



# Stratigraphy, biostratigraphy and C-isotopes of the Permian–Triassic non-marine sequence at Dalongkou and Lucaogou, Xinjiang Province, China

I. Metcalfe<sup>a,\*</sup>, C.B. Foster<sup>b</sup>, S.A. Afonin<sup>c</sup>, R.S. Nicoll<sup>d</sup>, R. Mundil<sup>e</sup>, Wang Xiaofeng<sup>f</sup>, S.G. Lucas<sup>g</sup>

<sup>a</sup> Asia Centre, University of New England, Armidale, NSW 2351, Australia

<sup>b</sup> Geoscience Australia, P.O. Box 378, Canberra, Australia

<sup>c</sup> Laboratory of Palaeobotany,<sup>1</sup> Palaeontological Institute, Moscow, Russia

<sup>d</sup> Department of Earth and Marine Sciences, Australian National University, Canberra, Australia

<sup>e</sup> Berkeley Geochronology Center, Berkeley, CA, USA

<sup>f</sup> Yichang Institute of Geology and Mineral Resources, CAGS, Yichang, People's Republic of China

<sup>g</sup> New Mexico Museum of Natural History and Science, 1801 Mountain Road N.W., Albuquerque, NM 87104-1375, USA

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## ABSTRACT

Measured lithostratigraphic sections of the classic Permian–Triassic non-marine transitional sequences covering the upper Quanzijie, Wutonggou, Guodikeng and lower Jiucayuan Formations at Dalongkou and Lucaogou, Xinjiang Province, China are presented. These measured sections form the framework and reference sections for a range of multi-disciplinary studies of the P–T transition in this large ancient lake basin, including palynostratigraphy, vertebrate biostratigraphy, chemostratigraphy and magnetostratigraphy. The 121 m thick Wutonggou Formation at Dalongkou includes 12 sandstone units ranging in thickness from 0.5 to 10.5 m that represent cyclical coarse terrigenous input to the lake basin during the Late Permian. The rhythmically-bedded, mudstone-dominated Guodikeng Formation is 197 m and 209 m thick on the north and south limbs of the Dalongkou anticline, respectively, and 129 m thick at Lucaogou. Based on limited palynological data, the Permian–Triassic boundary was previously placed approximately 50 m below the top of this formation at Dalongkou. This boundary does not coincide with any mappable lithologic unit, such as the basal sandstones of the overlying Jiucayuan Formation, assigned to the Early Triassic. The presence of multiple organic  $\delta^{13}\text{C}$ -isotope excursions, mutant pollen, and multiple algal and conchostracan blooms in this formation, together with Late Permian palynomorphs, suggests that the Guodikeng Formation records multiple climatic perturbation signals representing environmental stress during the late Permian mass extinction interval. The overlap between the vertebrates *Dicynodon* and *Lystrosaurus* in the upper part of this formation, and the occurrence of late Permian spores and the latest Permian to earliest Triassic megaspore *Otynisporites eotriassicus* is consistent with a latest Permian age for at least part of the Guodikeng Formation. Palynostratigraphic placement of the Permian–Triassic boundary in the Junggar Basin remains problematic because key miospore taxa, such as *Aratrisporites* spp. are not present. Palynomorphs from the Guodikeng are assigned to two assemblages; the youngest, from the upper 100 m of the formation (and the overlying Jiucayuan Formation), contains both typical Permian elements and distinctive taxa that elsewhere are known from the Early Triassic of Canada, Greenland, Norway, and Russia. The latter include spores assigned to *Pechorospores disertus*, *Lundbladispora foveata*, *Naumovasporea striata*, *Decussatisporites mulstrigatus* and *Leptolepidites jonkerii*. While the presence of Devonian and Carboniferous spores and Early Permian pollen demonstrate reworking is occurring in the Guodikeng assemblages, the sometimes common occurrence of *Scutasporites* sp. cf. *Scutasporites unicus*, and other pollen, suggests that the Late Permian elements are in place, and that the upper assemblage derives from a genuine transitional flora of Early Triassic aspect. In the Junggar Basin, biostratigraphic data and magnetostratigraphic data indicate that the Permian–Triassic boundary (GSSP Level) is in the middle to upper Guodikeng Formation and perhaps as high as the formational contact with the overlying Jiucayuan Formation.

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## 1. Introduction

Thick non-marine lacustrine/fluvial sequences that span the Permian–Triassic boundary are exposed in the Junggar Basin to

\* Corresponding author. Tel.: +61 2 67733499; fax: +61 2 677271361.

E-mail address: [imetcal2@une.edu.au](mailto:imetcal2@une.edu.au) (I. Metcalfe).

<sup>1</sup> Formerly.

the north of the Tien Shan (Bogda Shan) mountain ranges in Xinjiang Province, NW China (Fig. 1). These continuous sequences are well exposed in the Jimsar and Urumqi areas of Xinjiang and contain important Permian–Triassic transitional vertebrate faunas, conchostracan assemblages and palynofloras. In addition, the sequence preserves magnetostratigraphic and stable carbon isotope records. But despite extensive detailed stratigraphic logging and searches, no volcanic ash or clay beds were identified that could be used for isotopic dating. Studies of these sections were undertaken as part of a large international multi-disciplinary project that has studied marine, non-marine and transitional (marine–paralic–non-marine) Permian–Triassic sequences in China (Fig. 1 and see Metcalfe et al., 2001).

Our studies of the P–T transition in China have included lithostratigraphy, biostratigraphy (conodonts, palynomorphs, tetrapod vertebrates), radio-isotopic dating (U–Pb IDTIMS, U–Pb SHRIMP, Ar–Ar), palaeomagnetism (magnetostratigraphy), and chemostratigraphy (stable carbon isotope studies). These studies have been aimed at providing: accurate and precise ages for the Permian–Triassic boundary and mass extinction, constraints on the tempo of P–T extinction and recovery processes, correlation of marine and terrestrial Permian–Triassic boundary and mass extinction events, and constraints on the causative mechanism(s) of the greatest mass extinction of life on Earth. We here provide detailed measured stratigraphic sections for the Permian–Triassic boundary sequences at Dalongkou near Jimsar and Lucaogou near Urumqi, Xinjiang Province China and results of various studies of those sequences. Detailed results of palaeomagnetic studies, which utilise the same stratigraphic sections and use samples collected at the same time and recorded on the same sections presented here are being published elsewhere in this special issue (Glen et al.). Some results of the palynological studies relating to algae (Foster et al., 2002; Foster and Afonin, 2006) and abnormal pollen grains (Foster and Afonin, 2005) have already been published by our group.

## 2. The Permian–Triassic boundary global stratotype section and point (GSSP) in China: its age and relationship to the end-Permian mass extinction and key biostratigraphic, magnetostratigraphic and chemostratigraphic markers

The Permian–Triassic boundary (formally defined by the base of the Triassic) has been defined (Yin et al., 2001) and ratified by the IUGS at the GSSP level (base of Bed 27c) in shallow-marine strata in the quarry Section D, Meishan, Zhejiang Province, China (Figs. 1 and 2). The boundary is recognised by the first appearance of the conodont species *Hindeodus parvus* (Kozur and Pjatakova).

Traditionally, the Permian–Triassic boundary and mass extinction levels have been used interchangeably because in many sections around the world, the sudden change from typical Palaeozoic to typical Mesozoic faunas and floras coincided with a stratigraphic break or unconformity in both marine and terrestrial environments. Studies of complete sequences in recent decades has shown that there is in fact a zone in which some relict Palaeozoic organisms, following the end-Permian mass extinction, coexisted with new Mesozoic forms regarded as Triassic. The current placement of the Permian–Triassic boundary at the GSSP level at Meishan is at a slightly higher stratigraphic level than the main mass extinction level. The “Permian–Triassic” mass extinction thus occurs in the Upper Permian (upper Changhsingian stage) and is not temporally coincident with the Permian–Triassic boundary. The mass extinction should thus be more correctly termed the “latest Permian” or “end-Permian” mass extinction. It is important to take this into account when correlating other marine and non-marine sequences with the Meishan GSSP. The sharp negative  $\delta^{13}\text{C}_{\text{carb}}$  excursion at Meishan occurs above the main mass extinction level but below the currently defined P–T boundary GSSP level recognised by the first occurrence of *H. parvus* (Yin et al., 2001). This globally recognised  $\delta^{13}\text{C}_{\text{carb}}$  excursion also occurs in other sections in the latest Changhsingian, coincident with or immediately following the mass extinction and before the first appearance of *H. parvus*



Fig. 1. Locations of Permian–Triassic transitional sequences studied in China. See Fig. 3 for detailed locations of the Dalongkou and Lucaogou sections reported in this paper.

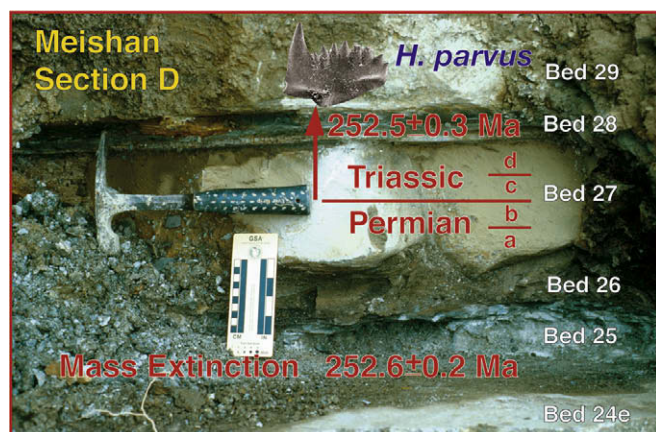


Fig. 2. Permian–Triassic boundary beds at the Meishan GSSP quarry Section D. The main mass extinction level occurs at the base of Bed 25.

(Yin et al., 2005). Many workers have failed to appreciate this and the mass extinction level and the negative carbon isotope excursion level recognised globally are still often presented as representing the Permian–Triassic boundary. Indeed, matters get worse when other data are interpreted relative to the “Permian–Triassic boundary” that has been misidentified due to misuse of “boundary” proxies (e.g. Steiner et al., 2003; de Kock and Kirschvink, 2004; Coney et al., 2007). The end-Permian mass extinction, negative carbon isotope excursion and P–T boundary are separated at Meishan, in the condensed sequence there, by only a few centimetres but these levels are separated elsewhere by metres or even hundreds of metres in more expanded sequences (see Fig. 13 below).

