Biostratigraphy of a Paleocene–Eocene Foreland Basin boundary in southern Tibet

Xiaoqiao Wan a,*, Xi Wang a, Luba F. Jansa b

a China University of Geosciences, 29 Xueyuan Road, Beijing 100083, China
b Department of Earth Sciences, Dalhousie University, Halifax, N.S., Canada

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Abstract This study of the Paleocene–Eocene boundary within a foreland basin of southern Tibet, which was dominated by a carbonate ramp depositional environment, documents more complex environmental conditions than can be derived from studies of the deep oceanic environment. Extinction rates for larger foraminiferal species in the Zongpu-1 Section apply to up to 46% of the larger foraminiferal taxa. The extinction rate in southern Tibet is similar to rates elsewhere in the world, but it shows that the Paleocene fauna disappeared stepwise through the Late Paleocene, with Eocene taxa appearing abruptly above the boundary. A foraminifera turnover was identified between Members 3 and 4 of the Zongpu Formation—from the Miscellanea–Daviesina assemblage to an Orbitolites–Alveolina assemblage. The Paleocene and Eocene boundary is between the SBZ 4 and SBZ 5, where it is marked by the extinction of Miscellanea miscella and the first appearance of Alveolina ellipsodalis and a large number of Orbitolites. Chemostratigraphically, the \( \delta^{13}C \) values from both the Zongpu-1 and Zongpu-2 Sections show three negative excursions in the transitional strata, one in Late Paleocene, one at the boundary, and one in the early Eocene. The second negative excursion of \( \delta^{13}C \), which is located at the P–E boundary, coincides with larger foraminifera overturn. These faunal changes and the observed \( \delta^{13}C \) negative excursions provide new evidence on environmental changes across the Paleocene–Eocene boundary in Tibet.

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1. Introduction

The Paleocene–Eocene (P–E) boundary is a major biostratigraphic boundary (~55.8 Ma) where many deep-sea benthic foraminifera species became extinct. Benthic foraminifera extinctions were up to 30%–50% of bathyal–abyssal species (Thomas, 1998). At lower abyssal depths above the Carbonate Compensation Depth (CCD), agglutinated foraminiferal faunas show similar changes in faunal composition, although no major faunal turnover occurred (Kaminski et al., 1996). The benthic foraminiferal extinction has been interpreted as occurring during a time of decreasing oceanic productivity, following a high
productivity period during the middle Paleocene (Aubry, 1998). Deep-ocean ostracodes also show a perturbation by shifting to an assemblage of smaller size and thinner-wall shells, even though there is no associated extinction (Steinbeck and Thomas, 1996). Many long-lived benthic foraminiferal species, which had been in existence from the Maastrichtian or Campanian, became extinct within about 10,000 years at the P–E boundary (Thomas, 1998). Post-extinction calcareous foraminifera according to Thomas (1998) are mainly dominated by small individuals, with thin walls, while calcareous agglutinants are absent. These faunas are of low diversity compared to the pre-extinction assemblages.

In continental marginal regions faunal changes including extinctions have been linked to low oxygen conditions, resulting from local high primary productivity of non-carbonate primary producers (Stupin and Muzyllov, 2001). Faunal as well as geochemical evidence supports the theory that high productivity was the major cause of anoxia and extinction (Gavrilov et al., 1997).

The P–E boundary marks the base of a pronounced negative oxygen isotope excursion, termed the Paleocene—Eocene thermal maximum (PETM; Zachos et al., 2003). The PETM, coincides with a pronounced negative carbon isotope excursion in both marine and terrestrial environments. The onset of this excursion is now accepted as the primary marker for the P–E boundary (Aubry et al., 2003).

Oxygen isotope data suggest that during the PETM the deep ocean water temperature abruptly warmed by 4 °C–5 °C, and high latitude surface waters warmed up from ~10 °C to ~20 °C (Stott, 1992). These temperature changes were concurrent with major changes in biological community structure recorded by planktonic and benthic foraminifers, calcareous nannoplankton, dinoflagellates and ostracodes. The changes were widespread, abrupt, and led to a nearly complete replacement of the pre-PETM fossil communities at this time (Bralower, 2002; Kelly, 2002).

