Discovery of the Paleocene-Eocene Thermal Maximum in shallow-marine Xigaze forearc basin, Tibet: implication to enhanced hydrological change

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Abstract

The Paleocene-Eocene Thermal Maximum (PETM, ~56 Ma) was one of the major global hyperthermal events in the geological past. The study of shallow-marine PETM records is crucial to understand the continental hydrological response to current global warming. This study presents the first detailed documentation of the PETM in the Xigaze forearc basin located along the northern active continental margin of the eastern Tethys Ocean, and illustrates the associated environmental and hydrological changes. Based on carbon-isotope stratigraphy, larger benthic foraminiferal biostratigraphy, and zircon U-Pb chronostratigraphy, the PETM event was identified within a siliciclastic unit in the middle of the largely carbonatic Jialazi Formation. Foraminiferal assemblages of Shallow Benthic Zone 4 are present below the siliciclastic unit, but are replaced by Shallow Benthic Zone 6 assemblages above the siliciclastic unit. High-resolution microfacies analysis indicates that the pre-PETM deposits consist of transgressive carbonate-ramp sediments followed by a sudden change to regressive syn-PETM siliciclastic rocks, followed in turn by renewed post-PETM carbonate-ramp deposition. Siliciclastic supply increased notably during the PETM, as indicated by the thickness of both sandstone and shale intervals, resulting in a temporary demise of the carbonate ramp. Provenance analysis does not indicate any major change in the source areas of terrigenous detritus through the early Paleogene. Increasing siliciclastic supply is thus chiefly ascribed to the intensification of seasonal precipitation and consequently increased hydrological circulation in the Gangdese arc during the PETM event.

Keywords：PETM; Tibet; Carbonate ramp; Gangdese arc; Carbon isotope excursion; Xigaze forearc basin

1 Introduction

The Paleocene-Eocene Thermal Maximum (PETM) was a short-lived global warming event that occurred around 56 Ma and lasted ~170-200 kyr, during which temperature rose between 5 °C and 8 °C in both surface and deep ocean waters (Zachos et al., 2001; Murphy et al., 2010; McInerney and Wing, 2011; Zeebe and Lourens, 2019). The onset of the PETM was marked by a large (ca. 3-6‰) negative carbon isotope excursion (CIE), caused by the injection of large amounts of ^{13}C-depleted carbon into the ocean-atmosphere system (Dupuis et al., 2003; McInerney and Wing, 2011). The source of such carbon emission is still debated, including the release of methane hydrates from the continental shelf (Dickens et al., 1997) or volcanic-gas emissions during emplacement of the North Atlantic Igneous Province (Jolley et al., 2002; Jones et al., 2019). The PETM, representing a typical hyperthermal event (Foster et al., 2018; Hu et al., 2020), caused severe climatic and environmental change, including oceanic acidification (Zachos et al., 2005) and anoxia (Nicolo et al., 2010), and extinctions and migrations in terrestrial and marine ecosystems (Kelly et al., 1998; Crouch et al., 2001; Scheibner et al., 2005; Wing et al., 2005).

As a potential ancient analogue to the present-day anthropogenic-forced climate change (Dickens et al., 1995; Thomas et al., 2002), the PETM has attracted much scientific attention during the last decades. Most articles dealing with the PETM event were focused on the source, volume, and/or triggering mechanisms of the greenhouse gas emission, on its oceanographic effects (e.g., acidification, carbonate-compensation-depth shoaling), on its duration, or on its biotic impact (McInerney and Wing, 2011). More recently, an increasing number of studies have tackled the changes in the hydrological cycle associated with the PETM. These studies have shown an abruptly and non-uniformly strengthened global hydrological cycle during the PETM (Bowen et al., 2004; Winguth et al., 2010; Carmichael et al., 2017, 2018). Precipitation increased at middle-high latitudes, whereas it decreased or remained unchanged at low latitudes but with increasing frequency of extreme precipitation events (Schmitz and Pujalte, 2007; Foreman et al., 2012; Giusberti et al., 2016). Most hydrological studies have focused on terrestrial settings (e.g., Bighorn basin, Wyoming, Foreman, 2014; Bass river, New Jersey, John et al., 2012; Nanyang basin, China, Chen et al., 2016; Dababiya, Egypt, Soliman et al., 2011), whereas studies on shallow-marine environments are few (e.g., middle-latitude Spanish Pyrenees; Schmitz and Pujalte, 2003; Pujalte et al., 2016; southern Tarim seaway, Li et al., 2020).

In this study, we provide new sedimentological, paleontological, and geochemical evidence from the Xigaze forearc basin, presently located in south Tibet and lying at subequatorial latitude along the northern active margin of the eastern Tethys Ocean during PETM time interval. Our principal aims are to document the PETM record in these shallow-marine, active-continental-margin sediments and to constrain environmental and hydrological changes across the PETM.

2 Geological setting

2.1 The Xigaze forearc basin

The Xigaze forearc basin is situated between the Yarlung Zangbo suture zone to the south and the Gangdese arc to the north, over a length of 510 km with a maximum width of 22 km (Fig. 1B). During the early Paleogene, the Xigaze forearc basin was located along the northern margin of the Tethys Ocean, an east/west-trending remnant subequatorial seaway in the northern hemisphere (Scotese, 2013). The forearc basin was filled mainly by Cretaceous deep-water turbidites and minor carbonates, and became a syn-collisional basin in the early Paleogene (Wang et al., 2012; Orme et al., 2015; Hu et al., 2016b), after middle Paleocene onset of the India-Asia collision (DeCelles et sl., 2014; Ding et al., 2017; Hu et al., 2015, 2017). Paleomagnetic data indicate that during the early Paleogene the Xigaze forearc basin was situated in the northern hemisphere between 12.9 ± 4.6° N (Yi et al., 2011) and 24.2 ± 5.9° N (Meng et al., 2012).

