Cps. oblonga* (Cadisch, 1932), *Praecalpionellites murgeanui* (Pop, 1974), in order of occurrence, and *Calpionella alpina*, *Remaniella cadischiana* (Colom, 1948), *Tintinnopsella carpathica* (Murgeanu & Filipescu, 1933), and *T. longa* (Colom, 1953). This assemblage indicates a Late Berriasian age (see Grün & Blau 1997, and references therein) which is confirmed by the presence, among the macrofossils, of the gastropod *Leviathania leviathan* (Pictet & Campiche, 1863) [see Decrouez in Zaninetti et al. 1988, where this taxon is described as *Leviathania sautieri* (Coquand, 1856); we adopt here the junior specific name *leviathan* in application of the I.C.Z.N. art. 59.3].

At the top of this succession a sharp lithologic change, probably corresponding to a short stratigraphic gap of the uppermost Berriasian, indicated by the absence of *Praecalpionellites dadayi* (Knauer, 1963) (see Grün & Blau 1997), marks the transition to grey-light brown micritic and slightly marly, commonly bioturbated limestones of about 2 m thickness, with *Calpionellites darderi* (Colom, 1934), *Praecalpionellites murgeanui*, *T. carpathica*, and *T. longa*, indicating an Early Valanginian age.

The following yellowish cherty limestones about 15 m thick (reported in the lower part of the first segment of the log in Fig. 3) contain ammonites, including *Olcostephanus* sp., *Busnardoites campylotoxus* (Uhlig, 1910), *B.* cf. *campylotoxus*, *Neocomites neocomiensis* (d'Orbigny, 1841), *Teschenites* cf. *teschenensis* (Uhlig, 1902) and *Neohoploceras* sp. juv. ? (Fig. 4). This assemblage indicates the Inostranzewi Zone of Reboulet et al. 2014 (= Campylotoxus Zone *auctt.*) of late Early Valanginian age (see Reboulet et al. 2009, 2014, and references therein). The microfossil assemblages are very rich in sponge spicules [reniform sterraster microscleres are particularly abundant, as in the entire interval of the log; Figs. 5 (s) and 6 (a)], associated with radiolarians, sparse benthic foraminifers [among

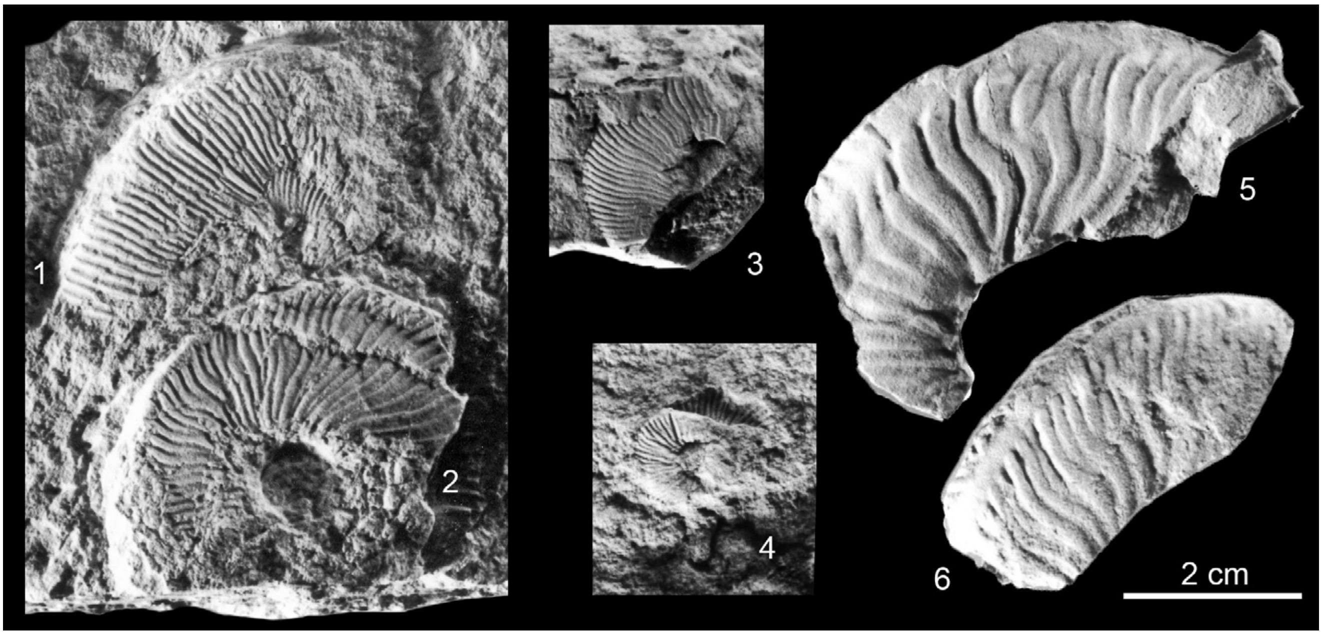

Fig. 4 - Ammonites from the upper Lower Valanginian (Inostranzewi Zone) (Schiriddè Limestone, S'Ozzastru area). 1) *Olcostephanus* sp.; 2, 3) *Neocomites neocomiensis* (d'Orbigny, 1841); 4) *Neohoploceras* sp. juv.?; 5) *Busnardoites campylotoxus* (Uhlig, 1910); 6) *Busnardoites* cf. *campylotoxus* (Uhlig, 1910).

which *Praedorothia praehauteriviana* (Dieni & Massari, 1966), *Protomarssonella hechti* (Dieni & Massari, 1966), *P. kummi* (Zedler, 1961), *Pseudotextulariella salevensis* Charollais, Brönnimann & Zaninetti, 1966, *Meandrospira favrei* (Charollais, Brönnimann & Zaninetti, 1966) [= *Glomospirella gaultina* (Berthelin) in Dieni & Massari 1966, p. 85, pl. I,fig. 1, and pl. IX, fig. 1], *Lenticulina crepidularis* (Roemer, 1842), *L.* spp., *Nodosaria* sp., *Rumanolina feifeli* (Paalzow, 1932), *Spirillina* sp., polymorphinids], and *Aeolisaccus inconstans* Radoičić, 1967 (Fig. 5). Local lamination and iso-orientation of elongate sponge spicules suggest episodic bottom current activity (Fig. 6b). For the overall Lower Valanginian interval we propose the formational name of **Schiriddè Limestone,** with type-section in the S'Ozzastru area (Figs 2B, 3).

This interval grades into grey marls (Fig. 2C) of about 30 m thickness, with laminated, sapropel-like layers (upper part of the first segment of the log in Fig. 3), passing to alternating dark-grey marls and marly limestones (yellowish where weathered) with iron sulfides and rich in organic matter and again with laminated, sapropel-like, intervals (second segment of the log, 25 m thick; Fig. 3). For this complex we propose the formational name of **Siniscola Marl**, with type-section again in the S'Ozzastru area (Fig. 3). The second segment is separated in the log as a member, which has been named **S'Ozzastru Member**. It should be noted that the estimated thickness of the Siniscola Marl is somewhat imprecise, due to the increasingly tectonized upper part of the succession, where the formation is in contact with the Eocene Cuccuru 'e Flores Conglomerate (Fig. 1) by means of the Nuoro fault system bounding the Mt Albo massif on the eastern side; as a result, the unit crops out in a limited, strip-like belt and is strongly sheared and steeply dipping (Pl. 1; Figs 2A, C).

