The appearance of an oxygen-depleted condition on the Capitanian disphotic slope/basin in South China: Middle–Upper Permian stratigraphy at Chaotian in northern Sichuan

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ABSTRACT

The global environmental changes related to the end-Guadalupian (Permian) extinction have been recently studied in various shallow-marine sections in the world; however, no previous stratigraphic research focused on the sequence deposited in a relatively deep disphotic zone. In order to investigate the environmental changes in the disphotic zone during that interval, lithostratigraphy and secular changes in the sedimentary environment were analyzed for the ca. 150 m thick Guadalupian–Lopingian (Middle–Upper Permian) carbonates at Chaotian in northern Sichuan, South China. The upper Guadalupian Maokou Formation and the Lower Lopingian Wujiaping Formation are mostly composed of bioclastic limestone of a euhypsic shelf facies and contain abundant shallow marine fossils, such as algae, corals, and fusulines. The topmost Maokou Formation (ca. 11 m thick) is unique in this section because it is composed of thinly bedded black mudstone/chert of a disphotic slope/basin facies with abundant radiolarians and ammonoids. The stratigraphic changes in litho- and bio-facies suggest a two-stepped transgression in the Capitanian followed by a great regression. The appearance of the oxygen-depleted condition on the Capitanian disphotic slope/basin in northern Sichuan is particularly important because it occurred clearly before the end-Guadalupian extinction event. It is also noteworthy that the oxygen-depleted seawater appeared for the first time on a continental margin significantly before the well-known P–TB shallow-marine anoxia.

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1. Introduction

The end-Permian mass extinction, the largest of the Phanerozoic (e.g., Sepkoski, 1984; Erwin, 1993, 2006; Hallam and Wignall, 1997; Alroy, 2010), was traditionally regarded as a single event. However, Jin et al. (1994) and Stanley and Yang (1994) pointed out that this event consists of two distinct extinctions; i.e., the first extinction at the Middle–Late Permian (Guadalupian–Lopingian) boundary (G–LB; ca. 260 Ma) and the second at the Permian–Triassic boundary (P–TB; ca. 252 Ma). As to the first end-Guadalupian extinction, the significance of the environmental changes in the Capitanian (Late Guadalupian) time has been more emphasized recently (e.g., Isozaki and Ota, 2001; Retallack et al., 2006; Isozaki, 2007, 2009a; Bond et al., 2010a,b). In addition to the extinction, several unique geologic phenomena in the Capitanian have been recognized: 1) the onset of prolonged deep-sea oxygen-depletion (superanoxia; Isozaki, 1997), 2) a significant sea-level drop (Jin et al., 1994; Haq and Schutter, 2008), 3) the onset of stable carbon isotope fluctuations during the Paleozoic–Mesozoic transition interval i.e., a high (up to +5‰) positive plateau interval named the Kamura event (Isozaki et al., 2007a), 4) one of the two Phanerozoic minimums in strontium isotope ratios (Veizer et al., 1999; MacArthur and Haworth, 2004; Korte et al., 2006; Kani et al., 2008), 5) the eruption of the Emeishan flood basalt in South China (e.g., Chung and Jahn, 1995; Courtillot et al., 1999; Zhou et al., 2002; Wignall et al., 2009; Xu et al., 2010), and 6) the onset of frequent geomagnetic polarity changes (Illawarra Reversal; Irving and Parry, 1963;
Isozaki, 2009b). Refer to Isozaki (2007, 2009a) for more details concerning these Capitanian geological events. These multiple phenomena may indicate that certain global-scale environmental changes likely have occurred in the Capitanian and eventually caused the end-Guadalupian extinction (Isozaki, 2007, 2009a), however, the detailed cause–effect relationships including kill mechanisms for the Guadalupian fauna are still in discussion (e.g., Bottjer et al., 2008).

Previous stratigraphic research on the late Guadalupian events was performed mostly on shallow marine carbonates with abundant fossil records in South China (e.g., Jin et al., 1998) and pelagic deep-sea cherts in Japan (e.g., Isozaki, 1997, Nishikane et al., 2011). The Middle–Upper Permian and the lowermost Triassic strata deposited on the South China craton have exceptionally well-preserved, continuous stratigraphic records that span the two important extinction-related intervals with abundant and diverse shallow-marine fossils (e.g., Zhao et al., 1981; Yang et al., 1987; Jin et al., 1998). For example, the detailed stratigraphy across the G–LB is analyzed at the best continuous section at Penglai in Guangxi, which is officially designated as the Global Stratotype Section and Point (GSSP) for the G–LB (e.g., Wang et al., 2004; Jin et al., 2006; Shen et al., 2007).

