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Late Eocene impact-induced climate and hydrological changes: Evidence from the Massignano global stratotype section and point (central Italy)

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ABSTRACT

The Eocene-Oligocene transition marks the passage from "greenhouse" conditions to an "icehouse state" with progressive global cooling starting in the early middle Eocene. The late Eocene is also characterized by a high concentration of extraterrestrial impacts, the effects of which, on living organisms and climatic changes, are still not understood. We carried out a high-resolution investigation on planktonic foraminiferal assemblages in an 8-m-thick segment of the Massignano global stratotype section and point for the Eocene-Oligocene boundary with the aim of assessing the effects that the impacts may have had on the environment and this group of organisms. The studied interval is punctuated by three late Eocene iridium-rich layers, several cosmic signatures, and enhanced levels of ^{3}He. The two lower closely spaced iridium anomalies are possibly linked to the Popigai and Chesapeake Bay impact events, respectively, whereas no particular impact event can be assigned to the third anomaly, even if it might be correlated with some large craters. Interpretation of data suggests that all the impacts had no abrupt, dramatic effects on planktonic foraminifera. However, acting as forcing factors, they induced some environmental perturbations and may have contributed to remarkable climate changes superimposed on the general late Eocene cooling trend. The Popigai and Chesapeake Bay impact events triggered significant changes in the water mass structure, in terms of stratification and trophic resources, associated with some climatic excursions that took place within chron C16n.1n and chron C15r and at the transition between planktonic foraminiferal zones P15 and P16.

The short-term warming pulse recognized after the Popigai impact might have been due to greenhouse effects produced by injection of CO_2 into the atmosphere and/or the release of methane hydrate after the impact itself. The dynamic between

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hydrological and climate changes across the impactoclastic layers as observed at Massignano displays different features at each impact event, probably due to the context in which each occurred in terms of impactor size, location, and target rocks. The relatively long duration of the enhanced cooling following the Chesapeake Bay impact suggests that this event induced a progressive cooling and triggered a feedback mechanism that sustained the initial impact-induced changes. Similar patterns of climatic excursions reported worldwide across the equivalent impact-ejecta horizons indicate that the impact-induced climate changes recorded at Massignano appear to be global in extent.

Keywords: planktonic foraminifera, impacts, hydrological changes, climate changes, late Eocene, Massignano, central Italy.

INTRODUCTION

The transition from Eocene to Oligocene marks one of the most important steps in Earth's climate history during the Cenozoic, when the "greenhouse" world of the early Eocene changed into the "icehouse" conditions of the early Oligocene (Miller et al., 1991; Zachos et al., 2001, for context). This transition was characterized by global cooling, changes in oceanic circulation, and stepwise faunal and floral turnovers and extinctions, together with a major tectonic reorganization and intensive volcanism (e.g., Boersma and Premoli Silva, 1991; Prothero, 1994; Coxall et al., 2000, 2005; Zachos et al., 2001; Prothero et al., 2003; Wade and Palïke, 2004; Tripati et al., 2005; Coxall and Pearson, 2007). The late Eocene (37.2–33.7 Ma) was also marked by a high concentration of impact craters, ejecta layers, and an elevated flux of interplanetary dust particles (e.g., Farley et al., 1998; Montanari and Koeberl, 2000). The two largest craters of the Cenozoic Era, Popigai (northern Siberia), 100 km in diameter, and Chesapeake Bay (offshore southeastern Virginia), 85 km in diameter, formed almost simultaneously at ca. 35 Ma and were accompanied by several others essentially isochronous small craters (e.g., Wanapitei and Mistastin, Canada, and Logoisk, Belarus). These craters may have been part of the same event (e.g., Koeberl et al., 1996; Bottomley et al., 1997; Montanari et al., 1998; Poag et al., 2004; Koeberl, 2007). Enhanced levels of ^{3}He clearly indicate that these impacts were related to an ~2.5-m.y.-long comet or asteroid shower, which produced impact debris (Farley et al., 1998; Tagle and Claeys, 2004; Claeys, 2007; Farley, 2007).

This multiple impact event is recorded in the Massignano section (northeastern Apennines, central Italy), the currently accepted global stratotype section and point (GSSP) for the Eocene-Oligocene boundary (Premoli Silva and Jenkins, 1993). Three Ir-rich layers occur, respectively, at 5.61 m stratigraphic level from the base (msl), from 6.15 to 6.18 msl, with a maximum value at 6.17 msl, and from 10.24 to 10.30 msl, with a maximum value at 10.28 msl (Montanari et al., 1993; Bodiselitsch et al., 2004), and they are associated with major spikes of extraterrestrial ^{3}He (Farley et al., 1998). The lowest Ir-rich layer contains high concentrations of shocked quartz grains (Clymer et al., 1995, 1996; Langenhorst and Clymer, 1996) and Ni-rich spinel crystals (Pierrard et al., 1998), and it has been interpreted as an impactoclastic horizon dated as old as 35.7 ± 0.4 Ma (Montanari et al., 1993). According to Bodiselitsch et al. (2004), the Ir anomaly at 5.61 msl is possibly linked to the Popigai impact, dated as old as 35.7 ± 0.2 Ma (Bottomley et al., 1997), and the Ir anomaly at 6.17 msl might be derived from the Chesapeake Bay impact, which has an inferred age of 35.2 ± 0.3 Ma to 35.5 ± 0.3 Ma (Poag et al., 2004). No particular impact event can be assigned to the highest Ir anomaly layer, even if it might be correlated with some large craters (e.g., Wanapitei, Mistastin, and Logoisk). The late Eocene multiple impact event may have played an important role related to the deterioration of the global climate at the end of the Eocene Epoch (Coccioni et al., 2000; Vonhof et al., 2000; Spezzaferri et al., 2002), even if the effects on the climate and on the biosphere are still not understood. The Massignano GSSP, which has well-established integrated stratigraphy and geochronology available for the entire section, provides a unique opportunity to explore whether the late Eocene impact events had any effect on the biosphere and climate with regard to short- and long-term changes. In this study, we focus on the late Eocene 4–12 msl segment of the Massignano section and investigate the response of planktonic foraminiferal assemblages to impact-induced environmental perturbations based on quantitative analyses and statistical treatment of data.

THE MASSIGNANO GSSP AND THE 4–12 MSL SEGMENT

The Massignano GSSP is exposed at 43°32′13″N, 13°35′36″E, in an abandoned quarry located along the provincial road of the Conero Regional Natural Park. The section is named after the nearby village of Massignano and is located at ~4 km northwest of the town of Sirolo, along the Adriatic coast of central Italy (Fig. 1). The marine sedimentary sequence at Massignano section spans the interval from the latest Eocene to the earliest Oligocene and belongs to the Umbria-Marche sedimentary succession, which represents the northwestern sector of the Neotethys Ocean. During the Eocene-Oligocene transition, in this region, the sediments were deposited in an upper to lower bathyal setting at a paleodepth of 1000–1500 m (Coccioni and Galeotti, 2003).

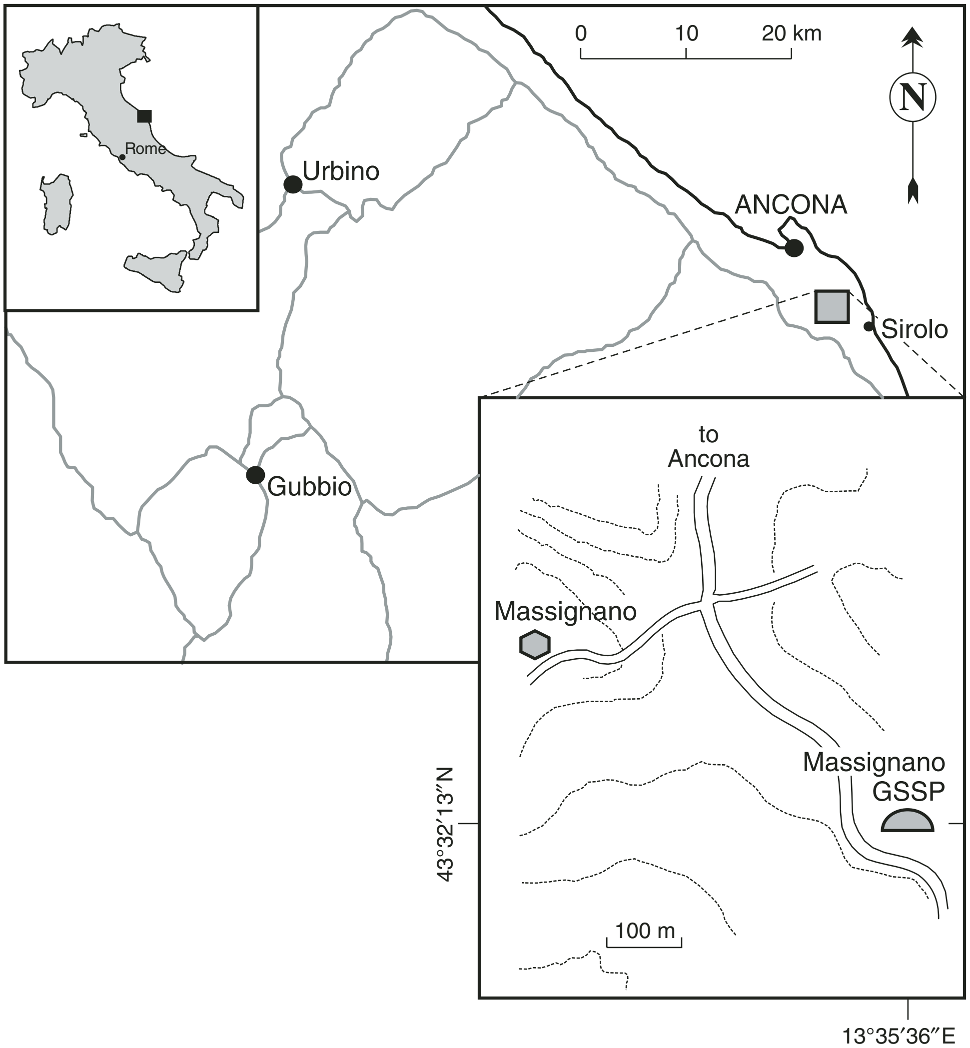

Figure 1. Location map of the Massignano global stratotype section and point (GSSP) for the Eocene-Oligocene boundary.

The exposed continuous and complete sequence is 23 m thick and consists of reddish and greenish-gray hemipelagic marls, calcareous marls, and marly limestones, rich in well-preserved benthic and planktonic microfossils. The sediments span the upper part of the Scaglia Variegata Formation (12 m thick) and the lower part of the Scaglia Cinerea Formation (11 m thick). In the lower to middle part of the section, they are intercalated with several biotite-rich volcano-sedimentary layers (Coccioni et al., 1988) (Figs. 2 and 3). In particular, the 4–12 msl segment contains five biotite-rich volcaniclastic layers at 5.25, 5.8, 6.5, 7.25, and 7.75 msl, respectively (Figs. 2, 4, and 5). Two reddish intervals interrupted by greenish-gray intervals occur in the studied segment. The sediments show evidence of only minor amounts of bioturbation (Coccioni et al., 1988), which appears limited to ~15–20 cm of vertical dispersion (Coccioni et al., 2000; Huber et al., 2001).

The Massignano GSSP spans the calcareous nannofossil zones NP18 to NP21 of Martini (1971) and CP15a to CP16 of Okada and Bukry (1980), the planktonic foraminiferal zones P15 to P18 of Blow (1969, 1979) and Berggren et al. (1995) and zones E14 to O1 of Berggren and Pearson (2006), the benthic foraminiferal zones BB4 to BB5 of Berggren and Miller (1989), and the dinocyst zones *Melitasphaeridium pseudorecurvatum* (Mps) to *Areosphaeridium diktyoplokus* (Adi) of Brinkhuis and Biffi (1993) (Coccioni et al., 1988, 2000; Brinkhuis and Biffi, 1993; Spezzaferri et al., 2002; Monechi et al., 2000) (Fig. 2). The Eocene-Oligocene boundary occurs at 19 msl where the planktonic foraminiferal *Hantkenina* and *Cribrohantkenina*, Eocene genera of the Hantkeninidae, both become extinct (Coccioni et al., 1988; Nocchi et al., 1988a; Premoli Silva and Jenkins, 1993).