Faunal extinction was coeval with a drop of 2.5% in carbon isotope (δ13C) values, in the oceanic environment, implying a large input of isotopically light carbon to the ocean and atmosphere (Bowen et al., 2001; Kaiho et al., 1996; Koch et al., 1992; Zachos et al., 2001). The decrease in both stable carbon and oxygen isotopic values imply that global perturbations occurred in deep ocean and shallow marine, as well as in terrestrial environments. This carbon isotope excursion started rapidly, within less than 10,000 years, but δ13C values returned to background values in several hundred thousand years (Katz et al., 1999; Norris and Röhl, 1999). These observations provide evidence for a large input of carbon to the ocean/atmosphere reservoir and of a transfer of carbon from the deep ocean to the atmosphere/surface ocean reservoir. In the bathyal—abyssal environments, benthic foraminifera extinctions commonly are associated with strong carbonate dissolution (Cocconi et al., 1994; Ortiz, 1995). Possible causes of the benthic extinction event at bathyal—abyssal depths could be numerous. According to Thomas (1998), it could be low oxygenation, either as a result of increased temperatures of bottom waters, or of oxidation of methane in the water column. A widely accepted explanation for the carbon isotope excursion (CIE) is a sudden dissociation of 12C-enriched marine gas hydrates (Dickens et al., 1997; Kaiho et al., 1996). The cause of the dissociation is not clear. One of the hypotheses suggests that rapid warming of the intermediate ocean waters was due to changes in the oceanic circulation patterns (Nunes and Norris, 2006). Kurtz et al. (2003), however, showed that the late Paleocene was a time of increasing terrestrial organic carbon-rich deposition and suggested that the δ13C excursion may have resulted from the burning of large peat deposits.

Because less information is available on Paleocene—Eocene boundary conditions from shallow marine environments, especially for larger benthic foraminifera, we aimed our study at the neritic environment of the northern margin of the India plate during the PETM period. Strata of appropriate age are exposed in southern Tibet, where we examined larger benthic foraminifera and made stable isotope analyses of samples collected across the P–E boundary. We documented changes in the relative abundances of benthic foraminiferal species across the P–E boundary and compared the benthic foraminifera with isotopic data. We attempt in this paper to define the extinction horizons and the P–E boundary in the Tibetan strata, and to examine the cause of larger foraminifera turnover, in the hope of providing stratigraphic constraints on PTEM studies in the future.

2. Sampling and methods

Two sections were measured and sampled (Fig. 1). The Zongpu-1 Section is a key section located about 1.5 km east of Gamba town. Samples for foraminifera study were collected from all 490 m of the Zongpu-1 Section (Fig. 2). A detailed sampling was made across the P–E boundary; 45 rock samples were collected at 20 cm intervals across 9 m of boundary strata in this section (Figs. 3 and 6). Because it proved difficult to extract foraminifera from lithified limestone in the section, this study is based on examination of foraminifera in thin sections.

All 45 P–E boundary samples were used for carbon and oxygen stable isotope analyses. For isotopic comparison, 20 additional samples were collected from a 4 m boundary interval in the second section, Zongpu-2, which is located 3.5 km east of Gamba town (Fig. 1). Isotopic compositions were measured on powdered bulk carbonate samples, using a Finnigan-MAT252 spectrometer and standard analytical techniques at the Geochemistry Lab of the National Oil and Gas Company of China. Carbon isotope ratios reflect changes in the carbon cycling, and the oxygen isotope data may show trends in paleo-temperature changes (Fricke et al., 1998; Shackleton and Kennett, 1975). However, as oxygen isotopes are affected by diageneric and sediment burial, they cannot be used to interpret seawater temperatures. They may, however, despite diagenetic changes retain evidence for changes in temperature, such as cooling or warming trends. For reference, the oxygen isotope data are kept together with carbon isotopes in the text (Tables 1 and 2).

