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Dating the onset and nature of the Middle Permian Emeishan large igneous province eruptions in SW China using conodont biostratigraphy and its bearing on mantle plume uplift models

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ABSTRACT

The Middle Permian Emeishan large igneous province of SW China has provided the quintessential example of the phenomenon of kilometre-scale pre-eruption domal uplift associated with mantle plume impingement on the base of the lithosphere. One key line of evidence is an interpreted zone of truncation of the platform carbonates belonging to the Maokou Formation that underlies the volcanic pile. Here we test this interpretation by conodont age dating the uppermost beds of the Maokou Formation in sections from Yunnan, Sichuan, Guizhou and Guangxi provinces, which span locations from the inner part of the igneous province to several hundred kilometres beyond its margins. The results show that eruptions began in the *Jinogondolella altudaensis* Zone (~263 Ma) of the Middle Capitanian Stage and greatly increased in extent and volume in the *J. xuanhanensis* Zone (~262 Ma). Pre-eruption uplift was muted, and most locations within the terrain and at many locations beyond its margins witnessed platform collapse (not uplift) with deep-water facies (radiolarian cherts and submarine fans) developing in the *J. altudaensis* Zone. The clearest evidence for an emergence surface occurs around the margins of the province in the *J. xuanhanensis* Zone. This is after the initial onset of eruptions and marks either a eustatic sequence boundary or a brief pulse of tectonic uplift contemporaneous with volcanism. As with recent studies on the basal volcanic successions of the Emeishan LIP, kilometre-scale plume-related domal uplift prior to Emeishan eruptions is not supported by these data; rather a more complex interaction between plume and lithosphere with minor localized uplift and subsidence is inferred.

1. Introduction

The Emeishan terrain in SW China is a relatively small large igneous province (LIP; 0.3–0.5 × 10^6 km^3; Ali et al., 2005), that has attracted considerable attention recently because its emplacement is thought to have triggered a Middle Permian (Capitanian) mass extinction (Courtillot, 1999; Wignall, 2001; Zhou et al., 2002; Wignall et al., 2009a,b). Additionally, it has recently been argued to show evidence for major plume-related uplift (≥1 km) immediately prior to eruption (He et al., 2003, 2006, 2009) and is commonly cited as an important example of this phenomenon (e.g., Campbell, 2005, 2007; Saunders et al. 2007). Ukstins Peate and Bryan (2008, 2009) have challenged this assertion by showing that voluminous mafic submarine hydromagmatic deposits are present close to the area of "considerable uplift" indicating that the initial stages of Emeishan volcanism was at or below the sea level.

Establishing when the first Emeishan Basalt flows were erupted is critical for evaluating both its link to the extinction and the potential limestone platform uplift prior to the volcanism. Unfortunately, although many U–Pb zircon-grain radiometric age-dates have now been obtained from the Emeishan (Fig. 1), they have limited value for constraining the onset of volcanism. First, most are from plutonic bodies intruded into rock units that sit well below

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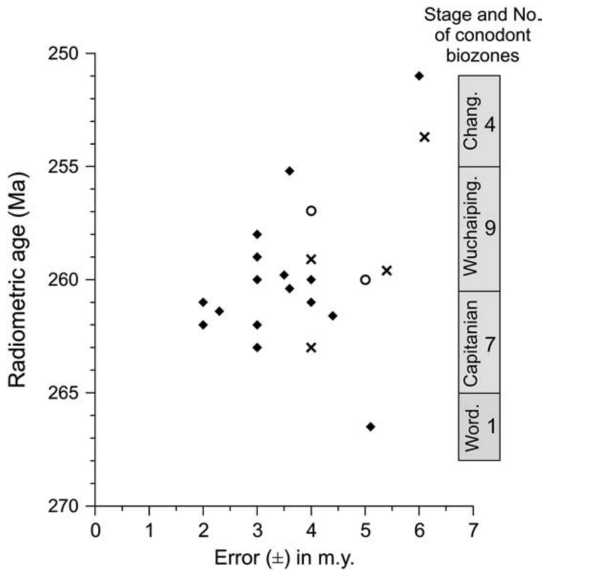


Fig. 1. Summary of the U–Pb radiometric age data for zircon grains from Emeishan LIP rocks (based on the recent compilation of Liu and Zhu, 2009). Diamonds: intrusive rocks; crosses: basaltic flows; circles: dated grains from the Xuanwei (Hsunwei) Formation which directly overlies the volcanic pile in the eastern part of the terrain. Note the typical ±3–5 m.y. errors. Also shown are the stages (see Menning et al., 2006) and the number of conodont biozones in each. Note that radiometric data from Ar–Ar studies have not been shown because they are commonly affected by overprinting (see Ali et al., 2004).

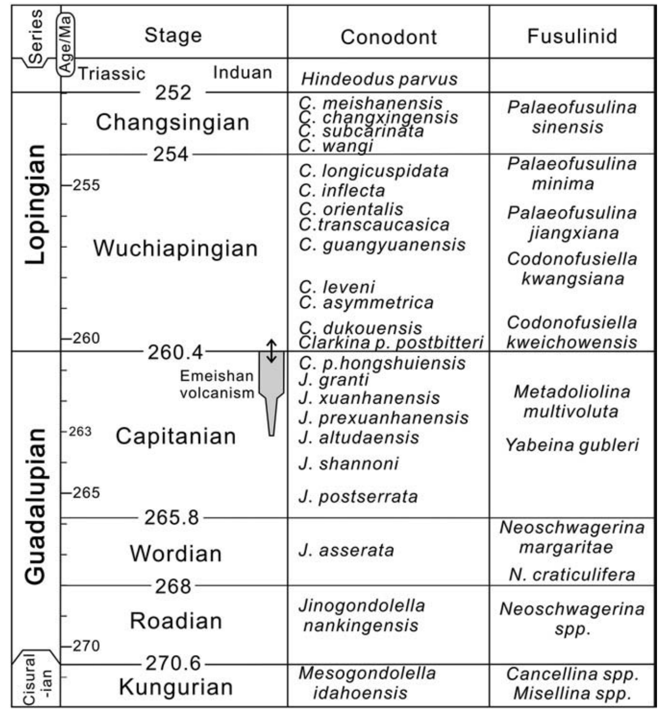


Fig. 2. Guadalupian–Lopingian (Middle–Late Permian) time scale. After Wardlaw et al., 2004; fusulinid data from Jin et al., 1994, 1998. Note the conodont genus *Clarkina* = *Neogondolella*.

the terrain, and thus have no stratigraphic context. Second, the error of each age-date is typically plus or minus 3–5 million years, which is too great to be useful in terms of linking the volcanism to the mass extinction event. Third, the dated zircons are from tuffs or volcaniclastic/sedimentary horizons that are often some distance above the lowermost basalts and thus only provide a minimum-age constraint.

Relative age dating offers an alternative solution. He et al. (2003, 2006) used fusulinid zones to demonstrate an end-Guadalupian Series (end-Capitanian Stage) age for the onset of volcanism. They also argued that the truncation of fusulinid zones in the centre of the province indicated that major uplift and erosion of the Maokou Formation had occurred immediately prior to the eruptions. However, the widespread presence of pillow lavas and mafic volcaniclastic deposits of hydromagmatic origin at the base of the volcanic pile argues for submarine eruption and magma–marine water interactions (Ukstins Peate and Bryan 2008, 2009). Biostratigraphic evidence for uplift is not clear cut. The fusulinid zonation scheme is coarse (only 3 zones have been defined for the entire Middle Permian) and such benthic fossils have potential problems of facies control. Conodont biostratigraphy is widely considered the best available zonation scheme due to its resolution (seven zones are present in the Capitanian Stage alone, see Fig. 2) and facies controls are much less of a problem for this pelagic group (e.g., Yin et al., 2001; Jin et al., 2006), and it is noteworthy that all of the stage boundaries of Permian System are defined by conodonts. To this end, we undertook detailed studies of several Maokou Limestone/Emeishan Basalt contact exposures (Fig. 3) to determine when volcanism started. Together with facies analysis on pre-volcanic Maokou marine sediments, we are able to evaluate the subtle base-level changes that are potentially in response to plume–crust interactions and determine whether significant uplift occurred preceded the volcanism. Furthermore, three biostratigraphic reference sections to the NE and ESE of the province were examined to provide information on the impact of the eruptions beyond the Emeishan margin.

2. Geological setting

The Emeishan Basalt Formation is widely distributed in the provinces of Sichuan, Yunnan and Guizhou in southwest China (Fig. 3), and is one of the most recently identified large igneous provinces — LIPs (Chung et al., 1998). By the standards of other LIPs, the Emeishan Province is relatively small; it covers 0.25–0.3×10^6 km^2, and has an estimated volume of c. 0.25×10^6 km^3, although the original volume could have reached twice this figure (Ali et al., 2005). In terms of eruptive volume estimates it sits between the Miocene Columbia-Snake River Province of northwestern USA (c. 2.35×10^5 km^3; Tolan et al., 1989), and the Oligocene Ethiopian Traps (c. 1×10^6 km^3; Hofmann et al., 1997). It is often difficult to constrain the full thickness due to faulting and possible stratigraphic repetition. The thicknesses of the basalt piles range from several hundred metres in western Guizhou (Liu and Xu, 1994) to as much as ~5 km in western outcrops in northern Yunnan.

