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Meng, Q, Xue, W, Chen, F et al. (7 more authors) (2022) Stratigraphy of the Guadalupian (Permian) siliceous deposits from central Guizhou of South China: Regional correlations with implications for carbonate productivity during the Middle Permian biocrisis. Earth-Science Reviews, 228. 104011. ISSN 0012-8252

https://doi.org/10.1016/j.earscirev.2022.104011

Stratigraphy of the Guadalupian (Permian) siliceous deposits from central Guizhou of South China: Regional correlations with implications for carbonate productivity during the Middle Permian biocrisis

Qi Meng [a], Wuqiang Xue [a], Fayao Chen [a], Jiaxin Yan [a,*] jaxy2008@163.com, Jiahua Cai [b], Yadong Sun [a, c], Paul B. Wignall [d], Ke Liu [e], Zhichen Liu [f], Deng Chen [f]

[a] *School of Earth Sciences, State Key Laboratory of Biogeology and Environmental Geology, China University of Geosciences, Wuhan 430074, China*

[b] *Institute for Environmental and Climate Research, Jinan University, Guangzhou 510632, China*

[c] *Humboldt Post-doc Fellow, University of Erlangen-Nuremberg, Schlossgarten 5, 91054 Erlangen, Germany*

[d] *School of Earth and Environment, University of Leeds, Leeds LS2 9JT, UK*

[e] *Wuhan Surveying-Geotechnical Research Institute Co., LTD. of MCC, Wuhan 430080, China*

[f] *102 Geological Party, Guizhou Bureau of Geology and Mineral Exploration and Development, Zunyi 563099, China*

[*] Corresponding author

Abstract

In comparison with the amount of study undertaken on the end-Permian mass extinction, the preceding Guadalupian mass extinction has received little investigation, even though it marks a significant biotic turnover associated with global environmental changes. During the earlier event, reef carbonate production shut down and was replaced by siliceous, mud-rich deposits (SRDs) in South China. However, changes in carbonate platform productivity during this epoch remain to be clarified. This paper presents sedimentological and conodont biostratigraphic investigations on the Guadalupian SRDs developed on the Yangtze Carbonate Platform (YCP) in central Guizhou. The findings are viewed in the context of Guadalupian sequence correlation of South China successions, which shows that the integrity of the YCP failed to match the platform tectonic evolution. The platform evolution saw the onset of major intra-platform depressions and the gradual onlap by SRDs along the platform margin. Stratigraphic correlation reveals that the platform experienced three phases of onlap by SRDs during the early Roadian, the late Wordian and the late Capitanian upwards. Platform carbonates re-expanded their extent following the first two phases, but not during the final phase. An evolutionary model is proposed for the Guadalupian carbonate platform, which follows the contemporaneous eustatic sea-level fluctuations. The partial drowning observed within the platform interior and increasing retreat along the platform margin could suggest an insufficient carbonate sediment supply shedding from the platform top during the Guadalupian. The variation in carbonate productivity raises our attention to the change in shallow-water carbonate factories, which is closely related to the fortunes of carbonate-secreting biota and environmental factors impacting the carbonate platform producers during the Guadalupian.

Keywords: Guadalupian, Platform shrink, Biocrisis, Carbonate factories, Conodont stratigraphy

1. Introduction

The Middle to Late Permian is a critical interval that saw major disturbances to environments and biota, exemplified by two mass extinctions, during the Middle Permian and at the end-Permian. Although the losses of the Middle Permian mass extinction used to be included in those of the end-Permian mass extinction, the former are now regarded as a distinct crisis (Jin et al., 1994; Stanley and Yang, 1994; Shi et al., 1999; Clapham et al., 2009; McGhee et al., 2013). The Middle Permian mass extinction is variously known as the Guadalupian extinction, end-Guadalupian extinction, Capitanian extinction, mid-Capitanian extinction, and the pre-Lopingian extinction etc. (Jin et al., 1994; Stanley and Yang, 1994; Shen and Shi, 2002; Bond et al., 2010a; Zhang et al., 2019b). Here we adopt the term "Guadalupian mass extinction".

The extinction event affected many groups including brachiopods, rugose corals, fusulinid foraminifers and alatoconchid bivalves (Jin et al., 1994; Isozaki and Aljinović, 2009; Wignall et al., 2009a). Records from the Global Stratotype Section and Point (GSSP) sections (Penglaitan, Tieqiao) of the Guadalupian – Lopingian Boundary (GLB) showed the losses spanned the late Capitanian to the earliest Wuchiapingian interval (Jin et al., 2006; Yin et al., 2007; Shen and Shi, 2009; Huang et al., 2018). In SW China (Guizhou, Yunnan, Sichuan provinces), the extinction has been regarded as the "mid-Capitanian mass extinction", corresponding to a major negative excursion of the $\delta^{13}C_{carb}$ value within the *Jinogondolella altudaensis* – *J. prexuanhanensis* conodont Zone, which severely affected photosynthetic and photosymbiont-bearing taxa: green algae and keriothecal-walled fusulinids (Wignall et al., 2009a; Bond et al., 2010b). A similar pattern is recorded in the coeval western Tethys, Panthalassa and the Boreal Realm (Ota and Isozaki, 2006; Isozaki and Aljinović, 2009; Bond et al., 2010a; Wignall et al., 2012; Bond et al., 2015; Zhang et al., 2019b).