2.2 Stratigraphy of the Cuojiangding area

Our study area is located near the glacier lake of Cuojiangding (Zhongba county, south Tibet, Fig. 1C) in the western part of the Xigaze forearc basin, where shallow-marine and deltaic strata of late Maastrichtian to earliest Eocene age are continuously exposed (Fig. 1D; Quxia and Jialazi formations, Qian et al., 1982). The Quxia and Jialazi formations crop out in a syncline separated by thrusts from the Gangdese arc to the north and from the Zhongba ophiolitic mélange to the south (Wan et al., 2001).

The Quxia Formation is 110-120 m-thick (Hu et al., 2016b). The lower part of the unit comprises greenish-gray and purplish-red mudrock with intercalated sandstone, whereas the upper part consists of numerous cyclothems of conglomerate, sandstone, and mudrock, testifying to fan-delta sedimentation in the earliest collisional stage. Based on the larger benthic foraminiferal (LBF) biostratigraphic ages of the underlying and overlying stratigraphic units, the depositional age is only loosely constrained as early-middle Paleocene (Danian-Selandian; Wan et al, 2001; Hu et al., 2016b).

The Jialazi Formation is ~200 m-thick, and conformably overlies the Quxia Formation (Hu et al., 2016b). The 80 m-thick lower part comprises sandy limestone and shale with intercalated sandstone, whereas the 50 m-thick middle part consists of thick-bedded to massive sandstone with mudrock interbedded with several tuff layers. The 70 m-thick upper Jialazi Formation again comprises shale and limestone with interbedded sandstone (Hu et al., 2016b). Foraminiferal assemblages changed from

Shallow LBF Zone (SBZ, Serra-Kiel et al., 1998) 3 and SBZ4 to SBZ5, indicating a late Paleocene to earliest Eocene age (Hu et al., 2016b). Tuff layers exposed in the middle Jialazi Formation yielded a weighted $^{40}Ar/^{39}Ar$ age of 56.3 ± 2.1 Ma (Ding et al., 2005). Zircon crystals contained in the tuffs yielded as a youngest group of U-Pb ages 54.9 ± 0.7 Ma and 55.7 ± 0.5 Ma (Hu et al., 2016b). The top of the Jialazi Formation is thus constrained as not older than the early Eocene (54 ±1 Ma). Kahsnitz et al. (2017) distinguished four carbonate microfacies in the Jialazi Formation (molluskan floatstone/rudstone, Nummulitidae wackestone/packstone, rhodolith wackestone/packstone, and Discocyclinidae-Nummulitidae floatstone/rudstone), which allowed identification of two events of deepening paleowater depths.

3 Sampling and methods

Samples were collected from the newly measured Goukou stratigraphic section (29° 55' 45.90" N, 84° 19' 14.70" E, 5233 m a.s.l.) and from the Quxia B section of Hu et al. (2016b) (29° 56' 19.20" N, 84° 19' 32.10" E, 5444 m a.s.l.) (Figs. 1 and 2). Sampling frequency is ~1 m for carbonates and variable for siliciclastic rocks. Overall, 95 carbonate samples were collected for biostratigraphy, microfacies and carbon-isotope analysis, and four tuffs and one sandstone were collected for zircon U-Pb dating and Hf isotope analysis.

3.1 Microfacies analysis

Microfacies analysis was based on fossil assemblages, textures, sedimentary structures, detrital minerals, and grain composition, observed in both outcrop and thin section. Carbonates and mixed carbonate-siliciclastic rocks (siliciclastic detritus <10% and >10%) were described based on Dunham (1962; integrated by Embry and Klovan, 1971) and Mount (1985), respectively. Environmental interpretation was carried out according to Beavington-Penney and Racey (2004) and by comparison with the standard microfacies and depositional model of Flügel (2010). To distinguish between siliciclastic and carbonate-related facies, we define the former as siliciclastic facies (SF) and the latter as carbonate facies (CF).

3.2 Carbon and oxygen isotopes

Carbon-isotope analyses were performed on 65 carbonate samples from the Quxia B section of the Jialazi Formation (see data in Table S1). Powdered samples were obtained by micro-drilling, taking care to avoid cement-filled veins and pores as well as bioclasts. The carbon and oxygen-isotope ratios of powdered samples were measured at the State Key Laboratory for Mineral Deposits Research, Nanjing University, using a Finnigan MAT Delta Plus XP mass spectrometer coupled with an in-line GasBench II autosampler. Samples were reacted with purified orthophosphoric acid at 70°C. Isotopic measurements were calibrated to Chinese national standard calcium carbonate sample GBW04405 ($\delta^{13}C_{VPDB}$ = 0.57‰ ± 0.03‰; $\delta^{18}O_{VPDB}$ = -8.49‰ ± 0.14‰). Data are expressed in standard delta notation (δ), as permil deviations from the Vienna Pee Dee Belemnite standard (VPDB). Duplicate measurements of standards yielded an analytical precision (1σ) of 0.05‰ for $\delta^{13}C$ and 0.07‰ for $\delta^{18}O$.

3.3 Zircon U-Pb dating and Hf isotope analyses

Zircon U-Pb ages of four tuff samples from the Jialazi Formation (data provided in Table S2) were conducted by Secondary Ion Mass Spectroscopy (SIMS) at the State Key Laboratory of Isotope Geochemistry, Guangzhou Institute of Geochemistry, CAS, China. U-Pb ages of zircon grains in the sandstone sample 09QXB21 (data provided in Table S4) were measured by LA-ICP-MS at the State Key Laboratory of Mineral Deposits Research, Nanjing University, following Jackson et al. (2004). Zircon U-Pb age calculations were made using Isoplot 4.15 (Ludwig, 2011). Hf isotope analysis of the four tuff samples (data provided in Table S3) were performed using Thermal Fisher Neptune Plus MC-ICP-MS coupled with a NewWave 193UC ArF laser-ablation system following Griffin (2000) at the Mineral Laser Microanalysis Laboratory, China University of Geosciences (Beijing). The 91500 standards were analyzed in every run, yielding $^{176}Hf/^{177}Hf = 0.282315 \pm 0.000021$ (2σ; n = 20), which is identical to the literature value of $^{176}Hf/^{177}Hf = 0.282313 \pm 0.000008$ (2σ, Blichert-Toft, 2008). Hf isotope data were processed using Iolite (Paton et al., 2011).