The Emeishan Basalt was erupted onto the Middle Permian platform limestones of the Maokou Formation upon which He et al. (2003, 2004) suggested that a karstic surface, with relief locally exceeding 200 m, was developed prior to infill by lava flows. In contrast, Jin and Shang (2000) suggested that the contact between the igneous succession and the limestone was conformable in the east, but unconformable in the west of the province. It is noted that low-relief karstic surfaces and decimeter thick clay/shale beds overlying the Maokou Limestone are well developed on the northern periphery of Emeishan in southern Sichuan (Lai et al., 2008).

The work of He et al. (2003) recognized an inner zone of the Emeishan LIP (centred in NE Yunnan Province, immediately north of Kunming) where >1 km of domal uplift and erosion of the Maokou Formation is suggested to have occurred prior to eruption (Fig. 3). This zone is concentrically surrounded by an intermediate zone where progressively less uplift occurred away from the centre of the province and the Maokou/Emeishan formation contact is considered unconformable. Beyond the area of volcanic rocks, called the outer

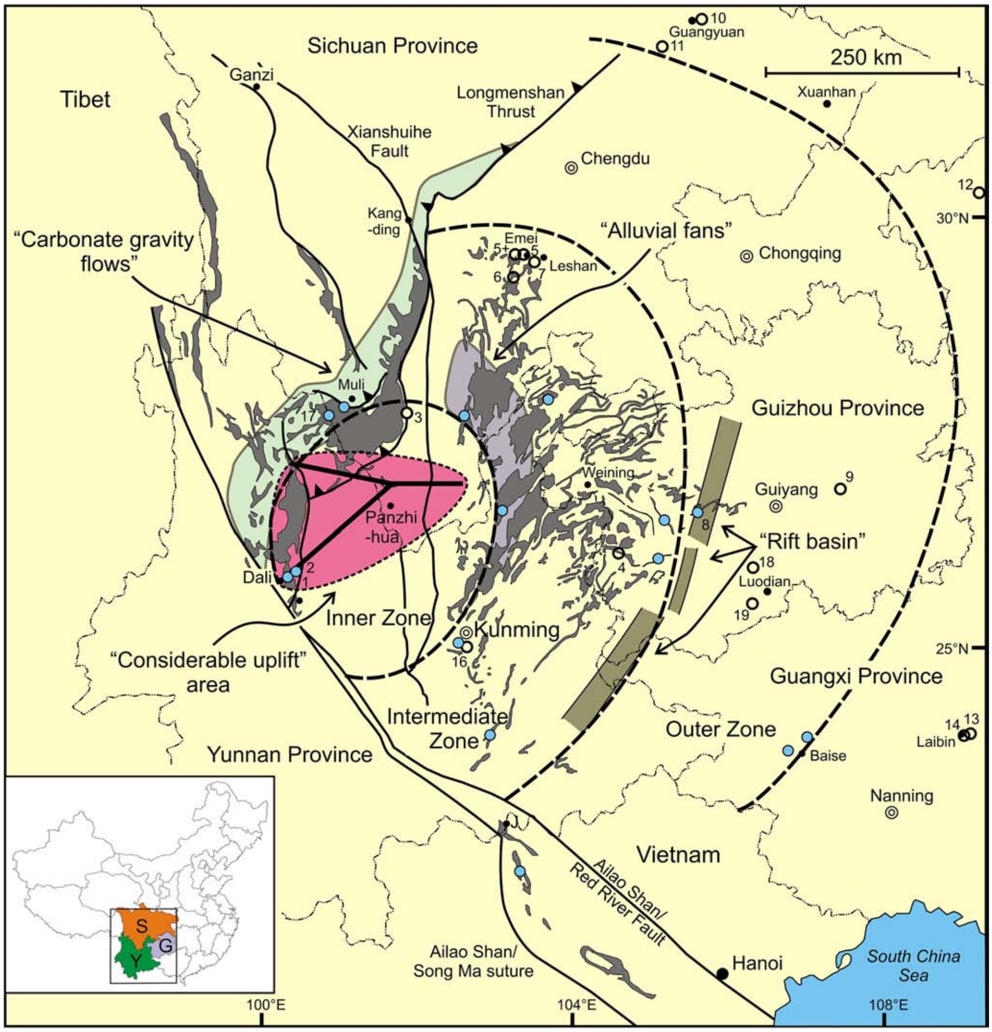


Fig. 3. Distribution of Emeishan Basalt and samples localities (modified from Ali et al., 2010); the "rift basin", "carbonate gravity flows", "alluvial fans" and three zones subdivision are after He et al. (2003, 2006); the pink area of "considerable uplift" is according to key sections of the fence diagram in He et al. (2003); blue dots represent marine eruptions localities where pillows, marine sediments intercalations and hydromagmatic volcanics/volcaniclastics are present. Data are from Lo et al. (2002), Hanski et al. (2004), Fan et al. (2004), Ukstins Peate and Bryan (2008), Wignall et al. (2009b), Ali et al. (2010) and this study; 1–19, samples localities see Table 1 for details. Note the same numbering in main text and table.

zone, the Mid-Upper Permian sedimentary succession is considered essentially conformable and a series of peripheral grabens were developed that received recognized carbonate gravity flows, the eroded material being shed from the elevated inner zone (He et al., 2006).

The age of the Maokou Limestones/Emeishan Basalts contact has been variously reported. He et al. (2003) argued that the Emeishan Basalts were primarily post-Maokou and that the lavas were therefore erupted around the Middle-Late Permian boundary (the Guadalupian–Lopingian boundary — GLB). However, as part of a geomagnetic investigation, Thomas et al. (1998) reported Capitanian fusulinids (*Neoschwagerina* sp., *Verbeekina* sp. and *Neomisellina* sp.) in an interflow limestone at Zhijin, western Guizhou, thereby rendering at least some of the flows older than the boundary. Similarly, Jin and Shang (2000) identified the Guadalupian fusulinids *Metadoliolina* sp. and *Neoschwagerina douvillei* in the limestones from the basal part of the basalt pile. Using magnetobiostratigraphic data from the northern and eastern parts of the Emeishan, Ali et al. (2002) suggested that the main Emeishan Basalt was at least two conodont zones (1–1.5 Ma) older than the GLB; an interpretation supported by a conodont study at the Xiong Jia Chang section, NW Guizhou Province (Wignall et al., 2009a). In this study we present a substantially expanded conodont dataset to demonstrate a clearly intra-Capitanian age for the onset of eruptions and discuss the nature of the Maokou/Emeishan contact.

3. Method and materials

Sections revealing the uppermost Maokou Formation and interflow limestones preserved within the Emeishan LIP, together with several sections located up to 600 km from the LIP's featheredge, were logged and sampled for conodonts and facies analysis. We have collected samples from the inner, intermediate and outer zones and present the results along this transect. A total of 204 samples were collected from 19 outcrops (Fig. 3; Table 1), each weighing between 2.5 and 6.5 kg. All samples were processed using the acetic acid dissolution procedure (see Jiang et al., 2007) and more than 2000 conodont specimens were obtained.

Conodont colour index (CAI) is a semi-quantitative thermometry. The colour changes are cumulative and irreversible (Epstein et al., 1977). We thus are able to monitor the subtle thermo alert of the conodont host rocks. It is noteworthy that some specimens have suffered considerable heating by lavas, as demonstrated by the large CAI variations from 4 up to abnormally high thermal alteration values of 7.

Table 1
Studied sections and their coordinates and related references.