In order to clarify the global environmental changes in Capitanian time, more information from the disphotic zone (usually deeper than 150 m) is needed, particularly in linking the extinction-related shallow-marine phenomena and the deep-sea oxygen-depletion event. Several studies report Guadalupian rocks of a relatively deep-water facies in South China (e.g., Kametaka et al., 2005; Sun and Xia, 2006), however, no previous studies have focused on the importance of the disphotic zone with respect to the Capitanian environmental changes. The late Guadalupian to early Lopingian strata of a relatively deep-water facies occur on the northwest margin of the South China craton (e.g., Li et al., 1989; Zhu et al., 1999; Wang and Jin, 2000; Fig. 1D). As these sequences may provide us with paleoenvironmental information different from that from commonly studied shallow-marine carbonates, the current study analyzed a detailed stratigraphy of the Guadalupian–Lopingian strata at Chaotian in northern Sichuan (Fig. 1A, D). On the basis of detailed observation both of the outcrops and drill core samples at Chaotian, we report on remarkable fluctuations in sea-level and in associated seawater redox on the relatively deep slope/basin of northwestern South China.

2. Geologic setting

During the late Paleozoic to early Mesozoic, South China was located on the eastern side of the supercontinent Pangea around the equator as an isolated continental block (e.g., Scotese and Langford, 1995; Muttoni et al., 2003; Fig. 1C). Thick shallow-marine shelf carbonates and terrigenous clastics were extensively deposited on South China during the Permian, which yield abundant and diverse shallow-marine fossils such as fusulines, brachiopods, mollusks, and rugose corals (e.g., Zhao et al., 1981; Yang et al., 1987; Jin et al., 1998). In northern Sichuan, along the southwestern edge of the South China craton, thick carbonates of a relatively deep-water facies accumulated in a slope/basin setting (Fig. 1D); e.g., the Shangsi section, once nominated as a candidate for the GSSP.
of the P–TB, showing a contrast to the typical shallow shelf sequences in other regions in South China (e.g., Li et al., 1989; Wignall et al., 1995; Lai et al., 1996; Jiang et al., 2011; Song et al., 2011; Fig. 1A).

We studied one such deep-water carbonate-dominant sequence at Chaotian (Latitude 32°37′ N, Longitude 105°51′ E), about 20 km to the north of Guangyuan city in northern Sichuan or about 60 km to the northeast of the above-mentioned Shangsi section (Fig. 1A).

This section displays continuous stratigraphy from the Middle Permian to lowermost Triassic in excellent exposures along a narrow gorge (Mingyuexia) of the Jialingjiang River, a branch of the Changjiang River (Figs. 1A, B and 2). The overall biostratigraphy of the Chaotian section was originally clarified on the basis of fusulines, conodonts and ammonoids (Zhao et al., 1978; Yang et al., 1987). Later, from the viewpoint of the two Permian mass extinctions,
The Wujiaping and Dalong formations are mainly correlated with the Wordian (Middle Guadalupian) to Capitanian (Upper Guadalupian). The upper 145 m thick part of the Maokou Formation is correlated with the bed. Recently, He et al. (2010) slightly modified the interpretation of Isozaki et al. (2004) into a volcaniclastic origin. Wangpo bed by identifying volcanic phenocrysts and He et al. (2007) obtained by deep drilling at Chaotian. As to the Permian–Triassic strata at Chaotian, we adopt the overall stratigraphic framework by Yang et al. (1987) and Isozaki et al. (2004, 2007b) including their stratigraphic subdivision and unit names. The Permian–Triassic rocks at Chaotian consist of the Guadalupian Maokou Formation, the Lopingian Wujiaping and Dalong Formations, and the lowermost Triassic Feixianguan Formation, in ascending order (Fig. 2). The Maokou Formation, over 200 m thick, is mainly composed of dark gray bioclastic limestone (packstone) partly containing black chert nodules, and yields abundant shallow-marine fossils, such as calcareous algae, ostracodes, and fusulines. The topmost part of the Maokou Formation, ca. 11 m thick, is composed of bedded black calcareous mudstone and black chert with ammonoids, small-sized brachiopods, radiolarians, and conodonts. The Wujiaping Formation, ca. 70 m thick, is mainly composed of dark gray bioclastic limestone (packstone) partly containing black chert nodules. At the base of the Wujiaping Formation, however, a ca. 2 m thick unique non-carbonate bed called the Wangpo bed occurs. Bioclastic packstone of the Wujiaping Formation contains shallow-marine fossils including brachiopods, calcareous algae, and conodonts. The Dalong Formation, ca. 26 m thick, is mainly composed of black mudstone and limestone (limite mudstone/wackestane), and contains conodonts and ammonoids. The Feixianguan Formation, over 30 m thick, is mainly composed of thinly bedded dark gray marl and light gray limestone (lime mudstone), and yields few conodonts, ammonoids, and brachiopods.

The Wangpo bed at the base of the Wujiaping Formation has been traditionally considered as a terrigenous shale bed that suggests a great regression in the western part of South China (Lu, 1956; Mei et al., 1994). Isozaki et al. (2004) first clarified the volcanic origin of the Wangpo bed by identifying volcanic phenocrysts and He et al. (2007) reported the SHRIMP U–Pb age of 260 ± 4 Ma for euhedral zircons from the bed. Recently, He et al. (2010) slightly modified the interpretation of Isozaki et al. (2004) into a volcaniclastic origin.

