Neritic isotope and sedimentary records of the Eocene–Oligocene greenhouse–icehouse transition: The Calcare di Nago Formation (northern Italy) in a global context

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From the Middle Eocene to Early Oligocene, the Earth experienced the most significant climatic cooling of the Cenozoic era. The Eocene–Oligocene transition (EOT) represents the culmination of this climatic cooling, leading to the onset of the Antarctic glaciation and, consequently, to the beginning of the present-day icehouse world. Whereas the response of deep-sea systems to this climate transition has been widely studied, its impact on the shallow-water carbonate realm is poorly constrained. Here, the sedimentary expression of the EOT in two shallow-marine carbonate successions (Nago and San Valentino, northern Italy) belonging to the Calcare di Nago Formation is presented. The chronostratigraphic framework was constructed by integrating litho-, bio-, and isotope-stratigraphic data (C and Sr isotopes), allowing to correlate these shallow-marine successions with pelagic sections in central Italy (Massignano), Tanzania (TDP Sites 12 and 17), and the Indian Ocean (ODP Site 744). Within several sections in northern Italy, including Nago and San Valentino, a Priabonian (Late Eocene) transgression is recorded. Oxygen isotopes of ODP Site 744 show a coeval negative shift of 0.4‰, suggesting a glacio-eustatic origin for this transgression. In the Nago and San Valentino sections, no prominent sequence boundary has been detected that would indicate a rapid sea-level drop occurring together with the positive shift in δ18O defining the EOT-1 cooling event. Instead, a gradual shallowing of the depositional environment is observed. At TDP Sites 12 and 17, the EOT-1 is followed by a negative shift in δ18O of around 0.4‰, which correlates with a relative deepening of the environment in the studied sections and suggests a melting pulse between EOT-1 and the Oligocene isotope event 1 (Oi-1). The positive δ18O shift related to the Oi-1 translates in San Valentino into a change in carbonate factory from a photozoan association dominated by larger benthic foraminifera, corals, and red algae to a heterozoan association dominated by bryozoans. The same bryozoan facies occurs in several Italian localities near the Eocene–Oligocene boundary. This facies is interpreted to represent an analogue of modern cool-water carbonates and results from a cooling pulse of at least regional scale, associated to the Oi-1 event.

1. Introduction

The Eocene–Oligocene transition (EOT) marks the transition between a greenhouse and an icehouse world. In deep-sea records, it is expressed by prominent positive shifts in stable oxygen- and carbon-isotope values (Kennett and Shackleton, 1976; Miller et al., 1991; Zachos et al., 1996, 2001; Coxall and Pearson, 2007; Coxall and Wilson, 2011). The positive shift of oxygen isotopes occurred in two steps: a precursor step (~0.5‰) leading to the EOT-1 event (Katz et al., 2008), and a second, more pronounced step (~1‰) leading to the Oligocene isotope event 1 (Oi-1), also referred to as the Eocene–Oligocene glacial maximum (EOGM). The duration of this shift is estimated at around 500 kyr (Coxall et al., 2005; Coxall and Pearson, 2007; Katz et al., 2008; Miller et al., 2008; Coxall and Wilson, 2011; Wade et al., 2012). An additional but less clear precursor step has been reported only from the St. Stephen Quarry in Alabama (Katz et al., 2008; Miller et al., 2009). The Eocene–Oligocene boundary (Global Boundary Stratotype Section and Point = GSSP) as defined at Massignano (Italy) by the last appearance of the planktonic foraminifer Hantkenina (Nocchi et al., 1988; Premoli Silva and Jenkins, 1993) occurs within the EOT. The GSSP falls within magnetochron C13r but cannot be correlated to any stable isotope event.

While it is widely accepted that large ice sheets first formed in Antarctica during the EOT, there is increasing evidence that smaller transient glaciations already occurred during the Middle and Late Eocene (Tripati et al., 2005; Peters et al., 2010; Dawber et al., 2011). However, fluctuations in oxygen-isotope composition measured in deep-sea sediments do not allow differentiating between changes in temperature, salinity, and ice-volume, and alone do not permit to

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reconstruct in detail the formation of ice-sheets in Antarctica. Shallow-marine successions, on the other hand, where changes in water depth affect sedimentation, can be used to reconstruct local and regional sea-level history. Where such relative sea-level changes occur together with global shifts in oxygen isotopes, they can be interpreted in terms of glacio-eustatic sea-level fluctuations linked to waxing and waning of glaciers and ice sheets. In this respect, neritic deposits have been studied on the New Jersey shelf (Pekar et al., 2002; Miller et al., 2005a, 2009), at St. Stephens Quarry in Alabama (Katz et al., 2008; Miller et al., 2008, 2009; Wade et al., 2012) (Fig. 1A), and in Priabona in northern Italy (Houben et al., 2012) (Fig. 1A and B). However, due to the scarce and often incomplete shallow-water records spanning the EOT, open questions regarding the climatic and oceanographic evolution during this time interval remain.

Neritic carbonates are an excellent proxy to monitor environmental changes: carbonate-producing organisms react sensitively to changes in the environment (e.g., water depth, light intensity, nutrient input, water temperature, hydrodynamic energy). However, the correlation of shallow benthic biozones with stable-isotope, magneto-, and calcareous-plankton stratigraphy from pelagic sections, where the EOT has been characterized, is not free of uncertainties (Brinkhuis, 1994; Brinkhuis and Visscher, 1995; Luciani et al., 2002; Cascella and Dinarès-Turell, 2009; Agnini et al., 2011). One problem is the relatively poor chronostratigraphic resolution attained by benthic foraminiferal
biostratigraphy. In addition, carbonate platforms declined rapidly during the Late Eocene and earliest Oligocene (Adams et al., 1986; Papazzoni and Sirotti, 1995; Kiessling et al., 2003; Nebelsick et al., 2005), thus offering fewer possibilities to constrain the EOT in shallow depositional environments.

Here, we present a dataset integrating lithofacies analysis with biostratigraphic and isotope-stratigraphic (C and Sr) data obtained from two Late Eocene to Early Oligocene shallow-marine sections belonging to the Calcare di Nago Formation (northern Italy, Fig. 1B,C,D). Based on these data a high-resolution correlation with pelagic successions from the Tethys (Massignano), Tanzania (TDP Sites 12 and 17), and the southern Indian Ocean (ODP Site 744) (Fig. 1A) is proposed. The aim of this study is to compare the lithofacies evolution observed in the shallow-marine sections with geochemical and micropaleontological data from the pelagic realm, and to characterize the EOT in the Calcare di Nago Formation. This study is relevant because it tracks the timing of the onset of the Antarctic glaciation in a carbonate neritic depositional environment and reveals how carbonate-producing biota reacted to a greenhouse–icehouse transition.