### 3. Dalongkou and Lucaogou non-marine sections, Xinjiang

Globally important non-marine P–T transitional sequences occur on the southern flank of the Junggar Basin in the foothills of the Tien Shan (Bogda Shan), Xinjiang Province, China (Fig. 3). Three of the best sections in this region are found at Dalongkou (north and south limb Dalongkou anticline sections), SSW of Santai and west of Jimsar, and the section at Lucaogou, ENE of Urumqi (Fig. 3). Indeed, the section at Dalongkou has been suggested as a candidate for the global non-marine Permian–Triassic boundary reference section (Liu, 1994). The Permian–Triassic transitional sedimentary rock sequence is represented by the Cangfanguo Group (see below) which was deposited in lacustrine/fluvial environments and which appears to record continuous sedimentation through the P–T transition with abundant fossils (including palynomorphs, algae, conchostracans, ostracods, bivalves, gastropods, and vertebrates). We here report results of work on two stratigraphic sections at Dalongkou, one on the north limb of the Dalongkou Anticline and one on the southern limb of the Anticline, and on a stream gully section at Lucaogou. Due to weathering, trenches had to be dug on the hillsides at Dalongkou to expose sufficiently fresh rock exposure for measurement and sampling (Fig. 3). The studied sections were measured in detail and stratigraphic logs are presented here.

### 4. Lithostratigraphy

The Upper Permian–Middle Triassic non-marine sedimentary sequence in the southern Junggar Basin is summarised in Fig. 4. Our studies at Dalongkou and Lucaogou covered the

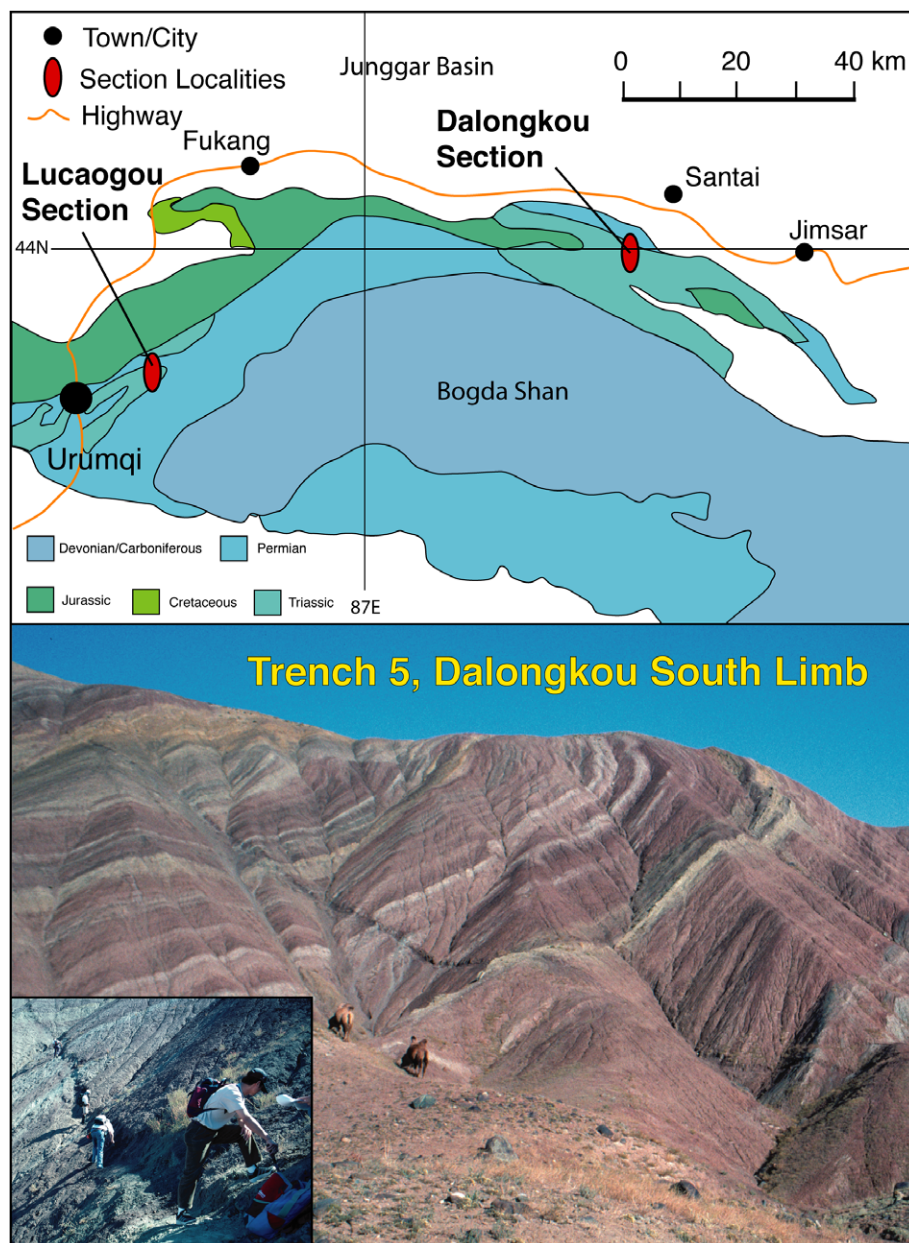
Cangfanguo Group, which comprises, from oldest to youngest, the Quanzijie, Wutonggou, Guodikeng, Jiucaiyuan, and Shaofanggou formations. In particular, our group focused on the P–T transition zone represented by the uppermost Wutonggou, Guodikeng and lowermost Jiucaiyuan formations. The Cangfanguo Group is underlain by the Upper Permian Hongyanchi Formation, and overlain by the Middle Triassic Karamay Formation. The entire sequence from the upper Wutonggou, through Guodikeng to lower Jiucaiyuan formations exposed in the north limb of the Dalongkou Anticline was measured in four artificially dug trenches for the upper part and in gully exposures for the lower part (Figs. 5 and 6). The same sequence was also measured on the south limb of the Dalongkou Anticline in an artificially dug trench (Trench 5, Figs. 7 and 8), and at the Lucaogou section in stream gully sections (Figs. 9 and 10). In addition, the upper part of the Quanzijie through Wutonggou to basal Guodikeng Formations were measured by Jacob's Staff (not to the level of detail of other sections) on the south side of the Dalongkou Anticline along the east side of the river (DLS Traverse – Fig. 11). GPS coordinates for these measured sections are given in Table 1.

The Upper Permian to Lower Triassic sequence of the southern Junggar Basin was deposited in a tectonically active foreland basin (Wartes et al., 2002) and represents lacustrine environments that evolved from shallow ephemeral lakes that frequently dried up completely, to deep stratified lakes that persisted for significant time periods, to freshwater lakes associated with extensive fluvial systems (Carroll, 1998). The Upper Permian–Lower Triassic sequence studied and reported here represents the later stage freshwater lake and fluvial environments of the basin.

The 121 m thick Wutonggou Formation at Dalongkou includes conglomerate, sandstone, siltstone and shale/mudstone deposited in fluvial, meandering stream, and lacustrine systems. Twelve sedimentary cycles, characterised by sandstone units ranging in thickness from 0.5 to 10.5 m that represent cyclical coarse terrigenous input to the lake basin during the Late Permian are recognised. These lake expansion–contraction and fluvial cycles probably represents changes in depositional regimes reflecting a complex interplay between mid-high latitude climate change and tectonics. Paleosols of the Wutonggou Formation indicate perennially wet soil moisture regimes in the lower part of the formation, and a humid environment with distinct seasonal variations in soil moisture budget for the upper part of the formation (Thomas, 2006).

The Guodikeng Formation measured in these sections comprises predominantly dark-grey, green and purple mudstones with minor siltstones and fine-grained green, buff and grey sandstones, and calcareous nodules (Figs. 5 and 6). Compared to the underlying Wutonggou Formation, there is a lack of significant coarse terrigenous input to the lake during deposition of the Guodikeng. The thickness of the Guodikeng Formation on the north limb of the anticline was measured as 197 m, and 209 m on the southern limb. Concentrations of conchostracans in calcareous nodular beds were noted at several horizons (see Figs. 5 and 6). The Guodikeng Formation at Lucaogou is thinner than at Dalongkou, with a measured thickness of 129 m. The sedimentological regime affected by climate and hydrological changes probably controlled the distribution of plants and algae, particularly ponding events and blooms. A major change at the close of Guodikeng Formation times occurred because the overlying Jiucaiyuan Formation has a basal conglomerate, and a cycle of sandstone bodies at the base (although less coarse than those of the Wutonggou). A significant drying of the climate during deposition of the Jiucaiyuan Formation is indicated by paleosols that include sub-surface calcium carbonate and pseudomorphs after gypsum suggesting net soil moisture deficiency (Thomas, 2006).





**Fig. 3.** Locations of the Dalongkou and Lucaogou Permian–Triassic transition sections in Xinjiang Province, China and photographs of part of the Dalongkou Anticline south limb showing Trench 5.

## 5. Biostratigraphy

### 5.1. Palynology

Palynomorphs have been studied from 83 productive samples from the upper Quanzijie, Wutonggou, Guodikeng and lower Jiuciyuan formations, collected from the continuously measured sections (Figs. 5–11). Details and taxonomy are being prepared for publication elsewhere by Foster and Afonin; discussed below are the key findings.

Yang et al. (1986) and Ouyang and Norris (1999) described plant microfossil assemblages from 15 samples from the Dalongkou section, and species reported as stratigraphically restricted have proved to be present throughout the Guodikeng Formation. Moreover, in addition to Upper Permian to Triassic spores and pollen recorded in our study, the Dalongkou and Lucaogou samples yield reworked Devonian, Carboniferous and Early Permian spores and pollen, derived from sediments of this age in the Bogda Shan to the south.

Yang et al. (1986) and Ouyang and Norris (1999) placed the P–T boundary at Dalongkou in the upper Guodikeng Formation between 30 and 50 m below the base of the Jiuciyuan Formation. Placement was based on changes in palynofloras and in particular the appearance of lycopod spores, particularly *Aratrisporites* spp., which are used widely elsewhere as a marker for the Early Triassic. *Aratrisporites* was not identified in our study, so we could not verify the previous reports of this genus. Despite this uncertainty, palynofloras, here assigned to Assemblage 3, from the upper 100 m of the Guodikeng Formation, are derived from a transitional flora; that is, one that includes pollen and spores that are common to both the Late Permian and Early Triassic in other localities in Russia and Europe. The transitional nature of the palynofloras has been noted previously by Ouyang (1991), Ouyang and Norris (1999), Lozovsky (1998), Afonin (2000) and Lozovsky et al. (2001), and is discussed below.