The first segment of the Siniscola Marl yielded some ammonites, among which *Neocomites neocomiensis* (d'Orbigny, 1841) and *Neocomites platycostatus* Sayn, 1907 (Fig. 7). Concerning the age assignment, *N. neocomiensis* is currently recorded from the uppermost part of the lower Valanginian (FOD in the Inostranzevi Zone) to the middle part of the upper Valanginian (LOD in the Peregrinus Zone, Nicklesi Subzone) and *N. platycostatus* from the upper part of the Inostranzevi Zone (FOD in the Platycostatus Subzone) to the Verrucosum Zone (LOD in the lower part of the Pronecostatum Subzone) (Reboulet & Atrops 1999; Reboulet et al. 2014). Consequently, the more restricted biostratigraphic interval is given by the occurrence of *N. platycostatus*. In conclusion, the ammonite fauna, integrated with the nannofossil content (see below), suggests that the Siniscola Marl may be attributed to the lower Upper Valanginian (Verrucosum Zone).

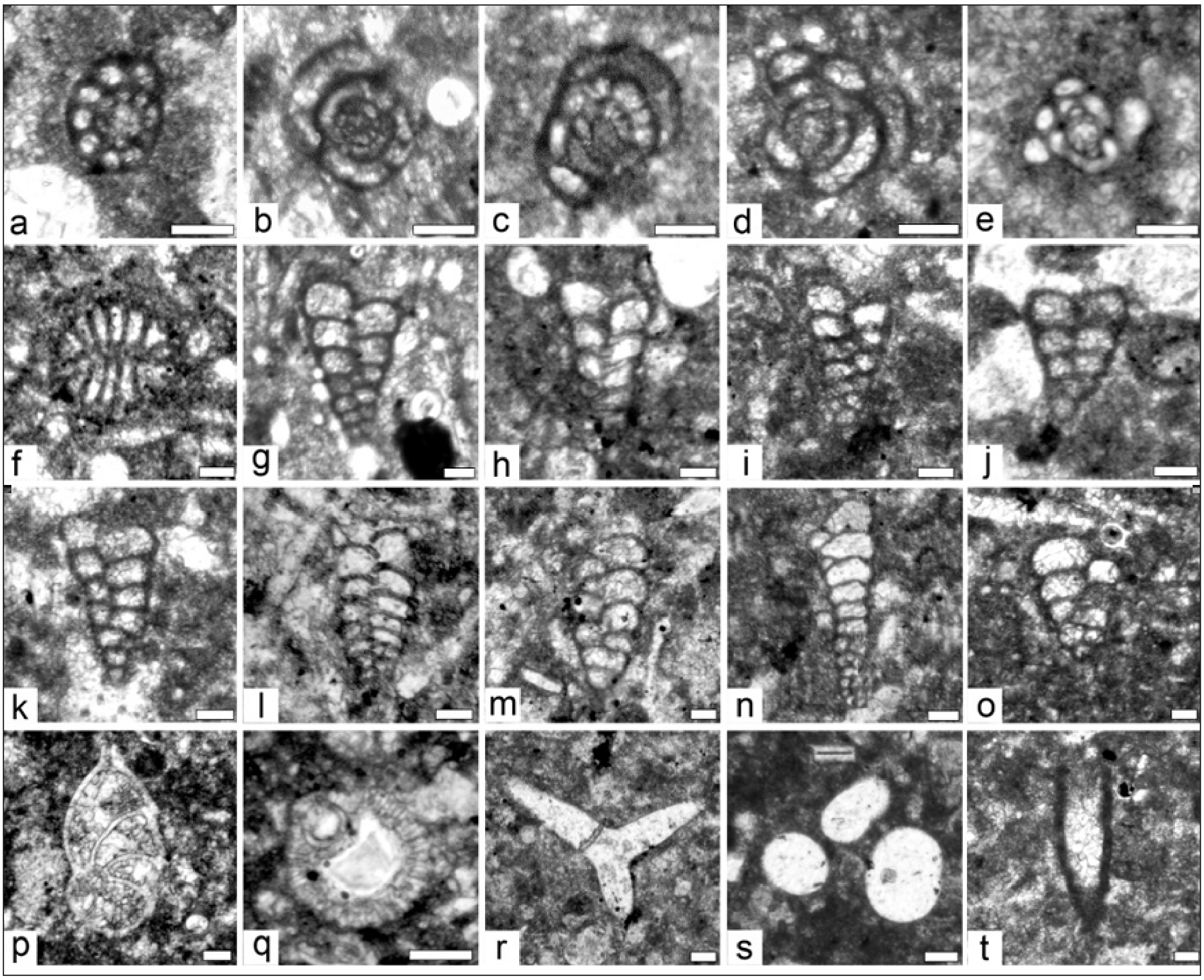

Fig. 5 - Microfossils from the Schiriddè Limestone (Early Valanginian) and Siniscola Marl (Late Valanginian) formations. a-e) *Meandrospira favrei* (Charollais, Brönnimann & Zaninetti), differently oriented sections of various morphotypes (beds 797, 835, 818, 808, 820). f) *Pseudotextulariella salevensis* Charollais, Brönnimann & Zaninetti, horizontal section (808). g) *Protomarssonella kummi* (Zedler), sub-axial section (835). h-k) *Protomarssonella hechti* (Dieni & Massari), sub-axial and oblique sections (803, 800, 820, 820). l-o) *Praedorothia praehauteriviana* (Dieni & Massari), variously oriented sections (810, 808, 797, 819). p) *Lenticulina crepidularis* (Roemer), equatorial section (797). q) radiolarian, oblique section (801). r) sponge microsclere of dichotriaene type (803). s) reniform sterraster microscleres (801). t) *Aeolisaccus inconstans* Radoičić, sub-axial section (820). Scale bars=50 µm.

The succession of the Siniscola Marl has been sampled in 1984 in the above mentioned place, where the rocks are quarried by the local cement factory. Since then, the quarrying activity has undergone changes which significantly modified the situation, particularly with the removal of the Quaternary cover. For this reason, the geological map presented in Pl. 1, illustrating the setting at the time of the 1986 field survey, is not completely updated.

The marls are intensely bioturbated, except in the finely laminated layers, and are particularly rich in very well preserved radiolarians [some of them determined by Aita (in Dieni et al. 1987, p. 148) and others figured by Dieni & Massari 1991, fig. 23 (c-i), and ascribed here to *Podobursa* sp. (c), *Pseudoxitus omanensis* Dumitrica in Dumitrica et al., 1997 (d), *Williriedellum peterschmittae* Schaaf, 1981 (e), *Cryptamphorella* cf. *conara* (Foreman, 1968) (f), *Pseudocrolanium puga* (Schaaf, 1981) (g), *Hiscocapsa zweilii* (Jud, 1994) (h), *Hemicryptocapsa* sp. (i)] and sponge spicules [some of them illustrated in Dieni & Massari 1991, fig. 23 (l-p), corresponding to microcriccorhabds (type a and b) of Desmospongia (l, m, p), dichotriaene type (n) and aster type (o)]. The entire assemblage of radiolarians and sponge spicules shall be described in a separate paper (Dumitrica P., Dieni I. & Massari F., in progress). Also the S'Ozzastru Member is very rich in sponge spicules, locally iso-oriented, and ra-