4 Facies and microfacies analysis

The ~60 m-thick Goukou section, exposing the lower part of Jialazi Formation, mainly consists of marl and limestone intercalated with sandstone. The ~200 m-thick Jialazi Formation, completely exposed in the Quxia B section, is described above in subsection 2.2. Seven carbonate facies (CF) and two siliciclastic facies (SF) document four distinct depositional environments (Fig. 3-6) in the measured Quxia B and Goukou sections.

4.1 Fan-delta front

4.1.1 SF1 Shale interbedded with sandstone

SF1 consists of grayish-black shale interbedded with thin-bedded and medium to fine-grained sandstone at the base of the Jialazi Formation, where the proportion of mudrocks increases upward (Fig. 2D). Sandstone beds, locally showing ripple lamination (Fig. 2E), contain moderately sorted and angular to rounded quartz and volcanic rock fragments cemented by calcite (Fig. 3A).

SF1 indicates a delta front environment, in which shale represents prodelta bottomsets and sandstones delta-front to sheet-sand deposition (Fig. 6; Miall, 1996).

4.1.2 SF2 Sandstone interbedded with shale

SF2 comprises thick-bedded sandstone interbedded with greyish-green shale or silty shale in the middle Jialazi Formation, and includes several thin tuff layers. Normal grading and parallel lamination are only locally observed at the base of sandstone beds. Body fossils and trace fossils are lacking. Sandstone beds are mainly 1-3 m-thick and may reach 5 m, whereas shale intervals are 2-3 m-thick; the shale/sandstone ratio is ~ 1 (Fig. 2F, 7). Sandstones contain poorly sorted and angular to rounded quartz grains, more feldspar than in SF1, and volcanic rock fragments set in calcite cement (Fig. 3B).

Massive to thick-bedded sandstone and silty shale indicate a sharp progradation relative to the lower Jialazi Formation, where sandstone beds are only10-50 cm-thick, and the shale/sandstone ratio is ~ 2 (Fig. 7). Graded bedding and parallel lamination indicate deposition by tractive currents in upper flow regime. SF2 documents a fan-delta environment reflecting rapid progradation after replacement of carbonate sedimentation (Fig. 6).

4.2 Lagoon

4.2.1 CF1 Sandy bioclastic wackestone

CF1 consists of black shale interbedded with sandy bioclastic limestone (wackestone) and sandstone in both lower and upper parts of the Jialazi Formation. Wackestones contain common (10-30%) bioclasts (mainly cm-sized gastropods, bivalves and miliolids, with green algae, echinoderms, and ostracods). Most mollusk shells are filled with sparry calcite and commonly display double-layer structure. Burrows occur (Fig. 3C). Fine- sand-sized, moderately sorted, locally rounded quartz and volcanic lithic fragments set in micritic matrix represent 5-20% of the rock.

Miliolids and gastropods suggest a restricted environment, and abundant bivalves indicate a shallow-water euphotic zone (Scholle and Ulmer-Scholle, 2003). Based on the abundance of micrite (testifying to low water energy), together with broken grains (indicating rapid transportation of coarse-grained terrigenous detritus presumably from a nearby river mouth), CF1 testifies to a restricted lagoon affected by terrigenous input (Fig.6).

4.3 Shallow marine

4.3.1 CF2 Sandy Ranikothalia floatstone/ rudstone

CF2 consists of black shale interbedded with sandy bioclastic limestone (mainly floatstone or rudstone) in the lower Jialazi Formation. Abundant (20-40%), mostly long and flat or locally broken foraminifera are mainly *Ranikothalia* with *Daviesina* and *Miscellanea* set in micritic matrix (Fig. 3D). Bivalves and echinoderm fragments also occur.

The high faunal diversity indicates a relatively open environment. The long and flat shape of *Ranikothalia* indicates near in situ deposition in a quiet environment, as suggested by abundant micrite (Baumgartner-Mora and Baumgartner, 2016) and some terrigenous input. CF2 may thus represent an open marine environment above fair weather wave base (FWWB, Fig.6).

4.3.2 CF3 Sandy LBF wackestone

CF3 mainly consists of gray-black silty shale with thin-bedded sandstone and bioclastic wackestone in the lower Jialazi Formation. Carbonate grains are mainly (10-25%) bioclasts, including foraminifera (*Ranikothalia* and *Assilina* with *Daviesina* and *Miscellanea*), bryozoans, bivalves, fragments of corallinacean algae, miliolids, and gastropods. Matrix includes both micrite and quartzose silt (Fig. 3E).

The high faunal diversity including *Assilina* and *Miscellanea* indicates a relatively open environment with well water circulation above FWWB (Fig. 6; Beavington-Penney and Racey, 2004; BouDagher-Fadel, 2018).

4.3.3 CF4 Bioclastic wackestone/ floatstone

CF4 mainly consists of medium-bedded bioclastic limestone (wackestone/ floatstone) interbedded with marlstone. Bioclasts are mainly foraminifera (*Discocyclina*, *Assilina*, *Ranikothalia*), bivalves, echinoderms, bryozoans, fragments of corallinacean algae, and intraclasts. The grain size is mainly 0.5-1 mm and can reach more than 2 mm. Poorly sorted grains are dispersed in micrite (Fig. 3F).

As indicated by highly diverse fauna including *Discocyclina* and *Assilina*, CF4 documents an open marine environment near the FWWB (Fig.6; Beavington-Penney and Racey, 2004; BouDagher-Fadel, 2018).