Ouyang et al. (2003) have provided a comprehensive palynological overview of Carboniferous and Permian strata from the North-

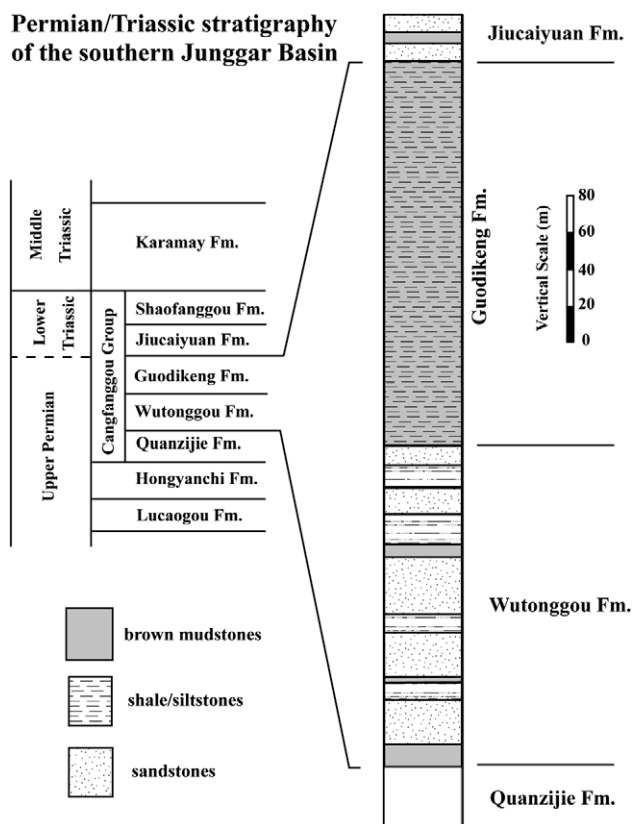


Fig. 4. Generalised Permian–Triassic stratigraphy of the southern Junggar Basin, Xinjiang Province, China.

ern Xinjiang Province and include range charts showing the distribution of taxa in the formations considered here (Quanzijie, Wutonggou, Guodikeng). They follow Ouyang and Norris (1999) in placement of the P–T boundary (see above), and discuss the correlation of the older Permian assemblages with those of the Cis Urals (Ouyang et al., 2003, pp. 473–479).

Our study of the late Permian–Triassic, is the most comprehensive to date with more than 100 species recorded, and assigned to three palynofloral assemblages (Afonin and Foster, 2005; and unpublished). In stratigraphic order, oldest to youngest, they are:

- Assemblage 1: *Tuberculatosporites homotubercularis* – *Potonieisporites* sp.Q; upper Quanzijie – lower Wutonggou Formations (two samples: 3406, 3395). The occurrence of *T. homotubercularis* Hou and Wang 1986 (= *Polypodioidites* sp.) provides a reasonable basis for correlation with palynofloras from the upper Chidru Formation, Wargal section, Salt Range (Balme, 1970), Western Australia (Liveringa Group), and the Luptyug Member, Vyatskian Horizon, of the Moscow Syncline (unpublished Afonin). Monosaccate pollen, including *Cordaitina uralensis* (Luber) Samoilovich 1953, are relatively common, but rare examples of *Scutasporites* sp. cf. *Scutasporites unicus* Klaus 1963, typical of the two younger assemblages, herald the changing nature of the gymnosperm flora. The latter, also occurs rarely in the Upper Permian of the Moscow Syncline, and is useful in interregional correlation (Foster and Jones, 1994; Ouyang, 1999; Ouyang et al., 1999).
- Assemblage 2: *Klausipollenites schaubergeri* – *Reduviasporonites chalcatus* – *Syndesmorion stellatum*; upper lower Guodikeng Formation, North Limb (seven samples: 3380, 3381, 3501 to 3504, 3361). In general, assemblages are dominated by the eponymous taxa; non-taeniate pollen grains and abundant algal

remains (see Foster and Afonin, 2006): palynomorphs may be strongly corroded. Gymnosperm pollen assigned to *Scutasporites* sp. cf. *S. unicus* is common in some samples, and other key pollen taxa, although rare, include: *Lueckisporites virkkiae* (Potonié and Klaus) Clarke 1965, *Lunatisporites transversundatus* (Jansonius) Fischer 1979, *Lunatisporites pellucidus* (Goubin) Helby ex de Jersey 1972, *Platysaccus queenslandi* de Jersey 1962, *Alisporites splendens* (Jizba) Foster 1979, and *Striatoabieites richteri* (Klaus) Hart 1964. Among spores making their appearance in this assemblage are *Leptolepidites jonkeri* (Jansonius) Yaroshenko and Golubeva 1991, *Limatulasporites fosulatus* (Balme) Foster and Helby 1979, and *Naumovaspore striata* Jansonius 1962. *N. striata* is more commonly found in Early Triassic assemblages, but is known from the latest Permian Moloma Formation of Russia (Yaroshenko and Gomankov, 1998). We conclude that the Assemblage 2 is of latest (post-Tatarian) Permian age. Supporting this conclusion is the coeval occurrences of algal blooms, particularly of *R. chalcatus* (see Krassilov, 2003). We note that *R. chalcatus* is also abundant in some samples from the acritarch-dominated but low diversity palynological assemblages recovered from the Changhsing Fm in the type marine Permian–Triassic boundary succession at Meishan (Ouyang and Utting, 1990).

- Assemblage 3: *Lundbladispore foveata* – *Pechorosporites disertus* – *Otyenisporites eotriassicus*. Upper Guodikeng Formation and lower Jiuciyuan Formations (74 samples). The majority of samples yielded palynofloras assignable to Assemblage 3. They have higher taxonomic diversity than the older assemblages, and in addition to the eponymous species, are dominated by lycopod spores and bisaccate taeniate pollen grains. Some species, including *Scutasporites* sp. cf. *S. unicus* (abundant in sample 3379, Fig. 6) and *A. splendens*, and the alga *R. chalcatus*, continue from Assemblage 2. *Ephedripites* sp. cf. *Ephedripites permasensis* Yaroshenko 1997 is abundant in one sample (3359, Fig. 5), other polyplicate taxa include *Decussatisporites mulstrigatus* Hou & Wang 1986, *Ephedripites scotti* (Jansonius) Yaroshenko and Lozovsky 1997, *Eucommiidites cathayensis* Ouyang and Norris 1988.

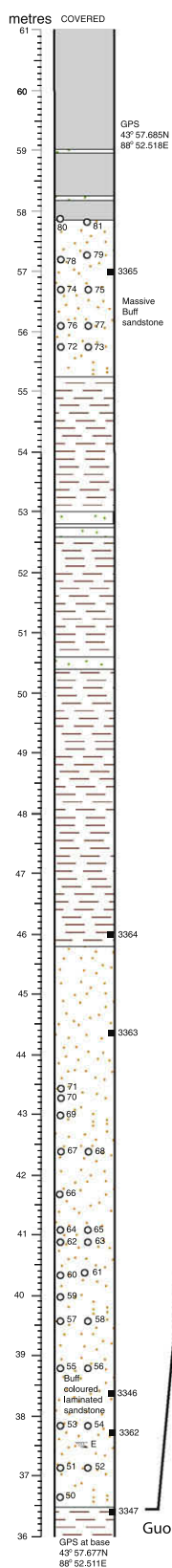
The absence of the alga *S. stellatum* is significant and distinguishes palynofloras assigned to Assemblage 3. Reworked spores from Devonian (common) and Carboniferous strata are often abundant and taxonomically diverse, and are a feature of palynofloras assigned to Assemblage 3 (see below). Reworked taxa include *Ahrsenisporites guerckei* (Horst) Potonié and Kremp 1954, *Anapiculatisporites largus* Playford 1971, *Ancyrospora incisa* (Naumova) Raskatova and Obukhovskaya 1993, *Ancyrospora* spp., *Convolutispora vermiformis* Hughes and Playford 1961, *Emphanisporites rotatus* McGregor 1961, *Geminospore* spp., *Grandispora parvula* Turnau 1999, *Knoxisporites literatus* (Waltz) Playford 1963, *Leiotriletes ornatus* Ischenko 1958, *Retusotriletes* spp., *Spelaotriletes* spp., *Stenozonotriletes clarus* Ischenko 1958, *Vallatisporites vallatus* Hacquebard 1957, *V. pusillites* (Kedo) Dolby and Neves emend. Byvsheva 1985, *Verrucosisporites nitidus* Playford 1964, and monosaccate pollen (Early Permian), *Parasaccites talchirensis* Lele and Makada 1972.

The distinctive megaspore *O. eotriassicus* Fuglewicz 1977 occurs within palynofloras assigned to Assemblage 3, from all sections (Fig. 13). It is a useful marker as it occurs in both marine and non-marine sections in Greenland, Italy, Russia, and Poland (Lozovsky et al., 2001; Foster and Afonin, 2005).

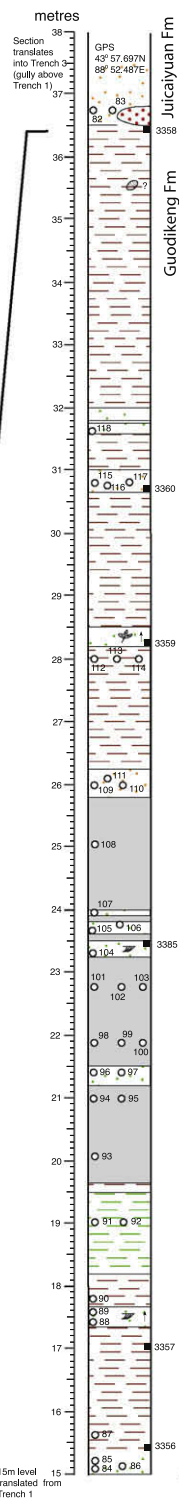
A significant feature of Assemblage 3 is the presence of a high percentage of abnormal pollen grains of *Alisporites*, *Scutasporites* sp. cf. *S. unicus*, and *K. schaubergeri*: they are also found at the same level in the Russian P–T sequences, suggestive of environmental stress on the parent plant during severe climatic conditions (atmospheric pollution, including

## Dalongkou North Limb

## Trench 3



## Trench 2



## Trench 1

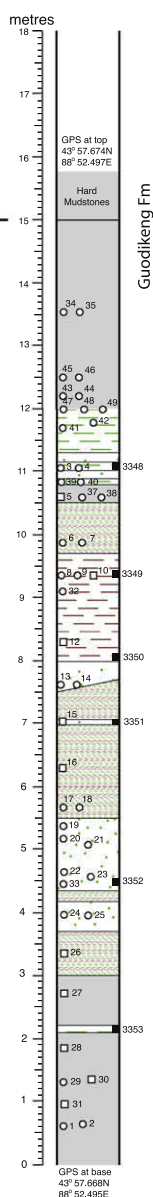


Fig. 5. Measured sections in Trenches 1, 2 and 3, Dalongkou north limb.