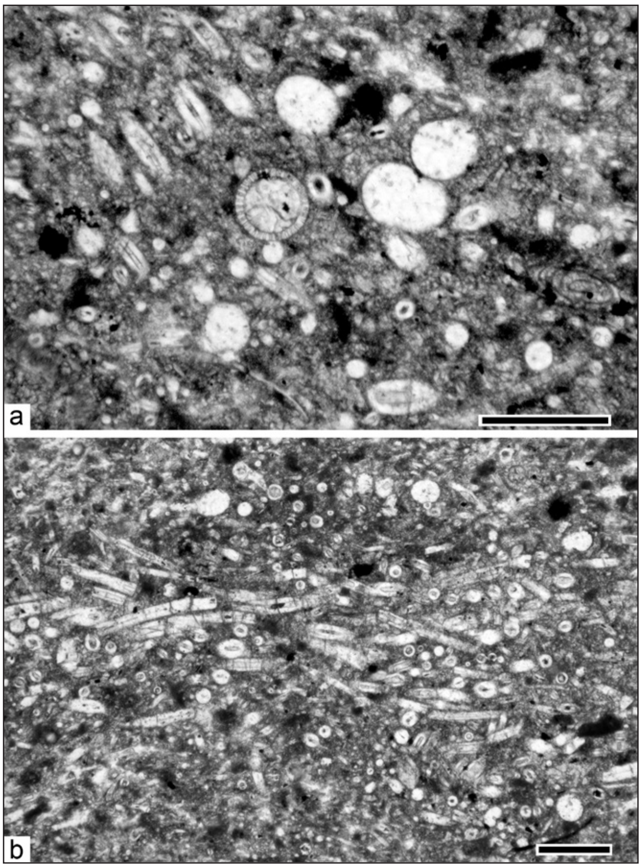

Fig. 6 - Microfacies dominated by sponge spicules in the Siniscola Marl, S'Ozzastru Member (marly limestone, bed 835). a) particularly common reniform sterraster microscleres associated with rare radiolarians. b) iso-orientation of elongate megascleres. Scale bars: 200 μm.

diolarians (Fig. 6).

Specific characteristics of the Siniscola Marl, such as the abundance of siliceous organisms and presence of sapropel-like laminites, indicate a relatively deep, probably upper slope, environment recording episodes of high organic productivity and probably dys- to anaerobic conditions.

B - Sas Mesas-Carchinarzos area

The Lower Cretaceous succession of Sedda Sas Mesas-Carchinarzos (Fig. 1) is interpreted as part of a palaeo-slide emplaced during the Middle Eocene mass wasting episode (Massari & Dieni 2014), possibly favoured by lubricating marly layers of the "Purbeckian facies", and involved in the strike-slip transpressional motion along the Nuoro Fault (Fig. 1). In spite of this complex history, the succession appears to be less disturbed than at the NE corner of the Mt Albo massif. The Berriasian is represented, like elsewhere, by a characteristic interval of commonly laminated and mud-cracked limestones, indicating deposition in restricted environments of very shallow-water and temporarily emergent inter- to supratidal flats with brackish to freshwater influences ("Purbeckian facies"), punctuated by marine episodes recorded by locally abundant nerineids and rudistids, among which *Matheronia rougonensis* Mongin, 1953, and *Hypelasma salevensis* (Favre, 1913). The topmost layers of the Berriasian succession contain the gastropod *Leviathania leviathan* indicating a Late Berriasian age. This interval is overlain by inferred Valanginian heterozoan foraminiferal limestones with *Pseudotextulariella salevensis*, *Meandrospira favrei*, *Pfenderina neocomiensis* (Pfender, 1938), and sponge spicules. In comparison with the S'Ozzastru succession, the lack of any deep-water marly facies is worth noting. The following layers, consisting of marly and locally cherty limestone, yielded abundant serpulid polychaetes, brachiopods [*Musculina sanctaecrucis* (Catzigras, 1948), *Belbekella rotundicosta* (Jacob & Fallot, 1913), *Lamellaerhynchia rostriformis* (Roemer, 1836), *Sellithyris deningeri* Dieni & Middlemiss, 1975, and *Sulcirhynchia renauxiana* (d'Orbigny, 1847); Dieni et al. 1975], echinoids [*Toxaster retusus* (Lamarck, 1816)], common non-rudistid bivalves (described by Dhondt & Dieni 1988), and rare ammonites, among which *Acanthodiscus radiatus* (Bruguière, 1789), zonal marker of the lowermost Hauterivian (Reboulet et al. 2014, and references therein). The Hauterivian succession is wholly comparable to that of the Orosei area described by Dieni & Massari (1963).

The Valanginian succession of Orosei

The Upper Valanginian whitish-pale yellow marls with very rare sponge spicules and radiolarians cropping out in the Orosei area (Badde Funtana Morta and eastern foot of the Mt Tuttavista) are bounded at the base by an unconformity recording a significant hiatus (gap encompassing the Lower Valanginian) and marked by erosion surfaces with epikarst formation. In this case the demise of the photozoan carbonate ramp combines a sequence boundary and a transgressive surface. The marls yielded some ammonites [among which *Olcostephanus* (*O.*) *nicklesi* Wiedmann & Dieni, 1968, sub-zonal marker of the Peregrinus Zone of the middle Upper Valanginian (see Reboulet et al. 2014, and references therein)], allowing the attribution to the Late Valanginian (Wiedmann & Dieni 1968). They reveal close analogies in facies and foraminiferal content with the Upper Valanginian succession of the NE