3. Geological overview

The Tethyan Himalaya, which lies between the High Himalaya Crystalline belt to the south and the Indus—Zangbo suture zone and the Lhasa continental block to the north (Fig. 1), consists primarily of Late Paleozoic to Early Eocene sedimentary rocks that were originally deposited along the northern margin of the India continent. During the Mesozoic, a relatively wide passive continental margin developed along the northern rim of the Indian continent. Northward subduction of the Neo-Tethyan oceanic crusts during the Cretaceous culminated in continental collision between the India and Lhasa blocks during the early Paleogene. The timing of the collision varies according to different authors, from the early Maastrichtian (Liu and Einsele, 1994; Willems et al., 1996) to the Eocene (Rowley, 1998; Searle...

In southern Tibet, the Tethyan Himalayas can be subdivided into two zones of different lithologic assemblages separated by the east-west striking Gyirong–Kangmar thrust (Burg and Chen, 1984). The northern zone is dominated by slightly metamorphosed deposits of outer shelf, continental slope and rise environments, whereas the southern zone is characterized by non-metamorphosed shallow-water shelf calcareous and terrigenous deposits, ranging from Paleozoic to Eocene in age (Wen, 1987; Willems et al., 1996).

This study concentrates on well-exposed Paleocene to Eocene sedimentary rocks in the Gamba area. The Zongpu sections studied here (Fig. 1) is located within the southern tectonic zone, where the youngest, well-documented marine strata of Paleocene–Eocene age are exposed (Wen, 1987; Willems et al., 1996; Xu, 2000). Because of an incomplete fossil record and a scarcity of stratigraphic studies the Paleocene–Eocene marine strata in the northern tectonic belt (Ding, 2003; Li et al., 2005) were not included in the present study.

4. Paleocene–Eocene stratigraphy

4.1. Lithology

The Zongpu sections contain a relatively complete succession of Paleocene and Eocene marine sediments. The sequence is composed of the Jidula, Zongpu and Zhepure Formations (Fig. 2) (Wan et al., 2002; Wen, 1974; Zhang, 1981). Willems (1993) established two stratigraphic systems for the “Tertiary” in southern Tibet, one for Gamba and the other for the Tingri area. In this study we use a single stratigraphic system as both Gamba and Tingri areas are part of the same sedimentary basin and have similar stratigraphic patterns.

In Gamba, the 177 m-thick Jidula Formation is a marine sequence dominated by yellowish-white, indurated, homogeneous sandstone, which is intercalated with sandy limestone beds in the middle and upper part of the formation. The exposure surface at the base of the formation is overlain by a 2–5 cm-thick yellowish-grey clay bed containing dispersed gravel. Foraminifera such as Rotalia hensoni Smout, Rotalia dukharni Smout, Smoutina crusi Drooger and Lockhartia haimei Davies were found in this clayey bed (Fig. 2). Ostracods and algae were found in limestone intercalations in the middle part of the formation. Collectively, the microfauna indicate a Danian age for the formation.

The overlying Zongpu Formation (223 m thick) consists of massive limestone (dolomitized in the lower section), nodular limestone, and minor amounts of calcareous marlstone and marl. In the field we subdivided this formation into four members, which from the bottom to the top are:

- **Member 1** is a medium bedded dark grey, partly dolomitized nodular packstone containing frequent dasycladacean algae, debris of Halimeda, and foraminifera.
- **Member 2** is a dark grey to blackish thickly bedded limestone sequence comprised of intercalated thick, dark colored wackestone and packstone beds. All beds are separated by laminae of marls. The foraminifera in this member belong to a Rotalia–Lockhartia assemblage, which indicates a Selandian age for the lower part of the formation (Wan, 1991). Limestones in Members 1 and 2 are intensively fractured and have a strong bituminous odor, indicating that they are enriched in organic carbon.
- **Member 3** is a dark-grey nodular limestone, wackestone and packstone composition, which in its uppermost 59 cm become a grey thick-bedded limestone breccia incorporating several thin layers of marl that lack larger benthic foraminifera. This marly bed was considered in the field to define an erosional surface. A larger foraminiferal fauna present is characterized by a Miscellanea–Daviesina–Oперculina assemblage, providing a Thanetian age for this member (Wan, 1991; Willems et al., 1995).
- **Member 4** is a grey massive and thick-bedded Alveolina packstone, with limestone beds separated by shaly intercalations 2–10 cm thick. This member forms a prominent ridge halfway up the northern flank of the Zongpu valley. It has a distinctive composition, characterized by masses of Alveolina, sporadically...
interspersed with Orbitolites (Wan, 1990). Member 4 was previously assigned to the overlying Zhepure Formation because its massive to thick-bedded texture is different from that of most nodular limestone in the underlying Member 3. However, it is distinct from overlying Zhepure shaly beds and we, therefore, assign it to Member 4 of the Zongpu Formation (Fig. 2).