The Massignano GSSP spans the interval from chron C16n.2n to chron C13n (Bice and Montanari, 1988; Lowrie and Lanci, 1994; Jovane et al., 2004, 2006, 2007, and this volume) (Fig. 2). K-Ar and Rb-Sr dating of biotite-rich layers in central Italian magnetostratigraphic sections (Montanari et al., 1988)

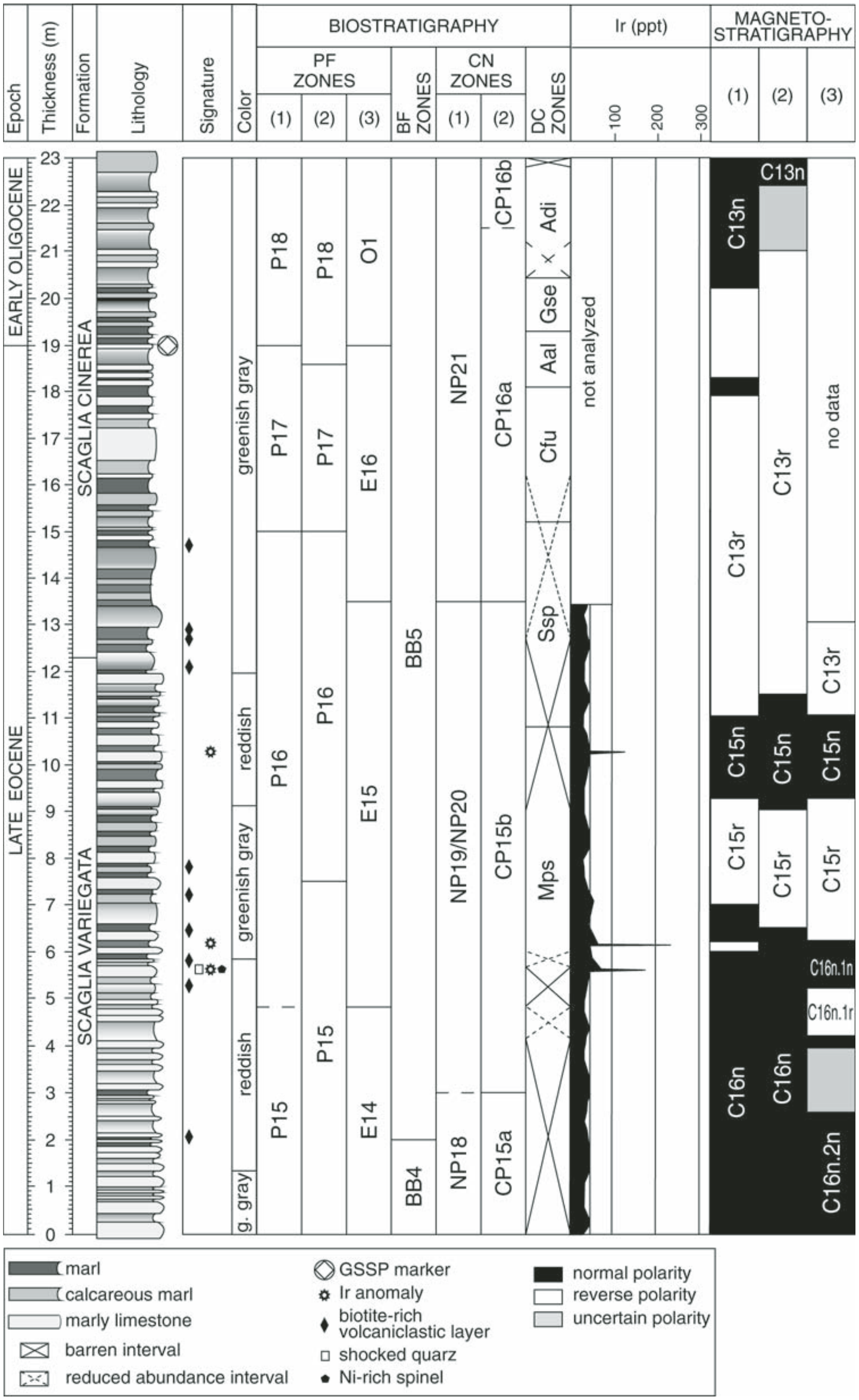

Figure 2. Synoptic stratigraphic scheme of the Massignano global stratotype section and point (GSSP). Lithostratigraphy is after Coccioni et al. (1988) and Jovane et al. (2004, this volume). Biostratigraphy is after Coccioni et al. (1988), Brinkhuis and Biffi (1993), Monechi et al. (2000), Spezzaferri et al. (2002), Coccioni and Galeotti (2003), and Van Mourik and Brinkhuis (2005). Different biozonal schemes are reported according to their authors. Planktonic foraminifera (PF): (1) Blow (1969, 1979), (2) Berggren et al. (1995), and (3) Berggren and Pearson (2006). Benthic foraminifera (BF): Berggren and Miller (1989). Calcareous nannoplankton (CN): (1) Martini (1971) and (2) Okada and Bukry (1980). Dinoflagellate cysts (DC): Brinkhuis and Biffi (1993). Signatures are after Coccioni et al. (1988), Montanari et al. (1988, 1993), Clymer et al. (1995, 1996), Langenhorst and Clymer (1996), Pierrard et al. (1998), and Jovane et al. (2004). Iridium profile is after Montanari et al. (1993) and Bodiselitsch et al. (2004). Magnetostratigraphy is after (1) Bice and Montanari (1988), (2) Lowrie and Lanci (1994), and (3) Jovane et al. (2004, 2007, this volume).

Figure 3. Outcrop photo of the 4–12 m stratigraphic level from the base (msl) segment at Massignano. Numbers correspond to the meters of the measured section. Asterisks mark locations of the impactoclastic layers at 5.61, 6.17, and 10.28 msl.

gave an estimated age of 33.7 ± 0.5 Ma for the Eocene-Oligocene boundary, which has been incorporated as a calibration point in the latest polarity time scales (Cande and Kent, 1992, 1995; Berggren et al., 1995). Additional results from ^{40}Ar/^{39}Ar dating of biotites (Odin et al., 1991) and U/Pb dating of zircons and monazites (Oberli and Meier, 1991) were combined with the earlier data to yield revised mean ages of 35.4 ± 0.3 Ma, 34.5 ± 0.3 Ma, and 34.2 ± 0.2 Ma for the biotite horizons at 7.25, 12.9, and 14.7 msl, respectively. A new magnetostratigraphic data set, combined with the results of the astronomical tuning of the Massignano section (Jovane et al., 2006), allowed researchers to refine the dates for the bio- and chronostratigraphic events previously recorded throughout the succession and to place the Eocene-Oligocene boundary at 33.7 Ma, 0.2 m.y. younger than in the time scale of Gradstein et al. (2004) (Jovane et al., 2007). Based on the magnetostratigraphy provided by Jovane et al. (2007) and the numerical ages from Cande and Kent (1995), the mean sedimentation rate of the 4–12 msl segment at Massignano is 0.73 cm/k.y. Accordingly, this segment would span ~1.1 m.y. (Figs. 4 and 5).

MATERIALS AND METHODS

High-resolution sampling of the 4–12 msl segment at Massignano GSSP was conducted at intervals of 5–10 cm, collecting a total of 87 samples. Samples were prepared using standard techniques for preparation of washed residues from marls, calcareous marls, and marly limestones. Samples were disaggregated with hydrogen peroxide and rinsed through a 40 µm sieve. The washed residues were additionally soaked in Desogen for several hours and then immersed in an ultrasonic bath for a few minutes. The final washed residues were dry-sieved into three different size fractions: 40–150 µm, 150–250 µm, and >250 µm. The residues were studied with a binocular microscope for their planktonic foraminiferal content. The 40 µm mesh sieve was used to retain very small species, such as small acarininids, chiloguembelinids, tenuitellids, and pseudohastigerinids, which have important paleoceanographic implications (e.g., Premoli Silva and Boersma, 1988; Spezzaferri, 1995; Pearson et al., 2006).

Quantitative analyses of planktonic foraminiferal assemblages were performed on the same samples previously used for biostratigraphy, following the standard method of Haq et al. (1977), and applied by Premoli Silva and Boersma (1989), Spezzaferri and Premoli Silva (1991), and Spezzaferri (1995). Generic and specific concepts adopted here are those of Premoli Silva and Boersma (1989), Spezzaferri (1994), and Pearson et al. (2006). Planktonic foraminiferal species were then assembled, whenever possible, in groups displaying phylogenetic affinities and environmental significance as reported in Premoli Silva and Boersma (1988) and Spezzaferri (1995). Grouping was necessary because the scarcity of specimens of some species sometimes prevented a correct paleoclimatic interpretation. Species and group abundances were then transformed to percentages of the total planktonic foraminiferal fauna, and percentage curves were constructed. Abundance data of planktonic foraminifers were double-square-root transformed (no standardization, no further species reduction) in order to limit the contribution of the most abundant, ubiquitous species with respect to the less abundant species and to simplify the interpretation of the data structure (Field et al., 1982).

Multivariate statistical analysis was performed with the software PAST version 1.67. The data were used in hierarchical agglomerative clustering based on Bray-Curtis similarity (Clifford and Stephenson, 1975). On the basis of the same similarity matrix, samples were ordered by nMDS (non-metric multidimensional scaling; Kruskal, 1977). The nMDS plot has no dimensions and no axes and can be arbitrarily scaled, located, rotated, or inverted; it gives the relationship of samples relative to each other.