According to the previous reports of fusulines, conodonts, and ammonoids (Yang et al., 1987; Isozaki et al., 2004, 2007b, 2008; Ji et al., 2007), the Permian rocks at Chaotian are dated as follows (Fig. 2): The upper 145 m thick part of the Maokou Formation is correlated with the Wordian (Middle Guadalupian) to Capitanian (Upper Guadalupian). The Wujiaping and Dalong formations are mainly correlated with the Wuchiapingian (Lower Lopingian) and with the Changhsingian (Upper Lopingian), respectively. On the basis of the above age data, Isozaki et al. (2008) tentatively assigned the biostratigraphically defined G–LB at the base of the Wujiaping Formation (Fig. 2). We follow the age assignment.

In this article, we report the detailed lithostratigraphy and total organic carbon (TOC) contents of the upper part of the Maokou Formation (ca. 145 m thick) and the lowermost part of the Wujiaping Formation (ca. 10 m thick) (Fig. 2). Moreover, on the basis of lithofacies descriptions and geochemical results, we discuss the secular changes in sea-level associated with redox of the sedimentary environments.

3. Lithostratigraphy

For lithofacies analysis in the laboratory, we analyzed more than 300 rock samples in total, collected from the outcrop and from drill core. Besides polished slabs for each specimen, about 600 thin sections were made for petrographic observation. We qualitatively evaluated relative abundance of each fossil taxon from both polished slabs and thin sections. The degree of bioturbation was determined using the ichnofabric index of Droser and Bottjer (1986). Figs. 6 and 7 show the detailed stratigraphic columns of the upper Maokou and lowermost Wujiaping formations with fossil occurrences.

3.1. The Maokou Formation

The main part of the upper Maokou Formation consists of ca. 130 m thick massive dark gray limestone, whereas the topmost ca. 11 m thick part is composed of thin alternations of black calcareous mudstone, black chert, and dark gray muddy carbonates (Figs. 2 and 6). In this article, we refer the main limestone part and the topmost mudstone part as the ‘Limestone Unit’ and the ‘Mudstone Unit’, respectively. Detailed description of lithofacies of the Mudstone Unit was mainly conducted using exceptionally fresh drill core samples collected from deeper than 150 m below the surface.

3.1.1. The Limestone Unit

The Limestone Unit is mainly composed of massive dark gray bioclastic limestone, partly containing black chert nodules of centimeter to decimeter size. Packstone dominates with a minor amount of wackestone, grainstone, and rudstone (Figs. 2, 3A, 4A–C and 6). This limestone yields abundant and various shallow-marine fossils such as calcareous algae, small foraminifera, and ostracodes, together with fusulines, rugose corals, bryozoans, crinoids, brachiopods, gastropods, bivalves, and trilobites. No significant change in fossil taxa and their relative abundance was observed throughout the Limestone Unit. Peloids of 200–400 μm in diameter occur frequently. Burrows of 2–7 mm in diameter are common but no particular alignment of burrows is recognized. Neither cross lamination nor grading structure is recognized in the Limestone Unit.

The Limestone Unit is divided into three subunits of distinct lithofacies: i.e., L1, L2, and L3, in ascending order (Figs. 2 and 6). The L1 subunit (ca. 60 m thick) is mainly composed of massive packstone containing abundant calcareous algae, small foraminifera, and ostracods.

The L2 subunit (ca. 40 m thick) mainly consists of massive packstone with minor amounts of grainstone (Fig. 4B), floatstone, and rudstone. The occurrence of floatstone and rudstone with abundant granule-sized particles is limited to the L2 subunit (Figs. 2 and 6). The L2 subunit contains many cm-sized black chert nodules that align nearly parallel to bedding planes. Abundant calcareous algae, crinoids, and large fusulines occur in the subunit.

The L3 subunit (ca. 35 m thick) is mainly composed of packstone with abundant bioclasts of calcareous algae, small foraminifera, and ostracods, and also peloids (200–400 μm in diameter). Angular dark brown phosphate grains (50–200 μm in diameter) are concentrated within the topmost part (ca. 30 cm thick) of the L3 subunit (Figs. 4C, 7).

3.1.2. The Mudstone Unit

The Mudstone Unit, ca. 11 m thick, is composed of more than 350 thin beds; i.e., thinly bedded black calcareous mudstone, black chert, black siliceous mudstone, dark gray limestone, and dark gray dolostone (Figs. 3B, E, F, 4D, E, and 7). The lithofacies boundary between the lower Limestone Unit and the upper Mudstone Unit is conformable without any sharp erosion feature (Fig. 3D). On the outcrop, calcareous mudstone and carbonate beds were mostly weathered into soft brown clayey material. Bedding planes in the Mudstone Unit are generally clear and almost all brachiopod shells are fragmented to the size less than
1 cm. This mode of occurrence likely suggests the allochthonous origin of the fossils. Several turbiditic muddy limestone layers have wavy bedding with normally graded brachiopod fragments. Bioturbation is completely absent in the Mudstone Unit and black calcareous mudstone layers show clear thin lamination (Figs. 3F and 4D). Framboidal pyrites of 5–10 μm in diameter occur abundantly and constantly throughout the Mudstone Unit.