4.4 Middle-outer ramp

4.4.1 CF5 LBF rudstone

CF5 mainly consists of thick-bedded bioclastic limestone (rudstone) in the upper Jialazi Formation exposed in the Quxia B section. Bioclasts are mostly (> 90%) foraminiers (*Discocyclina*, *Assilina*, *Nummulites*, and subordinate *Miscellanea*, *Operculina*, *Ranikothalia*, and *Daviesina*); echinoderm fragments occur. Foraminifera are locally oriented, with shell locally filled with greigite (melnikovite). The grains are mainly larger than 2 mm, and the whole is the grain supporting. Three sub-microfacies are identified based on the dominant foraminiferal group: 1) CF5 with *Nummulites*; 2) CF5 with *Discocyclina* and *Assilina* (occurring in the upper part of the Goukou section; Fig. 3H); and, 3) CF5 with *Nummulites*, *Discocyclina*, and *Assilina* (Fig. 3G).

Abundant and varied bioclasts indicate a relatively open environment. *Nummulites* thrive in waters 10-60 m-deep (BouDagher-Fadel, 2018), and the *Discocyclina-Assilina* assemblage points to a middle-outer ramp environment (Geel, 2000). Semi-oriented foraminifera and coarse-grained bioclasts suggest high-energy conditions, probably reached during storm events, and thus a middle-outer ramp environment between fair-weather and storm wave base (Fig. 6).

4.4.2 CF6 Mudstone

CF6 is mainly composed of thin-bedded structureless mudstone intercalated with black shale in the middle Jialazi Formation. Micrite includes tiny broken bioclastic particles and quartzose silt locally. (Fig. 3I).

Dominant mud and lack of bioclasts indicate a low-energy environment below the euphotic zone, possibly a middle-outer ramp setting below wave base (Fig. 6).

4.4.3 CF7 Bioclastic packstone

CF7 is mainly composed of gray thick-bedded bioclastic limestone interbedded with grayish-black thin-bedded marlstone in the lower to middle Jialazi Formation. Bioclasts (30-60%) are mainly red algae, rodophytes and coralline algae, *Distichoplax biserialis*, and LBF (*Daviesina*, *Discocyclina*, *Assilina*, *Miscellanea*, *Ranikothalia*, *Nummulites* and textulariids), with subordinate planktonic foraminifera (*Morozovella*), bivalves, echinoderms and bryozoans set in locally burrowed micrite (Fig. 3J).

Although the algal flora characterizes a euphotic intertidal - subtidal environment (Flügel, 2010; Sarkar, 2018), *Miscellanea* and *Discocyclina* indicate a middle-outer ramp environment (BouDagher-Fadel, 2018), and the presence of planktonic foraminifera with lack of terrigenous detritus confirms an outer ramp environment below wave base and free from terrigenous influence (Fig. 6). The algal assemblage is thus inferred to have been reworked in deeper waters.

5 Stratigraphic age

5.1. Larger benthic foraminifera biostratigraphy

In this subsection, we shall focus on the stratigraphic interval spanning the Paleocene-Eocene boundary. The adopted biostratigraphic scheme, based on LBF, follows Serra-Kiel et al. (1998), revised by Boudagher-Fadel (2018). The distribution of benthic foraminifera is shown in Figure 4 and Figure S1.

The LBF assemblages of the lower Jialazi Formation mainly consist of *Rotorbinella hermi*, *Miscellanea yvettae*, *Ranikothalia sindensis* (Fig. S1A), belonging to the upper SBZ3 (Hottinger, 2011). The boundary between SBZ3 and SBZ4 is defined by the first appearance of *Discocyclina sella* (Fig. S1G, BouDagher-Fadel et al., 2015). Within SBZ4, *Miscellanea yvettae*, *Ranikothalia sindensis*, *Ranikothalia nuttalli*, *Daviesina langhami*, *Nummulites globulus* are common. The boundary between SBZ4 and SBZ5 is defined by the first appearance of *Miscellanea miscella* (Fig. S1B, BouDagher-Fadel et al., 2015). Within the lower SBZ5, *Daviesina langhami*, *Assilina yvettae*, *Miscellanea julliettae* are common, and *Discocyclina seunesi, Assilina laminosa, Assilina subspinosa* also occur. Within the upper SBZ5 to early SBZ6, *Nummulites globulus*, *Discocyclina sella*, *Miscellanea miscella*, *Assilina dandotica* are common, and *Ranikothalia sindensis, Daviesina langhami, Assilina subspinosa, Daviesina tenuis* also occur. The SBZ6 is defined by the disappearance of *Miscellanea miscella* and and the appearance of *Daviesina salsa*, (Fig. S1C), *Daviesina ruida* (Fig. S1D), *Nummulites globulus* (Fig. S1E), *Discocyclina seunesi*, and *Assilina granulosa* are common within this zone, *Discocyclina dispansa* (Fig, S1Fa), *Assilina leymeriei* (Fig. S1Fb), *Discocyclina sella*, *Nummulites mamillatus*, and *Daviesina* spp. also occur.

According to LBF biostratigraphy, the lower Jialazi Formation belongs to upper

SBZ3-early SBZ5 (59.2 - 56.0 Ma; Thanetian) and the upper Jialazi Formation to SBZ5-SBZ6 (56.0 - 54.9 Ma; early Ypresian). The middle Jialazi Formation lacks fossils but can be safely assigned to SBZ5 by its stratigraphic position.

5.2 Zircon U-Pb ages from tuffs

The studied grayish and medium- to thin-bedded tuff layers consist mostly of glass shards (> 70%) with devitrified lithic fragments, minor angular quartz crystals, and locally altered mica. The cathodoluminescence image of zircon grains varies from regular hexahedron to plate columnar (Fig. 8A). Zircon-age distributions are dispersed rather than unimodal (Fig. S2), indicating mixing with detritus eroded from pyroclastic debris flows and/or older siliciclastic rocks. Nevertheless, the age of the young peak is normally distributed, and its weighted age thus constrains the age of magmatic activity. As for zircon grains in sandstones (Dickinson and Gehrels, 2009), inferences on (maximum) depositional age must be obtained from the age of the youngest age population, including zircons dated as 56.1 ± 0.8 Ma, 55.9 ± 0.8 Ma, 55.3 ± 0.9 Ma, and 55.1 ± 1.0 Ma (Fig. 8A, Table S5). The deposition of the middle Jialazi Formation can thus be determined both accurately and precisely as very close to the Paleocene/Eocene boundary (~56 Ma; Gradstein et al., 2012).