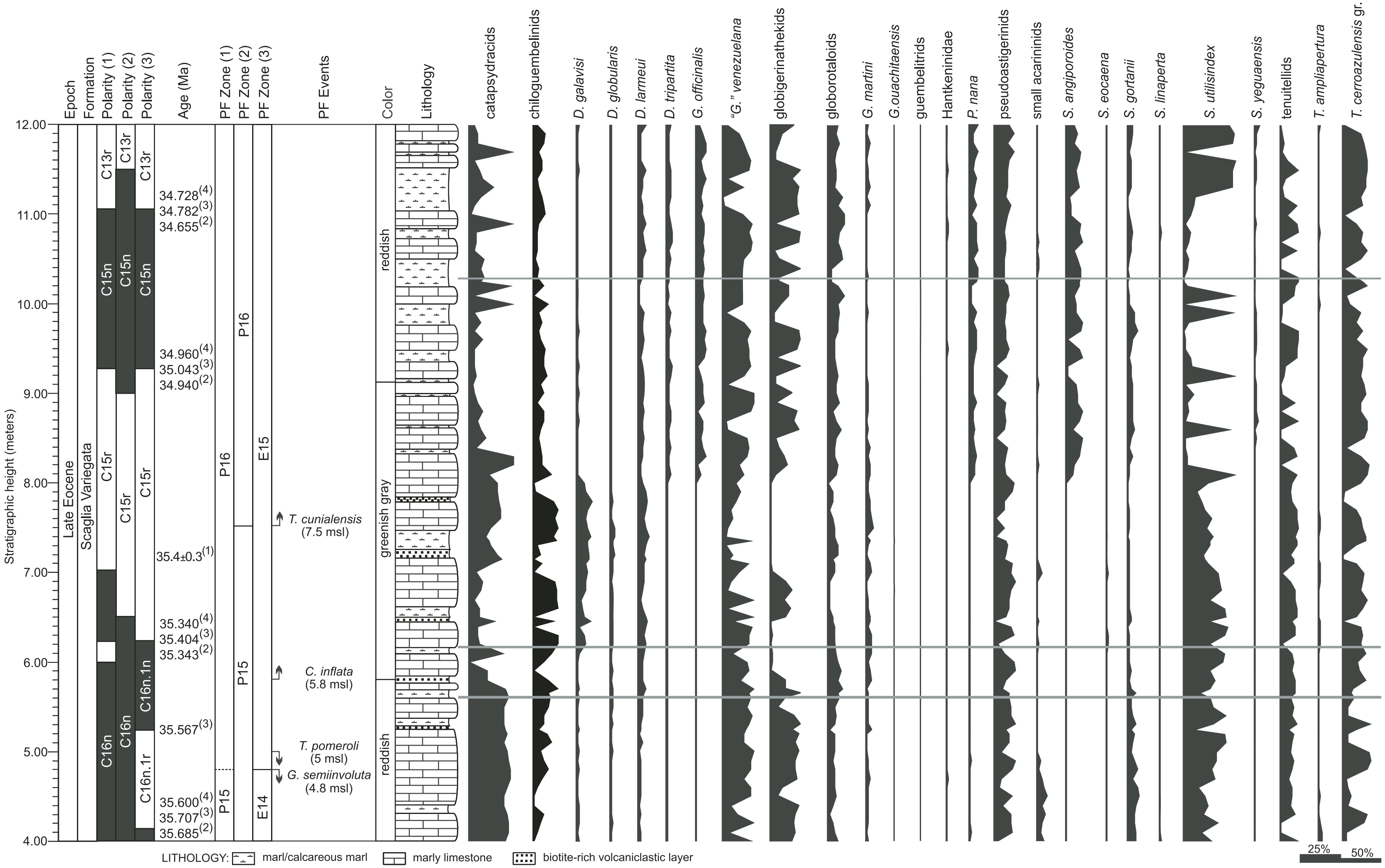

Figure 4. Stratigraphy of the 4–12 m stratigraphic level from the base (msl) segment at Massignano plotted against the relative abundances of planktonic foraminiferal species and groups. Magnetic polarity and planktonic foraminiferal (PF) zones are as in Figure 2. Numerical ages are after (1) Montanari et al. (1988), (2) Cande and Kent (1995), (3) Gradstein et al. (2004), and (4) Jovane et al. (2006). Thin gray lines mark the impactoclastic layers.

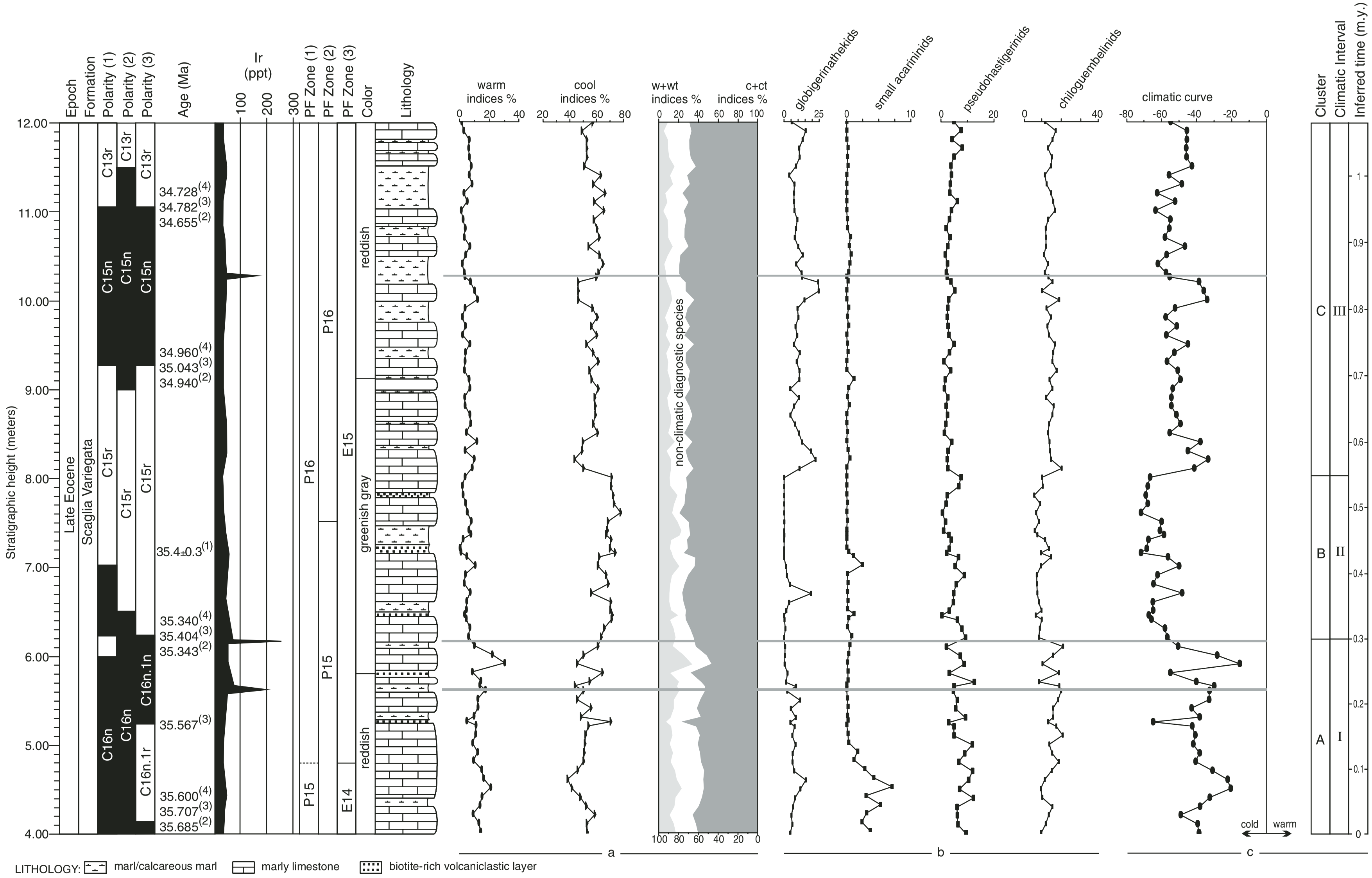

Figure 5. Stratigraphy of the 4–12 m stratigraphic level from the base (msl) segment at Massignano plotted against: (A) the percentage curves of warm and cool indices, warm + warm-temperate, and cool + cool-temperate indices; (B) abundance percentage curves of globigerinathekids, small acarininids, pseudohastigerinids, and chiloguembelinids; (C) climatic curve, clusters, and intervals with inferred time. The cool–cool-temperate index curve is reversed to emphasize the fluctuations in abundance of the non-climatic diagnostic species. Magnetic polarity, iridium profile, and planktonic foraminiferal (PF) zones are as in Figure 2. Numerical ages are as in Figure 4. Thin gray lines mark the impactoclastic layers.

RESULTS

Planktonic Foraminiferal Assemblages and Biostratigraphy

Planktonic foraminiferal assemblages are relatively well preserved, rich, and well diversified throughout the 4–12 msl segment (Fig. 4; Table 1). Specimens often display a slight calcitic overgrowth or infilling, which prevented reliable oxygen and carbon isotopes analyses; consequently, our observations are only based on assemblages.

The 4–12 msl segment is dominated by catapsydracids (*Catapsydrax dissimilis*, *C. unicavus*), chiloguembelinids, "*Globigerina*" *venezuelana*, globigerinathekids (*Globigerinateka index*, *G. luterbacheri*, *G. semiinvoluta*, *G. subconglobata*, *G. tropicalis*), globorotaloids (*Globorotaloides euvariabilis*, *Globorotaloides* sp. 1), pseudohastigerinids, *Subbotina utilizindex*, tenuitellids, and *Turborotalia cerroazulensis* group (*T. cerroazulensis*, *T. cocoaensis*, *T. cunialensis*, *T. pomeroli*), which occur throughout (Fig. 4; Table 1). *Globoturborotalita ouachitaensis*, Hantkeninidae (*Cribrohantkenina inflata*, *Hantkenina alabamensis*), *Subbotina eocaena*, *S. linaperta*, *S. yeguaensis*, *Turborotalia ampliapertura*, and guembelitrids occur discontinuously and in very low numbers. *Subbotina utilizindex* and *T. ampliapertura*, which according to Pearson et al. (2006) have their first occurrence (FO) in zone E15, are already present within the uppermost part of zone E14. *Dentoglobigerina tripartita*, *Globigerina officinalis*, *Paragloborotalia nana*, and *Subbotina angiporoides* are continuously present only from the upper half of the studied segment.

The bioevents identified in the studied segment are (from bottom to top): (1) last occurrence (LO) of *Globigerinatheka semiinvoluta* (4.8 msl), (2) LO of *Turborotalia pomeroli* (5 msl), (3) FO of *Cribrohantkenina inflata* (5.8 msl), and (4) FO of *Turborotalia cunialensis* (7.5 msl) (Coccioni et al., 1988, 2000; Spezzaferri et al., 2002) (Fig. 4).

The 4–12 msl segment spans the uppermost part of zone P15 to the upper part of zone P16 of Blow (1969, 1979), the upper part of zone P15 to the middle part of zone P16 of Berggren et al. (1995), and the uppermost part of zone E14 to the lowermost part of zone O1 of Berggren and Pearson (2006) (Figs. 2, 4, and 5).

Paleoenvironmental Remarks

The paleoclimatic significance and the paleoecology of Eocene planktonic foraminifera are generally deduced from their latitudinal abundance patterns, stable isotopic paleobiology, and from comparison of these parameters to those of modern counterparts.

The paleoecology (preferred biogeographical distribution and life position in the water column) and the paleoclimatic significance of the recognized species and groups are based on the abundant literature, largely following Douglas and Savin (1978), Boersma et al. (1979, 1987), Keller (1983), Poore and Matthews (1984), Nocchi et al. (1988b), Premoli Silva and Boersma (1988, 1989), Stott et al. (1990), Barrera and Huber (1991, 1993), Boersma and Premoli Silva (1991), Spezzaferri and Premoli Silva (1991), Pearson et al. (1993, 1997, 2001, 2006), Spezzaferri (1995), Van Eijden and Ganssen (1995), Pearson and Palmer (1999), Coxall et al. (2000), and Wade and Kroon (2002).

According to the aforementioned literature, we consider the following to be warm-water indices: the small acarininids, *Globoturborotalita martini*, Hantkeninidae, and the *T. cerroazulensis* group. As cool-water indices, we consider catapsydracids, *Globigerina officinalis*, globorotaloids, low-spired subbotininds (*S. angiporoides*, *S. eocaena*, *S. linaperta*, *S. utilizindex*), "*Globigerina*" *venezuelana*, and tenuitellids. We regard globigerinathekids, *P. nana*, and *T. ampliapertura* as cool temperate-water indices, and as warm-temperate water indices, we consider *Dentoglobigerina galavisi*, *D. larmeui*, and *D. tripartita*. Preferred geographic distribution and habitat and paleoclimatic significance of planktonic foraminiferal species and groups identified at Massignano are summarized in Table 2, and Figure 5 shows the abundance curves of the paleoclimatic indicators listed here.

The curve of the warm-water indices shows values ranging from 0.7% to 30.1%, decreasing upward. Positive peaks of 20.9%, 30.1%, and 12.1% occur at 4.5, 5.9, and 10 msl, respectively. The curve of the cool-water indices can be subdivided into three intervals: the first one from 4 to 6.1 msl, with average values around 52% and positive peaks of 70% and 64%, respectively, at 5.25 and 5.8 msl; the second one from 6.2 to 8 msl, with average values around 68%; the third one from 8 msl to the top of the studied segment, with average values around 56% (Fig. 5). The abundance curves of the warm and warm-temperate indices and of the cool and cool-temperate indices show a pattern similar to that obtained for the warm- and cool-water indices (Fig. 5). The abundance curve of the globigerinathekids shows higher numbers in the upper half of the studied segment, with average values increasing from 4% to 11%. Values range from 0% to ~25% at 10.1 msl (Fig. 5). The curve of the small acarininids shows values ranging up to 7.1%. Relatively higher abundances are observed from the bottom of the studied segment up to 4.9 msl, with a peak at 4.5 msl. From here upward, the small acarininids are absent or occur with very low numbers (Fig. 5). The curves of pseudohastigerinids and chiloguembelinids show marked fluctuations from the bottom of the studied segment up to 6.2 msl and an evidently opposite trend in the interval between the impactoclastic layers related to the Popigai and Chesapeake Bay impact events (Fig. 5).