No.	Name	Coordinates	Correlated reference
1	Shangcang Section	25° 48.503′ N, 100° 21.622′ E	He et al. (2003)
2	Wase Section	25° 49.291′ N, 100° 13.773′ E	–
3	Pingchuan Section	27° 40.779′ N, 101° 53.139′ E	Wignall et al. (2009a)
4	Pingdi Sections	26° 04.821′ N, 104° 35.421′ E	Lai et al. (2008)
5	Qing Yin Power Station	29° 34.671′ N, 103° 24.518′ E	Thompson et al. (2001); Ali et al. (2002); Lai et al. (2008)
6	Ebian Section	29° 16.139′ N, 103° 16.392′ E	Thompson et al. (2001; Ali et al. (2002); Lai et al. (2008)
7	Shawan Section	29° 26.100′ N, 103° 32.227′ E	–
8	Xiongjiachang Section	26° 29.042′ N, 105° 41.373′ E	Thomas et al. (1998); Wignall et al. (2009a)
9	Fuquan Section	26° 49.621′ N, 107° 26.787′ E	Bond et al. (2010)
10	Chaotian Section	32° 37.309′ N, 105° 51.626′ E	Isozaki et al. (2004); He et al. (2007); Lai et al. (2008)
11	Shangsi Section	32° 19.513′ N, 105° 27.032′ E	Sun et al. (2008); Lai et al. (2008)
12	Maoershan Section	30º 18.314′ N, 109º 18.416′ E	Xia et al. (2005); Zhang et al. (2008)
13	Penglaitan Section	23° 41.717′ N, 109° 19.267′ E	Mei et al. (1998); Jin et al. (2006); Wignall et al., 2009b)
14	Tieqiao Section	23° 42.733′ N, 109° 13.533′ E	Mei et al. (1998); Jin et al. (2006); Wignall et al. (2009b)
15	Xiangbo Section	24° 51.667′ N, 105° 11.482′ E	Fan et al. (1990)
16	Xishan Section	24° 57.13′ N, 102° 37.80′ E	Bond et al. (2010)
17	Lugu Lake Section	27° 41.711′ N, 100° 58.789′ E	Feng et al. (1994); Wignall et al. (2009a)
18	Gouchang Section	25° 52′ N, 106° 19′ E	Wignall et al. (2009a); Bond et al. (2010)
19	Houchang Section	25° 28′ N, 106° 18′ E	Bond et al. (2010)

4. Results

4.1. Inner zone location

4.1.1. Binchuan, Yunnan Province

The Emeishan Basalts reach their maximum thickness in the area around Binchuan County, in western Yunnan, and sections in the area have provided evidence for pre-eruption uplift and erosion of the Maokou Formation (He et al., 2003; Fig. 3). We examined one of the key contacts at Shangcang (Table 1) although there was 10 m of non-exposure between the uppermost limestones and basal basalt flow. Our conodont extraction from the topmost limestones yielded diverse taxa, namely *Streptognathodus*, *Idiognathodus*, *Gondolella* and *Hindeodus*, that surprisingly indicates a Late Carboniferous age. This suggests two possibilities: the entire Maokou Limestone (and several older Permian and Carboniferous formations) was eroded prior to Emeishan volcanism, or else there is a major (unmapped) fault at the contact.

4.1.2. Wa Se, Yunnan Province

The contact between the Maokou Formation and the overlying volcanics of the Emeishan LIP is well displayed in the hillside east of the village of Wa Se on the SW shores of Lake Erhai, Yunnan. The uppermost 10 m of Maokou limestones consist of packstones composed of calcareous algae, foraminifera and peloids. It is capped by an irregular truncation surface with a few centimetres of erosive relief and larger features (isolated hollows with 25 cm relief and 1 m diameter) draped by a silty shale with sponge spicules (Fig. 4a). The limestone immediately below this surface shows patchy recrystallisation to coarse, blocky ferroan calcite. The spiculitic shale is overlain by bright red mudstones with crystal shards, of volcanic origin, and the subsequent stratigraphy consists of alternations of pillow basalts (~25 m thick) and red mudstones (~5 m thick). Two bulk samples of the Maokou Formation were processed for conodonts but they failed to yield any specimens.

The Wa Se section provides an important record of the Maokou–Emeishan transition; the contact is ostensibly conformable albeit with an interesting history of palaeoenvironmental change in the sedimentary succession from immediately below the volcanics. An abundant calcareous algae flora indicates an initial phase of platform carbonate deposition in shallow-water. The subsequent truncation surface has been dolomitized suggesting emergence and dolomitization of the platform carbonates, probably in a supra-tidal setting. The hollows are interpreted to be dissolution basins (kamnitzas) often attributed to preferential dissolution around plant roots (Flügel, 2004). Subsequent flooding established fine-grained deposition indicating quiet water conditions with a siliceous sponge fauna suggesting relatively deep-water, implying substantial base-level changes. Alternatively, the shale could record a restricted lagoonal facies and therefore insubstantial base-level variations. However, as noted below, rapid deepening events are commonly recorded in many sections in SW China immediately prior to Emeishan volcanism.

4.1.3. Pingchuan, Yanyuan County, Sichuan Province

This section is located in Yanyuan County immediately east of the town of Pingchuan and it sits in the north of the “inner zone” of He et al. model (Fig. 3). However, in He et al. (2006) this location and its eponymous formation developed in a supposed rift graben is located in the outer zone where it received eroded products from the elevated inner zone. The section consists of a 90 m-thick pile of breccias, composed of varying proportions of dacite, basalt, limestone and occasionally shale clasts in a silty matrix, overlain by a thick, continuous succession of basalts. Clasts range from subrounded to angular and range in size from metre-scale blocks down to arenaceous grade material. Internal sorting or grading is not seen within the beds, and individual bed thicknesses range from 1.5 to 17 m. The breccia beds are separated by thinner beds (0.3–12.5 m thickness) of fine-grained strata: laminated, siliceous micrites, and spiculitic cherts containing goniatites. Unfortunately these finer strata did not yield conodonts. However, several typical *J. xuanhanensis* platform elements (Plate 1, fig. 20) were obtained from a limestone clast collected 72 m below the base of the first basalt. This is a significant finding because it indicates that limestones and basalts at least as old as the *J. xuanhanensis* Zone were present in the area and were eroded during the lifetime of the Emeishan volcanism.

The breccia beds of Pingchuan are interpreted to be the product of substantial debris flows shedding material down a submarine slope into a relatively deep-water setting suggested by the laminated, siliceous micrites. In some regards this interpretation in accordance with the model of He et al. (2003, 2006) who interpreted the Pingchuan Formation to be a submarine canyon fill in a narrow graben. However, this graben was placed in the outer zone where it was supposedly receiving only carbonate clasts derived from a deeply eroded and uplifted inner zone. Our conodont data indicates that Maokou platform carbonate deposition continued at least until the *J. xuanhanensis* Zone in the inner zone. The included volcanic clasts in the Pingchuan Formation also indicate that volcanism had begun in the Capitanian. Both these observations argue against any significant uplift prior to volcanism.

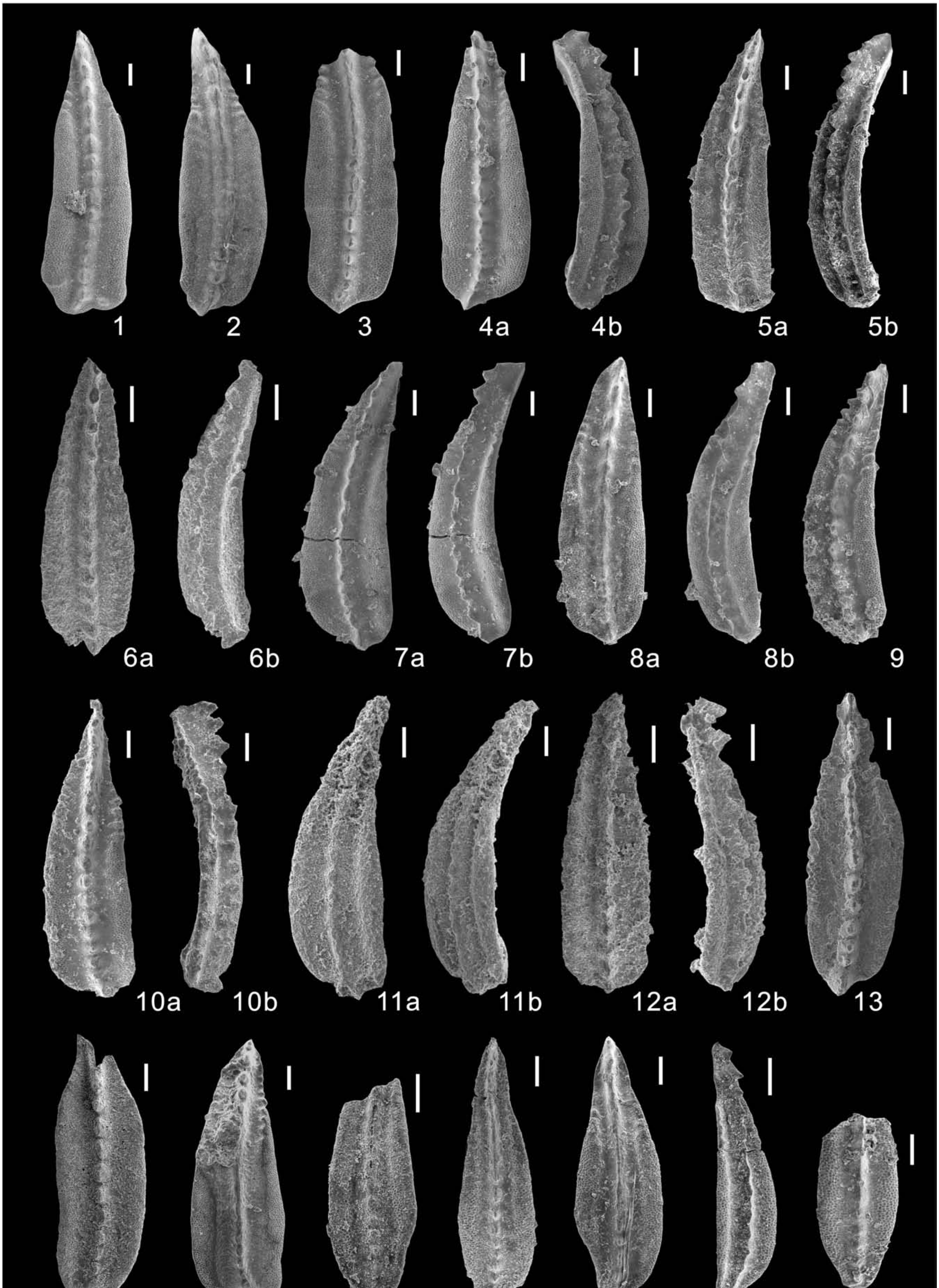
Fig. 4. Field photos showing the contact between the Maokou Formation and Emeishan Basalt Formation. a. Contact between the Maokou Formation and overlying volcanic clays at the base of the Emeishan LIP at Wa Se (25° 49.291′ N, 100° 13.773′ E), Lake Erhai, Yunnan Province. The right most dotted line marks a truncation surface interpreted to be a karstic surface developed towards the end of Maokou deposition and before the onset of volcanism at this location. David Bond (~1.8 m) for scale. b. The hollows of the karstic surface atop the Maokou Limestone are infilled by the Wangpo Bed, a yellow-orangey volcanogenic ash, Ebian section (29° 16.139′ N, 103° 16.392′ E). Mark pen (14 cm) for scale. c. Photograph showing the lower volcanic unit (right) sharply contacts with Maokou marine sediments (limestone and bedded chert atop) at Xiong Jia Chang (26° 29.042′ N, 105° 41.373′ E). The white arrow points to the contact. Haishui Jiang (~1.7 m) for scale.