2. Geological and stratigraphical setting

The studied sections are located near the villages of Nago and San Valentino, on the northeastern side of Lake Garda (northern Italy; Figs. 1B,C,D). Both successions belong to the Calcare di Nago Formation (Castellanir and Cita, 1969a), which was deposited on the western margin of the Lessini Shelf, a Tertiary carbonate platform superimposed on the Jurassic Trento platform (Bosellini, 1989; Luciani, 1989). According to Meullenkamp and Sissingh (2003), the Lessini Shelf was located within the subtropical belt at around 36° to 38°N during the Late Eocene. Sediments were deposited in open-marine conditions and are mainly composed of coralline red algae, larger benthic foraminifera, and corals (Luciani, 1989; Bassi, 1998; Bosellini, 1998).

Within the Nago section, Luciani (1989) recognized two large-scale shallow-up cycles, which in turn are superimposed by several smaller-scale cycles. Based on larger foraminifera biostratigraphy, the lower large-scale cycle (Luciani, 1989; Papazzoni and Sirotti, 1995) was interpreted to have been deposited in the Middle Eocene (Nummulites lyelli, N. biedini and N. variolarius/incrassatus zones, Bartonian), while the upper cycle was attributed to the Late Eocene (Nummulites fabianii s.s. zone, Priabonian). The interval studied in San Valentino was interpreted to have been deposited during the Early and Late Priabonian based on the recognition of the N. fabianii and N. retriatus zones (Castellanir and Cita, 1969a).

Although the Calcare di Nago Formation is mainly composed of shallow-water carbonates, scarce planktonic foraminifera allow refining the biostratigraphy. At the base of the Nago section (10 m; Fig. 2), Morozovella spinulosa was recognized, which indicates zones P14 (Berggren et al., 1995) and E13 (Berggren and Pearson, 2005). The occurrence of M. spinulosa is in agreement with the Bartonian age proposed by Luciani (1989) and Papazzoni and Sirotti (1995) for the lower part of the section.

In the upper part of the Nago section, Turborotalia cunialensis was found at metre 68, around metre 95, and at metre 140 (Fig. 2). T. cunialensis occurs within zones P16–17 (Berggren et al., 1995) and E15–E16 (Berggren and Pearson, 2005) in the latest Eocene. In the Massignano section, which includes the GSSP for the Eocene–Oligocene boundary, the first appearance datum (FAD) of this foraminifer occurs at metre 7.5 (Spezzaferri et al., 2002) and is dated at 35.2 Ma (Jovane et al., 2007), while the last appearance datum (LAD) occurs 40 cm below the Eocene–Oligocene boundary, dated at 33.765 Ma (Berggren et al., 1995). Although these data do not permit the placement in the Nago section of the Eocene–Oligocene boundary as defined in Massignano, they imply that the rocks occurring between metres 68 and 140 were deposited during the latest Eocene.

In San Valentino, orthophragmiums (Discocyclina and Asterocyclina) are present in almost the entire section, in contrast to other larger benthic foraminifera like Nummulites, Pissolinina, Heterostegina, Operculina and Spirorotalia that seem to be much more facies dependent. However, orthophragmiums decrease rapidly above metre 110.8 where tests are often filled with glauconite and abraded. The last debris of Discocyclina and Asterocyclina was found at 114.8 m, implying that these genera disappeared somewhere between 110.8 m and 114.8 m within bryozoan-dominated beds. Therefore, we tentatively place their extinction level at metre 113.5 where the last moderately well-preserved specimens are found. This extinction is followed by an interval of around 2 m containing up to 90% bryozoans (in absence of larger benthic foraminifera and red algae). Above this interval red algae, Nummulites, and Operculina reoccur.

The Eocene–Oligocene boundary in the Lessini Shelf and the Colli Berici areas is marked by the occurrence of bryozoan marls and limestones (Castellanir and Cita, 1969b; Ungaro, 1978; Setiawan, 1983; Barbin, 1988; Trevisani, 1997; Nebelsick et al., 2005). Interestingly, in three sections outcropping in the Priabonian region including the Priabonian type locality (Priabona, Buco della Rana and Bressana; around 40 km to the east of San Valentino), Setiawan (1983) reported the last occurrence of Discocyclina and Asterocyclina at the transition from larger benthic foraminifera to bryozoan dominated deposits (between microfacies units IV and V in Priabona, XI and XII in Buco della Rana and III and IV in Bressana). Similar as in San Valentino this extinction is followed by an interval completely dominated by bryozoans (up to 100%). Above this interval, Operculina and Nummulites reoccur. At the Val D'Ir (15 km north of San Valentino), Castellanir and Cita (1969b) found bryozoan marls and limestones corresponding to the transition between the Calcare di Nago and the Marine di Bolognano formations. Similar to the observations made in San Valentino and the Priabona area (Setiawan, 1983), they also report broken discocyclinids at the base of a bryozoan limestone, which then completely disappear. This implies that at least regionally the extinction of Discocyclina and Asterocyclina occurring together with the establishment of bryozoan beds as observed in San Valentino can be unequivocally used for correlation (Fig. 3).

Brinkhuis (1994) and Brinkhuis and Visscher (1995), by means of dinoflagellate cyst stratigraphy, were able to correlate the shallow-marine deposits of the Priabonian type locality (eastern Lessini Shelf, Fig. 1B) with pelagic sections of central Italy, including Massignano (Fig. 3). Following their correlation, the Eocene–Oligocene boundary as defined in Massignano occurs in the middle of the Priabonian type section. The bryozoan beds occur about 10 m above the transposed Eocene–Oligocene boundary in the upper part of the Gse and the lower part of the Adi dinoflagellate cyst zones (Brinkhuis, 1994). This implies that the bryozoan beds and extinction of discocyclinids in the Priabonian type locality occurred within the Early Oligocene, which is in agreement with observations made in Tanzania where the LAD of Discocyclina shortly postdates the Eocene–Oligocene boundary (Cotton and Pearson, 2011).