Climatic Curve

The climatic curve is derived from the algebraic sum of percent abundance of warm (positive) and cool (negative) indices as proposed by Cita et al. (1977) and applied by Spezzaferri and Premoli Silva (1991) and Spezzaferri (1995). In the biotite-rich volcaniclastic layers, only minor abundances of warm-water indices are observed (Fig. 5). Therefore, we consider the planktonic foraminifera–based curve to be reliable.

The climatic curve throughout the 4–12 msl segment allows us to recognize three discrete climatic intervals (CI-I to CI-III, from bottom to top) separated by two major shifts: the relatively warmest CI-I interval extends from 4 to 6.1 msl, with an average value of −38.1%, and shows two significant warmer pulses in the intervals from 4.4 to 4.7 msl and from 5.8 to 6 msl and two major cooler pulses at 5.25 and 5.8 msl; the relatively cooler CI-II extends from 6.2 to 8 msl, with an average value of –63.4%; the relatively warmer CI-III extends from 8.1 msl up to the top of the studied segment, with an average value of −50.6%, and shows two significant warmer pulses in the intervals from 8.1 to 8.4 msl and from 10 to 10.2 msl (Fig. 5).

The first two closely spaced impact events fall within the CI-I, and the last impact lies within the CI-III interval. Approximately 5.5 k.y. after the first impact (Popigai), some warmer conditions than those directly below the impact itself are recorded by the sample at 5.65 msl. Subsequently, climatic episodes alternate between, from bottom to top: (1) a cooling episode, up to 5.8 msl, therefore lasting for ~20 k.y., (2) a marked warming pulse, lasting for ~27 k.y., between 5.8 and 6 msl, and (3) a further cooling episode, lasting for ~23 k.y. from 6.0 msl up to the second impact (Chesapeake Bay). The relatively cooler CI-II interval that followed the Chesapeake Bay impact lasted for ~247 k.y. The last impact falls within a cooling episode occurring from 10 to 10.4 msl and lasting for ~55 k.y. About ~16 k.y. after this impact, a subsequent return to less cool conditions occurs that extends up to the top of the studied segment (Fig. 5).

Statistical Treatment of Data

Since patterns of community structures are often not readily apparent (Clarke and Warwick, 1994), we conducted statistical analyses of our data to better identify and characterize changes in planktonic foraminiferal assemblages, and to relate them to changing environmental conditions following the impact events.

At the 84% level of Bray-Curtis similarity, three clusters (A, B, and C in Figs. 6 and 7) were recognized. At the same level of dissimilarity, the differences among clusters were identified by the average abundance of species and/or groups. Cluster A differs from cluster B by higher percentages of chiloguembelinids, globigerinathekids, small acarininids, *Subbotina gortanii*, pseudohastigerinids and *T. cerroazulensis* group and lower percentages of catapsydracids, *D. galavisi*, and *Dentoglobigerina larmeui*. Cluster B differs from cluster C by higher percentages of catapsydracids, *D. galavisi*, *D. larmeui*, "*G.*" *venezuelana*, *S. utilizindex*, and lower percentages of chiloguembelinids, *G. officinalis*, globigerinathekids, *P. nana*, *S. angiporoides*, tenuitellids, and *T. cerroazulensis* group. Cluster C differs from cluster A by higher percentages of catapsydracids, globigerinathekids, globorotaloids, tenuitellids, *S. angiporoides*, and *P. nana*, and lower percentages of small acarininids, *S. utilizindex*, and *T. cerroazulensis* group.

The planktonic foraminiferal assemblages of cluster C can be considered relatively cooler than those of cluster A but relatively warmer than those of cluster B (Figs. 6 and 7). In fact, cluster A shows the lowest abundance of cool species and concurrently the highest abundance of warm species, such as small acarininids and *T. cerroazulensis* group. Moreover, cluster B has the highest abundances of catapsydracids, "*G.*" *venezuelana*, and *S. utilizindex* and the lowest abundance of *T. cerroazulensis* group and can be considered to be the coolest cluster of the studied segment.

The clusters recognized through the statistical treatment of our data fit well with the climatic intervals documented by means of the climatic curve (Figs. 5, 6 ,and 7). Remarkably, the line IMPACT 1, which corresponds to the Ir anomaly at 5.61 msl (Popigai impact), allows us to subdivide cluster A into subclusters A1 (pre-impact 1) and A2 (postimpact 1 to pre-impact 2), with only one exception, the sample found at 5.65 msl (Fig. 7). Subcluster A2 differs from subcluster A1 by lower abundances of catapsydracids, globigerinathekids, and small acarininids, and higher abundances of *D. larmeui*, globorotaloids, and *T. cerroazulensis* group. The most significant feature recognized through the multivariate statistical analysis in the nMDS plot is the line IMPACT 2, which corresponds to the Ir anomaly at 6.17 msl (Chesapeake Bay impact) (Fig. 7). This line clearly separates cluster A from cluster B, which correspond to CI-I and CI-II, respectively. The multivariate statistical analysis shows that significant changes in planktonic foraminiferal assemblages did not take place across the stratigraphical interval, including the Ir anomaly at 10.28 msl (line IMPACT 3) (Fig. 7).

DISCUSSION

High-Resolution Impact Stratigraphy, Geochronology, and Magnetochronology at Massignano: Implications for Global Correlation

On the basis of the available data from the Massignano GSSP, we provide here an updated magnetobiostratigraphic, geochronologic, and magnetochronologic constraint for the late Eocene impactoclastic layers at Massignano that will be useful for global correlation of equivalent horizons.

In terms of biostratigraphy, both the Ir anomalies at 5.61 and 6.17 msl, linked respectively to the Popigai and Chesapeake Bay impact events, fall in the uppermost part of planktonic foraminiferal zone P15 of Berggren et al. (1995), which is in the lowermost part of E15 of Berggren and Pearson (2006) and in the middle part of calcareous nannofossil zone NP19/20 of Martini (1971) (Figs. 2, 4, and 5). The Ir anomaly at 10.28 msl lies in the middle part of planktonic foraminiferal zone P16 of Berggren et al. (1995), which is the upper part of zone E15 of Berggren and Pearson (2006) and calcareous nannofossil zone NP19/20 of Martini (1971) (Figs. 2, 4, and 5).

As for magnetostratigraphy, different data are available. Following Bice and Montanari (1988) and Lowrie and Lanci (1994), the Popigai and Chesapeake Bay impacts fall within the uppermost part of chron C16n, whereas following Jovane et al. (2007), they are more precisely constrained within the upper half of chron C16n.1n. The last impact consistently lies within chron C15n (Figs. 2, 4, and 5).

TABLE 1. ABSOLUTE ABUNDANCES OF PLANKTONIC FORAMINIFERAL SPECIES AND GROUPS IN SAMPLES FROM THE 4–12 MSL SEGMENT AT MASSIGNANO

Stratigraphic level from base (m)	Chiloguembelinids	Catapsydracids	*D. galavisi*	*D. globularis*	*D. larmeui*	*D. tripartita*	*G. officinalis*	"*G.*" *venezuelana*	Globigerinathekids	Globorotaloids	*Gt. martini*	*Gt. ouachitaensis*	Guembelitrids	Hantkeninidae	*P. nana*	Pseudohastigerinids	Small acarininids	*S. angiporoides*	*S. eocaena*	*S. gortanii*	*S. linaperta*	*S. utilizindex*	*S. yeguaensis*	Tenuitellids	*T. ampliapertura*	*T. cerroazulensis* gr.	Total
12	112	112	10	3	35	20	91	103	52	50	5	0	0	0	46	48	0	42	4	30	2	42	26	92	11	12	948
11.9	162	127	1	0	7	5	29	22	144	39	3	0	0	4	17	72	0	56	0	24	0	72	10	106	0	26	926
11.8	143	123	0	0	8	9	43	39	121	52	15	0	0	4	22	39	0	40	0	18	2	64	5	112	0	46	905
11.7	127	85	1	0	13	8	43	51	104	35	5	0	0	4	45	78	2	80	0	10	2	102	0	101	0	54	950
11.6	146	113	3	0	13	13	16	67	106	25	1	0	0	0	28	48	1	62	0	19	0	85	6	128	1	62	943
11.5	144	142	10	2	41	17	31	61	83	38	7	0	1	10	23	37	2	52	0	23	3	71	9	91	0	68	966
11.4	112	166	0	0	22	16	45	101	34	39	6	0	0	4	39	39	0	61	0	19	0	63	5	120	0	58	949
11.3	130	238	5	0	22	25	27	36	77	28	23	0	0	1	32	37	0	68	0	21	0	71	3	117	0	64	1025
11.2	142	150	2	0	8	15	16	3	67	72	1	0	0	2	17	32	0	99	0	3	0	164	1	115	1	29	939
11.1	151	109	0	0	19	6	30	5	66	45	16	0	0	3	38	60	0	126	0	0	1	151	0	73	1	34	934
11	121	74	0	0	12	6	11	22	53	67	2	0	0	1	9	29	0	89	0	4	1	98	2	97	0	8	706
10.9	126	84	0	1	37	16	38	77	89	82	1	0	0	7	38	31	1	79	0	18	2	103	8	74	1	22	935
10.8	117	121	0	0	21	24	40	61	82	62	2	0	0	7	40	19	1	67	0	27	8	132	3	82	5	36	957
10.7	117	127	1	0	24	26	29	74	73	52	2	0	0	1	35	33	6	83	0	20	3	120	0	95	1	24	946
10.6	119	118	2	0	43	14	38	63	98	72	1	0	0	3	49	26	4	43	0	23	4	104	0	76	0	63	963
10.5	107	169	0	0	18	20	28	42	127	72	4	0	1	1	20	16	7	56	0	9	2	118	0	96	0	31	944
10.4	124	152	0	0	20	7	33	32	81	102	3	0	1	3	47	23	5	75	0	4	0	112	0	91	0	11	926
10.3	110	168	4	0	19	9	24	30	119	101	6	0	0	0	44	20	3	72	0	8	0	93	0	88	1	23	942
10.25	130	156	2	0	14	9	27	25	124	109	2	0	0	4	38	23	0	63	0	4	0	97	0	90	1	35	953
10.2	145	76	1	0	0	4	28	31	226	44	0	0	0	1	13	36	2	62	0	2	4	111	0	67	0	66	919
10.1	96	111	2	0	0	9	41	27	235	85	1	0	2	2	7	51	0	34	0	5	2	80	0	53	2	95	940
10	186	84	2	0	0	4	30	28	142	52	1	0	1	3	13	29	0	45	0	14	4	96	0	106	10	116	966
9.9	119	116	3	1	30	18	32	97	92	54	5	0	2	3	35	25	3	50	0	36	0	79	9	107	0	30	946
9.8	142	155	0	0	23	10	31	35	97	46	1	0	0	0	19	23	1	83	0	35	0	124	6	94	0	19	944
9.7	125	117	10	1	17	20	44	65	78	50	0	0	0	1	36	27	3	76	0	54	0	94	14	86	0	37	955
9.6	129	162	0	0	25	13	36	49	87	24	2	0	1	0	37	29	0	95	0	24	1	105	7	83	0	19	928
9.5	162	113	4	0	20	12	46	53	66	33	3	0	1	5	40	47	1	68	0	22	0	101	7	78	0	63	945
9.4	156	120	2	0	14	10	42	58	109	28	2	0	1	0	33	32	2	89	0	13	1	137	4	73	0	36	962
9.3	146	113	6	0	15	24	6	62	84	37	3	0	0	2	19	10	0	93	0	16	1	137	3	117	0	34	928
9.2	171	125	0	0	27	17	20	35	106	49	2	0	2	0	20	35	0	92	0	20	0	86	3	110	0	31	951
9.1	136	109	0	0	28	12	28	109	105	37	5	0	1	2	22	16	11	51	0	25	1	104	10	97	2	47	958
9	151	114	8	0	27	10	25	89	42	36	9	0	1	4	19	13	3	58	0	27	0	157	7	108	0	56	964
8.9	117	137	3	1	23	14	39	83	102	58	6	0	0	4	35	26	1	57	0	29	1	110	8	70	2	28	954
8.8	154	158	2	0	26	20	12	63	68	23	6	0	0	1	22	19	4	82	1	32	0	119	7	94	0	28	941
8.7	150	129	8	0	17	17	28	99	43	38	1	0	0	3	18	24	0	69	0	30	0	107	15	90	0	67	953
8.6	137	157	4	0	19	12	31	104	77	67	7	0	0	1	31	19	0	34	0	38	2	63	4	99	2	68	976
8.5	124	176	2	0	24	12	20	45	99	33	1	0	0	2	20	12	1	84	0	20	0	103	6	95	0	42	921
8.4	134	123	0	0	16	18	28	53	123	24	16	0	0	3	24	40	0	74	0	8	3	102	1	60	0	94	944
8.3	136	87	0	3	20	10	45	48	182	28	10	0	0	0	33	26	2	63	0	29	1	113	1	73	1	25	936
8.2	143	86	1	0	6	10	10	18	217	51	15	0	0	0	32	23	5	43	0	12	2	114	1	91	0	76	956
8.1	199	195	3	0	11	15	20	34	107	41	19	0	1	0	30	25	0	28	0	9	3	84	1	77	0	64	966
8	82	203	34	1	15	1	0	127	1	2	7	0	0	0	0	62	0	0	0	3	0	196	0	38	3	25	800
7.9	88	205	43	3	30	1	0	169	0	20	9	0	0	0	1	57	0	0	1	4	0	151	0	36	0	9	827