6 Provenance analysis

6.1 Data

U-Pb dating of 100 zircon grains from sandstone sample 09QXB21 of the Jialazi Formation yielded 85 concordant ages (Fig. 9). Most (75%) are Cretaceous, clustering between 132 and 66 Ma; Paleogene ages between 66 and 55 Ma are also common.

Analysis of 105 zircons from four tuff samples of the Jialazi Formation, all yielding concordant ages between 56 and 60 Ma, yielded negative and positive εHf (t) values, ranging between -4.0 and +4.5. No obvious correspondence is observed between zircon ages and εHf (t) values (Fig. 8B).

6.2 Interpretation

Our focus in presenting these data here is not to provide a detailed provenance analysis of Paleogene sandstone but to determine if significant changes in sediment source areas took place between pre-PETM and post-PETM sandstones.

Sandstones of the Jialazi Formation contain abundant andesitic to rhyolitic detritus derived from a volcanic arc. Based on south-directed paleocurrents (Ding et al., 2005), geochemistry of detrital Cr-spinel, and detrital zircon U–Pb spectra, their provenance from the active Gangdese arc was firmly constrained (Hu et al. 2016b). Provenance from the Gangdese arc in the southern part of the Lhasa block is supported further by the U–Pb zircon ages obtained in this study. Zircon ages show two major peaks (50-69 Ma, 80-120 Ma, Fig. 9), which are similar to results from the Lopu Kangri area (Orme et al., 2015) and compare well with those from Xigaze forearc sandstones (Wu et al., 2010; Aitchison et al., 2011; An et al., 2014). Differences from data reported in Hu et al. (2016b) include the lack of an age peak at ~75 Ma, for which three possible reasons are envisaged: 1) a bias induced by the depositional ages of the investigated samples and by the dominance of Eocene magmatism in the Gangdese arc; 2) a gap in magmatic activity around ~75 Ma in the source area of Cuojiangding sandstones; 3) unroofing processes with removal of rocks of this age from the Gangdese arc.

The tuffs were largely generated from penecontemporaneous Gangdese arc volcanism. The εHf (t) values more negative than in Gangdese arc rocks (Hou et al., 2015) may reflect mixing with country-rock material during zircon growth in the magmatic chamber, although external sources of volcanic ash transported across the stratosphere cannot be ruled out. Moreover, multimodal U-Pb zircon-age spectra (Fig.

S2) reveal mixing sources from diverse pyroclastic products of the Gangdese arc or possibly recycling of older siliciclastic or volcaniclastic rocks. All the available evidence indicates dominant provenance of tuff and sandstone layers from the Gangdese arc for the entire Jialazi Formation, with no manifest provenance change through time.

7 Carbon isotope stratigraphy

In the Quxia B section, the whole-rock inorganic-carbon isotope curve is well documented for the lower and upper parts of the Jialazi Formation, but not in the middle part where carbonate rocks are lacking (Fig. 4). Overall, the $\delta^{13}C$ values of pure carbonate range between -2.0‰ and +1.4‰, similar to the $\delta^{13}C$ record of Tethys Himalayan carbonates deposited on the southern margin of the eastern Tethys ($\delta^{13}C$ from -4.0‰ to +2.5‰ after Li et al., 2017, and from -4.5‰ to +2.9‰ after Zhang et al., 2017). This indicates that the Jialazi carbonates of the Cuojiangding area have chiefly retained their original carbon-isotope information. In the lower Jialazi Formation, $\delta^{13}C$ values range from -4‰ to +1.4‰, and finally cluster around +1.2‰. At the base of the siliciclastic middle Jialazi Formation, $\delta^{13}C$ abruptly decreases to the extremely negative value of -14.1‰. Above, in mixed carbonate-siliciclastic rocks of the upper Jialazi Formation, $\delta^{13}C$ values gradually increase again from -14.2‰ to -2.0‰ and eventually return back to pre-CIE levels clustering around 0‰. Although the carbon-isotope curve is incomplete, the negative excursion typical of the PETM is clearly documented, indicating that the siliciclastic rocks of the middle Jialazi Formation were deposited during the core of the PETM interval.

8 The PETM event in the Xigaze forearc basin

8.1 Diagenetic effects on carbon isotope curves

Dissolution and recrystallization of carbonate minerals during early diagenesis can significantly alter their carbon-isotope composition, usually resulting in a decrease of $\delta^{13}C$ and $\delta^{18}O$ values (Banner and Hanson, 1990). At higher temperatures during deep burial or metamorphism, the original $\delta^{13}C$ and $\delta^{18}O$ signal may be completely lost (Knauth and Kennedy, 2009). The carbon-isotope ratio is also affected by the diagenetic transformation of aragonite and high-Mg calcite into low-Mg calcite, or by occurrence of skeletal grains exhibiting non-equilibrium isotopic fractionation (Immenhauser et al., 2002).

Carbonate strata in the Jialazi Formation are mainly wackestone and packstone containing bioclasts preserving their original biological structure, and the matrix is mainly micritic calcite. However, microsparite is also common, indicating that some recrystallization took place during diagenesis (Ingalls, 2019). The extent of such transformation can be evaluated by checking the correlation of carbon and oxygen isotopes, because carbon and oxygen isotopes covary when seawater is mixed with atmospheric fresh water (Marshall, 1992; Knauth and Kennedy, 2009). The $\delta^{18}O$ *vs.* $\delta^{13}C$ correlation diagram in Figure 10 shows a relatively poor relationship between $\delta^{18}O$ and $\delta^{13}C$ ($r = 0.40$, $p < 0.01$, $n = 63$) and the distribution of $\delta^{18}O$ and $\delta^{13}C$ values is largely independent, indicating that, despite minor alteration, diagenetic homogenization did not occur. Apart from the extremely negative values recorded in the siliciclastic middle Jialazi Formation, which may be ascribed to dominantly terrigenous detritus supply and possible meteoric influx (Knauth and Kennedy, 2009), carbonate strata in the studied sections have undergone only limited diagenetic alteration, implying that isotopic information on the original seawater is largely retained. Notably, the result of this study is similar to that of Quade et al. (2020), who measured

$\delta^{13}C$ values in the same forearc- basin stratigraphic unit exposed near Lopu Kangri, and concluded that, contrary to the oxygen-isotope record, the carbon-isotope record has retained the original formation values. As shown in Fig. 10, the $\delta^{13}C$ values from Lopu Kangri overlap with our data, including the most negative value (-11‰), which encourages us to interpret even the most extreme negative $\delta^{13}C$ values obtained in our study, from the siliciclastic middle Jialazi Formation and immediately overlying strata, as original isotopic information.