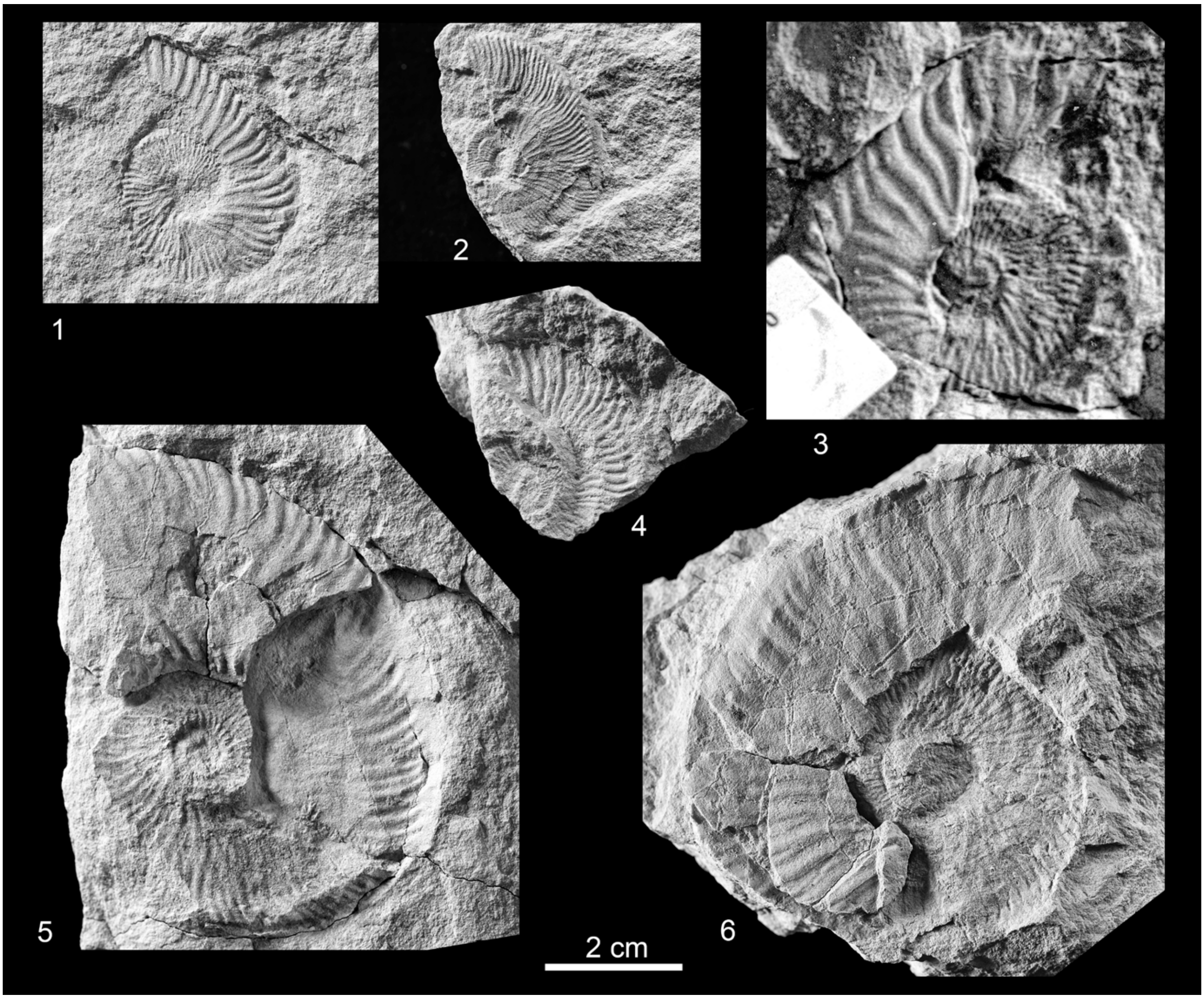

Fig. 7 - Ammonites from the lower Upper Valanginian (Verrucosum Zone) (Siniscola Marl, S'Ozzastru area). 1, 3, 4) *Neocomites platycostatus* Sayn, 1907 (MGP-PD 31531, 31534). 2, 5, 6) *Neocomites neocomiensis* (d'Orbigny, 1841) (MGP-PD 31532, 31535, 31533). The specimens are housed and catalogued in the Geological and Palaeontological Museum of the Padova University (MGP-PD).

Mt Albo area [Dieni et al. 1987; Dieni & Massari 1991, fig. 23 (a,b)] (see the log of fig. 2 in Wiedmann & Dieni 1968), and therefore may be regarded as lithostratigraphically equivalent of the Siniscola Marl.

The very rich foraminiferal assemblage (Dieni & Massari 1966) includes, among other taxa, *Praedorothia praehauteriviana*, *Protomarssonella hechti*, *P. kummi*, *Valdanchella miliani* (Schroeder, 1968) (= *Coskinolina* sp. in Dieni & Massari 1966, p. 109, pl. X, fig. 16), *Lenticulina crepidularis*, *L. guttata* (Ten Dam, 1946), *L. nodosa* (Reuss, 1863), *Citharina seitzi* Bartenstein & Brand, 1951, *Spirillina italica* Dieni & Massari, 1966, and *S. minima* Schacko, 1892.

Calcareous nannofossil assemblage composition of the marls (samples O1 - O5 in the Suppl. Table S1) is compatible with the age assignment based on ammonites. The content, scarce and poorly preserved due to common dissolution and overgrowth, is represented by rare *Cruciellipsis cuvillieri* (Manivit, 1966) Thierstein, 1971, *Cyclagelosphaera margerelii* Noël, 1965, *Diazomatolithus lehmanii* Noël, 1965, *Lithraphidites carniolensis* Deflandre, 1963, *L. pseudoquadratus* Crux, 1981, *Micrantholithus hoschulzii* (Reinhardt, 1966) Thierstein, 1971, *Nannoconus dolomiticus* Cita & Pasquarè, 1959, *N. steinmannii* Kamptner, 1931, *N. wintereri* Bralower & Thierstein in Bralower et al., 1989, *Rhagodiscus asper* (Stradner, 1963) Reinhardt, 1967, *Watznaueria barnesiae* (Black in Black & Barnes, 1959) Perch-Nielsen, 1968 and *Zeugrhabdotus erectus* (Deflandre *in* Deflandre & Fert, 1954) Reinhardt, 1965.

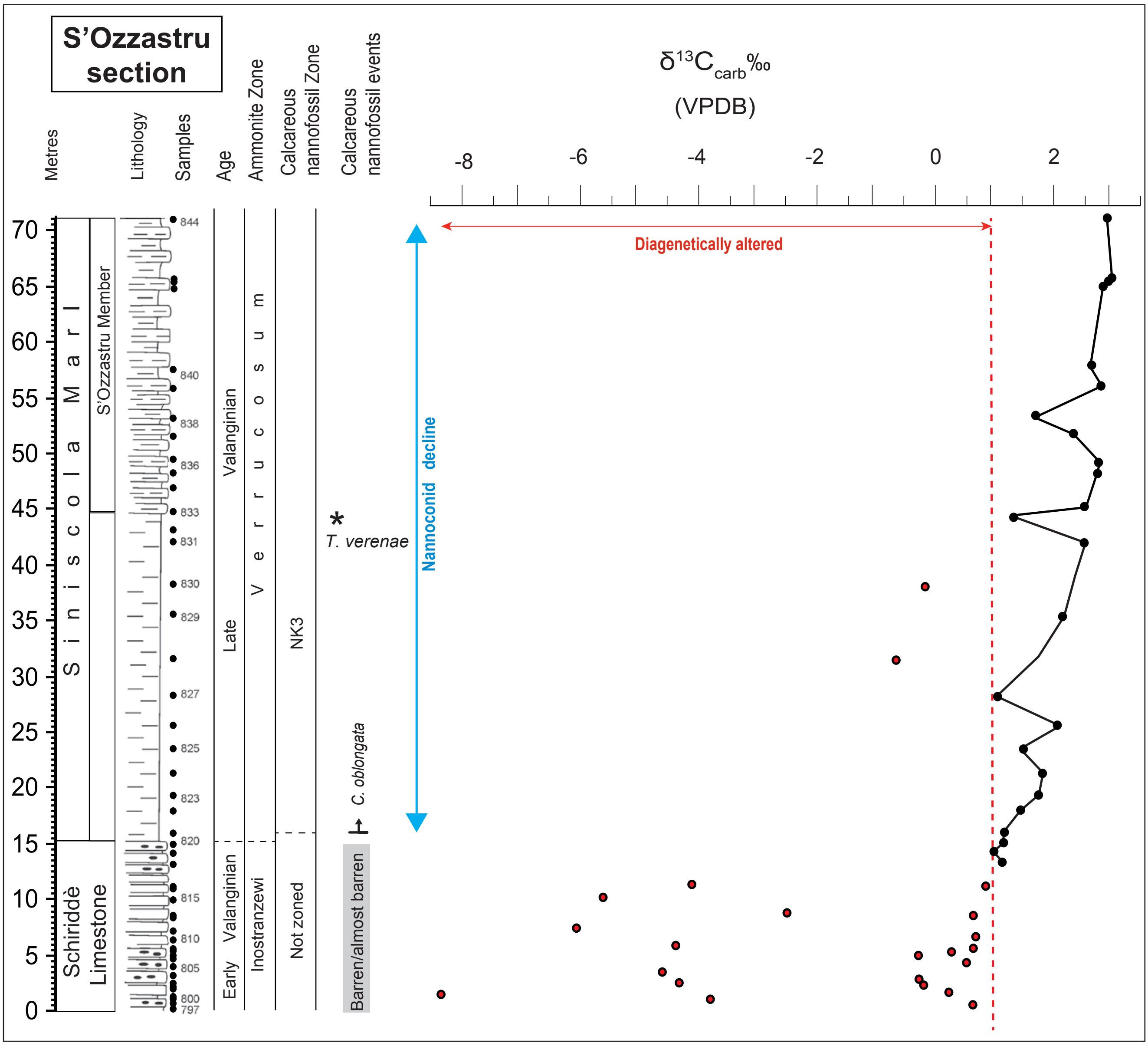

Fig. 8 - Litho-, chemo- and biostratigraphic characterization of the S'Ozzastru section. Calcareous nannofossil zonation after Bralower et al. (1995). *Tubodiscus verenae* was found in sample 833. C-isotope values below 1 ‰ are here considered to be altered by diagenesis.

Calcareous nannofossil biostratigraphy

A total of 52 samples (Fig. 3, Suppl. Table S1) were analyzed for nannofossil biostratigraphy. Specifically, 47 samples are from the S'Ozzastru section, and 5 samples are from the Siniscola Marl of Orosei (samples O1 - O5 in the Suppl. Table S1).