(*continued*)

TABLE 1. ABSOLUTE ABUNDANCES OF PLANKTONIC FORAMINIFERAL SPECIES AND GROUPS IN SAMPLES FROM THE 4–12 MSL SEGMENT AT MASSIGNANO (*continued*)

Stratigraphic level from base (m)	Chiloguembelinids	Catapsydracids	*D. galavisi*	*D. globularis*	*D. larmeui*	*D. tripartita*	*G. officinalis*	*"G." venezuelana*	Globigerinathekids	Globorotaloids	*Gt. martini*	*Gt. ouachitaensis*	Guembelitrids	Hantkeninidae	*P. nana*	Pseudohastigerinids	Small acarininids	*S. angiporoides*	*S. eocaena*	*S. gortanii*	*S. linaperta*	*S. utilizindex*	*S. yeguaensis*	Tenuitellids	*T. ampliapertura*	*T. cerroazulensis* gr.	Total
7.8	49	212	76	5	48	0	0	131	0	20	19	0	0	0	1	19	0	0	1	4	0	170	0	44	3	2	804
7.7	80	237	51	2	38	1	0	112	0	7	15	1	0	0	2	17	1	0	1	5	0	200	0	68	3	21	862
7.6	58	218	37	3	23	1	0	111	0	15	21	0	0	0	1	5	0	0	2	8	0	226	0	65	5	23	822
7.5	79	188	54	15	34	3	1	100	2	25	39	1	0	0	0	16	1	0	1	7	0	221	0	79	5	32	903
7.4	53	155	68	20	59	1	1	79	1	25	1	1	0	0	0	8	0	0	3	1	0	235	0	62	2	53	828
7.35	68	175	56	7	50	2	2	71	2	38	16	2	0	0	0	27	0	0	3	5	0	225	0	69	4	52	874
7.3	108	191	51	5	33	0	0	94	0	42	9	1	0	0	0	36	0	0	3	9	0	232	0	76	1	15	906
7.2	132	234	62	15	32	0	0	132	0	46	4	0	0	0	0	30	0	0	2	9	0	207	0	40	2	3	950
7.15	86	254	65	10	34	0	0	109	0	20	7	0	0	1	0	19	3	0	2	9	0	207	0	56	0	1	883
7.1	138	203	49	7	29	1	0	51	0	25	22	0	0	2	0	62	9	0	3	12	1	228	0	53	5	15	915
7	84	114	48	10	30	2	0	14	12	29	29	1	0	0	0	48	22	0	6	8	3	269	0	92	7	41	869
6.9	69	154	48	4	28	1	0	121	18	10	2	0	0	1	0	84	1	0	3	8	0	291	0	37	7	36	923
6.8	69	145	49	10	18	1	1	103	38	11	2	2	0	0	2	53	2	0	2	10	0	296	0	60	4	27	905
6.7	70	112	17	5	8	0	1	81	175	9	6	0	0	0	0	44	1	0	2	12	1	230	0	66	2	59	901
6.6	79	137	24	9	22	0	0	100	32	11	9	1	0	4	1	44	2	0	3	9	1	324	0	71	4	35	922
6.5	93	136	33	10	33	1	2	83	15	16	20	1	0	1	1	30	1	0	4	17	1	333	0	72	0	23	926
6.45	59	201	69	7	55	0	0	127	0	14	19	0	0	0	1	3	9	0	3	13	2	220	0	39	7	3	851
6.4	90	124	12	8	39	1	2	71	3	28	8	1	0	0	0	57	3	0	4	2	3	333	0	65	5	31	890
6.3	81	121	32	5	31	1	2	55	8	23	28	0	0	0	2	72	1	0	5	5	2	312	0	63	10	34	893
6.2	78	134	49	8	39	0	2	55	1	29	27	1	0	0	1	86	7	0	5	7	2	239	0	103	6	23	902
6.1	184	25	7	3	22	0	0	123	6	35	29	1	0	1	4	18	4	0	0	7	0	300	0	41	0	54	864
6	143	11	0	4	19	1	0	45	5	42	11	0	0	0	2	67	3	0	0	8	0	299	0	49	0	181	890
5.9	97	22	2	8	19	0	0	51	4	67	14	0	0	0	0	83	1	0	0	15	0	237	0	42	1	264	927
5.8	154	19	0	1	13	0	0	162	20	42	22	0	0	0	0	26	2	0	0	5	0	233	0	60	0	47	806
5.7	74	37	10	6	34	0	0	78	13	70	4	0	0	1	3	111	2	0	0	18	0	225	0	60	0	117	863
5.65	165	42	4	5	29	0	0	54	73	45	9	0	0	0	1	43	0	0	0	43	0	169	0	60	0	107	849
5.6	178	52	7	7	14	0	0	65	19	55	8	0	0	0	0	42	0	0	0	18	0	229	0	39	1	145	879
5.5	162	24	4	7	14	0	0	41	100	30	12	0	0	0	0	55	1	0	0	20	0	243	0	51	3	94	861
5.4	140	26	2	3	14	0	0	67	43	34	10	0	0	0	1	50	2	0	0	25	0	310	0	46	0	97	870
5.3	140	25	3	9	10	0	2	53	74	27	9	0	0	0	0	82	1	0	0	47	0	244	0	66	0	75	867
5.25	175	53	8	0	11	0	0	580	58	12	37	0	0	0	0	39	2	0	0	29	0	225	0	40	3	27	1299
5.2	143	43	3	1	2	0	0	33	61	22	3	0	0	0	0	41	0	0	0	30	0	281	0	52	1	87	803
5.1	179	25	4	0	15	0	0	33	51	26	1	0	0	0	0	42	2	0	0	26	0	306	0	46	1	89	846
5	122	47	9	7	3	0	1	78	71	26	4	0	0	4	1	106	3	0	0	20	0	258	0	32	4	71	867
4.9	136	52	3	2	9	0	2	60	54	18	10	0	0	0	5	80	15	0	0	34	0	258	0	51	2	82	873
4.8	159	58	7	4	8	0	0	62	51	28	12	0	0	1	0	57	9	0	0	48	0	227	0	42	1	58	832
4.7	122	57	1	0	5	0	0	33	58	25	16	0	0	4	0	98	22	0	0	23	0	228	0	20	2	80	794
4.6	92	37	1	0	8	0	0	53	123	21	11	0	0	0	0	84	33	0	0	48	0	147	0	43	1	83	785
4.5	81	25	2	3	9	0	0	57	99	30	6	0	0	0	1	61	60	0	0	60	0	195	0	43	5	111	848
4.4	87	28	6	0	3	0	0	109	65	13	10	0	0	0	1	105	25	0	0	44	0	212	0	35	0	91	834
4.3	124	40	5	0	4	0	0	134	51	8	10	0	0	0	0	50	42	0	0	34	0	174	0	54	2	57	789
4.2	120	54	6	6	4	0	0	76	46	17	11	0	0	0	0	56	27	0	0	35	2	269	0	91	10	42	872
4.1	104	49	4	2	8	0	0	104	50	23	4	0	0	0	7	56	20	0	1	42	1	180	0	89	20	87	851
4	83	33	8	0	10	0	0	66	37	41	5	0	0	0	1	82	31	0	0	39	5	226	0	75	18	84	844

TABLE 2. PALEOECOLOGY* AND PALEOCLIMATIC SIGNIFICANCE OF THE RECOGNIZED SPECIES AND GROUPS FROM THE 4–12 MSL SEGMENT AT MASSIGNANO[†]

Taxa/groups	Biogeographic distribution	Habitat	Paleoclimatic significance
Catapsydrax dissimilis	Cosmopolitan	Deep water	Cool
Catapsydrax unicavus	Cosmopolitan	Deep water	Cool
Chiloguembelinids	Mid-latitudes	Thermocline/oxygen minimum layer	Non–climate diagnostic
Dentoglobigerina galavisi	Low to mid-latitudes	Deep water	Warm temperate
Dentoglobigerina globularis	No data	No data	No data
Dentoglobigerina larmeui	Mid-latitudes	High thermocline	Probably warm-temperate
Dentoglobigerina tripartita	Low to mid-latitudes	High thermocline	Warm temperate
Globigerina officinalis	Low to mid-latitudes	Mixed layer	Cool
“*Globigerina*” *venezuelana*	Mid-latitudes	Deep water	Probably cool
Globigerinatheka index	Cosmopolitan, more abundant in high latitudes	Mixed layer	Cool temperate
Globigerinatheka luterbacheri	Common in mid-latitudes, very rare or absent in tropical regions	No data	Cool temperate
Globigerinatheka semiinvoluta	Low to mid-latitudes	Mixed layer	Cool temperate
Globigerinatheka subconglobata	Cosmopolitan	Mixed layer	Cool temperate
Globigerinatheka tropicalis	Mid- to low latitudes	No data	Cool temperate
Globorotaloides eovariabilis	High latitudes	No data	Cool
Globorotaloides sp. 1	Cosmopolitan	Probably deep water	Probably cool
Globoturborotalita martini	Mid- to low latitudes	Mixed layer	Warm
Globoturborotalita ouachitaensis	Mid- to low latitudes	No data	Non–climate diagnostic
Guembelitrids	Cosmopolitan	Probably upwelling	Probably cool
Hantkeninidae	Mid- to low latitudes	Mixed layer	Warm
Paragloborotalia nana	Low to mid-latitudes	Deep water	Cool temperate
Pseudohastigerinids	Low to mid-latitudes	Mixed layer to thermocline/oxygen minimum layer	Non–climate diagnostic
Small acarininids	Cosmopolitan, most common in low latitudes	Mixed layer	Warm
Subbotina angiporoides	Cosmopolitan, most common at high latitudes	Thermocline	Cool
Subbotina eocaena	Mid- to high latitudes	Deep water	Cool
Subbotina gortanii	Low to mid-latitudes	Deep water	Non–climate diagnostic
Subbotina linaperta	Cosmopolitan, most common at southern, high latitudes	Deep water	Cool
Subbotina utilisindex	Cosmopolitan, most common at high latitudes	Deep water	Cool
Subbotina yeguaensis	Low to high latitudes	No data	No data
Tenuitellids	Cosmopolitan, most common at high latitudes	Mixed layer	Cool
Turborotalia ampliapertura	Cosmopolitan	Mixed layer to deep water	Cool temperate
Turborotalia cerroazulensis	Cosmopolitan	Mixed layer to thermocline	Warm
Turborotalia cocoaensis	Cosmopolitan, more restricted to tropical latitudes than *T. cerroazulensis*	Mixed layer to thermocline	Warm
Turborotalia cunialensis	Cosmopolitan, more restricted to tropical latitudes than *T. cerroazulensis*	No data	Warm
Turborotalia pomeroli	Cosmopolitan, most common in northern mid-latitudes	Mixed layer	Warm

*Preferred biogeographical distribution and life position in the water column.