## Dalongkou North Limb

## Base Guodikeng Fm to Base Trench 4

## Trench 4

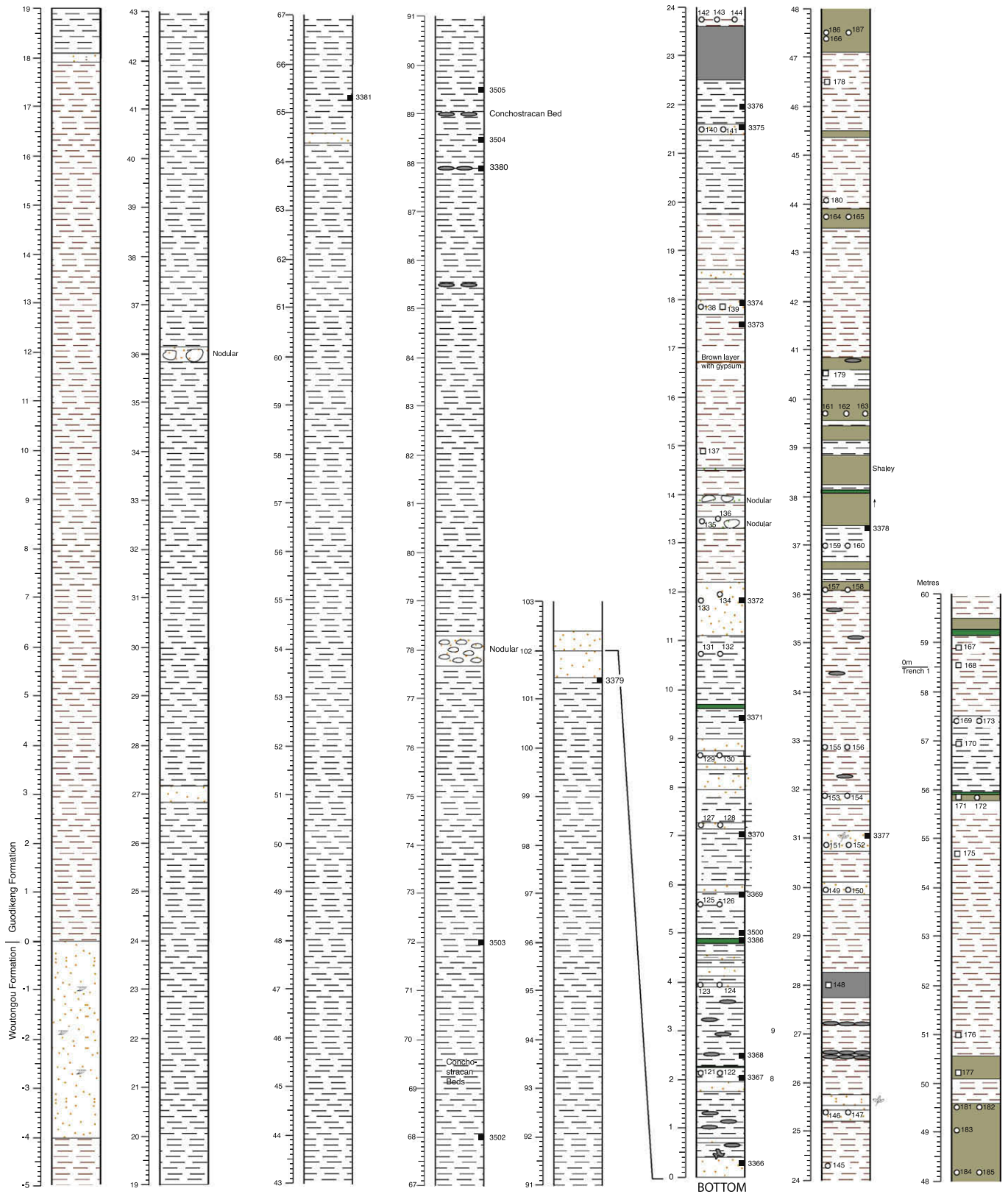


Fig. 6. Measured sections in Trench 4 and gulley section from base Guodikeng Fm to base Trench 4. Legend as for Fig. 5.

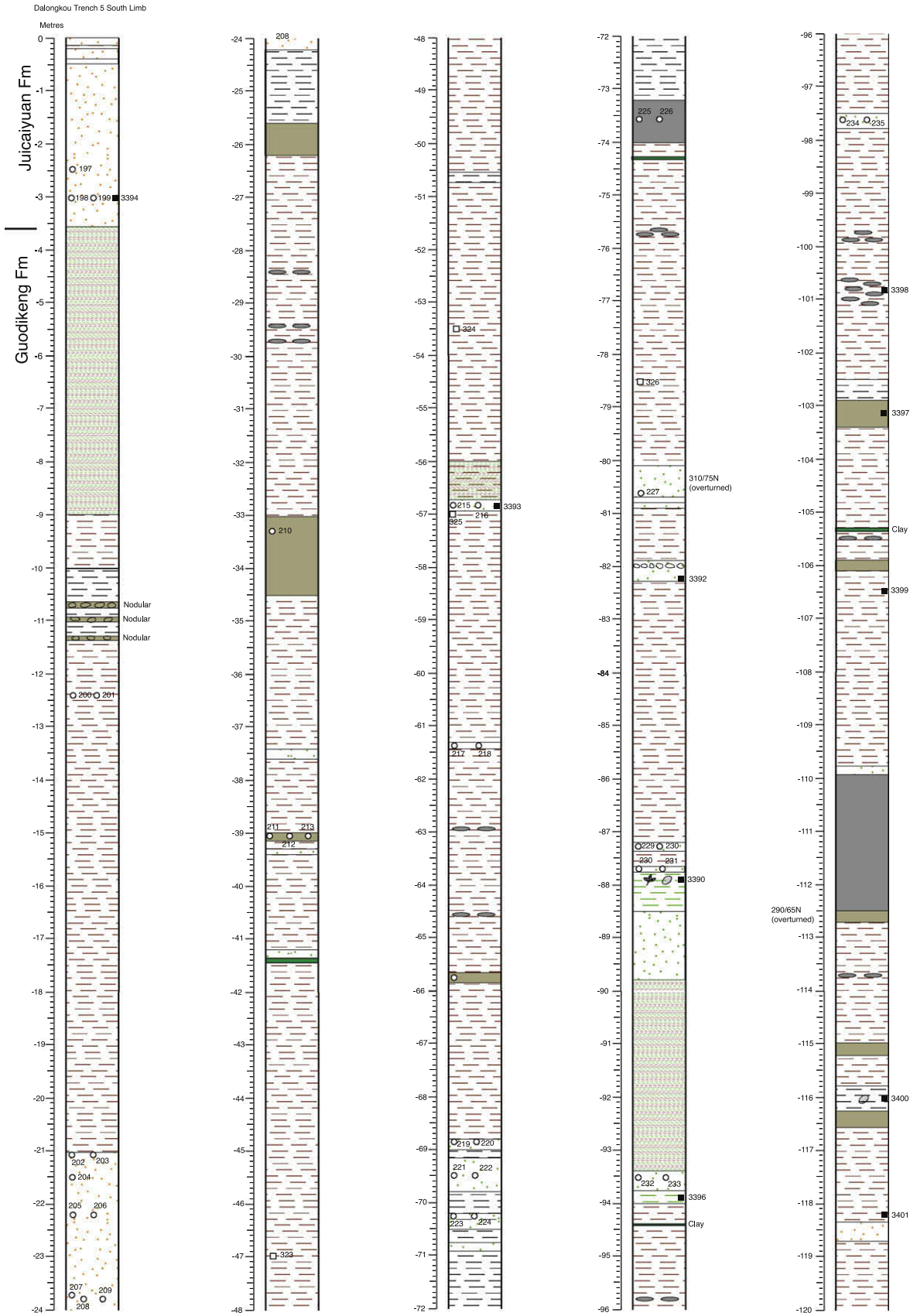


Fig. 7. Measured section in Trench 5, Dalongkou south limb (upper part). Legend as for Fig. 5.

increased UV radiation, caused by extensive volcanism) at the end-Permian mass extinction level (Foster and Afonin, 2005).

There is also a correlation between occurrences of abnormal pollen and negative carbon isotopic ( $\delta^{13}\text{C}_{\text{org}}$ ) excursions



Dalongkou Trench 5 South Limb

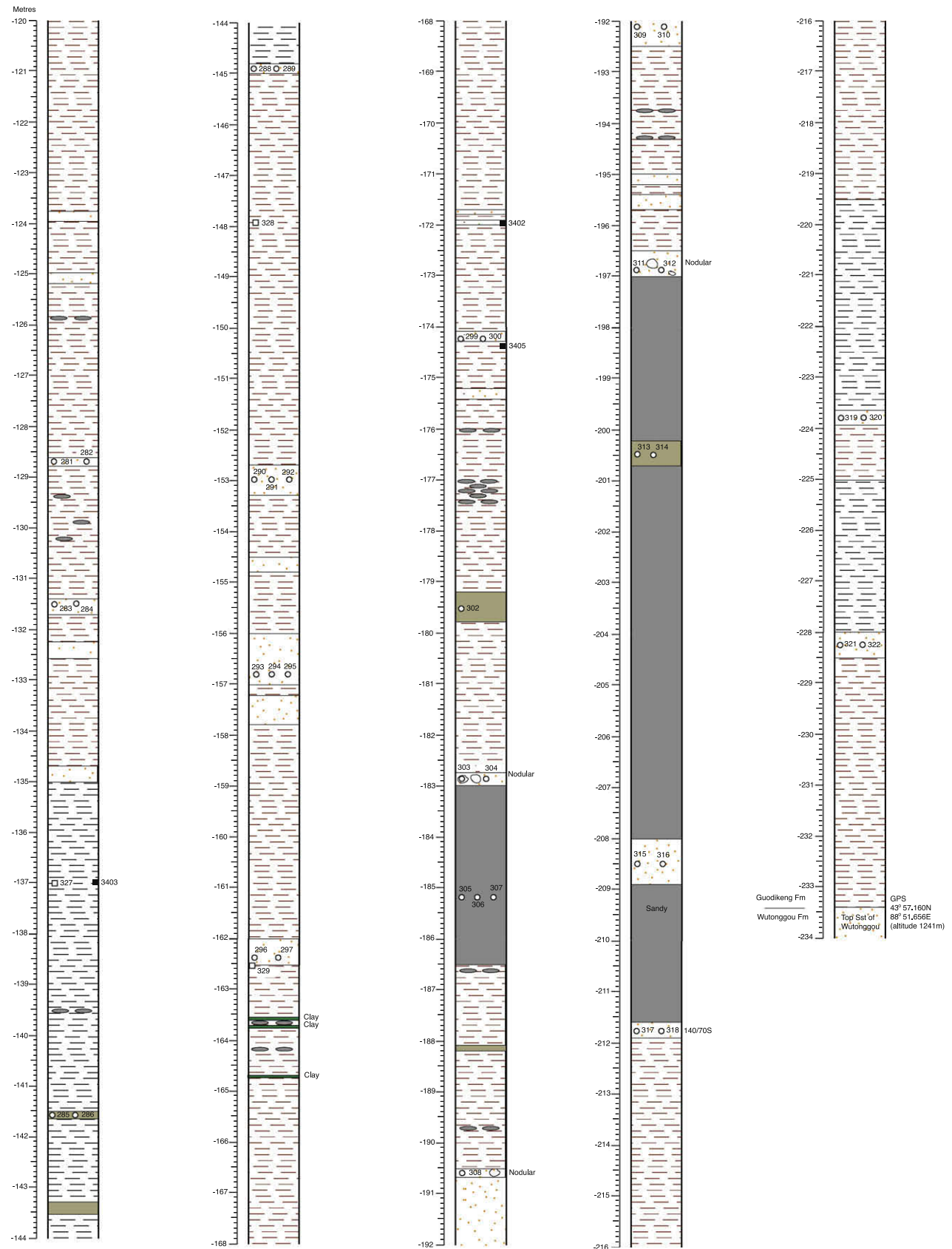


Fig. 8. Measured section in Trench 5, Dalongkou south limb (lower part). Legend as for Fig. 5.

reflecting drastic environmental change at both the local and global level.