The overlying Zhepure Formation contains greenish grey and red shale intercalated with thin-bedded sandy limestone yielding smaller benthic foraminifera. The top of the formation is truncated by a fault (Fig. 1).

**4.2. Biostratigraphy**

Benthic larger foraminifera are abundant in the Zongpu sections, especially in the Zongpu Formation. Fauna found in Members 1 and 2 of the Zongpu Formation—Rotalia dukharni Smout, Rotalia hensoni Smout, Rotalia orientalis Cushman and Bermudez, Rotalia sp., Lockhartia altispira Smout, Lockhartia conditi (Nut-tall), Lockhartia cushmani Applin and Jordan, Lockhartia diversa Smout, Lockhartia haimei Davies, and Smoutia crusi Smout—indicate a late Danian—Selandian age for the lower part of the formation.

A larger foraminiferal fauna is also present in Member 3, and includes Miscellanea miscella (d’Archiac and Haime), Miscellanea minor Sheng and Zhang, Miscellanea multicolumnata Sheng and Zhang, Miscellanea stampi (Davies), Daviesina khatiyahi Smout, Daviesina tenuis (Tambureau), Alveolina varians Hottinger, Alveolina ovulum (Stache), Operculina canalifera d’Archiac, Operculina patalensis Davies and Operculina subsalsa Davies. This microfaunal assemblage is widespread in the Tethys and is restricted to the Paleocene. It indicates a Thanetian age for the upper part of the Zongpu Formation. Miscellanea disappear toward the top of Member 3, and only some smaller benthic foraminifera remain.

Alveolina ellipsodalis (Schwager) first appears at the bottom of Member 4 and is followed by abundant species of Alveolina and Orbitolites. The dominant species are Orbitolites biplanes (Lehmann), Orbitolites contentinensis Lehmann, Operorbitolites gracilis (Lehmann), Alveolina ellipsodalis (Schwager), Alveolina cylindratus Sheng and Zhang and Alveolina oliviformis Sheng and

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**Figure 2** General columnar section of the Paleocene—Early Eocene sequence in Gamba 1. Packstone; 2. Marlstone; 3. Nodular wackestone and packstone; 4. Shale; 5. Sandstone.
5. Paleocene–Eocene boundary

5.1. Sediments and fossil zones across the P–E boundary

Foraminifera are the major microfossil group in sedimentary rocks of southern Tibet. In Gamba, 32 larger foraminiferal species belonging to 10 genera have been identified from Upper Paleocene and Lower Eocene strata (Fig. 3). Larger foraminifera show low diversity and high abundance. In abundance, the larger foraminifera are dominant, often accounting for 60%–80% of the rock sample, with the smaller benthos relatively fewer. The variations in diversity and abundance of the different groups are similar. The larger foraminiferal species show a clear evolutionary pattern that can be correlated with the Tethyan shallow benthic zones (SBZ) proposed by Serra-Kiel et al. (1998).

The P–E boundary is constrained by lithologic and faunal changes in the upper Zongpu Formation. The presence of larger benthic foraminiferal assemblages suggest that this formation was deposited in a neritic environment, probably a carbonate ramp, relatively close to land (Wan et al., 2002). For our high resolution biostratigraphic analyses, a 9 m thick interval from the upper part of Member 3 to the lower part of Member 4, which incorporates the P–E boundary, was further subdivided into 5 subunits (Fig. 3, right).

Subunit A (0–1.77 m) is a sequence of massive, dark-grey, thick-bedded foraminiferal wackestone and packstone. Foraminiferal fossils are dominated by larger species of Daviesina and Operculina. Lockhartia are also rich and dominated by L. haimei. It is a species occurring in SBZ 3–4 (Serra-Kiel et al., 1998), and marks the top of subunit A. According to its last appearance in the section, we suggest that subunit A is in the SBZ 4.