[†] Largely following Douglas and Savin (1978), Boersma et al. (1979, 1987), Keller (1983), Poore and Matthews (1984), Nocchi et al. (1988b), Premoli Silva and Boersma (1988, 1989), Stott et al. (1990), Barrera and Huber (1991, 1993), Boersma and Premoli Silva (1991), Spezzaferri and Premoli Silva (1991), Pearson et al. (1993, 1997, 2001, 2006), Spezzaferri (1995), Van Eijden and Ganssen (1995), Pearson and Palmer (1999), Coxall et al. (2000), and Wade and Kroon (2002).

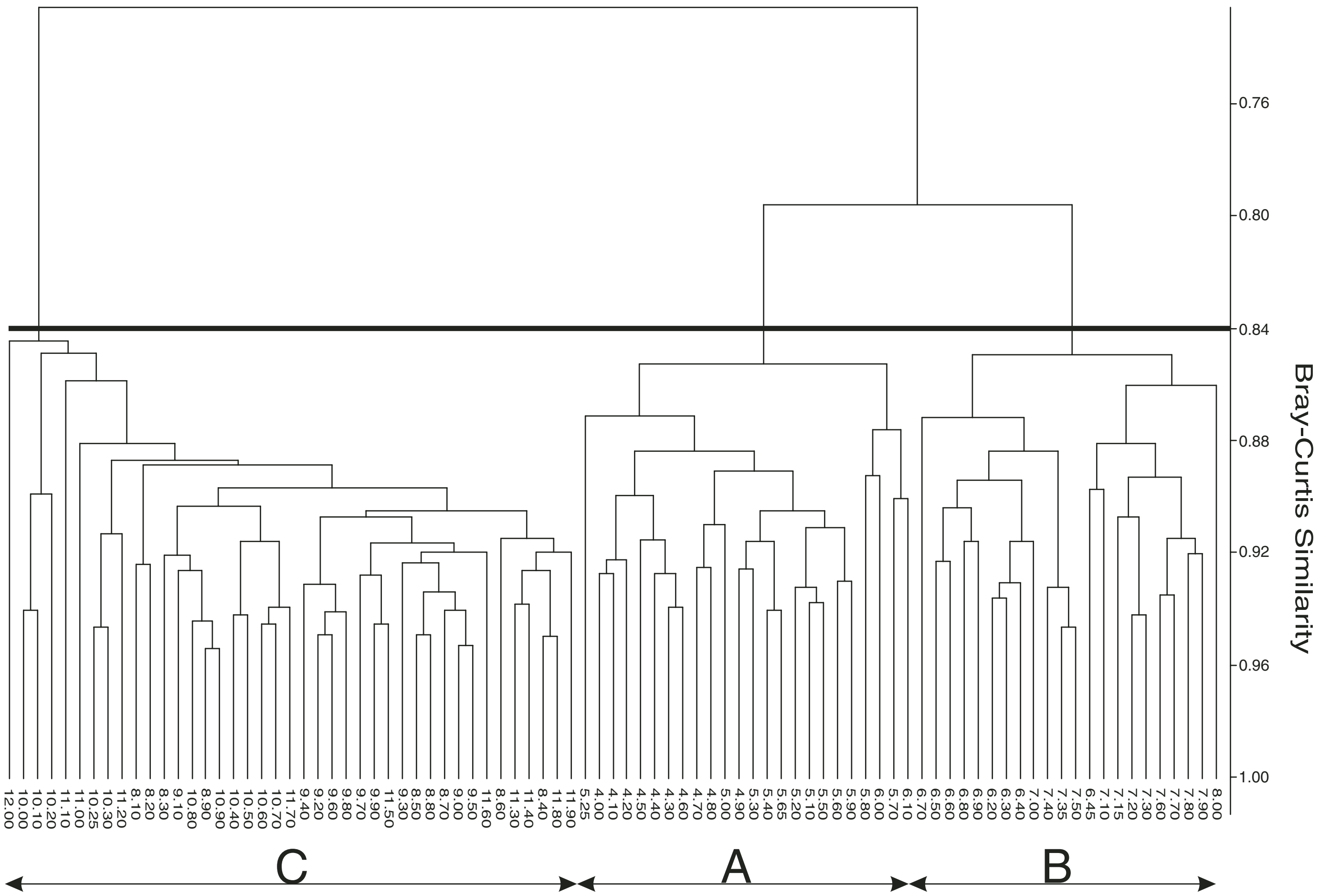

Figure 6. Classification of samples of the 4–12 m stratigraphic level from the base (msl) segment at Massignano using hierarchical agglomerative cluster analysis based on the Bray-Curtis similarity matrix. Three clusters (A, B, and C) were identified at the 84% level of Bray-Curtis similarity.

In terms of geochronology, an interpolated age of 35.7 ± 0.4 Ma for the impactoclastic layer at 5.61 msl was derived from radioisotopic dating performed in the volcaniclastic layers at Massignano (Montanari et al., 1993) (Figs. 4 and 5). On the basis of the magnetochronologic ages provided by Cande and Kent (1995), Gradstein et al. (2004), and Jovane et al. (2006) for chrons C16n.2r, C16.1n, and C15n (Figs. 4 and 5), the numerical ages for the Popigai, the Chesapeake Bay, and the unidentified impact events at Massignano would be 35.417–35.493 Ma, 35.346–35.411 Ma, and 34.778–34.894 Ma, respectively. On the basis of the mean sedimentation rate of 0.73 cm/k.y. (Jovane et al., 2007) and of the magnetochronologic ages, the time interval between the Popigai and Chesapeake Bay impacts would span ~77 k.y. on average.

A comparison of the high-resolution magnetobiostratigraphic and geochronologic data from the Massignano GSSP with those from Ocean Drilling Program (ODP) Holes 689B (Maud Rise, Southern Ocean) and 1090B (South Atlantic sector of the Southern Ocean), and the Chickahominy Formation in the Kiptopeke core hole (offshore southeastern Virginia) provides a valuable time constraint and global correlation for the late Eocene impact-ejecta horizons.

At ODP Hole 689B, the Antarctic impact-ejecta horizon that is interpreted to represent the clinopyroxene-bearing spherule-strewn field of the North America tektite strewn field, providing an ^{40}Ar/^{39}Ar date of 35.5 Ma (Glass et al., 1982; Vonhof et al., 2000), lies within chron C16n.1n (Florindo and Roberts, 2005). This horizon falls in the uppermost part of planktonic foraminiferal *Globorotaloides suteri* (AP12) zone (Stott and Kennett, 1990), which would correspond to the uppermost part of zone P15 of Berggren et al. (1995) and to the calcareous nannofossil NP19/20 equivalent zone (Wei and Wise, 1992; see also Berggren et al., 1995).

According to Pusz et al. (this volume), no distinct stratigraphic separation is found between the microtektite and microkrystite layers at Site 1090, recognized as its own layer (Liu et al., 2000; Kyte and Liu, 2002). Following the magnetostratigraphic age control provided by Channell et al. (2003), the Ir anomaly located 10 cm above the two stratigraphically coeval, but geochemically distinct, Chesapeake Bay impact structure microtektite

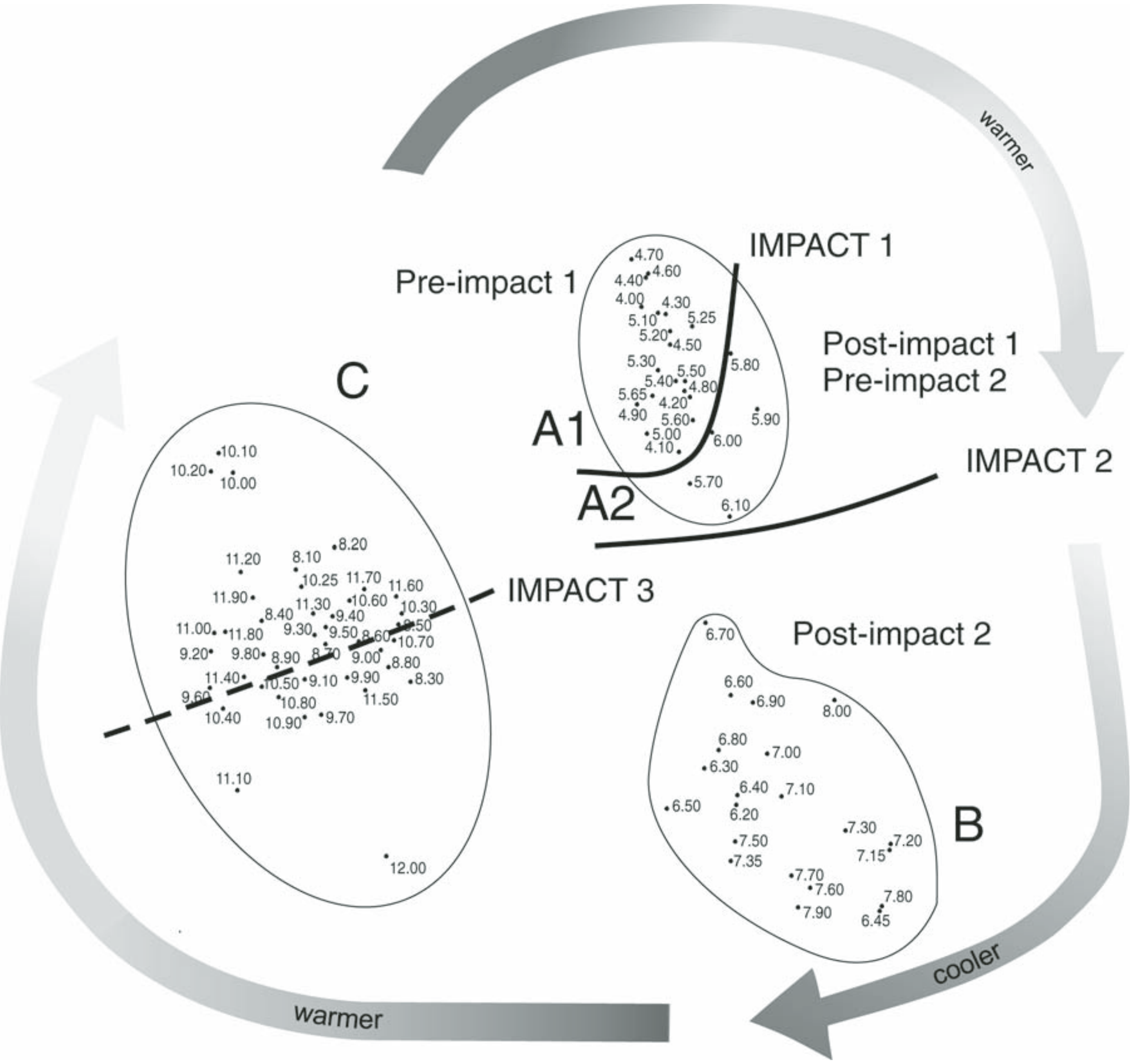

Figure 7. Interpretation of clusters identified in the dendrogram (Fig. 6) using non-metric multidimensional scaling (nMDS). The ordination by nMDS is an iterative procedure to represent “distance” of samples from a multidimensional space on the base of rank dissimilarities. The plot has no defined scale. The stress value (0.13) represents the distortion involved in compressing the data to a small number of dimensions; stress < 0.2 in 2D means a useful representation of data (Kruskal, 1977; Field et al., 1982). In this study, the line IMPACT 1, which corresponds to the Ir anomaly at 5.61 msl, allows us to subdivide cluster A into subcluster A1 (pre-impact 1) and subcluster A2 (postimpact 1 to pre-impact 2). The most significant feature recognized in the nMDS plot is the line IMPACT 2, which corresponds to the Ir anomaly at 6.17 msl. This line clearly separates cluster A from cluster B, corresponding to CI-I and CI-II, respectively. According to the multivariate statistical results, any significant change in planktonic foraminiferal assemblages took place across the Ir anomaly at 10.28 msl (dotted line IMPACT 3). The shaded bands with arrows indicate the temperature water gradient based on climate curve.

and Popigai microkrystite layers (both at 35.43 Ma) falls within chron C16n.1n, with an interpolated magnetochronologic age of 35.426 Ma (Miller et al., 2007; Pusz et al., this volume). This is consistent with published radiometric ages ranging from 35.3 to 35.7 Ma (Glass et al., 1982; Obradovich et al., 1989; Bottomley et al., 1997; Horton and Izett, 2005). According to the planktonic foraminiferal biostratigraphy provided at the ODP Site 1090 by Galeotti et al. (2002), the microtektite and microkrystite layers fall in the upper part of the *Globigerinatheka index*–*Globorotaloides suteri* (AP11–AP12) zones of Stott and Kennett (1990), which would correspond to the upper part of zones P15–P16 of Berggren et al. (1995) and in the middle part of zones NP19/20 of Martini (1971), following the nannofossil calcareous biostratigraphy established by Marino and Flores (2002).