8.2 The carbon isotopic record of the PETM

The $\delta^{13}C$ value ranges from -4‰ to +1.4‰ (-1.2‰ on average) in the lower Jialazi Formation, and suddenly drops to -14.1‰ at the base of the siliciclastic middle Jialazi Formation, marking a large negative carbon isotope excursion (CIE). The $\delta^{13}C$ value gradually increases from -13.6‰ to +0.6‰ in the upper Jialazi Formation, eventually reaching a stable value. According to the definition recommended by the International Commission on Stratigraphy (ICS; Aubry et al., 2007), the P/E boundary should be placed at the onset of the CIE. Although the central part of the CIE is not recorded in this study, the onset and recovery are well documented (Fig. 4), indicating that the massive siliciclastic rocks of the middle Jialazi Formation were deposited during the PETM event.

The age of the middle Jialazi siliciclastic interval exposed in the Cuojiangding area is also well constrained by both zircon chronostratigraphy and LBF biostratigraphy. The youngest zircon grains found in the interbedded tuffs are dated as ~56 Ma (Fig. 8, Table S5). Foraminiferal assemblages of SBZ4 are present below the siliciclastic unit and replaced by SBZ6 assemblages above it, which thus can be safely assigned to SBZ5 by its stratigraphic position (Fig. 4). These data allow us to pin-point the P/E boundary more firmly than generally defined by LBF in shallow Tethyan Ocean. In the lack of a suitably complete reference section, the P/E boundary was in fact originally placed by Serra-Kiel et al. (1998) at the SBZ5/SBZ6 boundary. Subsequently, it has been variously considered to correspond to the SBZ4/SBZ5 boundary, or to occur within SBZ5 or between SBZ4-5 and SBZ6 zones (Scheibner and Speijer 2008; Zhang et al., 2019). Our data (Fig. 4) document that the major LBF taxonomic turnover mostly took place between zones SBZ5 and SBZ6, and thus notably later than the isotopically defined P/E boundary. This confirms that there is inconsistency between the micropaleontological biostratigraphic boundary and that chosen by the ICS which is based on the isotopic record. This anomaly should be addressed by future work. The CIE amplitude in the Cuojiangding area reached -15.3‰, which is much larger than what recorded in shallow-marine settings elsewhere (from -2‰ to ca. -4.5‰; McInerney and Wing, 2011), possibly owing to a prominent effect of terrigenous detritus, restricted circulation, or meteoric influx (Knauth and Kennedy, 2009). Just like the study in northern Iraq, where the observed CIE amplitude reached -13‰, and it was ascribed to the effect of diagenesis (Al-Fattah et al., 2020). Alternatively, it may however even represent the original signal (Fig. 10; Quade et al., 2020), which would mean that more light carbon was released in the Xigaze forearc basin during the PETM than elsewhere.

8.3 Environmental changes across the PETM

High-resolution facies and microfacies analysis indicates two transgressive events in the Jialazi Formation, which is similar with the result in Kahsnitz et al. (2017). Both transgressive events are recorded in the Quxia B section (Fig. 4), whereas only the older one is testified in the Goukou section because of incomplete exposure of younger strata (Fig. 5). The first deepening event starts at the base of the Jialazi Formation, with fan-delta (SF1) to restricted low-energy lagoonal deposits (CF1) passing gradually up-section to open-marine rudstone with *Ranikothalia* (CF2), wackestone with *Assilina* and *Ranikothalia* (CF3), and floatstone with *Discocyclina*, *Assilina* and *Ranikothalia* (CF4). Above, middle-outer ramp mudstone and packstone with red algae, rodophyte and coralline algae, *Distichoplax biserialis* and LBF (CF6 and CF7) document the deepest paleowater conditions reached in the Cuojiangding area. Subsequently, a major forced regression is marked by the transition to thick-bedded or massive siliciclastic rocks (SF2) of the middle Jialazi Formation, documenting rapid progradation of terrigenous supply and continental runoff. The second deepening trend begins in the upper Jialazi Formation, with deposition of floatstone (CF5) with abundant foraminifera in a middle ramp environment.

The combination of facies analysis, LBF biostratigraphy, and zircon chronostratigraphy precisely constrains the evolution of sedimentary environments across the PETM event in the Cuojiangding area. The pre-PETM (lower Jialazi) deposits consist of transgressive carbonate-ramp sediments followed by a sudden change to regressive syn-PETM siliciclastic rocks (middle Jialazi), followed in turn by renewed post-PETM (upper Jialazi) carbonate-ramp deposition.

8.4 Hydrological changes during the PETM

8.4.1 The continental hydrological response in the Xigaze forearc basin

The sharp transition between the lower and middle Jialazi Formation, corresponding to the onset of the PETM event, marks a prominent environmental change that may have had multiple tectonic and/or climatic triggers. Although a pulse of tectonic activity cannot be ruled out, because the India-Asia collision was underway (Hu et al., 2016a), provenance analysis indicates no manifest change in the source of terrigenous detritus through the early Paleogene (Fig. 9). Conversely, the coincidence of enhanced siliciclastic supply during the PETM suggests that the temporary demise of the carbonate ramp and replacement with terrigenous deposits was determined by a pluvial episode. A concurrent factor may be represented by enhanced volcanic activity in the Gangdese arc, as documented by numerous tuff beds intercalated within the middle Jialazi Formation.