Based on the FAD of the characteristic Priabonian larger benthic foraminifera Nummulites fabianii, N. stellatus, and Spirorotalia carpatica, Papazzoni and Sirotti (1995) correlated the transgression occurring at metre 52 in the Nago section to the one recorded at the base of the Priabonian type section and to the one occurring between the Calcare Nummulitici and Marine di Priabona formations at Mossano (Fig. 1B). Luciani et al. (2002) have shown that the first occurrence of N. fabianii at the Calcare Nummulitici–Marine di Priabona transition in Mossano corresponds to the upper part of the planktonic foraminiferal zone P15 (Berggren et al., 1995) and the nanofossil zone NP18 (Martini, 1971) and, therefore, lies within the Late Eocene. Additionally, these authors noticed that 8 m above this transition the rocks are already referable to P16, which characterizes the latest Eocene. Therefore, the existence of a hiatus or of condensed levels at this transition in Mossano seems likely (Luciani et al., 2002).
This evolution resembles that in Nago, where the lower part is referable to the Middle Eocene and the upper part to the latest Eocene. Accordingly, an important stratigraphic gap must be postulated between the lower and upper part of the section (Fig. 2).

The Calcare di Nago Formation is overlain by the Marne di Bolognano Formation. Its base is Early Oligocene in age (Luciani, 1989), belonging to the P18 zone of Blow (1979) and Berggren et al. (1995). The planktonic foraminifer Globigerina tapuriensis occurs within the lower part of the Marne di Bolognano (Luciani, 1989). In the Massignano section, the FAD of G. tapuriensis is 2 m above the Eocene–Oligocene boundary (Coccioni et al., 1988).

3. Methods

The 150 m thick succession of the Nago section was logged and sampled in detail (Fig. 2). A sequence-stratigraphic interpretation was performed based on the principles described in Strasser et al. (1999) and Catuneanu et al. (2009). Thin sections of 210 samples were made for microfacies analysis and biostratigraphy. Six marl samples were wet-sieved, and the residue was analysed under a binocular for biostratigraphic purposes.

San Valentino is a complementary section that was sampled in order to cover the last appearance datum of Discocyclina. a) Miliolid dominated facies (Facies 6, Table 1), b) larger benthic foraminifera and coralline-algal dominated facies (Facies 1–5), and c) bryozoan packstone/grainstone (Facies 8).
missing in the Nago section. The lower part of the section is marly, tectonized, and partly covered. Therefore, the lower part of the section was sampled at a lower resolution (a total of 72 thin sections).

The carbon- and oxygen-isotope composition of 161 bulk samples of the Nago section and 59 of San Valentino was analysed (Fig. 2). Additionally, 27 sub-samples from different carbonate components on seven selected slabs from the Nago section were drilled and analysed in order to test the internal variability of stable isotope signatures within a single sample. Components drilled include coralline algae, discocyclinids, peloids, and micrite matrix. Isotope measurements were performed using a Finnigan MAT Delta Plus XL mass spectrometer equipped with an automated GasBench II at the Institute of Mineralogy and Geochemistry of the University of Lausanne (Switzerland). All results are reported in ‰ relative to the VPDB standard. The analytical reproducibility for three runs in Nago is better than ±0.08 ‰ for both $\delta^{13}C$ and $\delta^{18}O$. In San Valentino, the analytical reproducibility for three runs is better than ±0.1 ‰ for both $\delta^{13}C$ and $\delta^{18}O$.

For Sr-isotope stratigraphy (SIS), four oyster shells were collected from three different stratigraphic levels in the upper part of the Nago section. Samples were selected in the field and then prepared in the laboratory following the methods described in Frijia and Parente (2008) and Boix et al. (2011). The most delicate step when using SIS is to assess if the original marine Sr-isotopic signal is preserved in the samples and not altered by diagenesis. Oysters produce a primarily low-Mg calcite shell. Hence, they are more resistant to diagenetic alteration and have been proven to preserve pristine seawater Sr records in unaltered samples (DePaolo and Ingram, 1985; Jones et al., 1994; Schneider et al., 2009). In order to assess the preservation of the original shell microstructure, the samples were passed through a complete petrographic and geochemical screening by analysing the elemental (Mg, Sr, Mn, and Fe) composition of the shells. The micritic matrix of some samples was also analysed for comparison with the shells to constrain the diagenetic pathways. All geochemical analyses were performed at the Institute for Geology, Mineralogy and Geophysics of the Ruhr-University (Bochum, Germany). Concentrations of Mg, Sr, Fe, and Mn were determined through ICP-AES Thermo Fisher Scientific iCAP6500 Dual View (refer to Boix et al., 2011, for analytical procedures and reproducibility of replicate analyses).
Strontium-isotope ratios were analyzed on a Finnigan MAT 262 thermal-ionization mass spectrometer and normalized to an $^{87}\text{Sr}/^{86}\text{Sr}$ value of 0.1194. The long-term mean of modern seawater (USGS EN-1) measured at the Bochum isotope laboratory is $0.708159 \pm 0.000002$ (2 s.e., $n = 196$). The $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of the samples were adjusted to the value of 0.709175 of the USGS EN-1 standard, to be consistent with the normalization used in the compilation of the look-up table of McArthur et al. (2001; version 4: 08/04), which was used to derive the preferred numerical ages of the analyzed samples and the 95% confidence limits (see Steuber, 2003, for a graphical explanation). This look-up table is tied to the Geological Time Scale of Gradstein et al. (2004).

4. Results

4.1. Sedimentary evolution of the Calcare di Nago Formation

The Calcare di Nago Formation is represented in the Nago area by a 150 m thick limestone succession overlying volcanoclastic deposits. The first 52 m of the section (Figs. 2 and 4A) are characterized by massive beds of well-sorted miliolid- and acervulinid-dominated grainstones, occurring together with small coralgal patch-reefs in the uppermost part of this interval (photozoan association sensu James, 1997; Facies 6 and 7 of Table 1, Fig. 4C). These beds correspond to the uppermost part of this interval (photozoan association sensu James, 1997; Facies 6 and 7 of Table 1, Fig. 4C). These beds correspond to the uppermost part of this interval (photozoan association sensu James, 1997; Facies 6 and 7 of Table 1, Fig. 4C). These beds correspond to the uppermost part of this interval (photozoan association sensu James, 1997; Facies 6 and 7 of Table 1, Fig. 4C).