Subunit B (1.78–3.57 m) is a massive dark-grey nodular limestone intercalated with marls. Miscellanea are the dominant group of microfauna. Daviesina and Operculina are still frequent within this subunit. A. varians and D. tenuis first appear in this subunit. Both species define the SBZ 5 (Serra-Kiel et al., 1998), and therefore, this subunit belongs to SBZ 5.
Subunit C (3.58–4.17 m) is a grey thick-bedded limestone breccia incorporating several thin layers of marl. The breccia clasts are 5–20 cm in size (Fig. 4). This bed separates Member 3 from the overlying, thick-bedded Eocene Alveolina packstone. The breccia beds document a deformational episode tectonic background, which differs mostly from the wackestone and packstone composition of the Paleocene and may be indicative of a possible diastem separating Thanetian and Eocene strata (Fig. 4, between subunit C and D). However, in contrast to the underlying Paleocene carbonate platform, the lower part of the Eocene limestone has not been strongly tectonically fractured (Wan et al., 2002). According to its more or less continuous larger foraminiferal succession, this P–E contact surface is of less sedimentary break. Almost no larger foraminifera have been found from this subunit besides 4 species of Rotalia and Smoutina. Poorly preserved Discocyclina appears in the uppermost part of the subunit. As obscure of fossil evidence, this is looked as a transition at the P–E boundary, although we put the boundary between subunit C and D.

Subunit D (4.18–7.77 m) belongs to Member 4. It is a thick-bedded and massive grey Alveolina packstone, which overlies with a sharp contact the marlstone bed at the top of subunit C. The Alveolina packstone beds are intercalated with 3–5 cm thick intercalations of grey marl in the upper part of this subunit (Fig. 5). Alveolina ellipsodalis and two other local species first appear at the base of this subunit together with abundant Orpitotilites. A. ellipsodalis defines the SBZ 6 (Serra-Kiel et al., 1998). An important event is the extinction of Paleocene Miscellanea miscella. Two species of Orbitolites, three species of Alveolina and one species of Operitotilites are the first appearance of larger foraminifera. The overturn of larger foraminifera marks the P–E boundary, although it spans 59 cm in the section (Fig. 3).

Subunit E (7.78–9.07 m) is a sequence of greenish grey shale. We conducted our study only up to the base of this subunit, because it is almost barren of microfossils except for a few poorly preserved smaller foraminifera specimens. Fossil disappearance is attributed to lithological change, however strata above subunit E are not part of this report.

### Table 1 Stable isotope values from the Zongpu-1 Section.

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5.2. Benthic foraminiferal changes

Larger foraminifera which are sensitive to light-penetration and trophism changes (Drooger, 1983), form the dominant fossil group in the Gamba area, and show an evolutionary pattern. As shown in Fig. 3, the diversity of total foraminifera within subunit A is relatively high. In the Zongpu-1 Section, 22 species of larger benthic foraminifera were identified in lower subunit B, and their number was reduced in the upper part to 17. At the P–E boundary almost all larger foraminifera disappeared. All species begin to disappear progressively from the upper subunits in B. Most of the larger taxa are of limited range and become extinct toward the P–E boundary.

Subunit C is above the level of major extinction of Miscellanea, between subunit B and C. In addition to the extinction of Miscellanea, 9 species (4 genera) of foraminifera temporarily disappeared and only 4 species of foraminifera are present in subunit C. A few Miscellanea specimens were also found in this subunit. The extinction of benthic foraminifera occurred rapidly through 59 cm of sediment deposition, with a rapid loss of larger species. Only Operculina has a longer fossil record and higher abundance, and it also temporarily disappeared in subunit C. Some species of Rotalia and Smouthe existed throughout subunit C. Foraminifera reappear in the overlying subunit D, where larger taxa are all newcomers, except Operculina.