4.2. Intermediate zone locations

4.2.1. Pingdi, Liupanshui County, Guizhou Province

The Pingdi sections are located 60 km to the south of Liupanshui City in Guizhou Province close to the border with Yunnan. This location is towards the outer part of the intermediate zone of He et al. (2003). The sections consist of two exposures immediately north and south of the small town of Pingdi. Both show sections in the uppermost Maokou Formation overlain by a thick, uninterrupted pile of basalt lavas. The north Pingdi section consists of thick-bedded limestones yielding abundant calcareous algae and foraminifera overlain by 6 m of thin-bedded cherty micrites with a 1 m-thick intensely slumped wackestone at the base of the section. At the top of the section there is approximately 5 m of non-exposure before occurrence of the lowest exposures of deeply altered basalt. Fortunately, this obscured stratigraphy is seen in the nearby south Pingdi roadcut section where thin-bedded cherty micrites containing thin slumped horizons are seen. Petrographic examination of these reveals

Plate 1. Short bar for 100 μm, 'a' for upper view, 'b' for lateral view. All P1 elements, 1–13 from the Xiong Jia Chang Section, 14 from Gouchang Section, 15 from limestone block in Haichaohe; 16, 17 from Qin Yin Power Station, 18 from Shawan, 19 from Ebian Section, and 20 from Pingchuan Section. 1, 3, 4, 15: *Jinogondolella postserrata* (Behnken 1975), 1. XJ-2, 064025; 3. XJ-2, 0624028; 4. XJ-3,04002; 15. DL-2, 70001. 2, 8: *Jinogondolella* sp., 2. XJ-2, 0624012, 8. XJ-3, 064032. 5: Transitional form from *Jinogondolella postserrata* to *Jinogondolella shannoni*, XJ-3, 064028. 6, 7, 9, 10: *Jinogondolella shannoni* (Wardlaw 1994), 6. XJ-6a, 0668013; 7. XJ-3, 04001; 9. XJ-3, 04010; 10. XJ-3, 070013. 11–12: *Jinogondolella altudaensis* (Kozur 1992) 11. XJ-6b, 0668015; 12. XJ-6a, 0668011. 13: *Jinogondolella wilcoxi* (Clark and Behnken 1979), XJ-6a, 0668008. 14: *Clarkina postbitteri postbitteri* Mei and Wardlaw 1994, J, 04025. 16–20: *Jinogondolella xuanhanensis* (Mei and Wardlaw 1994), 16. QY-190, 70005; 17. QY-190, 70002; 18. LS-2, 70000; 19. Ebian-1, 04042; 20. PC-7, 01039.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common radiolarian tests and small, filamentous bivalve shells in laminated, organic-rich strata. The topmost 30 cm of the cherty micrite is interbedded with two 10 cm-thick, pale ash beds. These in turn are overlain by deeply altered basalt that forms the base of a ~200 m-thick volcanic pile. An assemblage with *J.* sp. and *Diplognathodus* sp. was recovered 2 m beneath the basal basalt unit, whilst *J. altudaensis*, *Jinogondolella postserrata*, *J.* sp. and *Sweetognathus* sp. were recovered from a sample 1 m below this contact. The conodont data indicate that the uppermost Maokou Formation around Pingdi belongs to the middle Capitanian *J. altudaensis* conodont zone.

The Pingdi sections are interpreted to record a late Maokou Formation history in which platform carbonates collapsed giving rise to deeper-water slope facies (slumped micrites) and ultimately basin-floor facies (laminated, cherty micrites). Such a facies is common in basinal facies throughout the Permian of South China (Wang and Jin, 2000) and it was into this setting that the first basalt flows were emplaced at Pingdi. It is difficult to quantify the amount of deepening recorded prior to eruptions but a value of several hundred metres is feasible (Wang and Jin, 2000).

4.2.2. Qing Yin Power Station Section, Mt. Emei, Sichuan Province

A series of sections in south and southwest of Sichuan Province around the Mount Emei area represents the type area for the Emeishan Province, although they occur in the intermediate zone because the Emeishan Basalt Formation here is only 250–300 m thick; a relatively thin development compared to more centrally located sections discussed above. The section is exposed next to the track that leads to the Qing Yin Power Station near Mt. Emei. The uppermost Maokou Limestones consist of packstones and they are sharply overlain by a yellow ash and a thin development of carbonaceous wackestones that sit below the basal lava flow (Fig. 5). The same sequence can also be seen in a nearby road-cut (Locality 5+, Fig. 3). The litho-, magneto- and biostratigraphy of this section are well known (Thompson et al., 2001; Ali et al., 2002; Lai et al., 2008) and this investigation builds on this previous work. Conodonts from the topmost carbonaceous wackestones are poorly preserved because they were largely destroyed through thermal alteration when the basal lavas were emplaced. Just a single *Hindeodus excavatus* specimen and one juvenile *Jinogondolella* were obtained from the contact interval. Abundant conodont specimens together with fish remains and ostracods were obtained from the top two metres of limestone. *Jinogondolella xuanhanensis* dominated assemblages from two samples 0.5 m and 1.4 m below the ash bed, whilst a sample from 2 m below this bed was dominated by *J. xuanhanensis* and *J. altudaensis*. All these data indicate that the limestones below the Emeishan Basalt belong to the late Capitanian *J. xuanhanensis* zone.

4.2.3. Ebian Section, Ebian Town, Sichuan Province

The Ebian section lies 40 km south of Mt Emei and the succession is similar to that seen at Qing Yin. At least 30 m of Maokou platform limestones are overlain by ~175 m of Emeishan Basalts (Thompson et al., 2001), and their basal contact can be clearly observed in a road-cut (Lai et al., 2008). An irregular surface is developed atop the Maokou Formation, and the hollows are infilled with yellow ash (Fig. 4b). Above this level, 0.8 m of carbonaceous limestones and coaly shales sit directly below the basal lava (Lai et al., 2008; Fig. 5). The intermingling of uppermost sediments and basalt probably records hot lavas flowing over and interacting with a wet sediment surface. One complete *J. altudaensis* specimen and several *J. xuanhanensis* specimens were obtained from a sample 1 m below the basal lava indicating that the upper Maokou Formation belongs to the *J. xuanhanensis* Zone.

The depositional history of the final stages of the Maokou Formation at Ebian appears closely comparable to that seen at Qing Yin and Wa Se. Thus, thick successions of platform carbonate, with typical shallow-marine fossils including calcareous algae, are terminated by a truncation surface showing hollows up to 0.5 m deep and ~1.0 m wide (e.g., Fig. 4b) interpreted to be kamnitza features. At Ebian and Qing Yin these are infilled with volcanic ash and the overlying coal-rich limestones suggest that shallow-water deposition had resumed immediately prior to the first eruptions.