Following the procedure described by Bown & Young (1998) a simple smear slide was prepared for each sample by powdering a few grams of rock with distilled water and mounting the dried suspension with the Norland Optical Adhesive. No ultrasonic cleaning and/or centrifuging were applied in order to retain the original biogenic and inorganic composition. Calcareous nannofossils were investigated with a polarizing light microscope (cross polarized, transmitted light and quartz lamina), at 1250x magnification. For each sample, nannofossils were semi-quantitatively estimated by examining at least 200 fields of view of the smear slide. The assemblages were typified for preservation and abundance as described below and reported in the range chart (Suppl. Table S1):

Preservation

G: good preservation = no evidence of dissolution and/or overgrowth, primary morphological characteristics only slightly altered, most specimens are identifiable to the species level. M: moderate = little evidence of dissolution and/or overgrowth is present, primary morphological characteristics are sometimes altered. MP: moderate/poor = evidence of dissolution and/or overgrowth is present, primary morphological characteristics are sometimes altered, fragmentation has occurred. P: poor = most specimens exhibit dissolution or overgrowth, primary morphological characteristics are sometimes destroyed, fragmentation has occurred.

Abundance

Total abundance was coded as follows: C: common, 10-20 specimens per field of view. CF: common/few, 5-9 specimens per field of view. F: few, 2-4 specimens per field of view. R: rare, 1 specimen per field of view. VR: very rare, less than 1 specimen per field of view.

Abundance of individual taxa was coded as follows: A: abundant, > 1 specimen per field of view. C: common, 1 specimen in 1-9 fields of view. CF: common/few, 1 specimen every 10 fields of view. F: few, 1 specimen in 11-29 fields of view. FR: few/rare, 1 specimen every 30 fields of view. R: rare, 1 specimen in 31-100 fields of view. VR: very rare, less than 1 specimen in more than 100 fields of view.

Stable isotope data

A total of 46 bulk samples has been analyzed for the oxygen and carbon isotope composition. They were drilled with a micro-drill choosing fresh surfaces and avoiding macrofossils or calcite veins. The samples were subsequently analyzed with a Finnigan GasBench II carbonate device connected to a Thermo Fisher Delta V PLUS mass spectrometer. The results were corrected with an internal laboratory standard MS2 (Carrara marble, $\delta^{13}C = 2.16‰$, $\delta^{18}O = -1.85‰$) and reported in the δ-notation relative to VPDB (Vienna Pee Dee Belemnite). The reproducibility of the measurements based on the standard was better than ±0.1‰ for both carbon and oxygen. The instrument is regularly calibrated with the international standards NBS19 and NBS18. The measurements were all carried out at the Geological Institute of ETH Zurich.

Results

Calcareous nannofossils

Throughout the S'Ozzastru section, calcareous nannofossil preservation is moderate to moderate-poor, with signs of overgrowth (Table S1). The most common taxa are illustrated in Fig. 9. Total nannofossil abundance, preservation and single taxa abundance are reported in the Suppl. Table S1. In the S'Ozzastru section, the interval comprised between samples 797 and 820 is characterized by 17 barren samples (797-805, 807-809, 812, 816, 818-820) and 4 almost barren samples (806, 810, 811, 813) which contain very rare specimens of *Watznaueria barnesiae*, *Zeugrhabdotus embergeri*, *Lithraphidites carniolensis*, *Nannoconus steinmannii* (Fig. 8). The barren/almost barren interval belongs to the Schiriddè Limestone. The following interval, corresponding to the Siniscola Marl, is characterized by higher nannofossil abundance, which progressively increases up to the top of the section.

The nannofossil assemblages show relatively low diversity (number of taxa) and the most common species encountered are *W. barnesiae*, *C. margerelii, Calcicalathina oblongata*, and *Z. embergeri*. Nannoconids are few to rare and narrow-canal forms (*N. steinmannii, N. dolomiticus, N. minutus*) are relatively more abundant than wide-canal species (*N. multicadus, N. wintereri*, *N. kamptneri*). Among pentaliths, *Micrantholithus obtusus* and *M. hoschulzii* are detected, being few to rare. Rare specimens of *Lithraphidites* cf. *L. pseudoquadratus* (Fig. 9) have been encountered in three samples (825, 838, 844). Sample 815 shows a relatively high content (common/few) of *L. carniolensis*. *Diazomatolithus lehmanii* is also present throughout the section although rare. Only in sample 833 rare specimens of *Tubodiscus verenae* (Thierstein, 1973) were identified (Fig. 9).

Nannofossil biostratigraphy

The stratigraphic scheme adopted is from Roth (1978) modified by Bralower et al. (1995). The only zonal marker species found at S'Ozzastru is *C. oblongata* (identified from sample 821 up to the top of the section). The First Occurrence (FO) of this species defines the base of Zone NK3 in the Early Valanginian and its extinction occurs at the end of the Hauterivian. *Tubodiscus verenae,* which has the FO in the middle Valanginian (basal part of Zone NK3) and the Last Occurrence (LO) in the latest Valangin-