Following the extinction at the top of Member 3 (subunit C) in the southern Tibet region, recovery began rapidly in the lower part of Member 4 (subunit D). The recovery occurred at the base of subunit D and is marked by appearance of Orbitolites–Alveolina fauna in both abundance and diversity. Abundant Operculina are also present in the lower part of Member 4. Total taxa includes 8 new and 13 reappearing “Lazarus” foraminiferal species.

All larger foraminifera disappeared from subunit E and only 3 poorly preserved species remained, largely as the consequence of a lithological change from packstone to shale. It is considered here that larger foraminifera are very sensitive to minor changes of light penetration. The arrival of more clay sediments into basin diminished the penetration of light, and thereby negatively affected environmental living conditions. The occurrence of larger foraminifera is, therefore, controlled by sedimentary conditions.

Carbon isotopes were measured in the two boundary sections of Zongpu-1 and Zongpu-2. The δ13C values from 45 samples of the Zongpu-1 Section vary from −6.1‰ to 1.5‰. An isotopic profile shows three negative trends in beds containing the P–E transition in the Zongpu-1 Section (Fig. 6, Table 1).

Trend 1: In subunit A, the δ13C values vary from 0.1‰ to 0.7‰, with an average value of 5 samples 0.3‰. A rapid negative shift occurs in the upper part of subunit A from 1.27 to 2.87 m of the measured profile. The δ13C values dropped from 0.7‰ to −6.1‰ in only 40 cm of the sedimentary profile and the low value was sustained throughout most of subunit B. The δ13C values rise to pre-excursion values of 1.5‰ in the uppermost part of the subunit. This negative shift of δ13C values is the most pronounced we encountered, but it was not accompanied by any change in foraminiferal diversity. During subunit A and subunit B time, both larger and benthic smaller foraminifera were abundant.

Trend 2: A brief negative excursion of δ13C occurred in the uppermost part of subunit C, with values declining sharply from 1.5‰ to −4.0‰ at the position of the biostratigraphic Paleocene/ Eocene boundary. The samples were collected from marl intercalations and strongly avoided the previously noted breccias. The decline forms a sharp peak on the δ13C curve from ordinary sediments (Figs. 4 and 6). It is associated with the extinction of benthic foraminifera; however, the extinction had already begun in the upper part of subunit B, where the δ13C has background values.
Trend 3: The $\delta^{13}$C values returned to background values in subunit D, with an average value of $0.3\%_{\text{o}}$ and an observed increase in foraminiferal diversity. A third negative excursion occurred in the upper part of this subunit, where the $\delta^{13}$C value drops to $-3.5\%_{\text{o}}$ and fluctuates from $0.4\%_{\text{o}}$ to $-2.6\%_{\text{o}}$.

The $\delta^{13}$C values in subunit E display a still stronger fluctuation between $-4.8\%_{\text{o}}$ and $-0.2\%_{\text{o}}$, with the minimum value in the basal part of this subunit.

The $\delta^{13}$C values from 20 samples of the Zongpu-2 Section vary from $8.6\%_{\text{o}}$ to $2.2\%_{\text{o}}$. In general, carbon isotopes tend to decrease from the Upper Zongpu to the Lower Zhepure Formations. The isotopic profile shows two negative trend excursions in the P–E transition beds (Fig. 6, Table 2). The lower trend of $8.6\%_{\text{o}}$ is covariant with the trend 1 of the Zongpu-1 Section; and the upper trend of $7.9\%_{\text{o}}$ is covariant with the trend 2 of the Zongpu-1 Section (Fig. 6). The upper part of the Zongpu-2 Section has not been measured for isotopes by the present work. Consequently, trend 3 of the Zongpu-1 Section has not been recognized in the Zongpu-2 Section.

The $\delta^{13}$C curves of the Zongpu sections reveal a long-term decline interrupted by 3 negative excursions of $\delta^{13}$C values. Interestingly, the most extensive excursion is trend 1 which preceded the P–E boundary, but it is not associated with a foraminiferal extinction. Despite considerable fluctuation of $\delta^{13}$C values, distinct and sharp negative spikes of $-4\%_{\text{o}}$ (Zongpu-1) and $-8.6\%_{\text{o}}$ (Zongpu-2) $\delta^{13}$C occur at the P–E boundary and are concurrent with benthic foraminifera extinction. The observations of a negative spike of trend 2, therefore, suggest that the negative carbon isotope excursions within member 4 of the Zongpu Formation represent the boundary between the Paleocene and Eocene. This line of reasoning follows that used for locating other boundaries in the rock record using the marine carbon isotope record, for example, the Paleocene–Eocene Thermal Maximum (Bowen and Bowen, 2008).