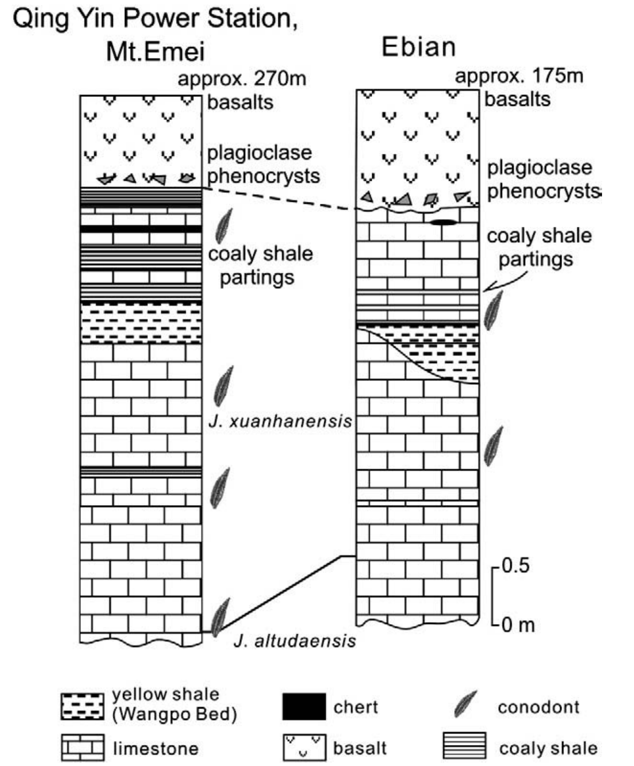


Fig. 5. Sedimentary log of the upper Maokou Formation and basal Emeishan lava flow at Qing Yin Power Station (29° 34.671′ N, 103° 24.518′ E) and Ebian (29° 16.139′ N, 103° 16.392′ E) sections (simplified from Lai et al., 2008).

4.2.4. Shawan Section, Leshan City, Sichuan Province

The Maokou Limestone is well exposed in a quarry in Shawan Borough, Leshan city (42 km east of Mt Emei). The uppermost Maokou Limestone consists of thick-bedded micrites and wackestones with common chert nodules and sporadic solitary corals. It is directly overlain by heavily weathered basalt flows. A 1 m-wide gap at the contact between the limestone and basalt, was obscured by landslipped detritus. Two limestone samples from within 3 m of the contact yielded abundant specimens of *J. xuanhanensis* indicating, once again, a *J. xuanhanensis* zonal age for the topmost Maokou Limestone.

4.3. Outer zone locations

4.3.1. Xiong Jia Chang, Zhijin County, Guizhou Province

The Xiong Jia Chang section in Zhijin County of western Guizhou lies near boundary of the intermediate and outer zones. It is key for evaluating the Emeishan eruption onset because the section includes an interflow limestone package lying about one-third of the way up the 260 m-thick Emeishan volcanic sequence (Wignall et al., 2009a). The succession consists of (in ascending order): a shallow-marine platform carbonates (algal packstones) that pass upwards into deeper-water siliceous limestones (micrites with sponge spicules, <5 m thick); these are sharply overlain by a 70 m-thick horizon of basaltic breccia interpreted to be the product of explosive phreatomagmatic activity (Fig. 4c; Wignall et al., 2009a). The breccia consists of basalt clasts with quenched glassy rims (now palagonite) to the clasts indicating emplacement in water

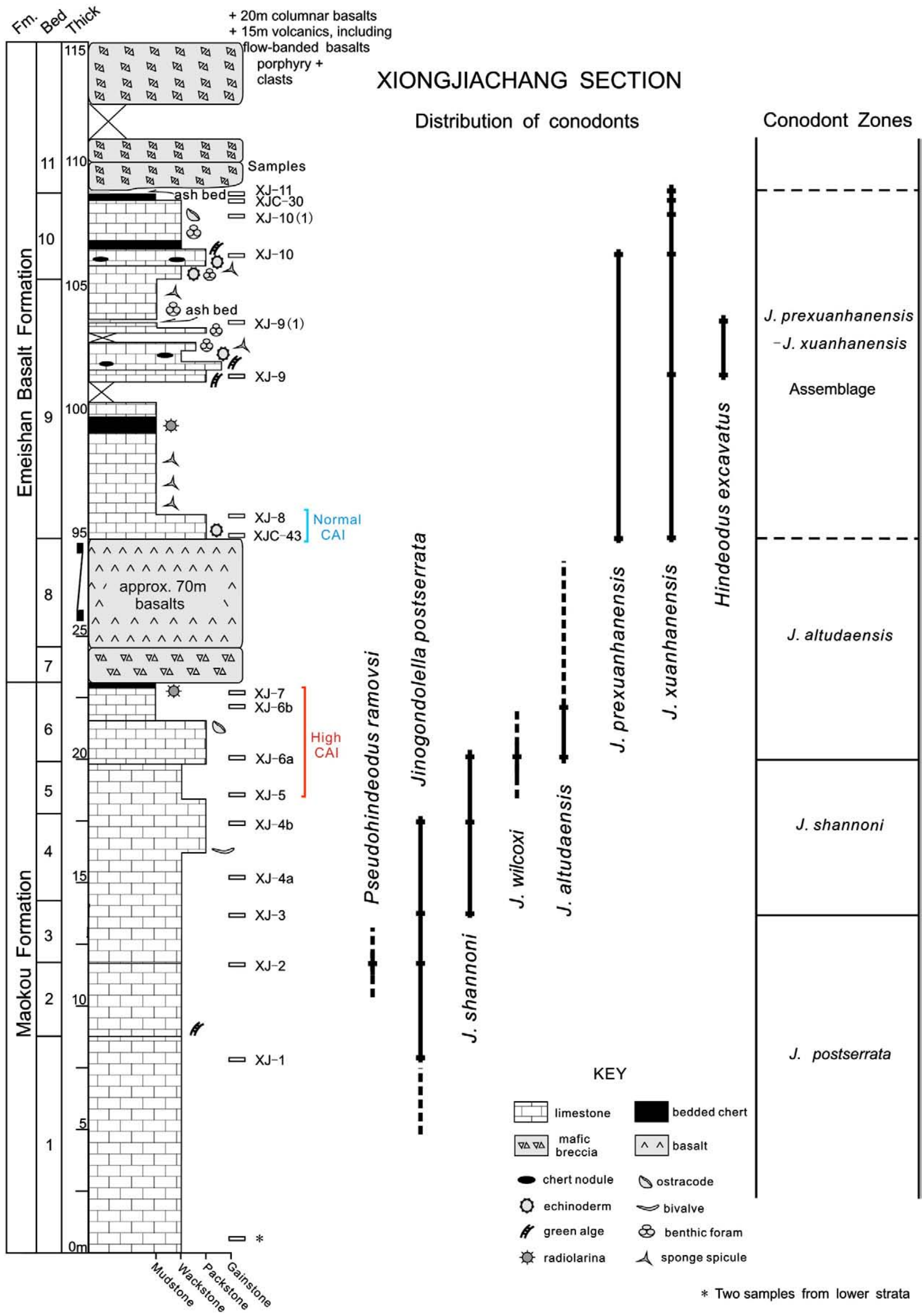
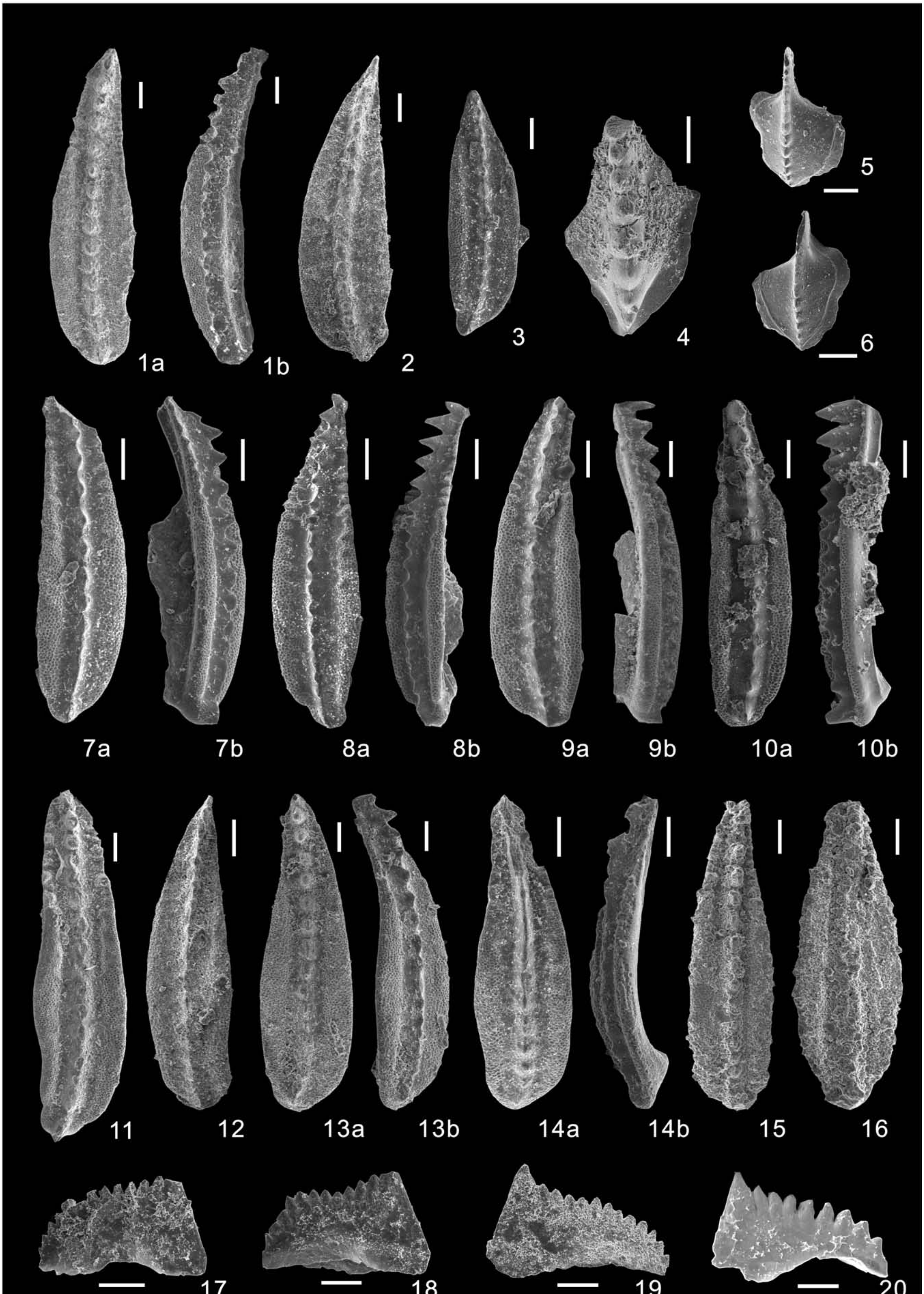