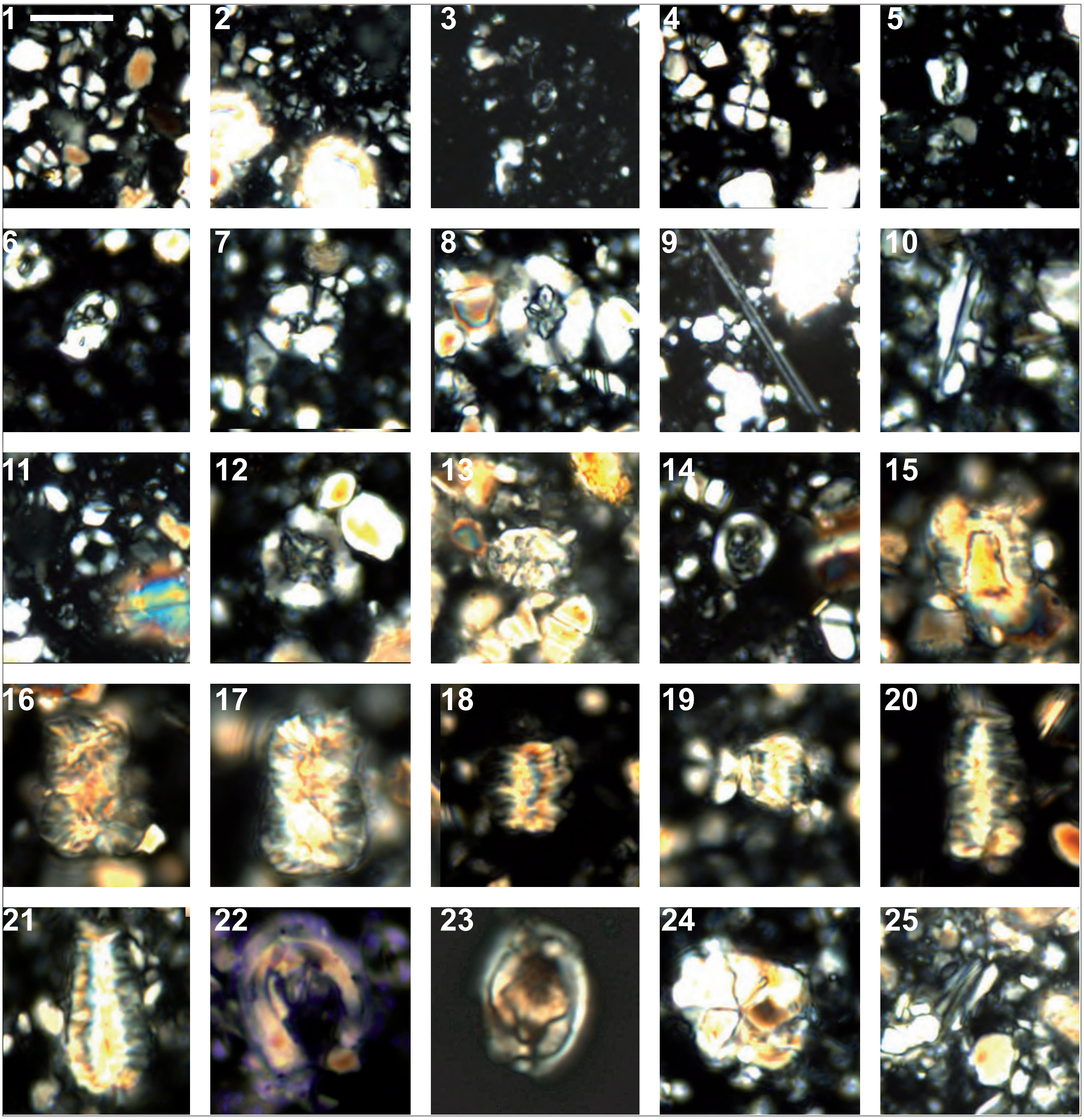

Fig. 9 - Calcareous nannofossils identified in the S'Ozzastru section. 1) *W. barnesiae* (sample 811). 2) *D. rotatorius* (838). 3) *Z. erectus* (838). 4) *C. margerelii* (844). 5) *R. pseudoangustus* (837). 6) *E. windii* (838). 7) *W. britannica* (837). 8) *C. cuvillieri* (821). 9) *L. carniolensis* (815). 10) *Lithraphidites* cf. *L. pseudoquadratus* (838). 11) *D. lehmanii* (844). 12) *Cretarhabdus* sp. (844). 13) *C. oblongata* (838). 14) *R. asper* (844). 15) *N. wintereri* (840). 16, 17) *N. multicadus* (844, 827). 18, 19) *N. minutus* (821, 839). 20) *N. dolomiticus* (828). 21) *N. steinmannii* (837). 22) *T. verenae* (833). 23) *Z. embergeri* (821). 24) *M. obtusus* (838). 25) *C. mexicana* (838). Scale bar (reported in micrograph 1) = 5 µm.

ian (marking the base of Zone NC4), was found very rare in sample 833, as above noted. The stratigraphic position of *C. oblongata* and *T. verenae* within the S'Ozzastru section indicates that the studied interval is comprised within nannofossil Zone NK3. This interpretation is supported by the absence of *Nannoconus bucheri,* which is a relatively common low-latitudinal taxon and has its FO in the latest Valanginian-earliest Hauterivian (Zone NC4) above the LO of *T. verenae* (Fig. 8). The absence of *Rucinolithus wisei* could possibly further constrain the studied interval within subzone NK3b but this species is usually rare in Tethyan sections, and therefore we decided not to use the absence of this taxon to subdivide the Zone NK3.

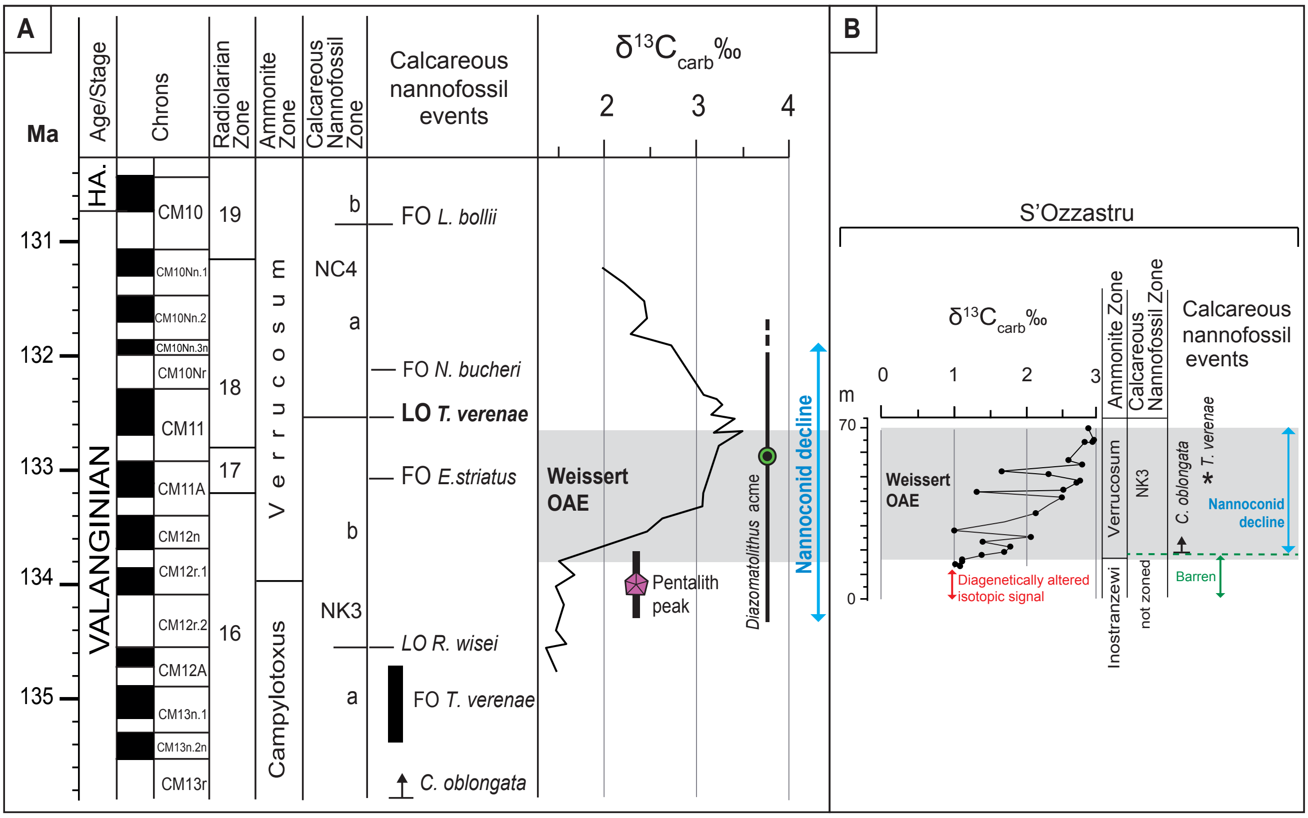

Fig. 10 - A) Bio-chrono-magnetostratigraphic scheme modified from Channell et al. (1995) and Weissert & Erba (2004). Numerical ages are based on time scales of Malinverno et al. (2012). Simplified C isotopic curve is after Weissert et al. (2008). Calcareous nannofossil zones after Bralower et al. 1995. B) Synthesis of S'Ozzastru section stratigraphic dataset compiled in this study. Campylotoxus Zone *auctt.* =Inostranzewi Zone of Reboulet et al. 2014. "Nannoconid decline" is *sensu* Erba and Tremolada (2004).

Stable isotope data

The measured oxygen isotope values range from -1.1 ‰ to -4.7‰, and the carbon isotope values range between -8.2‰ and +2.98‰ (Supplementary Table S2). C-isotope data measured in the lower part of the S'Ozzastru section (0 - 11 m) show high amplitude fluctuations and are marked by repeated shifts to low C-isotope values which do not correspond with measured pelagic carbon isotope values for the Berriasian-Valanginian (e.g. Channell et al. 1993; Hennig et al. 1999; Erba et al. 2004; Kujau et al. 2013; Charbonnier et al. 2013). Consequently, we do not consider C-isotope values below a threshold line of 1 ‰ (see Figs 8 and 10) since they most probably are affected by diagenetic processes. It is worth observing that the change of diagenetic regime occurs at the transition from cherty limestones (Schiriddè Limestone) to marls (Siniscola Marl). The C-isotope trend shows a remarkable positive shift from ca. 1 ‰ around the Schiriddè Limestone/Siniscola Marl boundary towards 2.98‰ at the top of the studied interval. The oxygen isotope data are not further considered in this study as the $\delta^{18}O$ values vary throughout the section and are affected by diagenesis; therefore, they cannot be interpreted as palaeotemperature data.