What is most intriguing about Assemblage 3 is that it contains many typical Late Permian pollen taxa (above).

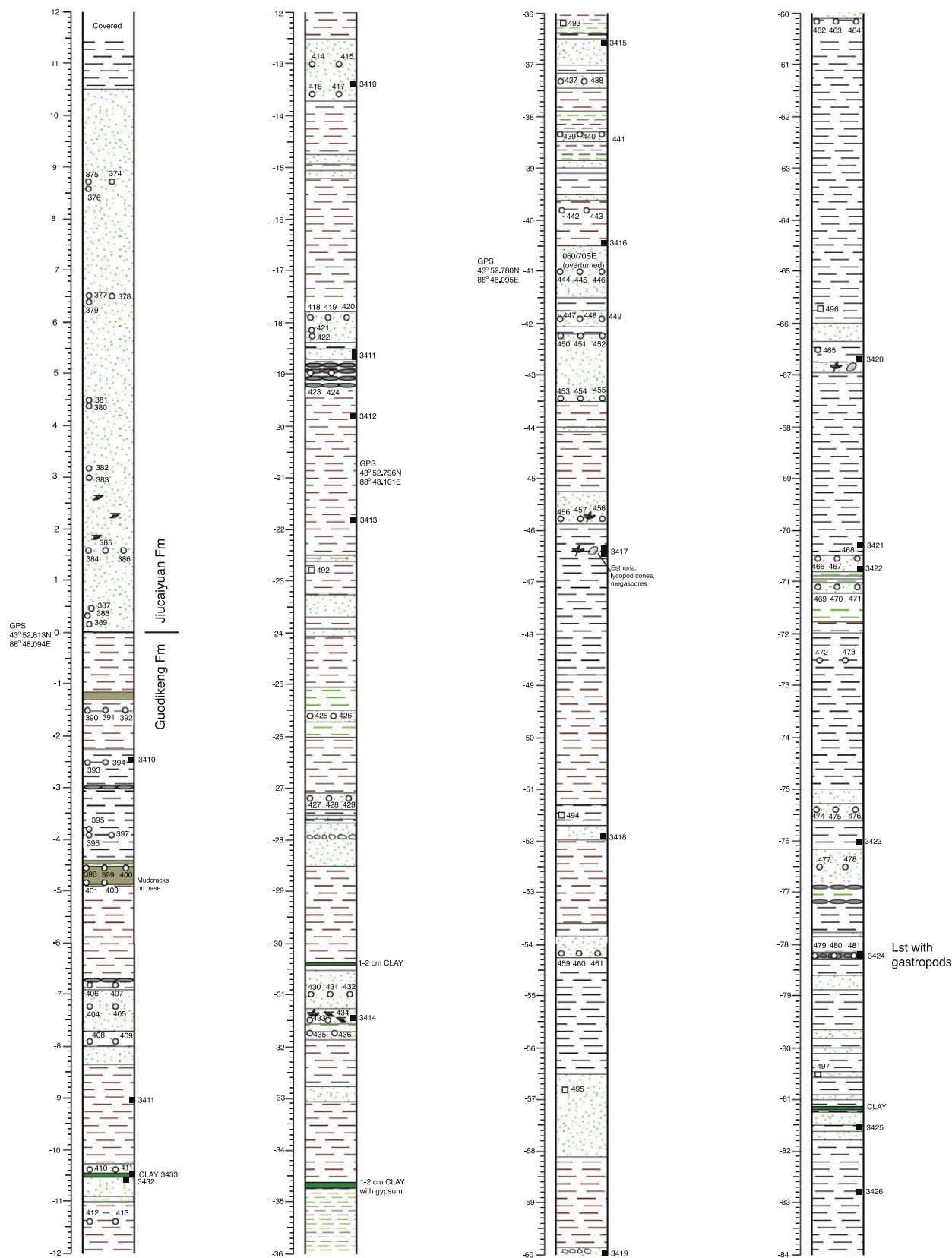


Fig. 9. Measured section in gully at Lucaogou (upper part). Legend as for Fig. 5.

And, while there are obvious reworked Devonian (common), Carboniferous, and Early Permian spores and pollen in Assemblage 3 palynofloras, the occurrences of *Scutasporites*

sp. cf. *S. unicus* and *K. schaubergeri*, and other typical Late Permian pollen are regarded as *in situ*, based on preservation, including the colour of the exine (lighter than of the re-

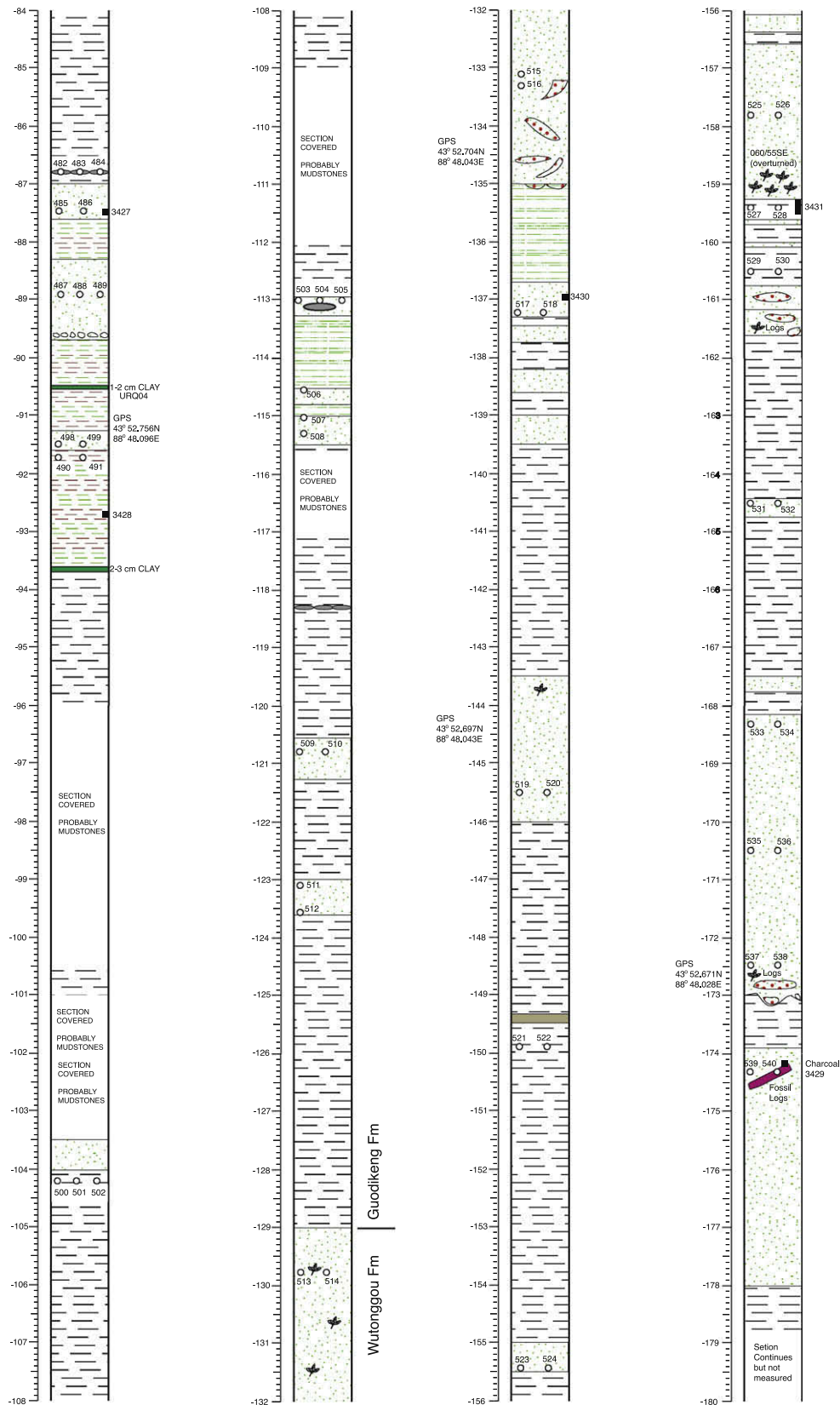
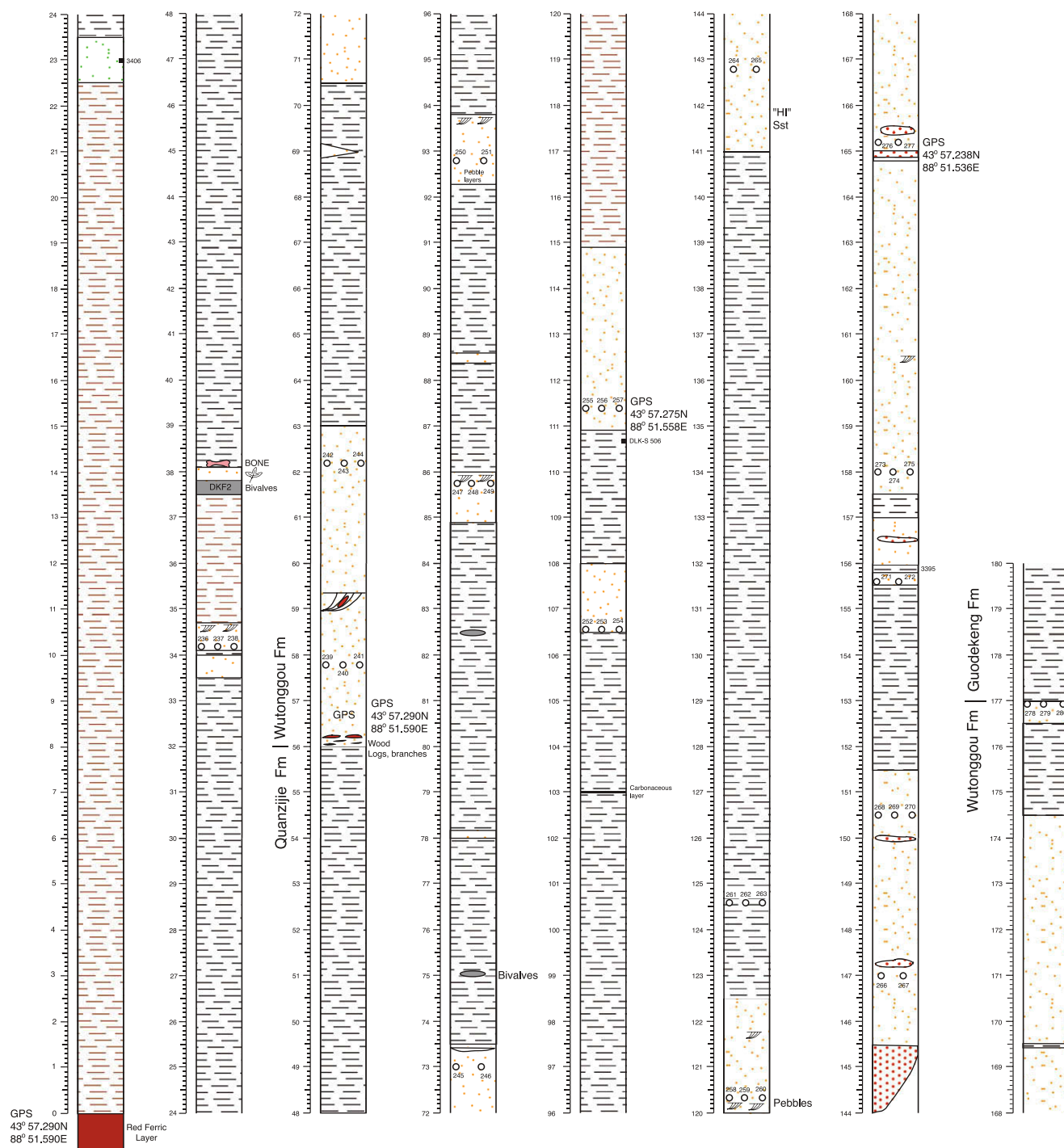