Fig. 6. Lithology and conodont ranges of the Xiong Jia Chang (26° 29.042′ N, 105° 41.373′ E). Note that only the conodont specimens from below the initial flow (Beds 5 and 6) show high CAI values, indicating that the upper part of the Maokou Limestone has been baked, whereas the specimens from lower beds of the intercalated limestone (base of Bed 9) have not been baked (they show normal CAI). This suggests that the lower volcanic unit (Beds 7, 8) is not an intrusion.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and this is further indicated by the presence of seafloor chert lithoclasts and fossils, including foraminifera and crinoids. The breccia is overlain by a 14 m-thick limestone unit recording deep-water anoxic conditions (finely laminated micrite with radiolarians) in its lower part, and shallow-water ramp conditions (algal, foraminifera packstones) in its upper part (Wignall et al., 2009a). Finally, this last unit is then overlain by further horizons of basaltic breccia and basalt lavas showing well developed columnar jointing. It is noteworthy that only the conodont specimens from Bed 5 and 6 shows high CAI value indicating that the upper part of the Maokou Limestone has been baked, whereas the specimens from lower most intercalated limestone (base of Bed 9) has only had minor heating and shows normal CAI. This suggests that the lower volcanic unit (Bed 7, 8) is not a sill or peperitic/hyaloclastitic margined intrusion (Fig. 6).

A complete succession of early Capitanian conodont zones is developed namely, *J. postserrata* zone, *J. shannoni* zone, *J. altudaensis* zone and late Capitanian *J. prexuanhanensis*–*J. xuanhanensis* assemblage zone can be recognized in the Xiong Jia Chang sequence (Fig. 6). The *J. postserrata* zone ranges from Beds 1 to 3 at the base of our studied section with *J. postserrata* and *Pseudohindeodus ramovsi* being the most common species. The top of this zone is defined by the first occurrence of *J. shannoni*. The *J. shannoni* zone ranges from Bed 3 to the basal part of Bed 6. The top of the zone occurs in the middle of Bed 6 and is marked by the first occurrence of *J. altudaensis*. The succeeding *J. altudaensis* zone ranges from the upper part of Bed 6 to the base of the first volcanic horizon. Specimens from this zone demonstrated very high CAI of 6–7, indicating significant thermal exposure. The *J. prexuanhanensis*–*J. xuanhanensis* assemblage zone occupies the intratrappean limestone beds (Beds 9 to 11). Only one typical *J. prexuanhanensis* was obtained in sample XJ-10 and several juvenile forms, together with *J. xuanhanensis* are found in sample XJC-43. *H. excavatus* is also present for a short interval in this zone.

Based on the fossil records of southern China and western Texas, Wang et al. (1998) and Lambert et al. (2002) suggested that there was no stratigraphic separation between *J. prexuanhanensis* and *J. xuanhanensis*. Lambert et al. (2002) further argued the two species exhibit such subtle ontogenetic variations that they were one species. Our study supports the first of these conclusions because we did not find stratigraphic separation between the two species in Xiong Jia Chang and other sections. However, *J. prexuanhanensis* (including the holotype) has a bigger, more highly erect cusp (Plate 2, fig. 10b; Mei et al., 1994a, Plate 2, figs. 12a, 16a, Plate 3, fig.1b) and a more rounded platform termination than *J. xuanhanensis*, suggesting that they are distinct species. Besides the gondolellids, *P. ramovsi* (Plate 2, figs. 5, 6), a long-ranging, globally distributed species was also found at the Xiong Jia Chang section.

4.3.2. Fuquan, Guizhou Province

The Fuquan Section, 80 km to the east of Guiyang city, is at the eastern-most margin of the Emeishan LIP where it represents the outermost edge of flood basalt development in the outer part of the outer zone. A 6 m-thick basalt lava is interbedded between platform carbonate (algal, foraminifera packstones) developments of the Maokou and Wuchiaping formations. Unfortunately, bulk samples from this section only yielded two indeterminate conodont fragments.

4.4. Beyond the outer zone

4.4.1. Chaotian and 11. Shangsi sections, Guangyuan City, Sichuan Province

The sections at Chaotian and Shangsi are located in the north margin of Sichuan Basin adjacent to the Longmenshan Thrust belt (Fig. 3). The region is ~400 km from the northern edge of the Emeishan LIP. Recent studies have shown that carbonate platform production shut down in the early Capitanian Stage around the *J. postserrata*-*J. shannoni* conodont zonal boundary (Sun et al., 2008). Open-marine carbonate productivity re-established in the Late Permian (Wuchiapingian Stage), but in the intervening interval there is an interesting succession of facies comprising (in ascending order) semi-restricted cherty micrites, deep-water cherts, a thick ash-like bed (the Wangpo Bed) and a thin development of coal (Lai et al., 2008). The cherty micrites belong to the *J. shannoni* Zone; unfortunately the overlying strata have not yielded conodonts (Lai et al., 2008). The origin and age of the Wangpo Bed is controversial. He et al. (2007) interpreted it to be the eroded products of the Emeishan volcanics, thus ascribing it a post-eruption age. However, others have considered it to be a major air-fall ash deposit (e.g. Isozaki et al., 2004; Isozaki and Ota, 2007), whilst Lai et al. (2008) suggested it may be the same thick ash horizon seen mantling the karstic surface in southern Sichuan sections (described above). The only conodont-dated development of the Wangpo Bed is in the Maoershan section of Hubei Province (see below) where it is seen to belong to the latest Capitanian *C. postbitteri hongshuiensis* Zone (Zhang et al., 2008). If this age is adopted for the Wangpo Bed at Chaotian then it suggests that there is a considerable hiatus at its basal contact at this location.

4.4.2. Maoershan, Hubei Province

The sections at Maoershan are located around 600 km to the northeastern edge of the Emeishan LIP. They consist of a series of small road-side quarries and a large road-side cliff section, and they display a fascinating range of Capitanian–Wuchiapingian strata. They have been the subject of intense conodont sampling that has established a refined conodont biostratigraphy (Xia et al., 2005; Zhang et al., 2008). These thereby provide a valuable comparison of sections seen nearer to the site of volcanism.