At metre 52, an abrupt change to facies dominated by larger symbiont-bearing benthic foraminifera and non-geniculate coralline algae is observed. The larger benthic foraminifera and associated corals also imply a photozoan association. However, when compared to the miliolid- and acervulinid-dominated grainstones below, there is a decrease in abundance of primarily aragonitic components, as evidenced by less micritic envelopes. These facies change corresponds to the base of the second large-scale cycle of Luciani (1989), which is characterized by a 40 cm thick interval containing siliciclastic grains. Above, a stepwise large-scale fining of components and an increase in micrite-supported fabrics, together with a thickening of beds, is observed. This evolution is accompanied by a change in dominance of larger benthic foraminifera, passing from a nummulitid/orthophragminid- to an orthophragminid-dominated association around metre 95. Lithologically, this evolution corresponds to an upward decrease in abundance of grainstones and rudstones of the coralline algal debris facies (Facies 5). Above metre 95, the deposits are dominated by thin coralline algal crusts and larger benthic foraminiferal wackestones/floatstones (Facies 2). Micritic/peloidal coral/sponge boundstones (Facies 3) occur intercalated in Facies 2. In the upper part of the section, between metres 125 and 134, coralgal patch-reefs (Facies 7), previously described by Bosellini (1998), are present. Voids within these coralgal framestones are filled with sediment containing abundant small miliolids, dasycladaceans, and gastropods, similar to the facies found below metre 52. From 134 m upwards, thick packages of Facies 5 dominate up to the top of the section.

The evolution of the Calcare di Nago Formation in San Valentino is similar to that observed in Nago, except for the presence of more abundant marly intervals (Fig. 2). The lower part of the section directly overlaying the volcanoclastic deposits was not sampled. The measured section begins at the transition between the miliolid and geniculate red-algal grainstone to non-geniculate coralline algal and larger benthic foraminiferal limestones. Similar to Nago, rocks within this transition contain siliciclastic components. In the middle part of the section (~50 m), sediments are composed of marls and marly limestones with intercalated coralline algal bindstones and extremely flat orthophragminds (Fig. 4D,E). At metre 110, a second major facies change occurs. Red-algal and larger benthic foraminiferal limestones (photozoan association, Fig. 4F) are replaced by bryozoan-dominated grainstones/packstones (heterozoan association sensu James, 1997; Facies 8, Fig. 4G).

4.2. Carbon and oxygen isotopes in Nago and San Valentino

Carbon and oxygen stable isotopes were measured throughout the Nago section (Fig. 2). However, only the Late Eocene interval (above 52 m) is taken into account because biostratigraphic data suggest the presence of a hiatus of uncertain duration around metre 52. The $\delta^{13}\text{C}$ curve can be subdivided into 8 chemostratigraphic segments (Fig. 2). Chemostratigraphic segment-boundaries are determined by turning points from positive to negative shifts.

The lowermost interval (segment C1) starts around metre 53. It is characterized by a negative shift displaying an amplitude of around 0.5‰, leading to a small spike at metre 57. Above this peak, there is a return to pre-shift values (mean + 1.00‰). Segment C2 is characterized at its base by a pronounced negative shift of 1.12‰, starting at around metre 70 and marking the base of a 25 m thick interval characterized by lighter $\delta^{13}\text{C}$ values (Fig. 2). Segment C2 can be subdivided into three sub-segments. Sub-segment C2a is characterized at its base by the negative shift leading to a peak of ~0.24‰ at 72.3 m. This is followed by a rise of the values to +0.42‰. At the base of segment C2b, the curve displays a decrease to ~0.22‰ at 77.7 m, followed by a gradual positive shift reaching a value of +0.66‰. Within this positive shift, segment C2c is characterized by a small negative spike with an amplitude of 0.4‰.

A prominent negative shift of ~1‰ characterizes the lower part of C3, where $\delta^{13}\text{C}$ values of ~0.20‰ are reached, corresponding to the last portion of the 25 m thick negative interval (Fig. 2). From this point, a stepwise shift (~95 to 120 m) to more positive values occurs. This shift is interrupted by a negative spike, defining segment C4. A second peak at the end of the positive trend characterizes segment C5. Segment C6 is defined at the base by the most positive values recorded in Nago (+2.15‰), followed by a rapid but low-amplitude (0.62‰) negative shift. The positive shift recorded above this peak displays a gentle slope leading to stable values (~2.0‰), making a plateau that persists for more than 10 m.

Segment C7 is characterized by a renewed negative $\delta^{13}\text{C}$ shift with an amplitude of ~1.00‰, followed by a plateau at around +1.00‰ up to the top of the Nago section.

Due to the poor outcrop conditions in the lower part of the San Valentino section and the resulting lower sampling resolution, the recognition of individual chemostratigraphic segments is more difficult. However, a negative shift in $\delta^{13}\text{C}$ similar to the one recorded in Nago marking the beginning of C1 occurs below metre 10. This is followed
4.3. Strontium isotope stratigraphy and numerical ages

The first sample analyzed (Sr, Table 2) comes from a bed containing red-algal and larger benthic foraminiferal grainstones of Facies 5 (58 m). The $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of 0.707790 translates into a numerical age of 34.39 Ma ($^{+/-}$ 0.76–0.4), corresponding to the latest Eocene (latest Priabonian). The second dated level at 93 m (Sr 2, Table 2) was sampled within an interval containing grainstones of Facies 5. It corresponds to a level occurring in the lower part of segment C3 of the carbon isotope curve. The Sr-isotopic value of 0.707800 gives a numerical age of 34.14 Ma ($^{+/-}$ 0.64–0.39), again corresponding to the latest Eocene. The third dated level (Sr 3, Table 2) occurs at 142 m within a grainstone of Facies 5, within segment C7 of the carbon isotope curve. The value of 0.707802 translates into an age of 34.09 Ma ($^{+/-}$ 0.59–0.37), corresponding to the latest Eocene, very close to the Eocene–Oligocene boundary placed at 33.9 Ma by Gradstein et al. (2004).

5. Discussion

5.1. Carbon-isotope values

The use of carbon-isotope chemostatigraphic analysis becomes a powerful tool for chronostratigraphic correlations between platform and pelagic sections (Jenkyns, 1995; Vahrenkamp, 1996; Ferreri et al., 1997; Grötsch et al., 1983; Marshall, 1992), aging of water masses within restricted environments (Immenhauser et al., 2002), and changing proportions of aragonite versus low-magnesium calcite within the rocks, as aragonite is naturally enriched in $^{13}$C if compared to calcite precipitated from the same fluid (Swart, 2008).