On the basis of lithofacies characteristics, the Mudstone Unit is divided into two subunits; i.e., the lower carbonate-rich M1 subunit (8 m thick) and the upper black mudstone-dominant M2 subunit (3 m thick) (Figs. 2, 6 and 7). The M1 subunit is mainly composed of alternations of black calcareous mudstone, black chert, and dark gray muddy limestone. This subunit yields abundant conodonts, brachiopods, small foraminifera, and ostracods, and brachiopod fragments are concentrated in several specific layers. Dark brown angular phosphate grains (50–200 μm in diameter) and phosphate nodules (1 mm–2 cm in diameter) occur particularly at the base of M1 (ca. 10 cm thick part) (Figs. 3D and 7).

The M2 subunit is mainly composed of alternations of black calcareous mudstone, black chert, and black siliceous mudstone. In contrast to M1, the carbonate content is relatively low and black calcareous mudstone is dominant (Fig. 7). The M2 subunit yields abundant radiolarians, ammonoids, and gastropods, and their occurrence is limited to several chert layers.

3.2. The Wujiaping Formation

The total thickness of the Wujiaping Formation at Chaotian is ca. 70 m. This study focused solely on the lowermost 10 m thick part of the Formation (Figs. 2, 6, and 7). This part is mainly composed of dark gray bioclastic limestone, whereas its base is the ca. 2 m thick Wangpo bed.

The basal Wangpo bed is composed of light gray felsic tuff and contains clay minerals, such as illite and montmorillonite, with apatite, zircon, plagioclase, and bipyramid quartz (Isozaki et al., 2004). On the outcrop, the Wangpo bed is altered to yellow-reddish brown bentonite. A ca. 1 m thick black mudstone bed (‘Heshan bed’ in He et al., 2010) is deposited on the Wangpo bed (Fig. 7). The overlying bioclastic limestone mainly consists of dark gray mud-poor packstone and gray grainstone (Figs. 3C and 4F), thus sparitic matrix dominates...
in the limestone. The bioclastic limestone yields abundant fragments of calcareous algae, accompanied by small-sized fusulines, smaller foraminifera, gastropods, ostracods, and bivalves, and peloids of 300–500 μm in diameter (Figs. 6 and 7). A few burrows of ca. 2 mm-diameter are recognized.

4. Total organic carbon content

Total organic carbon (TOC) contents of the rock samples (mostly from the ca. 25 m thick part across the G–LB) were measured. Powdered sample of 50–200 mg were treated with 1–2 ml of 10 M HCl, and were reacted completely by in an ultrasonicator for 1 h; the sample was then left for 24 h at room temperature. The residue was diluted by distilled water, centrifuged and dried at 80 °C for over 12 h. The remaining 100–1000 μg of residue was then placed into a tin cup for TOC measurement by an Elemental Analyzer at the Department of Environmental Chemistry and Engineering, Tokyo Institute of Technology.

Table 1 shows results of all measurements for 56 samples and Fig. 6 shows the stratigraphic change in TOC value. Fig. 7 illustrates an enlarged view of the ca. 25 m thick part across the G–LB with TOC values of 49 samples. Averaged TOC value in the Limestone Unit of the Maokou Formation is 0.21% (n=11). In contrast, those of the sedimentary rocks of the Mudstone Unit of the Maokou Formation are remarkably higher; i.e., 1.98% (black chert/siliceous mudstone: n=4), 9.85% (black calcareous mudstone: n=13) and 1.61% (dark gray carbonate: n=25), respectively. Averaged TOC value in the lowermost part of the Wujiaping Formation is 0.11% (n=3).

Within the Mudstone Unit of the Maokou Formation, the TOC values are systematically different according to their rock types (Fig. 7): TOC values of black calcareous mudstone are consistently higher than those of black chert/siliceous mudstone and dark gray carbonate. In addition, TOC values of black calcareous mudstone in the upper M2 subunit of the Mudstone Unit are relatively higher than those in the lower M1 subunit.

5. Depositional age

On the basis of the previously reported index fossils, such as fusulines, conodonts, and ammonoids (Yang et al., 1987; Xu, 2006;
Kuwahara et al., 2007, 2008; Isozaki et al., 2008; Lai et al., 2008), the geological ages of the upper Maokou and lowest Maowujiaping formations at Chaotian are constrained as follows (Fig. 2).