Platform carbonate facies dominate the Maoershan section with bryozoans and crinoids being the dominant bioclasts in the Maokou Formation. This style of deposition was interrupted briefly in the *J. altudaensis* Zone by a phase of spiculitic chert deposition. Carbonate deposition ceased for a more prolonged period in the late *J. granti* Zone because the top of the Maokou Limestone Formation is capped by an irregular truncation surface showing dolomitization of the underlying limestones extending to 1 m in depth. This is overlain by the "Kufeng Chert", a succession of thinly interbedded black chert and black shale beds are very organic-rich and are actively quarried as a source of "coal". The presence of *C. postbitteri honghshuiensis* Zone indicates a latest Capitanian age for this unit (Feng et al., 1997). The Kufeng Chert contains abundant goniatites and spherical concretions (that attain impressive diameters of 1 m or more) and is overlain by a 2 m-thick development of the Wangpo Bed, a pale yellow-brown claystone. In its lower part it contains angular fragments of black shale (presumably derived from the underlying beds) and abundant pyrite cubes and lumps. Zhang et al. (2008) report several horizons of coal

Plate 2. All P1 elements, scale bar for 100 μm. 4, 13 from the Pingdi Section; 14, 19 from the Ebian Section; the rest from the Xiong Jia Chang Section. 1, 3, 7–9, 11, 12: *Jinogondolella xuanhanensis* (Mei and Wardlaw 1994), 1. mature type (transition to *J. granti*), XJ-9, 0668019; 3. juvenile, XJ-10, 0668022; 7. XJ-10, 0668025; 8. XJ-10, 0668023; 9. XJ-11, 04003; 11, XJC-30, 070016; 12. XJC-30, 070024. 2, 16: *Jinogondolella* sp., 2. XJ-3, 064030; 16. XJC-43, 070001. 4: *Sweetognathus* sp., PG-2, 0670003. 5, 6: *Pseudohindeodus ramovsi* (Gullo and Kozur 1992), 5. XJ-2, 064026; 6. XJ-2, 064027. 10, 15: *Jinogondolella prexuanhanensis* (Mei and Wardlaw 1994), 10. XJ-10, 0668024; 15. XJC-43, 070002. 13, 14: *Jinogondolella altudaensis* (Kozur 1992), 13. PG-3, 0670014; 14. Ebian-1, 04013. 17, 18: *Hindeodus excavatus* (Behnken 1975), 17. XJ-9, 0668032; 18. XJ-9 (1), 0668028. 19, 20: *Hindeodus* ex gr. *minutus* (Ellison 1941) 19. Ebian-1, Ebian-1-8; 20. XJC-30, 070025.

from within this horizon although none was observed during our study. The Wangpo Bed is overlain by basal Wuchiaping limestones, called the Xiayao Limestone, which contain significant coal and silt material in their lower metres before passing upwards into "cleaner" bioclastic limestones.

Base-level changes within the Maoershan section were substantial. Shallow-water platform carbonate deposition was terminated by a truncation surface here attributed to karstification (although no distinct kamnitza-like features were present at this location). This was followed by rapid deepening and development of deep-water, organic-rich shales and cherts yielding abundant goniatites. Shallow conditions returned again with the Xiayao Limestone but it is not clear if the shallowing occurred at the base of this unit or the underlying Wangpo Bed. The eroded clasts of Kufeng Chert in the base of the Wangpo Bed may indicate that shallowing and erosion occurred before this ash bed was accumulated.

4.4.3. Penglaitan and 14. Tieqiao, Guangxi Province

The stratotype for the Guadalupian–Lopingian boundary occurs at Penglaitan, with the nearby Tieqiao section providing an ancillary stratotype. These sections record deposition in the Jiangnan Basin and are over 800 km east of the Emeishan LIP margin. Details of the palaeoenvironmental evolution are provided in Wignall et al. (2009b) and only the salient features are provided here for comparison with the depositional histories of sites closer to the LIP. Deep-water chert deposition dominates most of the Guadalupian–Lopingian interval with the exception of a late Capitanian carbonate horizon (the Laibin Limestone) which was deposited in a slope setting. This is interpreted to rest on a correlative conformity to an intra-*J. xuanhanensis* Zone sequence boundary (Wignall et al., 2009b). Evidence for mafic pyroclastic activity is seen, in the form of sub-mm sized basaltic pyroclasts within the Laibin Limestone of *J. granti* Zone, and these are attributed to contemporaneous Emeishan volcanism (Wignall et al., 2009b). Note there has been no search for pyroclasts in the cherts from below the Laibin Limestone.

5. Discussion

Our conodont data indicates that there is no evidence for substantial uplift of the Maokou Formation prior to the eruption of the Emeishan LIP in the mid-Capitanian Stage (Fig. 7; also see Thompson et al., 2001, Ukstins Peate and Bryan, 2008). There is however a considerable palaeoenvironmental change both within the region and elsewhere on the South China continent. Evidence for kilometre-scale domal uplift within the inner part of the province is notably absent from our 3 study locations in the inner zone. The Binchuan sections appear to be complicated by the presence of a large, previously undetected fault at the limestone/basalt contact whilst the Wa Se sections show a conformable contact. At Pingchuan carbonate platform deposition continued until at least the *J. xuanhanensis* Zone as indicated by the age of limestone clasts reworked in breccias beneath the main succession of basalt flows. This contrasts with claims for uplift and removal of all Capitanian limestones in this zone based on fusulinid data (He et al., 2006, 2007). The reworking of platform limestone clasts at Pingchuan suggests that either platform collapse or possibly uplift and erosion but this was during the eruptive history of the Emeishan Province because elsewhere (e.g. Pingdi, Xiong Jia Chang) volcanism had began before the *J. xuanhanensis* Zone (Fig. 7).

Rather than uplift, there is widespread evidence for rapid subsidence and development of deep-water cherts and cherty limestones prior to volcanism. This occurred during the *J. altudaensis* Zone of the mid-Capitanian Stage both within the area of the Emeishan Province and in far distant locations such as Maoershan and Penglaitan and elsewhere in southern South China (e.g., Nashui of Luodian, Mei et al., 2002). The development of basinal cherty facies on platform carbonates clearly represents a considerable deepening of the water column — this facies is typically found in Permian basins of South China and has been interpreted to form in water depths of at least 200 m (Wang and Jin, 2000; Shen et al., 2007).

The first pulse of volcanism – violent at Xiong Jia Chang and quiet at Pingdi – immediately followed collapse. This was followed by the re-establishment of shallow-water limestone deposition in many locations within the *J. prexuanhanensis*–*J. xuanhanensis* zones (~262 Ma). However, the carbonate platforms of the early Capitanian were not re-established to the same extent because sections, such as Chaotian in northern Sichuan, show the development of basinal facies at this time. However, shallow-water carbonates were still widespread and over large areas of the Emeishan LIP's periphery the oldest flows rest conformably on platform carbonates (e.g. Fuquan, Ebian, Qing Yin and Emeishan). A sequence boundary developed in the basal or early part of the *J. xuanhanensis* Zone (e.g. Ebian, Qing Yin, Tieqiao, Penglaitan sections) shortly before this main volcanic pulse (Fig. 7). This is either a phase of tectonic uplift prior to volcanism or it may have a eustatic origin (c.f. Wignall et al. 2009b).

5.1. Eustatic changes

A key issue further complicating the reliability of evidence for positive topography prior to Emeishan volcanism is the eustatic signal. He et al. (2003) argued for continuous sedimentation beyond their three erosional zones. Consequently, any evidence for subaerial exposure (e.g., pre-volcanic karstic surface) was interpreted to be in response to domal uplift (He et al., 2003, 2006). This is incorrect. The Guadalupian–Lopingian transition has been recognized as one of the largest first order lowstands of the entire Phanerozoic (Ross and Ross, 1985; Hallam and Wignall, 1999, Haq and Schutter, 2008). The regression terminated the Capitan reefs in west Texas and the Glass and Del Norte Mountains have their entire late Capitanian missing above the *J. altudaensis* zone (Ross and Ross, 1985; Wardlaw, 2000). In the Salt and Khizor ranges of Pakistan, the middle Capitanian rocks are capped by a major exposure surface (Wardlaw and Mei, 1999). In South China, a widespread hiatal and erosion surface is similarly well developed with continuous sections limited to deep-water basins as seen in the Laibin stratotype sections (Jin et al., 1994, 2006), although even here basinal facies are sharply overlain by slope carbonates in the *J. xuanhanensis* Zone (Wignall et al., 2009b). In Xiong Jia Chang, deep-water, radiolarian cherts are sharply overlain by shallow-water packstones within the *J. prexuanhanensis*–*J. xuanhanensis* zones. The occurrence of a sharp, base-level fall in the early *J. xuanhanensis* Zone in widely separated locations in South China (see Fig. 7) as well as Texas and Pakistan strongly argues for a eustatic regression.