Paleomagnetic data indicate that around the Paleocene/Eocene boundary the Xigaze forearc basin was located at low latitude in the northern hemisphere (Yi et al., 2011; Meng et al., 2012). Climate modelling suggests that a major increase in precipitation at low latitudes during the PETM is less likely than a significant increase in the intensity of seasonal precipitation (Carmichael et al., 2017, 2018). An increase in rainfall intensity would have led to increased weathering, erosion, runoff, and sediment-transport capacity, and consequent increase of the volume and grain size of terrigenous detritus (Armitage et al., 2011). Although other autocyclic or allocyclic factors including local tectonic uplift, paleo-drainage changes, or lateral shift of river mouths may be involved, the most likely explanation of drastic environmental change is an increase of precipitation intensity in the Gangdese source area during the PETM (Fig. 11). The restriction in vegetation growth and enhanced physical weathering during the PETM (Wing et al., 2005; McInerney and Wing, 2011), with consequent strong increase in the supply of fine-grained sediment during flood events (Schmitz and Pujalte, 2003; Pujalte et al., 2016), would explain the occurrence of mudrock intercalations as thick as 1-3 m within the thick-bedded sandstones of the middle Jialazi Formation.

8.4.2 Comparison with the Tethyan Himalaya

After middle Paleocene onset of the India-Asia collision, the Xigaze forearc basin became a syn-collisional (foreland) basin (Wang et al., 2012; Hu et al., 2016b). Lower

Paleogene strata deposited close to the suture zone underwent rapid tectonic uplift and erosion soon after deposition and were thus preserved only locally (e.g., Cuojiangding and Lopu Kangri areas; Orme et al., 2015; Hu et al., 2016b; Fig. 1C). This study of the Cuojiangding area represents the first detailed report of the PETM event in the Asian active continental margin. Similar environmental changes occurred in the Lopu Kangri area, situated ~30 km to the east of Cuojiangding and much closer to the Gangdese arc (Fig. 1C), where Orme et al. (2015) reported a change from carbonate platform deposits to siliciclastic rocks also probably related to climate and hydrological changes during the PETM.

In the Tethyan Himalaya, representing the southern margin of the eastern Tethys Ocean, the PETM event was identified and illustrated in the Tingri and Gamba areas by Zhang et al. (2013, 2017, 2019) and Li et al. (2017) (Fig. 1B). During the Paleocene, the Tethyan Himalaya was still part of the continental margin of India (Fig.1A), where carbonate deposition was widely developed in a warm low-latitude environment. Terrigenous influence in this external part of the foreland basin was negligible at this stage, because a remnant Tethyan seaway still existed to the north preventing the arrival of detritus from Asia whereas detritus from the Indian shield may have been trapped in the back-bulge basin to the south. This may be the reason why even an enhanced hydrological cycle during the PETM event could not be reflected in the sedimentary record by an increase in terrigenous detritus, which eventually reach the Tethyan Himalaya from Asia not before the mid-Ypresian (Garzanti et al., 1987; Najman et al., 2010).

8.4.3 Comparison with the southern Pyrenees

Prominent changes in sedimentary environments and hydrological cycle during the PETM were reported from various regions, although most studies focused on terrestrial environments (e.g., Bighorn basin, Wyoming, Foreman, 2014; Bass river, New Jersey, John et al., 2012; Nanyang basin, China, Chen et al., 2016; Dababiya, Egypt, Soliman et al., 2013) rather than on shallow-marine settings. One example worth comparing with the Xigaze forearc basin is provided by the southern Pyrenees, where bioclastic limestones with abundant benthic foraminifera were deposited both before and after the PETM event, whereas braided-river and deltaic siliciclastic sediments accumulated during the PETM (Pujalte et al., 2016, Fig. 12). Arid to semi-arid conditions prevailed in northern Spain both before and after the PETM, whereas water discharge increased by factors between 1.3 and 14 during the early phase of the PETM (Chen et al., 2018). The analogy with the Pyrenees thus lends support to the scenario of increased intensity of seasonal precipitation along the southern Asian margin, leading to increase in sediment supply to the Xigaze forearc basin and demise of the carbonate ramp during a pluvial stage triggered by the PETM event.

9 Conclusions

This study illustrates the stratigraphic record of the Paleocene–Eocene Thermal Maximum in the Xigaze forearc basin (south Tibet), representing the first detailed report of the PETM event along the Asian continental margin of the eastern Tethys and one of the few records of the PETM in active continental margins worldwide.

By coupling sedimentological, biostratigraphic, geochronological, and isotopic data, we reconstruct the sedimentary evolution and discuss changes in the hydrological cycle during the PETM. Our main results are:

1) a major negative excursion in the carbon-isotope record, combined with biostratigraphy and U-Pb ages of the youngest zircon grains in the interbedded tuff layers indicate that the siliciclastic rocks of the middle Jialazi Formation represent the

sedimentary record of the PETM event;

2) the pre-PETM (lower Jialazi) deposits consist of transgressive carbonate-ramp sediments followed by a sudden change to regressive syn-PETM siliciclastic rocks (middle Jialazi), followed in turn by renewed post-PETM (upper Jialazi) carbonate-ramp deposition;

3) both mud and sand fluxes increased during deposition of the middle Jialazi Formation, leading to the temporary demise of the carbonate ramp. This environmental change is mainly ascribed to a pluvial episode leading to increased intensity of seasonal precipitation and hydrological circulation in the Gangdese arc during the PETM event.

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Figures

Fig. 1. Geological and palaeogeographic maps. (A) Global paleogeographic setting during the early Paleogene, showing the southern Asian continental margin, the Tethyan Himalaya, and the Pyrenees (adapted from Scotese, 2013); (B) Simplified tectonic map of southern Tibet (redrawn from Hu et al., 2016b); GCT, Great Counter Thrust; STD, South Tibetan Detachment; MCT, Main Central Thrust; MBT, Main Boundary Thrust; MFT, Main Frontal Thrust; (C) Geological map of the Cuojiangding and Lopu Kangri areas (redrawn from the 1:250,000 geological map), showing the studied Quxia B and Goukou measured sections. The age of diverse units composing the Xigaze forearc-basin succession are given in the legend.