Fig. 10. Measured section in gully at Lucaogou (lower part). Legend as for Fig. 5.

worked elements), and abundance. Utting et al. (2004) have noted that reworking of Devonian and Carboniferous spores is common in Early Triassic assemblages from many parts

of the world, and suggest that some lycopod taxa used as indices for the Early Triassic, might be derived from the reworked palynofloras. Detailed discussion of their proposition





**Fig. 11.** DLS Traverse southwards along south limb of Dalongkou Anticline, east side of river. Zero metre is at top of red ferric weathering mudstones in upper part of Quanzijie Fm. Legend as for Fig. 5. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of the article.)

is outside the scope of this paper, but the lycopod taxa used by us are so far only known from latest Permian to early Triassic palynofloras.

Our study did not confirm the occurrence of *Aratrisporites* spp., but other lycopod taxa, namely *L. foveata* (a common component in most samples) and *Pechorosporites* spp. that define Assemblage 3, are known elsewhere from the basal Triassic (Yang et al., 1986; Yaroshenko et al., 1991). Yaroshenko (2006) considered that the *E. permasensis* – *P. disertus* assemblage from the Moscow Syncline, equated with palynofloras from the *Ophiceras commune* Zone, “though it probably also extends to a higher level in the Induan”, based on previous studies.

As noted above, many authors have recognised the transitional nature of the upper Guodikeng palynofloras: that is they contain species common to the Permian, and taxa that occur in younger Early Triassic assemblages. Transitional palynofloras, also with *O. eotriassicus*, have been recorded from a post-Tatarian succession in the Moscow Basin at Nedubrovo (Lozovsky et al., 2001; Foster and Afonin, 2005, Fig. 1).

Palynofloras assigned here to Assemblage 3 are found in the upper 100 m of the Guodikeng and in the lower Jiucayuan Formation. Whether or not the P–T boundary is placed at the base of Assemblage 3 is dependent on the real temporal ranges of the key taxa, such as *Pechorosporites* spp., and *L. foveata*, and the

**Table 1**

GPS locations for measured sections at Dalongkou and Lucaogou

		GPS at top	GPS at base
<i>Dalongkou north limb</i>			
TopP	Trench 3	43 57.685N, 88 52.51E	43 57.677N, 88 52.511E
	Trench 2	43 57.697N, 88 52.497E	
	Trench 1	43 57.674N, 88 52.497E	43 57.668N, 88 52.495E
	Trench 4		43 57.669N, 88 52.423E
Bottom	Gulley section	43 57.669N, 88 52.423E	
<i>Dalongkou south limb</i>			
	Trench 5	43 57.076N, 88 51.537E	43 57.160N, 88 51.656E
	DLS traverse		43 57.290N, 88 51.590E
<i>Lucaogou</i>			
	Gulley section	43 52.813N, 88 48.094E	43 52.671N, 88 48.028E

assumption that the first appearance of these taxa is synchronous. We require additional independent data to further test this. But this does not diminish the fact that the palynofloras represent a transitional flora: the lycopod genera represent pioneering plants, that later characterise younger Triassic assemblages, within a mature climax gymnosperm flora (Krassilov, 2003).

## 5.2. Vertebrates

In the Junggar Basin, Permian–Triassic vertebrate fossils are known from several localities near Jimsar north of the Tien Shan (Bogda Shan), including Xiaolongkou, Huangshanjie and Dalongkou. Similar and correlative vertebrate fossil assemblages are known from the Turpan basin south of the Tien Shan, especially at Taoshuyuan.

The original vertebrate fossil discoveries in the Junggar Basin were made at Dalongkou by P. Yuan in the late 1920s (Yuan, 1935). First documented by Yuan and Young (1934a, 1934b), these and later discoveries encompass Upper Permian vertebrate fossils assigned to the *Dicynodon* assemblage zone, and Lower Triassic vertebrate fossil assemblages assigned to the *Lystrosaurus* assemblage zone (e.g., Cheng, 1981, 1986; Lucas, 1998b, 2001, 2005, 2006). The section on the southern flank of the Dalongkou anticline is particularly significant to regional vertebrate biostratigraphy across the Permo–Triassic boundary as it yields both the *Dicynodon* and *Lystrosaurus* assemblages in stratigraphic superposition and documents the stratigraphic overlap of *Dicynodon* and *Lystrosaurus* (Cheng, 1981, 1986; Cheng and Lucas, 1993; Cheng et al., 1996).

In the Junggar Basin, the stratigraphic range of the *Dicynodon* assemblage zone is from the upper part of the Quanzijie Formation through the upper part of the Guodikeng Formation. This stratigraphic interval is more than 600 m thick, and vertebrate distribution is patchy and not prolific. Subsuming the vertebrate fossils from this entire stratigraphic interval into a single assemblage thus produces relatively coarse temporal resolution, but too few data are available to attempt vertebrate biostratigraphic subdivision of this interval.

*Dicynodon* assemblage vertebrates from the Junggar Basin are all dicynodonts, originally named *Jimusaria sinkiangensis* (Yuan and Young), *Jimusaria taoshuyuanensis* Sun; *Kunpania scopulosa* Sun, *Striodon magnus* Sun, *Turfanodon bogdaensis* Sun and *Dicynodon tienshanensis* Sun. Yuan and Young (1934a, 1934b), Sun (1973a, 1973b, 1978) and Cheng (1986) first published on these taxa, but now all of these genera are either considered synonyms of *Dicynodon* or *nomina dubia* (Cluver and Hotton, 1977, 1981; King, 1988; Lucas, 1998b, 2001, 2005). The only diagnostic Upper Permian dicynodont specimen from the Junggar Basin not assignable to *Dicynodon* is *Diictodon tienshanensis* (Sun). Sun (1973a) originally described the fossil as *D. tienshanensis*, but Cluver and Hotton (1977, 1981) reassigned this species to *Diictodon* (also see

King, 1988; Lucas, 1998, 2001, 2005). The tuskless skull, with its notched palatal rim, clearly belongs to the genus *Diictodon*. Therefore, in the Upper Permian strata of the Junggar Basin, only two dicynodont genera are known, relatively abundant *Dicynodon*, and rare *Diictodon*.

The *Lystrosaurus* assemblage in the Junggar Basin is derived from the upper part of the Guodikeng Formation and the overlying Jiucaiyuan Formation. *Lystrosaurus* is rare in the Guodikeng, but abundant in the Jiucaiyuan, and numerous species have been recognised (Yuan and Young, 1934b; Young, 1935, 1939; Sun, 1964, 1973b), though only two or three appear to be valid (Cluver, 1971; Colbert, 1974; Cosgriff et al., 1982; King, 1988; Lucas, 2001). Only the dicynodont *Lystrosaurus* has been documented from the Guodikeng Formation, but the Jiucaiyuan Formation also contains non-dicynodont taxa such as the proterosuchian “*Chasmatosaurus*” (= *Proterosuchus*) (Young, 1936), also known from the *Lystrosaurus* assemblage zone in South Africa (Lucas, 1998a).

No biostratigraphically significant fossil vertebrate records are known from the Guodikeng or Jiucaiyuan formations at the Lucaogou section. On the southern limb of the Dalongkou anticline, Cheng Zhengwu (personal communication to Lucas, 1996; also see Cheng, 1986; Li et al., 1986; Cheng and Lucas, 1993; Cheng et al., 1996) indicated in the field that the lowest occurrences of *Lystrosaurus* are ~160 m and 168 m above the base of the Guodikeng Formation. According to Cheng, the highest occurrence of *Dicynodon* is ~180 m above the base of the Guodikeng Formation, and there are more *Lystrosaurus* records stratigraphically higher in the Guodikeng Formation. This means there is a stratigraphic overlap of *Dicynodon* and *Lystrosaurus* at Dalongkou, as there is in the Karoo Basin of South Africa (e.g., Rubidge et al., 1995; Smith and Ward, 2001; Hancox et al., 2002; Retallack et al., 2003; Ward et al., 2005). On the northern limb of the Dalongkou anticline, no *Dicynodon* records are known from the Guodikeng Formation, but *Lystrosaurus* does have its lowest occurrence stratigraphically high in the Guodikeng Formation, at a horizon probably correlative to its record ~168 m above the Guodikeng base on the southern limb of the anticline.

Vertebrate paleontologists have long considered the lowest occurrence of *Lystrosaurus* to mark (or at least approximate) the base of the Triassic (cf. Lucas, 1998a). A vertebrate-based correlation of the lowest occurrence of *Lystrosaurus* on the southern limb of the Dalongkou anticline to the Karro basin of South Africa has thus been taken to indicate the PTB (based on vertebrate biostratigraphy) is in the upper part of the Guodikeng Formation at the level of the lowest occurrence of *Lystrosaurus* (e.g., Cheng, 1981, 1986; Cheng and Lucas, 1993; Cheng et al., 1996). More recent work in the Karoo Basin, however, indicates that the lowest occurrence of *Lystrosaurus*, which is in an interval of reversed magnetic polarity, must be older than the PTB (see Lucas, this volume). In the Karoo Basin, the highest occurrence of *Dicynodon* (about 40 m stratigraphically above the lowest occurrence of *Lystrosaurus*) is much closer to the PTB, though exact correlation is not certain.