According to the magnetobiochronology provided by Poag et al. (2003, 2004) for the continuous interval of the Chickahominy

Formation in the Kiptopeke core hole, which lies conformably above the Chesapeake Bay impact crater, the Chesapeake Bay impact would have taken place in chron C16n.2n and within planktonic foraminiferal zone P15 of Berggren et al. (1995), with an extrapolated age of 35.78 Ma. However, the new magnetostratigraphic data from the Massignano GSSP together with those from ODP Holes 689B and 1090B cast reliable doubt on the interpretation of the magnetostratigraphic record of the Kiptopeke core hole, which, as a consequence, needs to be refined.

Planktonic Foraminiferal Overall Trends during the Late Eocene

The end of the Eocene was a critical interval within the transition from the Eocene "greenhouse" world to the Oligocene "icehouse" world marked by significant evolutionary turnovers of marine biota and changes of oceanographic and atmospheric circulation patterns (Benson, 1975, 1990; Kennett, 1977; Miller et al., 1987, 1991, 2005; Lear et al., 2000, 2008; Zachos et al., 2001; Coxall et al., 2005; Pagani et al., 2005; Tripati et al., 2005; Coxall and Pearson, 2007). Through a study based on planktonic foraminifera assemblages from some Deep Sea Drilling Program (DSDP) sites in the low-latitude Atlantic (Site 363), Pacific (Site 292), and Indian (Site 219) Oceans and a high-latitude South Pacific site (Site 277), Keller (1983) observed that the Eocene to early Oligocene cooling proceeded by pulses separated by relatively stable climate conditions. These pulses were characterized by rapid cooling indicated by the oxygen isotope record, major faunal turnovers, and hiatuses. She identified one of the major pulses in the interval across the transition between planktonic foraminiferal zone P15 and P16 of Berggren and Miller (1988) and Berggren et al. (1995) and observed that this pulse was accompanied by successive replacement of warm middle Eocene surface planktonic foraminiferal species by cooler late Eocene intermediate- and deep-water species. According to Keller (1983), the disappearance of the surface dwellers suggests that in high- and low-latitude oceans, during the planktonic foraminiferal zone P15–P16 interval, significantly cooler temperatures prevailed in the upper 400–500 m of the water column. Boersma and Premoli Silva (1986) and Keller et al. (1992) pointed out that long-term trend of planktonic foraminifera evolutionary turnover from the middle Eocene into the Oligocene largely involved extinction of warm-water, tropical, surface-dwelling species.

Several authors have speculated about the causes that produced the late Eocene cooling and foraminiferal faunal reorganization. Keller (1983) and Keller et al. (1992) related the late Eocene cooling and water mass reorganization to significant ice-cap development and increased thermal gradients between the equator and the poles, and Berger et al. (1981) observed that cooling phases may have been enhanced by increased albedo. Boersma et al. (1987) explained the reorganization of the planktonic foraminiferal assemblage structures in zone P16 by a shoaling of the nutricline and oxygen minimum layer. Boersma and Premoli Silva (1991) and Hallock et al. (1991) attributed the variations in the structure of the planktonic foraminiferal communities from the late middle Eocene through the middle Oligocene to reduced temperatures, shoaling of the thermocline, and increased delivery of nutrients to oceanic surface waters. These changes produced a contraction of the trophic resource regional gradient and reduction of oligotrophic and warm surface-water habitats (Hallock, 1987). According to Coxall and Pearson (2007), the extinction at and near the Eocene-Oligocene boundary might therefore have been due to rapid environmental change and cooling, and it may also have been influenced by changing water mass stratification and patterns of biological productivity.

The discovery of some bolide-impact craters and associated breccias, ejecta horizons, and cosmic signatures (e.g., Montanari et al., 1993, 1998; Clymer et al., 1995, 1996; Langenhorst and Clymer, 1996; Poag and Aubry, 1995; Koeberl et al., 1996; Bottomley et al., 1997; Farley et al., 1998; Pierrard et al., 1998; Liu et al., 2000; Kyte, 2001; Kyte and Liu, 2002; Poag et al., 2002, 2003, 2004; Bodiselitsch et al., 2004) in the late Eocene has led to speculation that the late Eocene impact events may have played an important role related to the biotic turnover and the deterioration of the global climate at the end of the Eocene Epoch, even if their effects on the climate and on the biosphere are still not completely clear (e.g., Keller, 1986; Brinkhuis and Coccioni, 1995; Coccioni et al., 2000; Vonhof et al., 2000; Poag et al., 2002, 2003, 2004; Spezzaferri et al., 2002; Bodiselitsch et al., 2004).

Minor extinction episodes and assemblage shifts among planktonic (e.g., Keller, 1986; Coccioni et al., 2000; Spezzaferri et al., 2002) and benthic (Poag et al., 2004) foraminifera, dinocysts (Brinkhuis and Coccioni, 1995; Vonhof et al., 2000), and stable oxygen and carbon isotope excursions associated with the impact-ejecta horizons at several sites (Vonhof et al., 2000; Poag et al., 2002, 2003, 2004; Bodiselitsch et al., 2004; Pusz et al., 2006, 2007, this volume; Miller et al., 2007) have been taken as evidence for global-scale short- and long-term environmental disturbance related to the impacts. According to Coxall and Pearson (2007), it seems unlikely that the impacts had a major influence on Earth's biosphere at the Eocene-Oligocene boundary, and they have to be ruled out as a possible cause of the climatic transition. It is possible, however, that climatic disruption caused by atmospheric dust loading associated with the late Eocene impacts, coupled with an ice-albedo feedback mechanism that amplified impact-induced climatic cooling, may have contributed to global cooling in the run-up to the Eocene-Oligocene transition.

Late Eocene Impact-Induced Environmental Perturbations, Including Climate and Water Mass Structure Changes?: Evidence from the Massignano GSSP

Several lines of evidence from nannofossil and foraminiferal assemblages (e.g., Keller, 1983; Quilty, 2001; Persico and Villa, 2004; Coccioni et al., 2000; Spezzaferri et al., 2002), dinoflagellate cyst palynology (e.g., Brinkhuis, 1992; Coccioni et al., 2000), clay mineralogy (e.g., Diester-Haass et al., 1996; Ehrmann, 1997), oxygen isotopes (e.g., Kennett and Stott, 1990; Miller et al., 1992;

Vonhof et al., 2000; Bohaty and Zachos, 2003; Poag et al., 2003, 2004; Bodiselitsch et al., 2004; Pusz et al., 2007, this volume), and environmental magnetism (Sagnotti et al., 1998; Jovane et al., 2004) indicate that the major cooling event at the Eocene-Oligocene boundary was predated by distinct late Eocene cooling pulses, separated by relatively stable warming phases.

Our data show that the late Eocene impact events recorded in the 4–12 msl segment of the Massignano GSSP had no dramatic effects on planktonic foraminifera. However, acting as forcing factors, they induced some environmental perturbations and may have contributed to noticeable climate changes superimposed on the general late Eocene cooling trend.

Based on field evidence (Coccioni et al., 1988) and on the redistribution of extraterrestrial iridium (Huber et al., 2001) at Massignano, we exclude the effect of bioturbation and the resultant vertical displacement of biostratigraphic and paleoclimatic events. Accordingly, the paleoclimatic trend at Massignano can be considered uncorrupted.

The thermocline is the one of the main features responsible for partitioning of the water column and therefore of the vertical distribution of planktonic organisms, especially foraminifera. Either generally warmer surface water or the presence of a deeper thermocline quantitatively favors shallow and warm-water dwellers, while a shallower thermocline quantitatively favors deeper-dwelling and cool-water forms (e.g., Kennett and Barker, 1990).

The generally high abundance of cool- and deep-water species, the low abundances of warm and temperate forms, and the climatic curve from the 4–12 msl segment indicate that the studied interval was characterized by generally cool conditions (Fig. 5), a relatively shallow thermocline, as deduced by the generally high abundances of cool- and deep-water dwellers, and consequently a reduced surface-layer thickness.

Within a time of global cooling, however, abundance variations of the recognized paleoclimatic species and groups document different climate conditions throughout the studied segment at Massignano. Relatively warmer conditions (CI-I) occurred from the base of the studied segment up to the Chesapeake Bay impact event, then relatively coolest conditions (CI-II) persisted for a few hundreds of thousands of years, and finally relatively less cool conditions (CI-III) developed up to the top of the segment (Fig. 5). The statistical treatment of our data corroborates the paleoclimatic trend detected by the fluctuations of the planktonic foraminiferal climatic indices (Fig. 7). Based on our observations, we interpret lines IMPACT 1 and IMPACT 2 in the nMDS plot to represent the temperature gradient (Fig. 7), and we relate the gaps between subcluster A1 and subcluster A2 and between subcluster A2 and cluster B to a prevailing warming pulse and a cooling episode after the Popigai and Chesapeake Bay impact events, respectively.

Both the Popigai and Chesapeake Bay impacts liberated energy, probably equivalent to some 10 million megatons of TNT (trinitrotoluene) (Montanari and Koeberl, 2000). These impacts may have had some short-term effects on the environment by injecting dust into the atmosphere, inducing acid rain, cooling, and inhibiting photosynthesis (e.g., Gerstl and Zardecki, 1982; Toon et al., 1994; Kring, 2000).

A slight decrease in abundance of globigerinathekids, which have a symbiotic life strategy (Boersma et al., 1987; Pearson et al., 1993), took place just after the Popigai impact and might have been related to increased water turbidity and/or inhibition of photosynthesis following the impact event itself, which prevented the proliferation of algal symbionts (Figs. 4 and 5). Acarininids are also known as symbiont-bearing species (Coxall et al., 2000), but they are very rare throughout the studied interval (Figs. 4 and 5), and it was not possible to verify their decrease in abundance after the impact events. The recognized distinct opposite trends of the low-oxygen-tolerant chiloguembelinids and pseudohastigerinids may indicate enhanced competition between them during times of environmental stress. There was a slight reduction in the number of chiloguembelinids after the second impact event, which may reflect a very moderate enrichment of O_2 in the water masses (Figs. 4 and 5).

At Massignano, an increase in productivity in surface waters across the Popigai and Chesapeake Bay impact intervals persisting for ~137 k.y. might be inferred from a strong abundance increase of the cosmopolitan, calcareous dinoflagellate species *Thalassiphora pelagica*, which is interpreted to be a cold-water or productivity indicator (Coccioni et al., 2000). In fact, this species is known to be associated with anoxic sediments (e.g., Köthe, 1990) and with the major cooling event recorded worldwide in the early Oligocene (Brinkhuis and Biffi, 1993; Zachos et al., 1996).

Changes in ventilation and flushing at the seafloor, reflecting significant reorganization of the water mass structure after the Popigai and Chesapeake Bay impact events, might be indicated by sudden changes in sediment color (e.g., Coccioni et al., 2000; Spezzaferri et al., 2002) (Figs. 2–5).