Sediments in Nago and San Valentino were deposited in open-marine conditions. Therefore, it is believed that no important aging of water masses occurred. The vertical facies evolution in both sections is highly heterogeneous, containing intervals with different proportions of primary aragonitic components (corals), but no correlation between facies and stable isotope trends is observed (Fig. 2). The low covariance ($R^2 = 0.0889$ in Nago and $R^2 = 0.0848$ in San Valentino) between the carbon and the oxygen isotopes in the sections (Fig. 5) can be interpreted as indicating negligible diagenetic overprint (Menegatti et al., 1998; Burla et al., 2008). Therefore, it is interpreted that the
\[ \delta^{13}C \text{ and } \delta^{18}O \text{ values from carbonate bulk rock samples of the Nago and San Valentino sections.} \]

Table 2

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<th>Sr (ppm)</th>
<th>Mg (ppm)</th>
<th>Fe (ppm)</th>
<th>Mn (ppm)</th>
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<th>2 s.e. (10^-6)</th>
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<td>CN-4 M</td>
<td>Na</td>
<td>Nago</td>
<td>matrix</td>
<td>389870</td>
<td>307</td>
<td>4061</td>
<td>260</td>
<td>33</td>
<td>0.707851</td>
<td>0.000006</td>
<td>0.707867</td>
<td>P</td>
<td>0.707867</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

5.2. Strontium isotope stratigraphy

The age of the upper part of Nago section has been constrained by means of Sr-isotope stratigraphy (SIS), a powerful tool for dating and correlating shallow-marine carbonate deposits (Steuber et al., 2005, 2007; Frijia and Parente, 2008; Boix et al., 2011).

Trace element analysis of the oyster samples (Table 2) reveal Mg concentrations in the range of low-Mg calcite shells and Sr concentrations in the range of extant and well-preserved fossils oysters (Schneider et al., 2009). The Mn concentrations for all samples are below the 50 ppm cut-off limit of Jones et al. (1994) for well-preserved ostreids.

Trace element values of the shells contrast with those of the matrix, which yields much higher concentrations of Fe, Mg, and Mn associated with lower Sr (Table 2). Furthermore the \( \delta^{87}Sr/\delta^{86}Sr \) ratio of the matrix is more radiogenic than in oysters from the same level, suggesting that the diagenetic fluids had a higher \( \delta^{87}Sr/\delta^{86}Sr \) signal than the original seawater.

Finally, the pattern of changes of \( \delta^{87}Sr/\delta^{86}Sr \) throughout the section and the consistency of the Sr isotope composition within an individual stratigraphic level (Sr.1, Table 2) argues against diagenetic alteration, which may otherwise lead to unsystematic changes in the Sr isotope composition (McArthur, 1994; McArthur et al., 2004; Steuber et al., 2005; Frijia and Parente, 2008). Consequently, it is considered that all our samples preserve their pristine \( \delta^{87}Sr/\delta^{86}Sr \) composition.

5.3. Platform-to-basin correlation

The closely spaced Nago and San Valentino sections can be correlated by bio-, chemo-, and lithostratigraphy, supporting a coeval evolution of bothsuccessions. Concerning the carbon isotopes, the most complete deep-sea records show that the latest Eocene is characterized by a shift to more negative values (Poag et al., 2003) prior to the positive shift defining the EOT (Fig. 6). This is well recognized within the \( \delta^{13}C \) curves of Massignano, ODP Site 744, and TDP Sites 12 and 17, where a period of 0.5 to 1 myr of relatively stable values is followed by an interval of more unstable values, showing several negative excursions.

The here proposed correlation between our platform sections and pelagic records has been achieved using biostratigraphic data together with Sr-isotope stratigraphy (SIS), a powerful tool for dating and correlating shallow-marine carbonate deposits (Steuber et al., 2005; Steuber et al., 2005; Frijia and Parente, 2008). Consequently, it is considered that all our samples preserve their pristine \( \delta^{87}Sr/\delta^{86}Sr \) composition.
Fig. 6. Tentative chemostratigraphic correlation between the studied shallow-water successions and three deep-water sections in the Tethys realm, the southern Indian Ocean, and Tanzania. A) Bulk rock stable carbon-isotope curves of the Nago and San Valentino sections with chemostratigraphic segments. Black stars mark the position of Sr-samples. Numerical ages derived from strontium isotope stratigraphy are tied to the chronostratigraphic chart of Gradstein et al. (2004) where the Eocene–Oligocene boundary is placed at 33.9 Ma. The occurrence of Morozovella spinulosa (M.s.), Turborotalia cunialensis (T.c.) (white stars), and last appearance datum (LAD) of Discocyclina are shown. Planktonic foraminifera (PF) zones follow Berggren and Pearson (2005), shallow benthic zones (SBZ) Serra-Kiel et al. (1998). B) Carbon-isotope curve of the Massignano section measured on bulk rock (Bodiselitsch et al., 2004; Brown et al., 2009), displaying the position of the Global Stratotype Section and Point (GSSP) defined by LAD of Hantkenina at metre 19 (Nocchi et al., 1988; Premoli Silva and Jenkins, 1993). Also shown is the range of Turborotalia cunialensis (T.c.) after Cocconi et al. (1988), Dinoflagellate cysts (DC) after Brinkhuis and Biffi (1993); Calcareous nannoplankton (CN) zones after Martini (1971); PF zones after (1) Berggren et al. (1995) and (2) Berggren and Pearson (2005). Magnetostratigraphy is after (3) Lowrie and Lanci (1994) and (4) Bice and Montanari (1988). C) Stable carbon-isotope curve of ODP Site 744 measured on benthic foraminifera (Cibicidoides spp., Zachos et al., 1996). Dashed part of the curve corresponds to a 50 cm thick interval which was not sampled because of severe drilling disturbance. Age model (5) is based on conversion to Site 1218 (Coxall and Wilson, 2011). D) Composite carbon-isotope curve of TDP Sites 12 and 17 (measured on planktonic foraminifera Turborotalia ampliapertura) plotted against CN and PF zones and the LAD of Hantkenina (Pearson et al., 2008). LAD of Discocyclina is based on Cotton and Pearson (2010). Numerical ages are according to Berggren et al. (1995) and Berggren and Pearson (2005), placing the Eocene–Oligocene boundary at 33.7 Ma.
At Massignano the transition from the Gse to the Adi zone is found around metre 21 (Brinkhuis and Biffi, 1993), roughly coinciding with the base of magnetostratigraphic C 13 n, which further permits a correlation to ODP Site 744 where magnetostratigraphic data are available (Fig. 6). Numerical ages derived from strontium isotope stratigraphy are not absolute ages but are derived from a look-up table tied to the Geological Time Scale of Gradstein et al. (2004) (see methods chapter), which places the Eocene–Oligocene boundary at 33.9 Ma. Therefore, the numerical ages in Nago are not directly comparable to the ones in the pelagic sections presented in Fig. 6 where the Eocene–Oligocene boundary is dated at 33.7 Ma according to the time scale of Berggren et al. (1995) and Berggren and Pearson (2005). However, taking into account the different ages of the Eocene–Oligocene boundary depending on the timescale used and the error margins of the "preferred age" obtained with SIS (Table 2), the upper interval of the Nago section can be confidently constrained to the latest Eocene and the EOT.