It thus might be tempting to argue that the highest occurrence of *Dicynodon* on the southern limb of the Dalongkou anticline is equivalent to the PTB. In the Karoo Basin, in the stratigraphic interval in which *Lystrosaurus* and *Dicynodon* overlap, *Lystrosaurus* is rare. Above that interval it becomes extremely abundant (Smith and Ward, 2001). Similarly, on the southern limb of the Dalongkou anticline, *Lystrosaurus* is rare in the upper Guodikeng Formation but becomes extremely abundant in the overlying Jiucaiyuan Formation. In the Karoo, this acme of *Lystrosaurus* begins very close to or just above the PTB, and a correlation based on this acme suggests that the PTB at Dalongkou is very close to the base of the Jiucaiyuan Formation. Thus, vertebrate biostratigraphy reliably tells us that the PTB at Dalongkou is either very high stratigraphically

in the Guodikeng Formation or perhaps very close to or at the Guodikeng–Jiucaiyuan contact.

### 5.3. Conchostracans

Conchostracan assemblages are common in the Wutonggou, Guodikeng and Jiucaiyuan formations of the P–T sequence in the Junggar Basin and coquina levels of high abundance are observed, particularly in the lower Guodikeng Formation. Liu (1994) recognised 11 conchostracan assemblages for the non-marine Late Permian and Triassic of China. At Dalongkou, Liu (1994) recognised three conchostracan assemblages for the upper Wutonggou, Guodikeng and lower Jiucaiyuan Formations interval. These assemblages are, from oldest to youngest, the *Polygrapta* Assemblage, *Falsisca-Cornia-Cyclotunguzites* Assemblage, and the *Falsisca-Cyclotunguzites* Assemblage. Liu (1994) placed the P–T boundary at the base of his *Falsisca-Cornia-Cyclotunguzites* Assemblage, which is approximately 40 m above the base of the Guodikeng Formation on the northern limb of the Dalongkou anticline. Kozur (1998) discussed the conchostracan assemblages and ostracod faunas reported from Dalongkou, and together with his own field observations, compared these with other sections in China and with the Tunguska Basin, Siberia and the Germanic Basin. Kozur (1998) suggested that forms assigned to the genus *Cornia* in the *Falsisca-Cornia-Cyclotunguzites* Assemblage by Liu should be classified as the “typical Permian” genera *Megasitum* and *Bipemphigus* and he also reported another Permian genus, *Tripemphigus*, as being present in this assemblage. Kozur went on to suggest that the P–T boundary based on conchostracans and ostracods should be correlated approximately with the level of the recorded last appearance of the vertebrate *Dicynodon* at Dalongkou.

## 6. Chemostratigraphy

### 6.1. Carbon isotopes

Stable carbon isotope data from Permian–Triassic transitional sequences worldwide suggest that a marked negative  $\delta^{13}\text{C}$  excursion occurs at or close to the main end-Permian mass extinction level and a short distance below the currently defined Permian–Triassic boundary at Meishan, China (e.g. Baud et al., 1989; Holser et al., 1989; Morante, 1996; Wignall et al., 1998; Jin et al., 2000; de Wit et al., 2002; Korte et al., 2004; Krull et al., 2004; Riccardi et al., 2007).  $\delta^{13}\text{C}$  data for marine carbonates indicate a single negative excursion at the extinction level, whereas some  $\delta^{13}\text{C}$  data from organic kerogen material in non-marine terrestrial sequences appear to record multiple excursions at this level (e.g. de Wit et al., 2002).

Stable carbon isotope analyses of organic matter in palynology samples collected from the Dalongkou and Lucaogou sections were undertaken. The  $\delta^{13}\text{C}_{\text{org}}$  values are given in Table 2 and shown in Fig. 12. Values vary from –22.37 to –34.65 and despite our data set being rather coarse, there is no obvious single negative excursion that could be interpreted as representing the Late Changhsingian global negative excursion seen in the marine environment. Multiple excursions in the isotope curves are however seen to occur within the Guodikeng Formation (Fig. 12) similar to the multiple excursions observed in the P–T transitional zone of the non-marine sequences of Gondwana (de Wit et al., 2002). Correlation between the negative excursions documented on the North and South limbs, Dalongkou and at Lucaogou is not obvious, although the negative excursion points at 3353, 3390 and 3423 (Fig. 12) are potentially correlatable. In the lower to middle part of the Guodikeng Formation, isotopically light carbon  $\delta^{13}\text{C}$  corresponds at several levels with blooms of the algae *Reduviasporonites* (Foster et al., 2002) and abundant occurrence of conchostracans and the

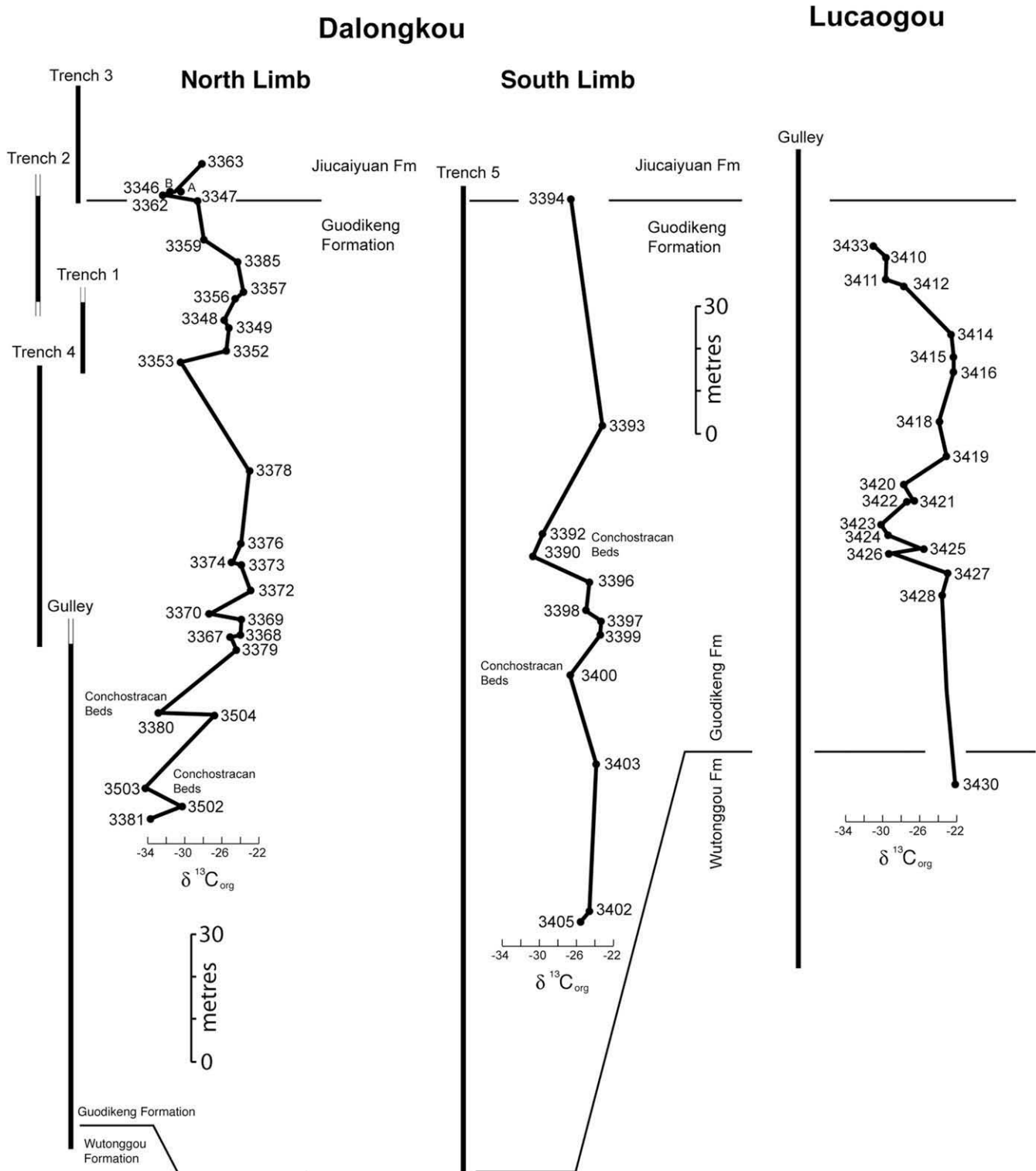
**Table 2**

$\delta^{13}\text{C}$  values obtained from Dalongkou and Lucaogou. For precise sample locations see Figs. 5–11

Sample number	Lithology/description	$\delta^{13}\text{C}$
<i>Dalongkou north limb carbon isotope data</i>		
3346	Greenish, silty sandstone	–30.87
3347	Green sandstone with plant fragments	–28.64
3348	Dark grey shale	–25.67
3349	Chocolate brown shale	–25.20
3352	Green fine-grained sandstone	–25.49
3353	Dark grey shale	–30.40
3356	Pale grey mudstone with carbonaceous flecks	–24.52
3357	Chocolate brown mudstone	–23.68
3362	Buff coloured laminated sandstone	–32.37
3363	Fine-grained sandstone with plant debris	–28.07
3367	Greenish shale	–24.95
3368	Chocolate brown mudstone	–23.97
3369	Seasonal “varved” mudstones	–23.87
3370	Seasonal “varved” mudstones	–27.27
3372	Grey green fine silty mudstone	–22.87
3373	Black clay	–23.88
3374	Fine-grained calcareous sandstone with nodules	–24.77
3376	Grey green laminated shale	–23.83
3378	Grey mudstone with plant stems	–23.00
3379	Seasonal “varved” mudstones	–24.40
3380	Estheria-rich mudstones	–32.70
3381	Estheria-rich mudstones with plant remains	–33.61
3385	Feldspathic sandstone	–24.24
3501	Tan nodular mudstone	–34.65
3502	Grey mudstone	–30.30
3503	Dark grey calcareous siltstone coquina	–34.23
3504	Dark grey calcareous siltstone coquina	–26.70
<i>Dalongkou south limb carbon isotope data</i>		
3390	Grey green mudstone with plant fragments and Estheria	–30.61
3392	Green grey sandstone	–29.42
3393	Sandy layer in chocolate brown shales	–23.22
3394	Fine-grained grey sandstone	–26.60
3396	Dark grey micaceous mudstone	–24.50
3397	Nodular fine-grained sandstone	–23.32
3398	Argillaceous grey siltstone	–24.92
3399	Claystone overlying a fine sand	–23.37
3400	Mudstone with rare Estherids	–26.61
3402	Very fine-grained green sandstone	–24.67
3403	Grey mudstone	–23.75
3405	Grey mudstone	–25.55
<i>Lucaogou carbon isotope data</i>		
3432	Grey mudstone	–31.09
3410	Fine-grained green sandstone	–29.56
3411	Chocolate brown mudstone	–29.63
3412	Sandstone with plant fragments	–27.76
3414	Grey silty sandstone	–22.50
3415	Dark grey silty mudstone	–22.37
3416	Fine grained sandstone with concretions	–22.38
3418	Weathered greenish sandstone	–23.86
3419	Grey shale	–22.99
3420	Laminated green sandstone with plant fragments	–27.77
3421	Dark grey siltstone	–26.61
3422	Fine-grained green sandstone with plant fragments	–27.37
3423	Laminated grey mudstone	–30.11
3424	Limestone with gastropods	–29.20
3425	Grey sandstone	–25.90
3426	Grey siltstone	–28.91
3427	Green fine-grained sandstone	–22.68
3428	Green/grey sandstone	–23.56
3430	Dark weathered shale/mudstone	–22.17
3433	Claystone with rare plant fragments	–31.32

green alga *Botryococcus*. These levels appear to represent freshwater ponds that have dried up. Negative carbon isotope excursions also correlate with occurrences of abnormal pollen. This suggests atmospheric climatic perturbations were occurring at the times of the C-isotope negative peaks resulting in severe climatic conditions (atmospheric pollution, including increased UV radiation,