Quantitative changes in the planktonic foraminiferal assemblages indicate a slight warming pulse ~5 k.y. after the deposition of the impactoclastic layer related to the Popigai impact. This discrete pulse is followed by alternating cooler and warmer intervals. However, as highlighted by subcluster A2, the interval between the Popigai and Chesapeake Bay impacts, lasting for ~77 k.y., was characterized by prevailing warmer condition compared to the conditions encountered prior to the first impact (Figs. 5 and 7).

The Chesapeake Bay impact took place within a following cooling episode. Bodiselitsch et al. (2004) documented a negative shift in bulk carbonate $\delta^{18}O$ and $\delta^{13}C$ close to the Ir-rich level at 6.17 msl compared to the continuously decreasing trend over the entire Massignano section. According to these authors, despite the fact that oxygen isotope values at Massignano are clearly diagenetically influenced, the negative excursion of $\delta^{18}O$ might provide evidence of a short warming pulse. After the Chesapeake Bay impact, an enhanced cooling (CI-II) interval started (Fig. 5; cluster B in Figs. 6 and 7), as testified by the increase in abundance of cool and cool-temperature water species and groups. This change in abundance occurred along with a concurrent decrease in abundance of warm and warm-temperature water species and groups, which reveals a contraction of the mixed and surface layers, and consequent decrease of niches in the warmer surface water. This relatively long-term cooling episode lasted for ~247 k.y. and was accompanied by a shallowing of the thermocline, which occurred in the interval spanning the transition between planktonic foraminiferal zones P15 and P16 of Berggren et al. (1995), which is in the lower part of zone E15 of Berggren and Pearson (2006), and it may have been related to the cooling pulse identified by Keller (1983) in the same biostratigraphic interval.

Oxygen isotope records at ODP Hole 689B indicate accelerated climatic cooling, possibly combined with increased surface-water productivity deduced by $\delta^{13}C$ data and calcareous dinocyst abundances, coinciding with the two nearly contemporaneous impact events (Vonhof et al., 2000). The relatively long duration of the cooling pulse that followed the impact events led Vonhof et al. (2000) to suggest that a feedback mechanism, such as a global albedo increase due to extended snow and ice cover, sustained the initial impact-induced changes. We agree with this hypothesis to explain the enhanced cooling following the Chesapeake Bay impact and corresponding to interval CI-II, where a rapid increase of $\delta^{13}C$ and $\delta^{18}O$ values occurs (Bodiselitsch et al., 2004) together with an increase in abundance of *Thalassiphora pelagica* (Coccioni et al., 2000).

The enhanced cooling terminated within the middle part of chron C15r and the lowermost part of planktonic foraminiferal zone P16 of Berggren et al. (1995), which is in the lower part of zone E15 of Berggren and Pearson (2006) (Fig. 5). From there on, climate conditions intermediate (CI-III) between the warmest conditions (CI-I) and the coolest conditions (CI-II), as recorded through the studied segment, developed and persisted (Fig. 5), as also evidenced by cluster C (Figs. 6 and 7).

The third impact, which lies within the CI-III interval, led to an enhanced cooling that was followed after ~15 k.y. by a subsequent return to less cool conditions, which extend up to the top of the studied segment (Fig. 5).

However, this impact does not appear to have produced significant environmental perturbations, as evidenced by the unimportant fluctuations of the planktonic foraminiferal assemblages (Figs. 5 and 7). According to Bodiselitsch et al. (2004), negative excursions in bulk carbonate $\delta^{18}O$ and $\delta^{13}C$ occur close to the Ir-rich level at 10.28 msl, which might indicate a sudden warming pulse caused by this impact event. Again, on the basis of carbon and oxygen isotope ratio data, it seems that seawater temperature did not remarkably change in between the two last Ir-enhanced intervals at Massignano (Bodiselitsch et al., 2004). However, the average $\delta^{13}C$ values differ somewhat between the two regions, which could indicate an incursion of colder, more vigorous bottom waters and an increase of biomass and productivity during this time span as seen in the deep-water benthic foraminifera faunal turnover recognized in the same interval by Coccioni and Galeotti (2003).

Based on this line of evidence, the climatic effects of the late Eocene impact events as documented at Massignano therefore appear to be not univocal. In fact, it seems that the Popigai impact was followed by a prevailing warming pulse, whereas the other two impact events were followed by an enhanced cooling.

It is well known that the environmental and the biological effects of an impact event must be evaluated in terms of impactor size as well as in the context in which it occurs in terms of location and target rocks (e.g., Kring, 2000). The Popigai and Chesapeake Bay impactors are estimated to have been 5 km (Tagle and Claeys, 2004) and 3.2 km (Collins and Wünnemann, 2005) in diameter, respectively. The Popigai impact occurred on the continent, whereas the Chesapeake Bay impact took place on the continental shelf. The target rocks of the Popigai structure are generally granitic gneisses overlain by sandstone and carbonates (Bottomley et al., 1997) produced by the impact of an ordinary chondrite body (Masaitis and Raikhlin, 1986). The Chesapeake Bay impact structure is developed in a mixed-target substrate composed of granitoids and metasedimentary rocks overlain by dominantly siliciclastic, sedimentary rocks (Koeberl et al., 1996; Poag et al., 2002). According to Bodiselitsch et al. (2004), the Ir anomaly at 10.28 msl may be also derived from an impact into a continental shelf, similar to the Chesapeake Bay impact. Kent et al. (2003) suggested that release of methane hydrates from mechanical disruption of sediments as a result of an impact could cause a greenhouse effect, which is revealed by negative excursions in the carbon and oxygen isotope records. In view of this, Bodiselitsch et al. (2004) speculated that the warm pulses at 6.17 and 10.28 msl could be due to the release of large amounts of seafloor methane hydrate during and after the last two impact events that induced the negative excursions in $\delta^{13}C$ in both Ir-enhanced regions. A similar hypothesis has been evaluated by Pusz et al. (2007, this volume) on the basis of the significant $\delta^{13}C$ negative excursion in both bulk and benthic foraminiferal carbonate coincident with the impact-ejecta horizons at ODP Site 1090 and DSDP Site 612. According to our data, a release of methane large enough to induce a greenhouse scenario can be evaluated only for the Popigai impact, which was followed by a prevailing warming pulse.

On the basis of the relationship among the Chesapeake Bay and Montagnais impacts and the oxygen isotope (inferred paleotemperature) and extraterrestrial ^{3}He curves, Poag et al. (2002, 2003, 2004) stated that the general cooling trend from the middle Eocene to the early Oligocene was interrupted by long- and short-term warming episodes. These episodes, possibly triggered by a multiple impact event, would have released greenhouse gases into the atmosphere during the late Eocene comet or asteroid shower and would have been superimposed on an overall cooling trend. Poag et al. (2002, 2003) proposed that this impact-generated greenhouse warming may have postponed a pending late Eocene mass extinction until the early Oligocene. More specifically, Poag et al. (2003, 2004) recognized three distinct subpulses of warm climate, W-1 to W-3, that followed the Chesapeake Bay impact and are indicated by lower oxygen and carbon isotope ratio values. According to these authors, the warm pulses correlate with an interval of increased ^{3}He at Massignano. In particular, subpulse W-1 correlates with chron C16n.2n and the lower part of chron C16n.1r and falls within zone P15 of Berggren et al.

(1995), and subpulse W-2 coincides with chron C16n.1n and the lower two-thirds of chron C15r and lies in the transition between planktonic foraminiferal zones P15 and P16 of Berggren et al. (1995). However, as reported in "High-Resolution Impact Stratigraphy, Geochronology, and Magnetochronology at Massignano: Implications for Global Correlation," the correlation of these subpulses with the magnetostratigraphy at Kiptopeke core hole has to be refined.

At Massignano, Bodiselitsch et al. (2004) correlated the Ir anomalies at 6.17 and 10.28 msl, respectively, with warm subpulses W-1 and W-2 of Poag et al. (2003, 2004). Following this and according to the refined magnetobiostratigraphy provided by Jovane et al. (2007, this volume), the warm subpulse W-1 would fall in the uppermost part of chron C16n.1n and in the upper part of planktonic foraminiferal zone P15 Berggren et al. (1995), which is in the lower part of zone E15 Berggren and Pearson (2006). Warm subpulse W-2 would lie in the middle part of chron C15n and in the lower part of planktonic foraminiferal zone P16 Berggren et al. (1995), which is in the middle part of zone E15 Berggren and Pearson (2006) (Figs. 4 and 5). However, according to the magnetobiostratigraphy currently available for the Massignano GSSP and our climatic curve, correlation of the warm subpulse W-2 of Poag et al. (2003, 2004) with the warm pulse recognized at Massignano in the interval between 8.1 and 8.4 msl and well constrained in the middle part of chron C15r and in the lowermost part of planktonic foraminiferal zone P16 of Berggren et al. (1995) (Fig. 5) would be more consistent. Moreover, it fits well with the prominent negative value of $\delta^{18}O$ recorded at 8.2 msl by Bodiselitsch et al. (2004).

The main climatic episodes documented at Massignano by the abundance variations of the planktonic foraminiferal paleoclimatic indices are largely mirrored by the alternating relatively warm and humid intervals with cold and dry intervals recognized in the late Eocene on a global scale through high-resolution environmental magnetic studies carried out at CIROS-1 drill core from McMurdo Sound (Antarctica) (Sagnotti et al., 1998) and at Massignano (Jovane et al., 2004). The existence of alternating warm and cool intervals at the Eocene-Oligocene transition has also been revealed through high-resolution paleoclimatic analyses of calcareous nannofossils at the Southern Ocean ODP Holes 689D (Maud Rise, Weddell Sea, Antarctica) and 744A (Kerguelen Plateau, Antarctica) (Persico and Villa, 2004). In particular, from the uppermost part of chron C16n.2n up to the lowermost part of chron C13r, three warm intervals alternate with two cool intervals. These intervals are well constrained by the magnetostratigraphy provided by Roberts et al. (2003) and Florindo and Roberts (2005). We can point out that in terms of magnetobiostratigraphy, these intervals fit very well with the climatic intervals we recognize at Massignano, including the cool intervals detected at the transition between chron C16n.1r and chron C16n.1n, which would be correlated with the cooling recorded between 5.20 and 5.30 msl.

The consequences of the late Eocene impacts seem to be much smaller than those seen following the Chicxulub impact at the Cretaceous-Paleogene boundary (e.g., Alvarez et al., 1980; Kring, 2007), which led to the well-known mass extinction event (e.g., Sepkoski, 1986).

Further high-resolution studies, possibly on millennial time scales, from high-quality sediment archives should provide reliable data sets to document and to improve understanding of impact-induced biotic and climatic changes and to more accurately correlate them with marine and continental records.

CONCLUSIONS

The three late Eocene impact events recorded in the Massignano GSSP had no abrupt, dramatic effects on planktonic foraminifera. However, the high-resolution quantitative study and the multivariate statistical analysis revealed noteworthy changes in the assemblages occurring after the impacts. Acting as forcing factors, these impacts induced some environmental perturbations and may have contributed to significant climate changes superimposed on the general late Eocene cooling trend. These climate changes displayed different features at each impact event, probably due to the context in which each occurred (e.g., target rocks).

In particular, the first two closely spaced impacts related to the Popigai and Chesapeake Bay craters revealed significant changes in the water mass structure, in terms of stratification and trophic resources, associated with some warming and cooling episodes that took place within chron C16n.1n and chron C15r and at the transition between planktonic foraminiferal zones P15 and P16 of Berggren et al. (1995).

The prevailing warming pulse recognized after the Popigai impact might have been due to greenhouse effects produced by injection of CO_2 into the atmosphere and/or the release of methane hydrate after the impact itself. The relatively long duration of the enhanced cooling following the Chesapeake Bay impact suggests that this impact induced a progressive cooling and triggered a feedback mechanism that sustained the initial impact-induced changes. Similar patterns of climatic excursions reported across the equivalent impact-ejecta horizons in the Pacific, Indian, Atlantic-Caribbean, and Southern Oceans, as well as in Antarctica, indicate that climate changes recorded at Massignano seem to have had a global, rather than local or regional, significance.

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