Comparison of the δ13C profiles at Nago and San Valentino with the curves of the deep-sea sites shows a remarkable similarity in the pattern of isotopic fluctuations. Key features used for the correlation include a prominent negative peak (chemostratigraphic segment C3), occurring in the uppermost part of the negative interval (C2a–C3). Following this peak, a prolonged positive shift superimposed by a negative spike corresponding to segment C4 is recognized in all localities. Another key feature to correlate the shallow and basinal sections is a negative shift occurring at the base of C7 followed by a short plateau and a prominent positive shift characterizing the upper part of segment C7 (the positive shift is missing in Nago because the outcrop ends exactly where it would be expected).

The LAD of Discocyclina and other larger benthic foraminifera including a number of species of Nummulites in Tanzania occurs between 102.14 and 99.55 mcd, shortly postdating the Eocene/Oligocene boundary at 102.28 mcd (Cotton and Pearson, 2011). Although in San Valentino the LAD of Discocyclina appears to be somewhat later than the extinction of the genus in Tanzania, it occurs in segment C7 at a similar position relative to the carbon curve, providing additional support for the proposed correlation.

According to bio- and magnetostratigraphy, the Massignano section was deposited coevally with the other sections. However, the correlation of the δ13C curve is not straightforward, most probably due to a moderate diagenetic overprint (Cocioni et al., 2000; Vonhof et al., 2000; Bodiselitch et al., 2004) and/or to locally different oceanographic conditions.

Below metre 13, a prominent negative shift is recorded in Massignano. In Nago, San Valentino, and Tanzania, only a relatively small peak is seen at this position.

A prominent positive carbon-isotope shift has been globally recognized (e.g., in ODP Sites 744 and 1218 and in DSDP Site 522), just above the base of magnetostratigraphic C13n. This shift is not clearly developed in Massignano. The positive shift recorded at metre 20 is produced by only one value (Bodiselitch et al., 2004), which is below the base of C13n. Therefore, it does not correspond to the above mentioned positive shift.

Segment C6 in Nago is represented by a thick (15 m) interval with relatively constant positive δ13C values. The corresponding intervals in TDP Sites 12 and 17 and Massignano do not show this pattern of prolonged constant values. Within the Nago section, this interval corresponds mostly to a coralgal framestone, interpreted to have been deposited at a higher sedimentation rate compared to the rest of the deposits. Therefore, it is likely to represent a local phenomenon.

The carbon-isotope records of the Nago and San Valentino sections show shifts with higher amplitudes than the coeval pelagic records. Similar observations have been made by other researchers who have reported higher amplitudes in shallow-marine sections when compared to pelagic ones (Jenkyns, 1995; Vahrenkamp, 1996; Grötsch et al., 1998; Wissler et al., 2003; Burla et al., 2008; Huck et al., 2011). This has been explained by arguing that shallow waters are less stable in their isotopic composition and more sensitive to a variety of factors including local changes in productivity or organic matter reservoirs that may amplify open oceanic patterns in δ13C.

Although debated, the most common explanation for the globally recognized positive excursion of carbon isotopes around the Oi-1, found also in the upper part of segment C7 in the shallow-marine San Valentino section, is a global increase in export and burial of organic carbon in the deep-sea (Salymo and Zachos, 1999; Coxall and Pearson, 2007; Miller et al., 2009; Coxall and Wilson, 2011).

5.4. Record of the Late Eocene and the EOT sea-level and climate changes in the Calcare di Nago Formation

The platform-to-basin correlation proposed in this study makes it possible to compare palaeoecological and stratigraphic observations made in a shallow-marine depositional system with oceanographic events recorded in the deep-sea realm (Fig. 7).

Based on the changes in relative water depth of deposition as inferred from the facies analysis, a large-scale sequence-stratigraphic analysis of the Nago and San Valentino succession was performed. Large-scale shallowing-deepening trends are interpreted as transgressive-regressive sequences sensu Embry (1993).

Below metre 52 the facies succession is monotonous and dominated by thick, massive packages of shallow-water grainstones of Facies 6 and coral framestones. Based on facies evolution, no large-scale shallowing or deepening trend can be recognized. However, meteoric cements occur in the uppermost 10 m of this interval. Therefore, it can be postulated that at least the upper part represents a shallowing-up. The widespread meteoric cementation occurring within rocks composed of Facies 6 directly overlain by deposits dominated by larger benthic foraminifera and non-geniculate coraline algae (Figs. 2 and 7 and Table 1) implies a marine transgression onto an interval characterized by episodes of subaerial exposure. A similar change in facies is recorded in San Valentino. Therefore, the surface displaying this facies change is interpreted as a transgressive surface superimposed on a sequence boundary. The episodes of subaerial exposure postulated here are interpreted to be related to the regressive phase recognized on top of Bartonian deposits in the western Colli Berici (Ungaro and Bosellini, 1965; Ungaro, 1969) and the Lessini Shelf (Frascari, 1963), where it favoured the establishment of brackish conditions and, in some cases, ended with emersion. Therefore, as proposed by Papazzoni and Sirott (1995), the transgression recorded in Nago and San Valentino is interpreted to represent the so-called Priabonian transgression (Frascari, 1963).