5.1. The Maokou Formation

5.1.1. The Limestone Unit

The occurrence of fusulines dominated by Neoschwagerina cf. kueischei from the base of the L1 subunit indicates a Wordian (Middle Guadalupian) age, whereas the one dominated by Lepidolina guiberti from the upper L2 subunit indicates a Capitanian (Late Guadalupian) age (Isozaki et al., 2008; Fig. 2). Therefore, the ages of both the L1 and L2 subunits are the Wordian to Capitanian, whereas the precise Wordian–Capitanian boundary horizon is unknown.

On the other hand, the conodont Jinogondolella postserrata occurs at the topmost part of the L3 subunit (Fig. 2). This indicates that the topmost part of the L3 subunit belongs to the J. postserrata Zone of the early Capitanian age (Isozaki et al., 2008; Lai et al., 2008; Figs. 2 and 5). Thus, the age of the entire L3 subunit is no doubt early Capitanian.

5.1.2. The Mudstone Unit

J. postserrata and J. shannoni occur both from the basal and the topmost parts of the M1 subunit of the Mudstone Unit (Isozaki et al., 2008; Lai et al., 2008; Fig. 2). The M1 subunit, therefore, belongs...
to the J. shannoni Zone of the early–middle Capitanian (Fig. 5). The occurrence of radiolarians of the Pseudoalbaillella longtanensis–Pseudoalbaillella globosa Assemblage from M1 supports this age assignment (Kuwahara et al., 2007).

On the other hand, from the upper M2 subunit, ammonoids of genus Paraceltites, indicating the Roadian (Early Guadalupian) to Wuchiapingian (Early Lopingian) age, and genera Cibolites and Altudoceras, ranging from the Wordian–Wuchiapingian occur (Isozaki et al., 2008; Fig. 2); however, as these ammonoids are long-ranging taxa, the detailed age of the M2 subunit is not fully constrained. As no conodont characterizing the middle–late Capitanian; i.e., those of the J. altudaensis Zone. J. prexuanhanensis Zone, J. granti Zone, and Clarkina postbitteri hongshuensis Zone, occurs at Chaotian, it is suggested that the upper Capitanian rocks are possibly missing (Isozaki et al., 2008; Lai et al., 2008; Fig. 5).

5.2. The Wujiaping Formation

The basal Wangpo bed was directly dated ca. 260 ± 4 Ma by SHRIMP U–Pb dating of zircons (He et al., 2007; Fig. 2). The black mudstone immediately above the Wangpo bed contains no index fossils, whereas the overlying bioclastic limestone yields abundant small fusulines, Codonofusiella and Reichelina, without any large ones such as Neoschwagerina and Lepidolina (Isozaki et al., 2008; Figs. 2, 6, and 7). No conodonts occur in the Wujaping bioclastic limestone. Although Codonofusiella and Reichelina appear first in the Guadalupian, their dominance characterizes the Wuchiapingian age (e.g., Sheng et al., 1984). The lowermost part of the Wujaping Formation at Chaotian is, thus, likely correlated with the Wuchiapingian (Isozaki et al., 2008). This assignment is concordant with the U–Pb age of the basal Wangpo bed.

Accordingly, Isozaki et al. (2008) tentatively assigned the biostratigraphic G–LB at the bottom of the bioclastic limestone of the lowermost Wujaping Formation at Chaotian (Figs. 2, 6 and 7), and this paper follows this.

5.3. Extinction horizon

At Chaotian, the shallow-marine taxa apparently disappear at the lithofacies boundary between the Limestone Unit and the Mudstone Unit of the Maokou Formation (the L3/M1 boundary; Figs. 6 and 7). The overlying M1 subunit of the Mudstone Unit, however, belongs entirely to the early–middle Capitanian (the J. shannoni Zone; Isozaki et al., 2008; Lai et al., 2008; Figs. 2 and 5); the L3/M1 boundary is not likely a major extinction horizon. Thus, the disappearance of the shallow-marine taxa across this facies boundary represents an apparent result of a facies-controlled fossil occurrence, as discussed later. The real extinction horizon at Chaotian is probably within or at the top of the upper M2 subunit of the Mudstone Unit, although the precise horizon is not yet clarified due to the poor occurrence of index fossils. Judging from the widely known hiatus at the base of the overlying Wangpo bed in western South China, the actual extinction horizon is expected possibly at the top of the M2 subunit at Chaotian (Fig. 7).

As to the timing of the extinction of the Guadalupian fauna, Bond et al. (2010a,b) recently proposed that the so-called end-Guadalupian extinction actually occurred in the mid-Capitanian (at the top of the J. altudaensis Zone; Fig. 5), significantly before the G–LB, on the basis of fusulines and chemotaxonomic data from Yunnan and Guizhou. In the Tieqiao section next to the GSSP in Guangxi, however, the occurrence of Lepidolina sp., a taxon that went extinct in the end-Guadalupian extinction, was reported from the upper J. granti Zone (latest Capitanian; Jin et al., 2006; Shen et al., 2007; Fig. 5). This indicates that the large-tested fusulines did not disappear at the top of the J. altudaensis Zone as Bond et al. (2010a,b) suggested, but survived much longer at least up to the two higher conodont zones. Moreover, even in their own study of the Gouchang section in Guizhou, Wignall et al. (2009) and Bond et al. (2010a) reported the occurrence of Neoschwagerinidae indet. from a stratigraphic horizon significantly above the proposed extinction interval. As Clapham et al. (2009) recently commented, the so-called end-Guadalupian extinction might occur in a gradual manner rather than as a rapid event (also refer to Yang et al., 2004). Under these circumstances, the recognition of a main extinction event in the mid-Capitanian needs further research.