The timing of the event in South China is debated. For example, most large and morphologically complex species in the fusulinid families Schwagerinidae and Neoschwagerinidae disappeared before the mid-Capitanian, and were replaced by an assemblage of smaller fusulines in the latest Guadalupian which persisted into the Wuchiapingian (Fan et al., 2020; Shen et al., 2020). Assessing global diversity databases suggests the supposed extinction event was a more protracted diversity decrease that spanned the entire early-middle Guadalupian (Clapham et al., 2009; Groves and Wang, 2013). In contrast, major environmental changes in the Capitanian Stage, including the eruption of the Emeishan Large Igneous Province (ELIP) in SW China (Zhou et al., 2002; Wignall et al., 2009a), shallow-marine habitat loss caused by great regression (Hallam and Wignall, 1999; Chen et al., 2009; Arefifard, 2017; Wei et al., 2017), marine anoxia (Isozaki, 1997; Zhang et al., 2015), potentially oceanic acidification (Weidlich, 2002a; Clapham and Payne, 2011) and global climate cooling (Isozaki et al., 2007), may have contributed to a short duration mass extinction. It is clearly desirable to examine the geological evidence for changes throughout the Guadalupian to decipher the pattern and mechanism of the Guadalupian mass extinction.

Carbonate platforms provide valuable geological archives that result from the interplay of carbonate production and accumulation, and are intrinsically related to changes in the environmental and biotic conditions (*e.g.*, Kiessling et al., 2003; Godet, 2013; Phelps et al., 2015). Carbonate platforms were widely distributed in tropical-temperate zone during the earliest Guadalupian (Fig. 1A) (Kiessling et al., 2003; Pomar and Kendall, 2008; Moore and Wade, 2013). However, during the late Guadalupian, the Permian Chert Event saw chert accumulation greatly expand at the expense of carbonate deposition (Murchey and Jones, 1992; Beauchamp and Baud, 2002). Thus, platform carbonates are overlain by cherts in the Guadalupian Series of western North America (Murchey and Jones, 1992; Beauchamp and Baud, 2002; Matheson and Frank, 2020a), Far East Russia (Ueno et al., 2005) and Transcaucasia (Leven, 1998). The main contributors to Permian carbonate platform formation were foraminifers, calcareous algae, along with microbes and reefbuilders (Kiessling et al., 2003). Studies of shallow-water carbonate platforms show a long-term decline in their area from the late Carboniferous to the Middle Triassic (Kiessling et al., 2000; Kiessling et al., 2003), although their coarse stratigraphic resolution prevents analysis of the Guadalupian biocrisis. On the other hand, reef compilations at stage resolution have revealed a gap in reef evolution across the Guadalupian to Lopingian interval (Weidlich, 2002b). The decline in global reef carbonate production (assessed according to their numbers, dimensions and debris potential) has been calculated as nearly 90% (Flügel and Kiessling, 2002). The succeeding early Lopingian was a period of reef recovery (Huang et al., 2017; Wang et al., 2019). Long-term marine acidification (Beauchamp and Grasby, 2012) or climate cooling (Kani et al., 2018) were suggested as the cause of the decline of carbonate production. However, the process of the reduction in carbonate extent, as well as its implications for the coeval biotic and environmental evolution remains unclear.

The Yangtze Carbonate Platform (YCP) in South China is one of the tropical platforms that developed during the Guadalupian. It occurred in a relatively simple and well defined tectonic context, although the ELIP erupted in the region towards the end of the Guadalupian. The YCP is well preserved and includes surrounding marginal and basinal facies. In contrast with marine sections elsewhere, where Guadalupian sequences record a major hiatus (*e.g.*, Wignall et al., 2012), South China preserves a relatively complete Guadalupian succession that includes the GSSP of the GLB (Jin et al., 2006). Extensive investigations of fossil records in the Guadalupian strata have provided a high resolution biostratigraphic framework allowing cross-platform correlation.

The main objective of this paper is to evaluate the Guadalupian evolution of the YCP in the context of contemporaneous biotic and environmental evolution. We begin with sedimentological description and conodont stratigraphy of the Guadalupian siliceous deposits from the central part of Guizhou Province, that developed in intra-platform depressions on the YCP. We then construct correlation transects from platforms to basins, to investigate the fluctuations of the carbonate platform extent during the Guadalupian. Finally, a model for the evolution of the YCP is proposed using a sequence stratigraphic framework.