Discussion

In pelagic records the Weissert OAE corresponds to the nannofossil Zone NK3 (e.g. Bersezio et al. 2002; Erba et al. 2004; Sprovieri et al. 2006; Duchamp-Alphonse et al. 2007; Bornemann & Mutterlose 2008; Mattioli et al. 2014 Greselle et al. 2011; Charbonnier et al. 2013; Morales et al. 2015), and extends from the upper part of the ammonite Campylotoxus Zone *auctt.* through the Verrucosum Zone (e.g. Channell et al. 1993; Hennig et al. 1999; Erba et al. 2004; Greselle et al. 2011). The highest values of the positive anomaly of the Weissert OAE usually coincide or shortly predate the LO of the nannofossil *Tubodiscus verenae* (Lini 1994; Bersezio et al. 2002; Erba et al. 2004; Sprovieri et al.

2006; Duchamp-Alphonse et al. 2007; Bornemann & Mutterlose 2008; Greselle et al. 2011; Mattioli et al. 2014; Charbonnier et al. 2013; Morales et al. 2015) and are recorded in the Verrucosum Zone.

In the S'Ozzastru section, the positive CIE (from ca. 1 to 2.9 ‰) detected within the Siniscola Marl is interpreted to be indicative of the Weissert OAE CIE (Figs 8, 10). In the studied section, the CIE corresponds to the nannofossil Zone NK3 and to the uppermost part of the ammonite Inostranzewi Zone and the Verrucosum Zone. The lowermost 15 metres of the section, within the Schiriddè Limestone, are attributed, on the basis of the presence of *Busnardoites campylotoxus*, to the upper Valanginian Inostranzewi Zone (= Campylotoxus Zone *auctt.*). In the same interval, nannofossils are absent or very rare and no marker species were detected. The carbon isotope values are below 1 ‰, being lower compared to average $\delta^{13}C$ values (1 to 1.5 ‰) recognized in the Lower Valanginian within the Campylotoxus Zone *auctt.* (e.g. Channell et al. 1993; Lini 1994; Hennig et al. 1999; Bersezio et al. 2002; Erba et al. 2004; Sprovieri et al. 2006; Duchamp-Alphonse et al. 2007; Kujau et al. 2012; Greselle et al. 2011; Charbonnier et al. 2013; Kujau et al. 2013) and, consequently, interpreted here to be affected by diagenesis. The altered C-isotope signal in the Schiriddè Limestone prevents the certain identification of the onset of the CIE at S'Ozzastru section. The absence of any decrease in the C-isotope values on top of the studied section, suggests that also the interval following the CIE climax is not recorded. The intense shearing along the Nuoro fault system resulted in the suppression of the upper part of the Valanginian succession, due to the tectonic contact with the Middle Eocene Cuccuru 'e Flores Conglomerate (Pl. 1 and Fig. 2A).

The assemblage composition of calcareous nannofossils agrees with the stratigraphic characterization of the section: the relatively low nannoconid abundance found in the S'Ozzastru section possibly corresponds to the "nannoconid decline" interval (Fig. 10) recognized in several pelagic sequences of the lower Valanginian extending from prior to the CIE to the end of the positive CIE (e.g. Channell et al. 1993; Bersezio et al. 2002; Erba & Tremolada 2004; Erba et al. 2004; Greselle et al. 2011; Melinte & Mutterlose 2001; Duchamp-Alphonse et al. 2007; Bornemann & Mutterlose 2008; Mattioli et al. 2014). In some of these pelagic records, the onset of the "nannoconid decline" coincides with a peak in abundance of pentaliths and a progressive increase in abundance of *D. lehmanii* that became more common during the Weissert OAE (Bersezio et al. 2002; Erba & Tremolada 2004; Erba et al. 2004; Duchamp-Alphonse et al. 2007; Bornemann & Mutterlose 2008). At S'Ozzastru section, the onset of the CIE is not identified. The stratigraphic interval marked by the CIE is instead characterized by *D. lehmanii* and pentaliths but without peaks in abundance.

Sedimentological information suggests that the transition from the Berriasian "Purbeckian facies" to the marly lithology of the Siniscola Marl in the S'Ozzastru section reflects a significant environmental change. A first modification in depositional conditions with probable evidence of increased humidity combined with enhanced freshwater input occurred in the Berriasian "Purbeckian-facies" interval. This fact, together with the first appearance of biosiliceous components (radiolarians and sponge spicules) in the Lower Valanginian chert-bearing Schiriddè Limestone, and predominance of these components in the following Siniscola Marl, suggests that the Weissert Event was preceded by a phase of progressive environmental change starting in the Berriasian and culminating in the Late Valanginian. This indicates that the environmental change occurred in steps and that it started well before the positive CIE. As commonly observed elsewhere (e.g. Kuhn et al. 2005; Chatalov et al. 2015), palaeoenvironmental stress and biotic crisis led to the progressive decline and subsequent demise of the Tithonian-Berriasian carbonate factory, accompanied by a shift from photozoan to heterozoan communities.

The change of sedimentary and diagenetic regime at the transition from cherty limestones (Schiriddè Limestone) to marls (Siniscola Marl) (Fig. 8), marks a significant increase in fine-grained terrigenous input, inferred to be driven by increased runoff and terrestrial weathering resulting in a change in the carbon cycle (e.g. Weissert 1990). The particular abundance of the biosiliceous component in the marls is thought to represent the consequence of enhanced fertility of marine waters triggered by chemical weathering and continental runoff under humid climate conditions (Baumgartner 2013; Celestino et al. 2017).

The lithological evidence, paralleled by the geochemical and micropalaeontological findings, is in agreement with other records (e.g. Channell et al. 1993; Lini et al. 1992; Weissert et al. 1998; van de Schootbrugge et al. 2000; Bersezio et al. 2002; Erba et al. 2004; Föllmi et al. 1994, 2006; Erba & Tremolada 2004; Duchamp-Alphonse et al. 2007; Bornemann & Mutterlose 2008; Barbarin et al. 2012; Charbonnier et al. 2013; Kujau et al. 2013; Mattioli et al. 2014; Martinez et al. 2015).

In addition, the presence of layers characterized by sapropel-like fine lamination in the Upper Valanginian marls probably indicates periodic anoxic conditions. This feature seems to contrast with the absence of evidence for widespread anoxic conditions during the Weisert event in the western Tethys. The only laminated layers which appear to have been formed under anaerobic depositional conditions are the Lower Valanginian "Barrande" levels at Vergol (eastern France), which predate the carbon-isotope excursion (Westermann et al. 2010) and a Valanginian black shale level 68m above the base of the Maiolica formation at the locality Torre de Busi (S. Alps, Italy, Weissert et al. 1979). A change in Valanginian oceanography of the Tethys – Atlantic Ocean is indicated by a change in pelagic sedimentation from white nannofossil limestone to alternating black shale – limestone in the North American Basin, Western Central Atlantic (e.g. Bernoulli 1972).

The change in pelagic facies in the Western North Atlantic precedes the C-isotope excursion and it may indicate a profound change in palaeoceanography. The Valanginian marks the onset of a circumequatorial current system accompanied by strong upwelling (Hotinski & Toggweiler 2003).

Taking into account the probable upper slope depositional setting of the Siniscola Marl, we suggest that intermittent anoxic or suboxic pulses occurred within expanded oxygen-minimum zones in mid-water settings. In general, it may be observed with this regard that anoxic conditions, at least as periodic pulses, are favoured in marginal seas, situated in close proximity to sources of continental organic matter. In this regard, Westermann et al. (2010) stressed the capacity of the Valanginian terrestrial cover to produce a surplus in terrestrial Corg, locally exported into the oceans.