6. Discussion

On the basis of the newly described lithologic and geochemical characteristics, we discuss the depositional environments of the upper Maokou and lowermost Wujiaping formations at Chaotian, with particular focus on the secular changes in sea-level and in associated seawater redox of the relatively deep slope/basin in northwestern South China.

6.1. Sedimentary environment

The abundant bioclasts of sand-granule size (e.g., algae, corals, brachiopods, and bryozoans) together with lime-mud (Figs. 3A, 4A–C and 6), without any indication for extremely shallow environment such as cross laminations or evaporites, in the Limestone Unit of the Maokou Formation (L1–L3) indicate that these rocks were mainly deposited on a continental shelf probably below the storm wave base (generally 50–80 m deep). In addition, the abundant occurrence of algae and large fusulines indicates the deposition within a euphotic zone (usually less than 150 m deep; Fig. 7).

In contrast, the Mudstone Unit of the Maokou Formation is dominated by very fine-grained siliceous rocks that are enriched in radiolarians and ammonoids without shallow-marine fossils (e.g., corals, fusulines, and calcareous algae). As radiolarians and ammonoids are generally dominant in the open ocean with relatively great water depths, their occurrence indicates that the Mudstone Unit was deposited on a relatively deep slope/basin setting off the outer shelf margin. Making precise estimates for ancient water-depth is difficult; however, the complete absence of fossils of photosynthetic organisms (e.g., algae and large fusulines) suggests that the depositional depth was clearly below the euphotic zone; in other words, the sedimentary environment was in the disphotic zone (probably over 150 m deep; Fig. 7). However, Lai et al. (2008) reported a different interpretation as to the sedimentary environment of the Mudstone Unit; therefore we discuss it in detail separately in the next section. On the other hand, Zhao et al. (1978) reported the occurrence of land plant debris from the upper black mudstone bed, immediately above the Wangpo bed, in the lowermost Wujaping Formation at Chaotian. This suggests that the sea-level should have been extremely
low to receive plant debris (Fig. 7), although the ancient water-depth of the depositional site of the Wangpo bed is not well constrained. The upper Wujiaping bioclastic limestone is mainly composed of dark gray mud-poor packstone and gray grainstone with sparite and lesser amount of lime-mud matrix, containing shallow-marine fossils such as calcareous algae and fusulines (Figs. 3C, 4F, 6 and 7). These
characteristics indicate that this limestone was deposited on a relatively shallow high-energy shelf in the euphotic zone (Fig. 7).

6.2. Topmost Maokou Formation: deep or shallow?

There is a remarkable disagreement in interpretation between Isozaki et al. (2008) and Lai et al. (2008) as to the sedimentary environment of the Mudstone Unit of the Maokou Formation, even though both studied exactly the same section (Chaotian). Isozaki et al. (2008) interpreted the Mudstone Unit as sediments of a deep-water slope/basin facies, whereas Lai et al. (2008) regarded it as sediments deposited in an extremely shallow-water setting, such as lagoon/tidal flat (Fig. 7). The present study positively supports the former. Key criteria to solve this disagreement are in the following aspects of the faunal assemblage of this unit (Figs. 6 and 7); i.e., 1) the abundant occurrence of radiolarians and ammonoids (this study; Kuwahara et al., 2007; Isozaki et al., 2008) that are common in an open-ocean setting with much greater water depth but not in an extremely shallow-water lagoon; 2) the dominance of gondolellid-form conodonts (Isozaki et al., 2008; Lai et al., 2008) suggesting a relatively deep-water facies (e.g., Clark, 1981; Wang and Wang, 1997); and 3) the complete absence of fossils of shallow-water photosynthetic organism (e.g., algae and large fusulines). These characteristics clearly indicate the deposition of the Mudstone Unit in a deep-water environment. In addition, the occurrence of turbidite beds intercalated within the Mudstone Unit also supports the deep-water interpretation because it implies the development of a slope to generate gravity flows. We can conclude, therefore, that the topmost Maokou Formation no doubt recorded a remarkable deepening episode in the Capitanian continental shelf in northwestern South China.