5.2. A plume uplift model?

He et al. (2003, 2006, 2009) postulated that the evolution of the Emeishan Province accords well with the model of an impacting large plume head in which surface uplift, and associated erosion, is the result of dynamic and buoyant uplift of lithosphere caused by an ascending mantle plume. This conclusion is based on the interpretation that a clear uplift signature is recorded in the sedimentary successions that underlie the basalts, and that stratigraphic thinning and the isopachs of the Maokou Formation outline a subcircular region, thus strongly indicating long-wavelength (~400 km wide) domal uplift prior to eruption of the Emeishan Basalts (He et al., 2003). However, this simple plume head model does not explain our observations of platform collapse and deep-water sedimentation, nor previous observations of phreatomagmatic-style volcanism early in the Province's history (Ukstins Peate and Bryan, 2008). If there was uplift it must have been restricted to local areas within the inner zone (see Ali et al. (2010) and Fig. 3). Critically, even in the "inner zone" where more than 1 km pre-volcanic domal uplift was proposed,

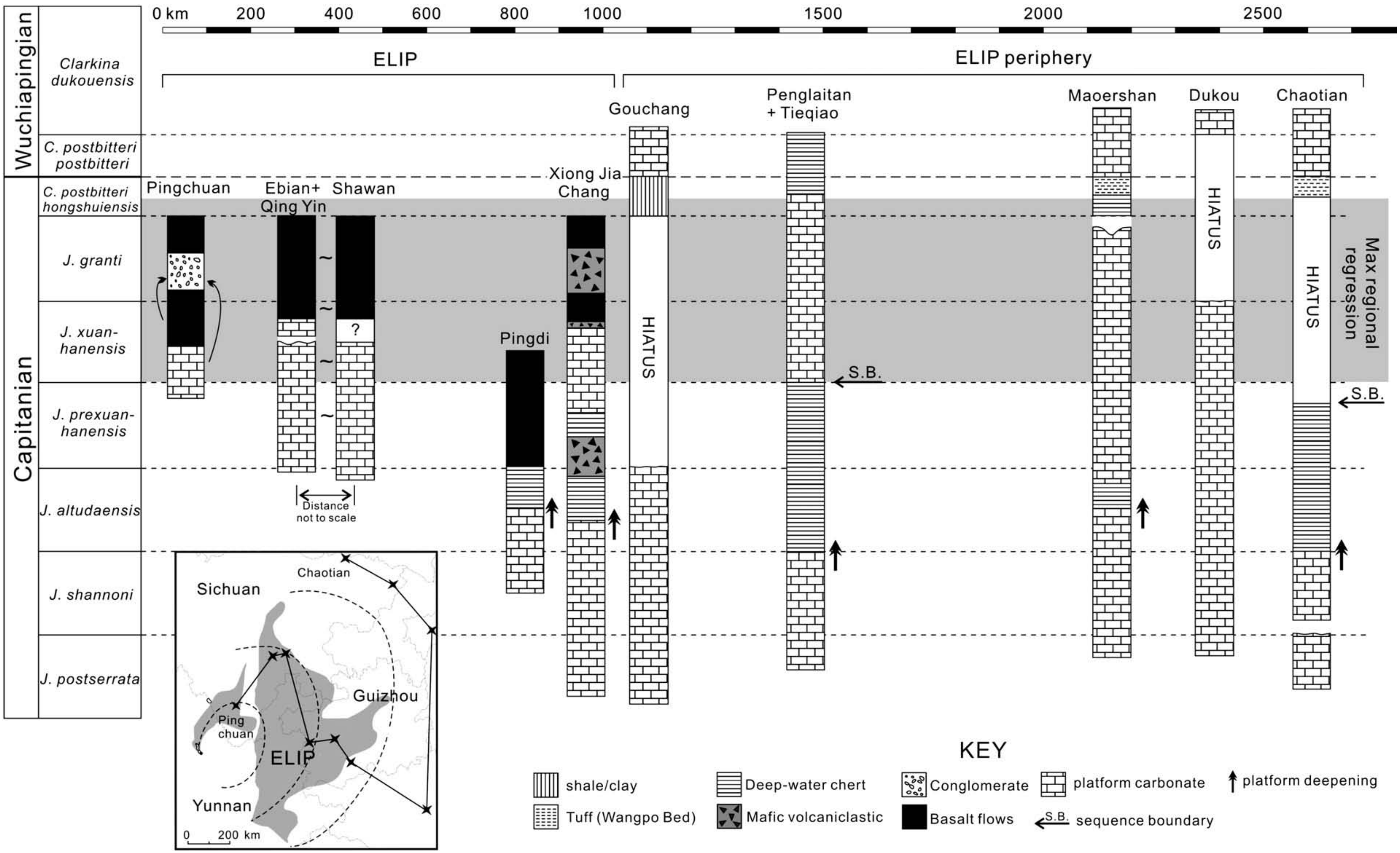

Fig. 7. Correlation chart for sections with robust conodont age control arranged along a E–W–N transect, from left to right, from the inner part of the Emeishan large igneous province, to the province margin to sections several hundred kilometres beyond the province margin in South China (Chaotian and others). The arrows in the Pingchuan column indicate reworking of material into the conglomerates of the Pingchuan Formation. Features to note are the widespread development of deep-water, radiolarian-bearing chert facies in the *J. altudaensis* Zone shortly before the first phase of volcanism, indicating the pre-volcanic platform collapse (deepening). The arrow marked s.b. (sequence boundary) in the Penglaitan/Tieqiao column may indicate either a phase of uplift in South China (note short hiatuses within the *J. xuanhanensis* Zone in some sections) or a eustatic regression. The grey background represents the maximal regional eustatic regression. Data of Gouchang and Dukou (Xuanhan) sections are from Bond et al. (2010) and Mei et al. (1994a, b). The figure is horizontally to scale except the distance between Ebian/Shawan sections; sections are not to vertical scale.

pillow lavas and intercalated limestone are seen in Wa Se and Binchuan areas, arguing for a submarine eruptive initiation (description above and Ukstins Peate and Bryan, 2008). In effect, the local pre-volcanic subsidence (shown as carbonate platform collapse) seen at Wa Se and elsewhere (see Fig. 7) is more consistent with the recent model of Leng and Zhong (2010).

The oldest known basalt flows with well-calibrated conodont data belong to the *J. altudaensis* zone and are seen at Xiong Jia Chang and Pingdi. But widespread eruptions only developed later. This is also seen in other LIP (e.g., Paraná-Etendeka) where the first eruptions were small and increased in magnitude and volume with time (Jerram et al., 1999). For other LIPs, for example in the North Atlantic Igneous Province (NAIP), rapid, transient elevation changes are seen in high-resolution sedimentary records in certain parts of the basin both prior to and during flood volcanism (Jones et al., 2001; Ukstins Peate et al., 2003; Saunders et al., 2007). This is because plume–lithosphere interactions are conditioned not only by the dynamics of the plume head, but also by the interaction of the plume with the rheology and structure of the overlying lithosphere (Burov and Guillou-Frottier, 2005; Jerram and Widdowson, 2005). It is noteworthy that Jerram et al. (2009) have shown significant thicknesses of hyaloclastite in Faroe-Shetland basin of the NAIP, which also occur onshore in Greenland, pointing to significant parts of the area not having significant uplift, and therefore complicating the simple plume model in the NAIP example. The domal uplift model may become significantly modified resulting in regions of crustal dilation and contraction. In such cases, uplifts and subsidences of several hundred metres can produce a series of narrow basins (Burov et al., 2007); this system of highs and lows then evolves into an intra-basin topography with wavelengths of several hundred kilometres with attendant patterns of erosion and deposition.

We suggest that the development of the Maokou Limestone, and pattern of deposition of eroded debris during the early stage of the Emeishan eruptions is inconsistent with the model of broad, regional-scale, domal uplift argued by He et al. (2003, 2006) (also see Ukstins Peate and Bryan, 2008; Ali et al., 2010). Rather, the volcano-tectonic evolution of the region is more complex, and is better considered as a plume–lithosphere interaction which is controlled not only by the thermal dynamic of the plume head, but also by the nature of the lithosphere of SW China.

6. Conclusion

The eruption of the Emeishan large igneous province in SW China was preceded by a phase of partial platform carbonate collapse at the start of the Middle Capitanian *J. altudaensis* Zone (~263 Ma). Thus, in many sections both within the province and beyond it, shallow-water limestones of the Maokou Formation are overlain by deeper-water basinal facies. In more stable platform areas, shallow-water carbonate deposition persisted until the *J. xuanhanensis* Zone (~262 Ma) until a regression of possible eustatic origin caused carbonate deposition to cease. The subsequent onset of eruptions saw the widespread occurrence of mafic, phreatomagmatic-style explosions that span at least the *J. altudaensis*–*J. granti* Zones (~263–260 Ma). Though explosive, the first eruptions were probably in small volume and very localized around *J. altudaensis* zone. It greatly increased in volume and extent later because over large area of the terrain limestones of *J. xuanhanensis* zone were then blanketed by extensive lavas.

Evidence for substantial kilometre-scale, pre-volcanic, plume-related uplift cannot be confirmed and requires further investigation of sections in the Emeishan Province's inner zone. However, it is clear that any uplifted area must have been considerably smaller than previously reported because, over large areas of the province, the initial flows rest conformably on a variety of Maokou sedimentary facies.

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