Astronomical calibration of the Valanginian Weissert episode in the succession of the Vocontian Basin (south-eastern France) evidenced that the precession was the main driver of the deposition of the limestone-marl couplets (Charbonnier et al. 2013; Gréselle & Pittet 2010). Taking into account this conclusion, it may be speculated that the alternation of bioturbated and laminated intervals in the marls and of marly limestone and marl in the upper part of the S'Ozzastru section were driven by precession.

Concluding, although the geochemical and micropalaeontological data are somewhat limited by diagenetic alteration at the base of the succession and nannofossil species richness is relatively low, the entire dataset collected for the S'Ozzastru section constitutes an important contribute to the stratigraphic and palaeoceanographic characterization of the Weissert OAE. This is the first evidence of the Weissert OAE in Sardinia which, during the Valanginian, had a peculiar palaeogeographic position, being located south of the Vocontian Basin, between the North Atlantic and the Tethys oceans, where other reference sites for the Valanginian are located (namely Polaveno and La Breggia sections in Southern Alps; Fig. 11). The palaeoposition of Sardinia could explain similarities with lithological facies of the Vocontian Basin and analogies with nannofossil assemblage composition found in the Atlantic sites (DSDP Site 534 and DSDP Site 603) where no peaks in abundance of *D. lehmanii* were identified (Bornemann & Mutterlose 2008), contrarily to the Tethyan sections.

Conclusions

This study allowed the stratigraphic characterization of the Valanginian S'Ozzastru section in the central-eastern part of Sardinia and highlighted, for the first time, the presence of a positive carbon isotope anomaly which corresponds to the Valanginian Weissert OAE.

The isotopic signal of the Schiriddè Limestone of late Early Valanginian age, at the base of the section includes several samples diagenetically altered and calcareous nannofossils are absent or extremely rare. The positive CIE corresponds to the Siniscola Marl of Late Valanginian age, being comprised within nannofossil Zone NK3 and the ammonite Verrucosum Zone, where rare specimens of *T. verenae* were encountered. The end of the CIE is not recorded possibly due to the increasingly tectonized upper part of the succession, near the faulted con-

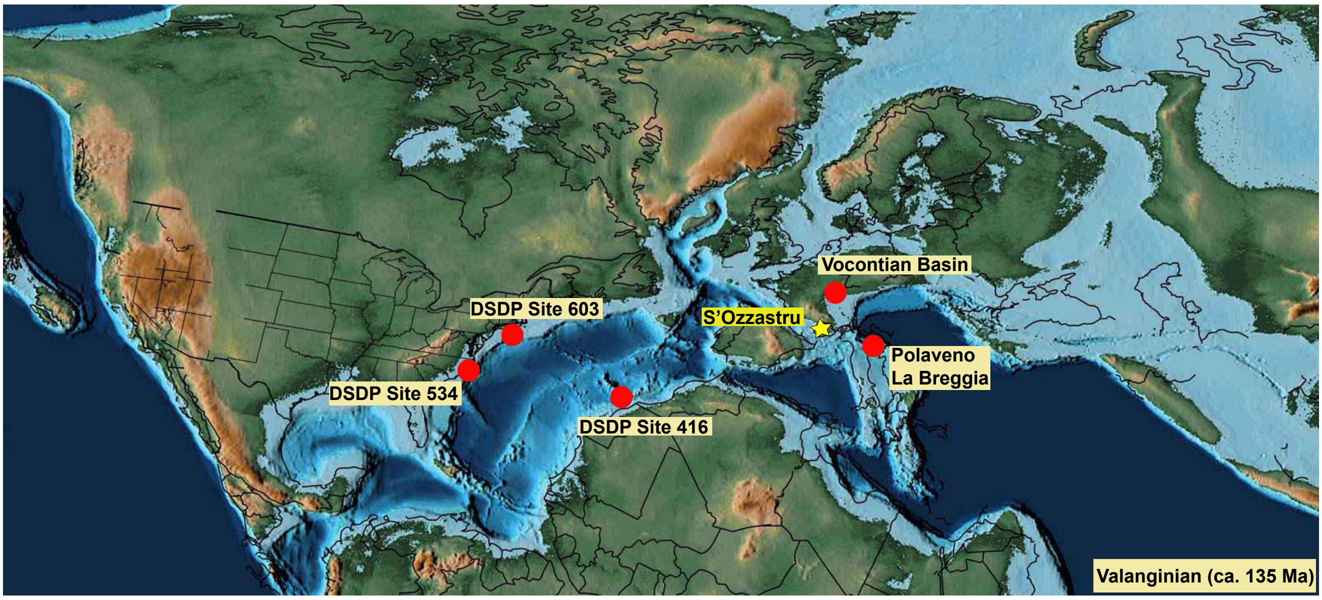

Fig. 11 - Palaeomap at ca. 135 Ma modified from Scotese (2016) PaleoAtlas. Yellow star represents the palaeoposition of the S'Ozzastru section with respect to other reference sites for the Weissert OAE.

tact with the Cuccuru 'e Flores Conglomerate. The nannofossil assemblages detected within the Siniscola Marl are thought to record the "nannoconid decline interval" which globally marks the Weissert OAE in pelagic sequences. The CIE at S'Ozzastru corresponds to a change in lithology towards higher terrigenous content, likely reflecting a shift in the direction of more humid conditions identified in other coeval sections spanning the Weissert OAE. The new data collected for the S'Ozzastru section constitute an important documentation of the Weissert OAE in Sardinia and contribute to frame the Valanginian deposits of Sardinia in a wider palaeogeographic and palaeoceanographic context.

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Taxonomic list of calcareous nannofossils present in the S'Ozzastru section

Biscutum constans (Górka, 1957) Black in Black and Barnes, 1959
Calcicalathina oblongata (Worsley, 1971) Thierstein, 1971
Conusphaera mexicana Trejo, 1969
Cretarhabdus sp. Bramlette & Martini, 1964
Cretarhabdus angustiforatus (Black, 1971) Bukry, 1973
Cruciellipsis cuvillieri (Manivit, 1966) Thierstein, 1971
Cyclagelosphaera margerelii Noël, 1965
Diazomatolithus lehmanii Noël, 1965
Discorhabdus rotatorius (Bukry, 1969) Thierstein, 1973
Eiffellithus windii Applegate & Bergen, 1988
Lithraphidites carniolensis Deflandre, 1963
Lithraphidites cf. *L. pseudoquadratus* Crux, 1981
Micrantholithus hoschulzii (Reinhardt, 1966) Thierstein, 1971
Micrantholithus obtusus Stradner, 1963
Nannoconus dolomiticus Cita & Pasquarè, 1959
Nannoconus kamptneri Brönnimann, 1955
Nannoconus minutus Brönnimann, 1955
Nannoconus multicadus Deflandre & Deflandre-Rigaud, 1959
Nannoconus steinmannii Kamptner, 1931
Nannoconus wintereri Bralower & Thierstein in Bralower et al., 1989
Percivalia fenestrata (Worsley, 1971) Wise, 1983
Rhagodiscus asper (Stradner, 1963) Reinhardt, 1967
Rhagodiscus pseudoangustus Crux, 1987
Staurolithites sp. Caratini, 1963
Tubodiscus verenae Thierstein, 1973
Watznaueria barnesiae (Black in Black & Barnes, 1959) Perch-Nielsen, 1968
Watznaueria britannica (Stradner, 1963) Reinhardt, 1964
Zeugrhabdotus embergeri (Noël, 1959) Perch-Nielsen, 1984
Zeugrhabdotus erectus (Deflandre in Deflandre & Fert, 1954) Reinhardt, 1965

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