Fig. 2. Outcrop photographs. (A)-(B) full view of the measured Goukou and Quxia B sections, showing the contacts between the Qubeiya, Quxia, and Jialazi formations; (C) first limestone layer marking the base of the Jialazi Fm.; (D) delta-front sandstone in the lower Jialazi Fm.; (E) ripple lamination in the lower Jialazi Fm.; (F) thick-bedded sandstone intercalated with shale in the middle Jialazi Fm.

Fig. 3. Representative microfacies: (A) SF1 Shale interbedded with sandstone; (B) SF2 Sandstone interbedded with shale; (C) CF1 Sandy bioclastic wackestone; (D) CF2 Sandy *Ranikothalia* floatstone/rudstone; (E) CF3 Sandy LBF wackestone; (F) CF4 Bioclastic wackestone/floatstone; (G)-(H) CF5 LBF rudstone; (I) CF6 Mudstone; (J) CF7 Bioclastic packstone.

Fig. 4. Stratigraphic log of the Jialazi Formation in the Quxia B section, showing microfacies, relative paleo-water depths, larger-benthic-foraminifera distribution

and zones, and whole-rock carbonate $\delta^{13}C$ curve. FWWB, fair-weather wave base. SBZ: Shallow Benthic Zones of Serra-Kiel et al. (1998), revised by Boudaugher-Fadel (2018). Time scale after Gradstein et al. (2012).

Fig. 5. Stratigraphic log of the Jialazi Formation in the Goukou section, showing microfacies and relative paleo-water depth. Legend as in Fig. 4. FWWB, fair-weather wave base.

Fig. 6. Depositional model for the Cuojiangding area in the late Paleocene to earliest Eocene, indicating the relative paleogeographic position of the microfacies types identified mainly with reference to the depositional model of Flügel (2010).

Fig. 7. Comparison of the thickness of single sandstone and shale intervals in the pre-PETM, PETM, and post-PETM intervals across the Paleocene/Eocene boundary,

Fig. 8. (A) U-Pb zircon ages of tuff beds intercalated in the middle Jialazi Formation calculated by the "youngest detrital zircon" routine of Isoplot (Ludwig, 2011). Representative cathodoluminescence images of zircons are shown; (B) εHf(t) values *vs*. U-Pb ages of zircons from tuff beds. Data base for the Lhasa block compiled by Hou et al. (2015).

Fig. 9. Relative probability plot of U–Pb detrital-zircon ages. Data sources: Quxia and upper Jialazi Fm. after Hu et al. (2016b); Xigaze forearc strata after Wu et al. (2010), Aitchison et al. (2011), An et al. (2014), Orme et al. (2015), W. Huang et

al. (2015), Orme and Laskowski (2016), and Li et al. (2017); Jialazi Fm. near Mt. Lopu Kangri after Orme et al. (2015).

Fig. 10. Cross plots of whole-rock carbonate carbon versus oxygen isotopes ($\delta^{13}C$ vs. $\delta^{18}O$) from the Jialazi Fm. Circles are data of this study, rectangles are data from Lopu Kangri after Quade et al. (2020).

Fig. 11. Paleogeographic and environmental evolution in the Cuojiangding and Lopu Kangri areas across the Paleocene-Eocene interval.

Fig. 12. Comparison between the Quxia B section (this study) and the Brecha de Arazas section in the southern Pyrenees, Spain (Pujalte et al., 2016). Lithofacies, Shallow Benthic Zones, and carbon-isotope curves are shown. SBZ in the Pyrenees section are from Serra-Kiel et al. (1998), SBZ in this study are from Serra-Kiel et al. (1998), revised by Boudaugher-Fadel (2018).

Supplementary material

Table S1 Carbon and oxygen isotopic values obtained from the Jialazi Formation.

Table S2 U-Pb ages of tuff beds in the Jialazi Formation.

Table S3 Zircon Hf isotopic values of tuff beds in the Jialazi Formation.

Table S4 U-Pb ages of detrital zircons in sandstone samples 09QXB21 from the middle Jialazi Formation.

Table S5 Zircon chronostratigraphy places robust constraints on the depositional age (Ma). [a] YDZ = age calculated by the "youngest detrital zircon" routine of Isoplot (Ludwig, 2011); [b] YSG = youngest single detrital zircon age with 1δ uncertainty;

[c] YPP = youngest graphical detrital-zircon-age peak on an age-probability plot or age-distribution curve; [d] YC1σ (2+) = weighted-mean age (±1σ incorporating both internal analytical error and external systematic error) of youngest cluster of two or more grain ages overlapping in age at 1σ; [e] YC2σ (3þ) = weighted mean age (± 1σ incorporating both internal analytical error and external systematic error) of youngest cluster of three or more grain ages overlapping in age at 2σ (Dickinson and Gehrels, 2009).

Fig. S1 Larger benthic foraminifera in the Jialazi Formation. Scale bars: Figs 1, 4-7 = 1 mm.; Figs 2-3 = 0.5 mm. A. *Ranikothalia sindensis* (Davies), middle to late Thanetian, SBZ3-SBZ4, sample 17QXB32; B. *Miscellanea miscella* (d'Archiac and Haime), late Thanetian, SBZ5, sample 17QXB45; C. *Daviesina salsa* (Davies and Pinfold), earliest Ypresian, SBZ6, sample 17QXB70; D. *Daviesina ruida* (Schwager), earliest Ypresian, SBZ6, sample 17QXB73; E. *Nummulites globulus* Leymerie, earliest Ypresian, SBZ6, sample 17QXB71; F. a) *Discocyclina dispansa* (Sowerby), b) *Assilina leymeriei* (d'Archiac and Haime), earliest Ypresian, SBZ6, sample 17QXB78; G. *Discocyclina sella* (d'Archiac), late Thanetian, SBZ5, sample 17QXB47.

Fig. S2. Probability plots for U-Pb zircon ages from tuff beds intercalated in the middle Jialazi Formation.