6.3. Multiple sea-level changes

Lithofacies changes abruptly at the boundary between the Limestone Unit and the overlying Mudstone Unit of the Maokou Formation (the L3/M1 boundary; Figs. 3D and 6). Across this boundary, the rocks become much more siliceous, along with a remarkable size decrease of matrix particles from sand to clay. In addition, the biofacies change dramatically across the boundary; i.e., radiolarians and conodonts become remarkably abundant, whereas shallow-marine fossils, such as calcareous algae, fusulines, and corals, disappear totally (Figs. 6 and 7). These synchronized changes in litho- and bio-facies do not appear to record the global extinction event but a local ecological change along with the deepening of the depositional setting; i.e., the above-mentioned change from a shallow shelf to a deep slope/basin (Isozaki et al., 2008). The complete disappearance of calcareous algae, large fusulines, and corals across the L3/M1 boundary strongly suggests the sedimentary environment shifted concordantly from euphotic to disphotic conditions, as those fauna were probably dependent on photosymbiosis. Although the L3/M1 boundary appears sharp, there is no evidence for remarkable erosion both at outcrop and in drill core (Fig. 3D). A faint erosion may have been accompanied the transgression, however, all the observations suggest a rapid sea-level rise. The concentration of phosphate grains and nodules strictly around this lithofacies boundary (Figs. 3D, 4C and 7) might be related to the abrupt sea-level rise.

The lithofacies change at the upper M1/M2 subunit boundary within the Mudstone Unit is also noteworthy (Fig. 7). From the M1 to M2, carbonate decreases in volume ratio, whereas black calcareous mudstone increases. In addition, the composition of fossil taxa...
changes again dramatically across the M1/M2 boundary. Brachiopods become very rare, whereas radiolarians become remarkably dominant, suggesting that the sea-level has risen again across the M1/M2 boundary and the depositional site has become much deeper.

On the contrary, the lithofacies change between the topmost Maokou Formation and the overlying Wujiaping Formation records a remarkable sea-level drop (Fig. 7); the black mudstone bed immediately above the Wangpo bed was probably deposited on an extremely shallow environment to receive plant debris. Next, the overlying Wujiaping bioclastic limestone indicates a slight sea-level rise again up to the carbonate deposition level. The above results document that the sea-level has changed drastically in the Capitanian to early Wuchiapingian at Chaotian in northern Sichuan. In particular, this study identifies for the first time a two-stepped sea-level rise followed by a sharp drop (Fig. 7).

A similar trend in lithofacies change from bioclastic limestone to siliceous rock in the topmost Maokou Formation was reported elsewhere in South China: e.g., the Shangsi section, ca. 60 km to the southwest (Li et al., 1989; Yan et al., 2008; Xie et al., 2008; Fig. IA), the Dukou and Nanjiang sections in northeastern Sichuan (Mei et al., 1994; Zhang et al., 2008), the Maershan section in western Hubei (Xia et al., 2005; Zhang et al., 2008), and the Xiongjiajiang and Gouchang sections in western and central Guizhou (Wignall et al., 2009). These concordant observations suggest that a sea-level rise occurred extensively in western South China during the Capitanian. The two-stepped sea-level rise recognized in the present study needs to be tested in those sections also. In contrast, outside South China, such a record of sea-level rise in the Capitanian is quite rare. The Izawaki Limestone in Northeast Japan, positioned along the eastern margin of South China in the Permian (e.g., Maruyama et al., 1997; Isozaki et al., 2010), records a similar transgression signature (Shen and Kawamura, 2001). Judging from these data, we consider that the Capitanian sea-level rise may not represent global eustasy but record local subsidence of sedimentary basins restricted to South China, possibly related to regional tectonics within South China.

On the other hand, the end-Guadalupian great regression has been widely recognized on a global scale (e.g., Jin et al., 1994; Haq and Schutter, 2008), as a remarkable unconformity developed across the G-LB in many sections across the world (e.g., Mei and Wardlaw, 1996; Hallam and Wignall, 1997, 1999). The remarkable sea-level drop around the G-LB recognized at Chaotian and other sections in South China no doubt corresponds to the end-Guadalupian eustatic regression.

The sea-level changes in western South China in the Capitanian to early Wuchiapingian can be summarized as follows (Fig. 7): 1) a considerable deepening during the Capitanian, 2) a rapid shallowing around the G-LB, and 3) a second moderate deepening in the early Lopingian. The unique tectonic setting of South China, isolated from Pangea and other continental blocks during the Permian, likely allowed the continental shelf domain to remain relatively low with respect to the sea-level. Accordingly, South China accumulated the exceptionally continuous late Guadalupian sediments, particularly those of a deep-water facies in northern Sichuan (Zhu et al., 1999; Wang and Jin, 2000).