### 6. Discussion on foraminiferal extinction at P–E boundary

The foraminiferal extinction at the P–E boundary has been recognized at many oceanic sites (Miller et al., 1987; Pardo et al., 1997; Thomas and Shackleton, 1996). Benthic foraminiferal extinctions have also been recorded in marginal and epicontinental basins (Speijer and Wagner, 2002). The extinction period is associated with a negative shift in oxygen-isotope values that signify an increase in bottom-water temperatures. Similarly, as oxygen isotopes drop at the P–E boundary, carbon isotopes do the same. Oceanic carbon isotope $\delta^{13}$C values dropped by $2.5\%_{\text{o}}$ at the onset of the P–E boundary (Bains et al., 1999; Bralower et al., 1997; Kennett and Stott, 1991), implying a large input of isotopically negative carbon to the ocean and atmosphere reservoir.

Biostratigraphic study of the P–E boundary in southern Tibet has revealed that the foraminiferal extinction occurring during the Late Paleocene was not a sudden event, but occurred stepwise. It is seen as a general decline in abundance, which begins in subunit A (Daviesina fauna). Rapid decline of diversity is notable at the top of subunit B (Miscellanea fauna). Low diversity then spreads throughout subunit C, where in its uppermost part the number of larger foraminifera drops almost to zero. A fluctuation of species

**Figure 6** Comparison of stable carbon and oxygen isotope curves between Zongpu-1 and Zongpu-2 sections (see lithological key given in Fig. 3).
Biostratigraphy of a Paleocene–Eocene Foreland Basin boundary

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diversity occurs within subunit D, with a renewed decline in larger foraminifera at the middle part of this interval. All larger foraminifera temporarily disappeared in subunit E, where its greenish grey shale lithology from muddy or turbid water provided an inhospitable sedimentary environment for them.

The larger foraminiferal extinction rate in southern Tibet is similar to that recorded in deep-sea deposits in other parts of the world, where analyses suggest that the Paleocene fauna disappeared gradually through the Late Paleocene, but that Eocene taxa appeared abruptly above the boundary. The earliest Eocene fauna were substantially different from the Paleocene fauna at the ordinal taxonomic level. The number of modern orders represented in the latest Paleocene is ~12% of the total assemblage, while immediately following the P–E boundary, it had increased to ~48% (Gunnell, 1998). The pattern of stepwise extinction of the Miscellanea–Daviesina assemblage and abrupt appearance of Orbitolina–Alveolina assemblage in the southern Tibet sections we studied are comparable with the faunal changes across the P–E boundary.

The total disappearance of larger foraminifera and apparent low diversity in subunit C of the Zongpu sections may reflect changes in environmental conditions for benthos in the neritic sea during the P–E boundary time. After the Paleocene the number of foraminifera rapidly increased, especially for larger ones that show a turnover of species. Among 21 species, 8 first appeared at the base of subunit D. In this interval, abundance and diversity of larger foraminifera reached their maximum, indicating a rapid faunal recovery. Diversity fluctuated in subunit E, coinciding with environmental changes as recorded by changes in lithology. Diversity is higher in the Alveolina packstone and lower in the shale beds.

When we compare changes in foraminiferal diversity with changes in δ13C values, the first negative shift of 6.1‰ in δ13C value (Trend-1), marks the top of subunit A. δ13C maintains a negative value of −5‰ to −4‰ δ13C level across the upper part of subunit A and subunit B, before returning to pre-excision values between subunits B and C. The second negative shift of −4‰ δ13C (Trend-2) occurs at the P–E boundary and is accompanied by a larger foraminiferal turnover. The third and final δ13C excursion (Trend-3) has a fluctuation pattern with the initial drop of −2.6‰ δ13C in the middle part of subunit D. Because the Gamba area may have been located closer to land, more terrigenous sediments and fresh water from the continent could arrive in the basin, and be the cause of fluctuating δ13C values. The varying input of nutrients from the land would clearly affect bioproductivity.