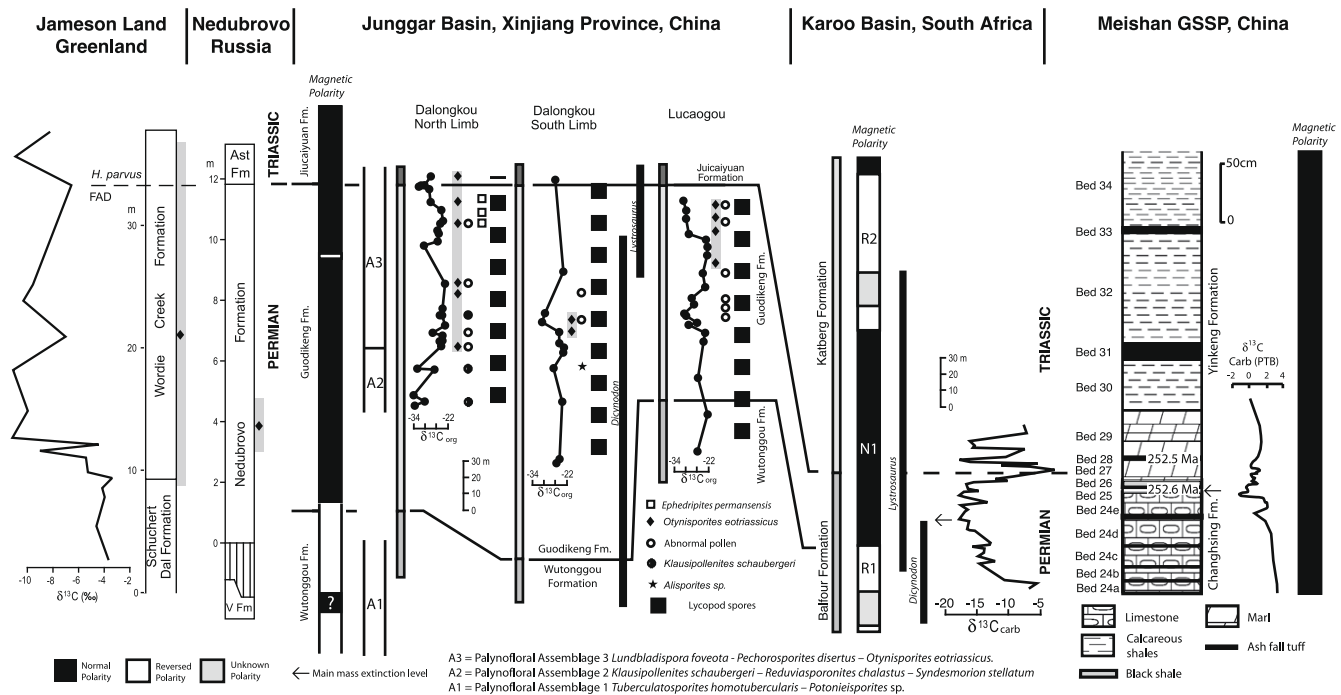
**Fig. 12.**  $\delta^{13}\text{C}_{\text{org}}$  isotope curves for the north and south limbs, Dalongkou, and Lucaogou sections. For locations of individual samples see Figs. 5–11.

caused by extensive volcanism) at the end-Permian mass extinction level. These perturbations of the atmosphere may well also reflect a delayed response to earlier regressive events and may have included methane influx from the release of gas hydrates.

In addition, recent preliminary studies of sulfur isotopes in our samples from Dalongkou record a rapid negative shift in sulfur isotopes in the upper Guodikeng Formation that indicates transfers of oceanic derived  $\text{H}_2\text{S}$  to the terrestrial ecosystems of the Junggar Basin (Newton *et al.*, 2007).

## 7. Discussion

The placement of the Permian–Triassic boundary in the Dalongkou and Lucaogou P–T sequences is equivocal. The distribution of vertebrates, which includes an overlap of *Dicynodon* and *Lystrosaurus* similar to that observed in the Karoo Basin of South Africa, suggests the P–T boundary at Dalongkou should be placed at the last appearance datum (LAD) of *Dicynodon* or higher. Palynofloras however, suggest a lower placement of the P–T boundary in the



**Fig. 13.** Correlation of the Dalongkou and Lucaogou non-marine sequences (this paper and Foster and Afonin, 2005) with other important P-T transitional sequences in Jameson Land, Greenland (Looy et al., 2001; Twitchett et al., 2001; Wignall and Twitchett, 2002; Foster and Afonin, 2005), Nedubrovo, Russia (Lozovsky et al., 2001), Karoo Basin, South Africa (MacLeod et al., 2000; De Kock and Kirschvink, 2004), and the Meishan Section D GSSP, China (Yin et al., 2001; Nicoll et al., 2002). See text for further explanation.

middle part of the Guodikeng Formation based on correlations with Russian sections. The level of the well-established negative C-isotope excursion seen in the marine environment (at or slightly above the mass extinction but below the GSSP P-T boundary level) cannot be determined from our  $\delta^{13}C_{org}$  data, which shows multiple negative excursions in the Guodikeng Formation that correspond to algal and conchostracan blooms and occurrence of abnormal pollen.

The magnetostratigraphy recorded at Dalongkou and Lucaogou shows a reversed to normal transition slightly above the Wutonggou-Guodikeng Formations boundary. The remaining Guodikeng and lower Jiuciyuan Formations are of normal polarity (with a possible short reversal in the upper Guodikeng represented by only one sample). The reversed to normal transition in the basal part of the Guodikeng is here correlated with the reversed to normal transition seen globally in many sections in the latest Changhsingian slightly below the main mass extinction level. Unfortunately, the magnetostratigraphy of the GSSP section at Meishan, South China is equivocal with two magnetostratigraphies (Li and Wang, 1989; Zhu and Liu, 1999) showing contrasting and inconsistent results. The magnetostratigraphy reported by Zhu and Liu (1999) places the Permian-Triassic boundary and FAD of *H. parvus* within a zone of uncertain polarity that equates to Bed 27 at the GSSP. Other sections in South China, e.g. Shangsi (Heller et al., 1988; Steiner et al., 1989; Glen et al., 2009), Zhongzhai (Metcalfe and Nicoll, 2007; Glen et al., 2009) and other parts of the world (e.g. Szurliés et al., 2003; Scholger et al., 2000; De Kock and Kirschvink, 2004) show the P-T boundary within a normal magnetic zone. It is therefore impossible, based on currently available information, to correlate precisely the P-T boundary (GSSP Level) in the P-T transitional sequences at Dalongkou and Lucaogou. However, the boundary must occur somewhere between the middle of the Guodikeng Formation and the basal Jiuciyuan Formation. The most likely position of the GSSP boundary level is close to the Guodikeng - Jiuciyuan formational boundary.

## 8. Conclusions

The P-T transitional non-marine lacustrine/fluvial sequences at Dalongkou and Lucaogou have been measured and logged, including the uppermost Quanzijie, Wutonggou, Guodikeng and lowermost Jiuciyuan formations.

The Wutonggou Fm is 121 m thick at Dalongkou and with its 12 coarse sandstone (with subordinate conglomerates) units represents a dominantly fluvial environment interpreted as a response to climate change, hydrological regimes and active foreland basin tectonics in the Junggar Basin. Palynology and conchostracan biostratigraphy suggests a Late Permian age.

The Guodikeng Formation is 197 m, and 209 m thick on the Dalongkou anticline northern limb and southern limb, respectively, and 129 m thick at Lucaogou. The formation comprises predominantly dark-grey, green and purple mudstones with minor siltstones and fine-grained green, buff and grey sandstones, and calcareous nodules deposited in a sedimentological regime starved of coarse terrigenous input, probably as a result of climatically induced hydrological changes and lake expansion in the latest Permian.

The Jiuciyuan Formation represents a further major depositional change and has a basal conglomerate, and a cycle of sandstone bodies at the base (although less coarse than those of the Wutonggou). Significant drying of the climate during Jiuciyuan Formation times is indicated.

Faunas and palynofloral assemblages of the Guodikeng Formation are of an Upper Permian-Lower Triassic transitional nature. The lower part of the Guodikeng is interpreted on palynological and conchostracan grounds to be of Upper Permian age. The upper Guodikeng can be interpreted as either Lower Triassic (based on palynofloras) or Upper Permian (based on vertebrates and conchostracans). Although plant fossils were not studied by our group, Yang et al. (1986; summary, p. 229) noted that they occurred "only in the lower and middle parts of this formation." and that the

“plant fossil assemblage also exhibits features of the late Angara flora (latest Permian)”.

Multiple  $\delta^{13}\text{C}_{\text{org}}$  negative excursions occur in the Guodikeng Formation. There is no obvious single negative excursion that could be interpreted as representing the Late Changhsingian global negative excursion seen in the marine environment. The recorded multiple excursions are similar to those observed in the P–T transitional zone of the non-marine sequences of Gondwana (de Wit et al., 2002) and correspond at several levels with blooms of the algae *Reduviasporonites* (Foster et al., 2002) and abundant occurrence of conchostracans and the green alga *Botryococcus* and occurrence of abnormal pollen. This suggests atmospheric climatic perturbations were occurring at the times of the C-isotope negative peaks resulting in severe climatic conditions (atmospheric pollution, including increased UV radiation, caused by extensive volcanism and oceanic degassing of  $\text{H}_2\text{S}$ ) at the end-Permian mass extinction level. These perturbations of the atmosphere may well also reflect a delayed response to earlier regressive events and may have included methane influx from the release of gas hydrates.

In the Junggar Basin, biostratigraphic data and magnetostratigraphic data indicate that the Permian–Triassic Boundary (GSSP Level) is somewhere in the middle to upper Guodikeng Formation and perhaps as high as the formational contact with the overlying Jiucuiyuan Formation.

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