A 0.4‰ negative shift in the oxygen-isotope values at ODP Site 744 (Fig. 7) correlates with segment C1 in Nago. Interestingly, this negative shift coincides with the onset of the Priabonian transgression in Nago and San Valentino. As no major phase in Alpine tectonics is thought to have occurred at this time in the Southern Alps (Doglioni and Bosellini, 1987), it is postulated that the transgression was associated to a glacio-eustatic sea-level rise. It represents therefore additional evidence for the existence of at least medium-scale continental ice sheets before the beginning of the Eocene–Oligocene transition.

Variations of 0.4‰ would translate into ca. 40 m of sea-level change, following the Late Pleistocene δ18O/sea-level calibration (0.11‰/10 m) of Fairbanks and Matthews (1978). However, Miller et al. (2005b) stress that the average isotopic composition of past ice sheets is unknown, and that ice sheets during greenhouse intervals must have had higher δ18O values than in Late Pleistocene time. Melting of ice sheets with higher δ18O values has less influence on the mean ocean δ18O value per unit sea-level change than melting of ice with lower δ18O values (Miller et al., 2005b). Thus, the δ18O fluctuation observed at ODP Site 744 could account for a sea-level rise with an amplitude even higher than 40 m.
Fig. 7. Oxygen-isotope curves of ODP Site 744 (measured on Cibicidoides spp.) and composite TDP Sites 12 and 17 (measured on Turboratalia ampliapertura) showing the range of the EOT and the position of the EOT-1 and Oi-1 events, plotted against the sequence-stratigraphic and lithostratigraphic evolution in the shallow-marine Nago and San Valentino sections. Correlation and comparison is based on carbon-isotope chemostratigraphy (dashed lines, Fig. 6). Sequence-stratigraphical elements are: transgressive surface (TS), transgressive deposits (TD), maximum-flooding zone (MFZ) and regressive deposits (RD). Arrows indicate large-scale facies changes within the sections: (1) miliolid grainstone to larger benthic foraminifera and coralline algal dominated facies (Priabonian transgression in Nago and San Valentino), (2) larger benthic foraminifera and coralline-algal dominated facies (photozoan association) to bryozoan packstone/grainstone (heterozoan association) in San Valentino. Box A: negative δ¹⁸O shift occurring together with a transgression in Nago and San Valentino. Box B: positive δ¹⁸O shift related to the EOT-1 event in Tanzania. Box C: negative δ¹⁸O shift between the EOT-1 and Oi-1 events implying warming and sea-level rise, and leading to transgressive deposits in Nago and San Valentino. Boxes D and E: positive δ¹⁸O shifts resulting in the Oi-1 isotope event in ODP Site 744 and Tanzania, respectively.
Between metres 52 and 95 in the Nago section there is a stepwise decrease in dominance of thick packages of grainstones and rudstones of Facies 5, leading to a thick interval dominated by wackestones and boundstones of Facies 2 and 3. This stepwise change, responding to a decrease in energy of the depositional environment, is interpreted to be the expression of a long-term deepening. This is supported by the change in larger foraminiferal assemblages, passing from a nummulitid- and orthophragmimid- to orthophragmimid-dominated association (Hottinger, 1997). The deepest interval marking the maximum-flooding zone (MFZ) is found around 95 m in Nago. In San Valentino, the MFZ is represented by a marly interval (around metre 50) characterized by loose boundstones constructed by thin red-algal crusts (<1 mm thick) and the dominance of very flat orthophragmimids (Figs. 2 and 4D,E). Studies dealing with recent TDP Sites 12 and 17 (Step 1, Pearson et al., 2008) has a duration of a regressive trend.

In San Valentino does not show a clear facies pattern as in they represent the uppermost interval of the regressive deposits. Facies 6, which suggest a shallow water depth (Hallock, 1979; Hallock and Glenn, 1986; Hottinger, 1997; Beavington-Penney and Racey, 2004; Cosovic et al., 2004), supporting a low energy, relatively deep environment of deposition in San Valentino.

The interval above the MFZ, representing the regressive deposits, is characterized by a stepwise shallowing of the environment ending with the establishment of massive coralligenous facies between metres 125 and 134 in Nago (Facies 7, Table 1). Smaller coral communities involved in the construction of boundstones are found throughout the Nago section within Facies 2 and 3. However, they differ in taxonomic composition, growth forms, and percentage and proportion within the framework (Bosellini, 1998), as well as in their matrix, which suggests deeper environments of formation. The coralligenous facie between metres 125 and 134 are interpreted represent the shallowest facies within the Priabonian part of the Nago section. Sediments filling voids in the coralligenous facies contain abundant small milolids, dasyycladaceans, and gastropods (Facies 6), which suggest a shallow depositional environment. It is here interpreted that the coralligenous facies could grow because falling sea level placed the depositional environment within the inner and most proximal middle ramp. Therefore, they represent the uppermost interval of the regressive facies evolution in San Valentino does not show a clear facies pattern as in Nago, therefore it cannot be used to support the occurrence of this regressive trend.

The positive oxygen-isotope shift leading to the EOT-1 as seen in TDP Sites 12 and 17 (Step 1, Pearson et al., 2008) has a duration of at least 100 kyr and develops between C4 and C6 (Fig. 7; missing in ODP Site 744 due to core distortion). Interestingly, in Nago it does not correlate with a sharp sequence boundary but falls within an interval characterized by a stepwise shallowing of the depositional environment between the MFZ and the coralligenous facies. It is probable that this shallowing in Nago is related to a lowering of sea level caused by waxing of continental ice sheets, as evidenced by the positive oxygen-isotope shift leading to the EOT-1 in Tanzania. This gradual lowering of sea level could also explain why it has until now been difficult to locate a sequence boundary related to this event in other localities (Miller et al., 2009). The positive oxygen-isotope shift corresponds to a temperature drop of 2.5 °C (Katz et al., 2008; Miller et al., 2009) and a sea-level fall of 20 to 30 m in St. Stephens Quarry (Alabama) and Priabona (Italy), respectively (Katz et al., 2008; Houben et al., 2012).

Following the EOT-1, a ~0.4% positive 818O shift occurring at the transition between C6 and C7 is recorded in Tanzania (Fig. 7; missing in ODP Site 744 due to core disturbance). This shift closely coincides with a relative deepening of the environment in Nago. The deepening is recorded above the coralligenous facies, which suddenly stopped growing at around 134 m and is replaced by rocks of Facies 2 and 5 deposited on the middle and outer ramp. Given that the negative oxygen-isotope shift and the deepening trend in Nago seem to be coeval, it is here postulated that a small deglaciation pulse must have occurred between the EOT-1 and Oi-1.