A variety of factors that have been considered as causing the extinction of foraminifera were decreased oxygenation and increased corrosiveness of deep and intermediate waters, changes in surface water productivity affecting deep ocean nutrient availability, an increase in bottom and surface oceanic water temperature, a pronounced increase in pCO2 in both atmosphere and oceanic waters and major change in ocean circulation (Nunes and Norris, 2006; Stott, 1992; Thomas, 1998). Data from southern Tibet show that it is unlikely that the first benthic foraminifera extinction event is associated with any change in temperature regime, or directly related to a change in bioproductivity. Higher organic carbon in subunits A and B document high surface productivity, as indicated during our field studies by strong bituminous odors in their sediments. This enrichment in organic carbon is probably the result of the development of dysoxic conditions below the bottom water–sediment interface. The broad negative δ13C excursion (Trend-1) during deposition of subunit B can not be explained by a major change in bioproductivity. However, a lithologic change from backstone to nodular limestone intercalated with marls, which characterizes subunit B, is broadly correlative with the carbon isotope event and thus presumably results from a change in the depositional environment. The δ13C (Trend-2) response is delayed when compared to the foraminiferal extinction, as evidenced by the fact that the isotope anomaly is recorded only in the upper half of the foraminiferal extinction event.

The biotic extinctions suggest different causes of foraminiferal extinctions and of δ13C negative excursions at the P–E boundary. Trend-1 occurred during a period of high surface productivity and is concurrent with a lithologic change to more marly strata composition. Trend-2 is located at the P–E boundary within the upper half of the foraminiferal extinction event. The third decrease in foraminiferal diversity, Trend-3 in the Early Eocene is also reflected by the δ13C negative excursion. It is concurrent with a lithologic change from backstones to calcareous shale, which encloses rare nammosaurs and a few small foraminifera. This may indicate a transgressive event and development of an inhospitable environment for large foraminifera.

Major biotic changes involve extinction of benthic foraminifera and diversification of planktonic foraminifera (Crouch et al., 2001; Kelly, 2002). Our data from southern Tibet provide a more complicated picture than shown by deep-sea site studies, presumably because of the foreland location of the Tibetan sedimentary sequence. A shallow-water carbonate platform or a carbonate ramp covered the southern Tibet area during the Late Paleocene–Early Eocene. The sedimentary sequence exposed in the Gamba area was deposited in a very shallow environment, not too distant from a landmass (Wan et al., 2002). The Zongpu-I Section records lithological changes in the upper Zongpu Formation during the Late Paleocene and Early Eocene. The loss of benthic foraminifera occurs at the P–E boundary and theoretically should be a result of the Paleocene–Eocene thermal maximum, although more detailed work is required to document this.

7. Conclusions

The Zongpu Formation in the Gamba area of southern Tibet yields abundant larger foraminifera. The formation has been subdivided into four lithological members. In the 9 m of strata studied in detail that bracket the P–E boundary, foraminifera show a stepwise disappearance below and a rapid recovery above the boundary bed. A foraminiferal turnover was identified between Members 3 and 4 of the Zongpu Formation, from the Miscellanea–Daviesina assemblage to Orbitolites–Alveolina assemblage. The Paleocene and Eocene boundary lies between SBZ 4 and SBZ 5, where it is marked by the extinction of Miscellanea miscella and the first appearance of Alveolina ellipsodalis and a large number of Orbitolites.

The extinction rate for larger foraminiferal species is 46% at the boundary section. This southern Tibet rate is similar to that in the rest of the world, with the Paleocene fauna disappearing gradually through the Late Paleocene, but with an Eocene taxa appearing abruptly above the P–E boundary.

The δ13C values within the boundary strata show three negative excursions—Trend-1 within the Late Paleocene, Trend-2 at the Paleocene–Eocene boundary, and Trend-3 in the Early Eocene. Trend-2 is a chemostratigraphic expression of the boundary, and compared to the foraminiferal extinction.
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