6.4. Capitanian oxygen-depletion on the disphoric slope/basin

The Limestone Unit of the Maokou Formation at Chaotian was probably deposited on a shallow-marine oxic shelf with highly active of benthic organisms, on the basis of the dominant occurrence of shallow-marine fossils (e.g., calcareous algae and corals) and frequent bioturbation. In contrast, the overlying Mudstone Unit completely lacks bioturbation (Figs. 3F, 4D, 6, and 7). This remarkable shift in bioturbation mode suggests that benthic animal activities were suppressed probably by a decrease in dissolved oxygen in the bottom water. In addition, the Mudstone Unit is intercalated with many organic and laminated black calcareous mudstone layers (Figs. 3F and 4D), and yields abundant 5–10 μm-sized pyrite frambooids (this study; Isozaki et al., 2008; Lai et al., 2008). TOC values of the sedimentary rocks in the Mudstone Unit are remarkably high up to 10 wt.% (Figs. 6 and 7). The Mudstone Unit was probably deposited under oxygen-depleted conditions, as already pointed out by Isozaki et al. (2008) and Lai et al. (2008), on the basis of lithologic and geochemical characteristics that are common in black shales (e.g., Wignall, 1994). The abundant occurrence of allochthonous benthic fossils, such as brachiopods and gastropods, in the Mudstone Unit is not contradictory to these sedimentary signatures for oxygen-depleted conditions; indeed, those fossils are highly fragmented and mainly occur in turbidite beds. Tests of planktonic radiolarians were likely delivered directly from the shallower water column to the deep oxygen-depleted slope/basin.

It is noteworthy that TOC values of black calcareous mudstone in the M2 subunit are extremely high (10–16 wt.%; Fig. 7), because this suggests an intensification of reducing condition in accordance with additional deepening across the M1/M2 boundary. Other possible interpretations include the decrease of sedimentation rates by the deepening and an increase of primary productivity in the shallower euphotic zone, although we cannot test this hypothesis with the present data.

The sedimentary environment at Chaotian probably shifted from a shallow oxic shelf to a deep oxygen-depleted slope/basin during the early-middle Capitanian. At Shangsi, the emergence of oxygen-depleted conditions accompanied the early-middle Capitanian deepening (Ma et al., 2008; Fig. 1A). This suggests that oxygen-depleted conditions developed extensively at least in northern Sichuan.

In contrast, signs of oxygen-depletion are absent in the overlying Wujiaping Formation at Chaotian; instead oxic conditions predominate with abundant shallow-marine fossils (e.g., calcareous algae and fusulinids) and minor bioturbation (Figs. 6 and 7). The sedimentary environment returned to the oxic shelf across the G-LB, likely in response to the sharp sea-level drop. Again, a similar lithofacies change occurred elsewhere in Sichuan (e.g., Li et al., 1989), implying that a common sedimentary/redox history was shared throughout western South China.

As to the cause of the appearance of oxygen-depletion on the Capitanian disphoric slope/basin at Chaotian, several explanations are possible. For example, it might reflect a local stagnation of seawater owing to restricted basin geometry. Otherwise, primary productivity in the surface ocean might have increased to lower the dissolved oxygen level in seawater, and led a vertical expansion of the oxygen minimum zone, in a similar scenario to a recently proposed one for the P-TB (Algeo et al., 2010, 2011). From a different viewpoint, the detected oxygen-depletion may correspond to the onset of the long-term oxygen-omaly in the superocean of the Paleozoic–Mesozoic transition (superanoxia; Isozaki, 1997, 2009a). At any rate, the development of the oxygen-depletion in the disphoric zone at Chaotian adds another unique feature to the list of uncommon geological phenomena in Capitanian time (Isozaki, 2007, 2009a). In particular, it is significant that the oxygen-depleted condition appeared clearly before the end-Guadalupian extinction event. It is also noteworthy that the Capitanian oxygen-depleted condition detected in this study occurred significantly before the P-TB shallow-sea anoxia that was reported from various areas around the world (e.g., Hallam, 1991; Wignall and Hallam, 1992). In other words, it may record the first development of oxygen-depleted seawater on a continental margin in the Paleozoic–Mesozoic transition interval, and this needs to be checked elsewhere. In order to clarify the global environmental changes in a disphoric zone with respect to the end-Guadalupian extinction, further research is also needed on disphoric slope/basin sediments from other regions, using detailed litho- and bio-facies analyses as performed in the present study.
7. Conclusion

This study provides detailed descriptions of lithofacies of the upper Guadalupian to the lowermost Wuchiapingian rocks at Chaotian in South China, and discusses secular changes in sea-level and redox across the Guadalupian–Lopingian boundary (G–LB). The following new results were obtained:

1. In the early–middle Capitanian (Late Guadalupian), sea-level rose significantly in two steps to change the sedimentary environment from a euphotic shelf to a disphotic slope/basin. In turn, a large sea-level drop followed around the G–LB and the sedimentary environment returned to a shallow euphotic setting.

2. The Capitanian deepening was likely due to a tectonics-related local subsidence of the sedimentary basin, whereas the G–LB shallowing might reflect a global regression at the end of the Guadalupian.

3. In association with the abrupt sea-level rise in the Capitanian, an oxygen-depleted condition appeared in the disphotic slope/basin in northern Sichuan, clearly before the end-Guadalupian extinction; it might record the first development of oxygen-depleted seawater on a continental margin in the Paleozoic–Mesoozoic transition interval, significantly before the well-known P–TB shallow-sea anoxia.

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