A eustatic sea-level fall is interpreted to have occurred around the Oi-1 (see Miller et al., 2005, for discussion). Based on backstripping on the New Jersey shelf, a sea-level drop between 55 and 105 m is assumed (Kominnz and Pekar, 2001; Miller et al., 2005a). Houben et al. (2012) suggest that a sea-level fall of 50 to 60 m, recorded on top of the bryozoan beds in Priabona, was related to the Oi-1. However, no physical evidence for a sequence boundary is found in the corresponding interval in the Nago Formation.

In the pelagic records, the positive 818O shift leading to the Oi-1 occurs at the end of the negative δ13C isotope plateau of segment C7, prior to the positive shift of carbon-isotope values (Fig. 7). At San Valentino, this position is found at the base of the bryozoan limestones (Fig. 4G), which roughly coincides with the LAD of Discocyclina (Fig. 2). Unfortunately, the outcrop in Nago ends exactly where the bryozoan beds would be expected. However, exactly the same evolution is observed at other localities in the Lessini Shelf, where bryozoan marts and limestones comprising the level with the extinction of discocyclinids occur just above the Eocene–Oligocene boundary (Fig. 3, Castellarin and Cita, 1969b; Ungaro, 1978; Setiaewan, 1983; Barbin, 1988; Brinkhuis, 1994; Trevisani, 1997). The fact that this change in carbonate factory occurred coevally at several localities and that it was accompanied by the extinction of discocyclinids, suggests that it corresponds to an important change in environmental conditions. Bryozoan-dominated deposits can occur below the photic zone (Nebelsick et al., 2005) and this environmental change could thus have been related to a sea-level rise. However, these beds appear simultaneously at several localities in Italy and at different palaeo-water depths. In addition, a simple deepening cannot be responsible for the extinction of discocyclinids, which were important carbonate producers throughout the Eocene and are present in virtually all samples in San Valentino and Nago below the bryozoan beds.

Based on Mg/Ca ratio of benthic foraminifera from St. Stephens Quarry, Wade et al. (2012) propose that a global cooling of ~2 °C occurred around the Oi-1. Quantitative analyses of warm-oceanic vs. cool-oceanic dinocyst distribution in the Massicore resulted in a qualitative sea-surface temperature proxy for the EOT (Houben et al., 2012; Massicore was drilled ca. 100 m away from the Massigano section; Montanari et al., 1994). This proxy shows that a cooling maximum occurred within the upper part of the Gse dinocyst zone, which coincides with the lower part of the bryozoan beds in Priabona (Brinkhuis, 1994) and, consequently, in San Valentino (Fig. 3). According to our stratigraphic scheme this corresponds to the interval of the Oi-1 and thus provides additional evidence for a cooling of ocean waters in the Mediterranean region.

Most of the modern bryozoan carbonates are being produced in cool-water environments (e.g., Mediterranean and southern Australia; James et al., 1997; Pedley and Carannante, 2006; Betzler et al., 2010). Taking into account that a cooling has been reported to occur during the Oi-1, it seems reasonable to interpret the deposition of the bryozoan beds as being mainly driven by a cooling of ocean waters. However, Hafar et al. (2006) described an example of modern bryozoan-dominated deposits forming under eutrophic conditions in warm waters, showing that eutrophic conditions can favour the establishment and dominance of heterotrophic organisms. As minor reductions in surface temperature can represent a several-fold increase in trophic resources (Hallock et al., 1991), it is probable that the relative cooling was accompanied by an increase in nutrient input into shallow environments due to changes in ocean circulation. Moreover, Miller et al. (2009) argue that the formation of a large ice sheet in Antarctica resulted in major perturbations of the thermohaline circulation and ocean upwelling, which in turn could have been responsible for nutrient input into the shallow-marine settings. Thus, although the formation of bryozoan-dominated deposits is here interpreted to be mainly related to cooling of the ocean waters occurring at the Oi-1, it is probable that
nutrients also played a role in the formation of bryozoan facies and in the extinction of oligotrophic larger benthic foraminifera.

6. Conclusions

We present for the first time an integrated biostratigraphic, lithological, and geochemical study of shallow-water carbonates spanning the Late Eocene–earliest Oligocene in the Tethys realm. Calibrated by strontium-isotope stratigraphy and biostratigraphic tie-points, a carbon-isotope stratigraphic correlation between the two studied shallow-water sections (Nago and San Valentino) and three pelagic sections from the Tethys (Massignano), southern Indian Ocean (ODP Site 744), and Tanzania (composite TDP Sites 12 and 17) is proposed. This correlation allows to precisely locate important cooling and glacio-eustatic events occurring at the Eocene–Oligocene greenhouse-icehouse transition and to study their expression in shallow-marine environments.

The Priabonian transgression identified in Nago and San Valentino coincides with a 0.4‰ negative oxygen-isotope shift at ODP Site 744. As no important phase in Alpine tectonics is documented at this time in the Southern Alps, the transgression is interpreted to be associated to a deglaciation pulse producing a glacio-eustatic sea-level rise. This observation confirms previous studies suggesting that smaller transient glacialities in Antarctica occurred already during the Middle and Late Eocene.

The positive oxygen-isotope shift leading to the EOT-1 event is represented in the studied sections by a gradual shallowing of the environment rather than a sharp sequence boundary, which could indicate a gradual build-up of continental ice sheets. The EOT-1 event is followed by a negative shift of around 0.4‰ that corresponds to a relative deepening of the environment in Nago, suggesting a melting pulse between the EOT-1 and the Oi-1.

The positive δ18O shift leading to the Oi-1 in pelagic records translates in San Valentino into a facies change from red-algal and larger benthic foraminiferal limestones to bryozoan limestones. These bryozoan beds containing the LAD of Discoscina are also found in other Italian localities just above the Eocene–Oligocene boundary. They are interpreted to result from a regional or global cooling pulse occurring at the Oi-1 and are therefore believed to be an analogue of modern cool-water carbonates.

The paucity of documented shallow-marine carbonate deposits spanning the EOT may be partly due to the difficulties to precisely date Late Eocene–Early Oligocene successions. In this sense, a way to overcome this problem and to obtain insight into shallow-marine processes during this important interval of palaeoecological and palaeoceanographic changes is the use of a multi-proxy approach combining facies analysis, chemostratigraphy, and biostratigraphy as proposed in this study.

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