2. Geological setting in South China

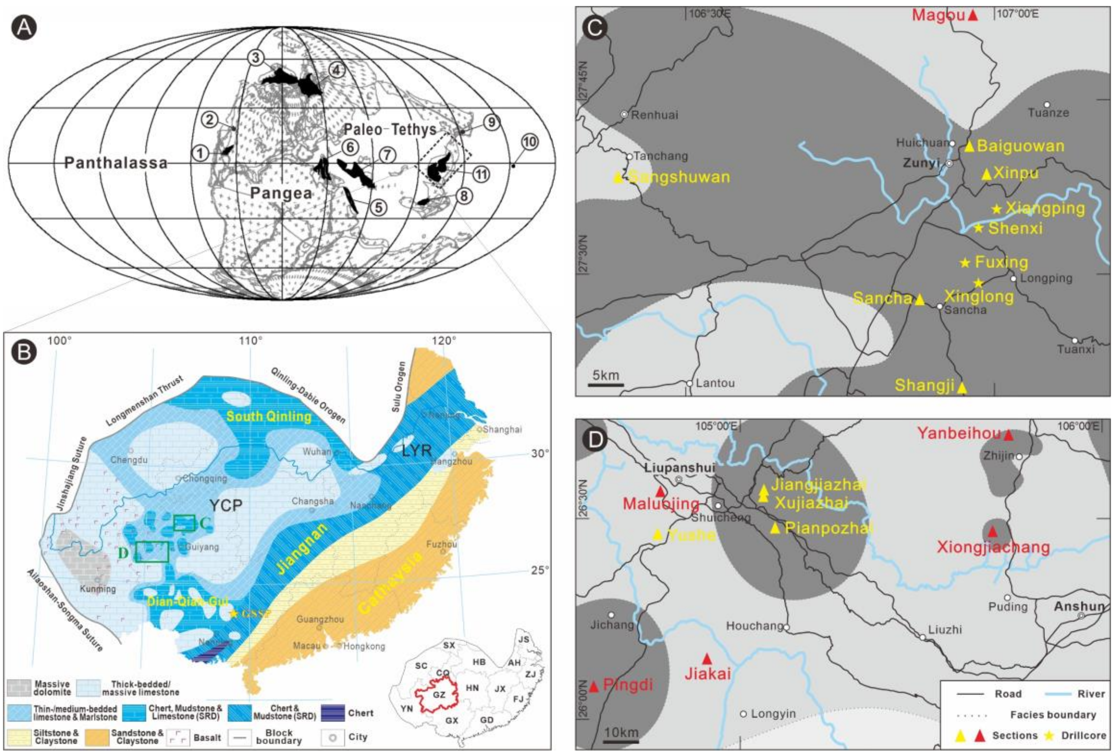

Fig. 1. (A) Middle Permian palaeogeographic map of the world (modified after Kiessling et al., 2003). Locations of Guadalupian carbonate platforms are shown in black. 1 - Delaware Basin; 2 - Phosphoria Sea; 3 - Sverdrup Basin; 4 - Barents Sea; 5 - Arabian platform; 6 - Transcaucasia; 7 - Cimmeria; 8 - Pha Nok Khao platform; 9 - Far East Russia; 10 - Akasaka paleo-atoll; 11 - YCP (this study). (B) Palaeogeographic map of South China during the Middle-Late Guadalupian (modified after Liu and Xu, 1994; Ma et al., 2009; Chen et al., 2018). The two green squares show

the studied Zunyi area (C) and Liupanshui – Anshun area (D) from central Guizhou. Provinces: AH - Anhui; CQ - Chongqing; FJ - Fujian; GD - Guangdong; GX - Guangxi; GZ - Guizhou; HB - Hubei; HN - Hunan; JS - Jiangsu; JX - Jiangxi; SC - Sichuan; SX - Shaanxi; YN - Yunnan; ZJ - Zhejiang. Note that the South China block has rotated ~80° clockwise since the Middle Permian. (C, D) Locality maps of study sections and drillcores. The Magou, Maluojing, Jiakai, Yanbeihou sections are the research sites of Regional Geological Survey Report (1971). The Pingdi and Xiongjiachang sections are the research sites of Sun et al. (2010).

The South China craton is bounded by the Ailaoshan – Songma suture zone, the Jinshajiang suture, the Longmenshan Thrust and the Qinling – Dabie – Sulu orogen, from the southwest and northwest to the north respectively (Fig. 1B) (Wan, 2012; Metcalfe, 2013; Cawood et al., 2018). During the Permian, the plate was situated in the equatorial eastern paleo-Tethys with the Panthalassa Ocean to its east (Fig. 1A) (Kiessling et al., 2003; Scotese, 2014). It comprised the northwest Yangtze Block and the southeast Cathaysia Block, amalgamated along the Jiangnan suture during the Early Paleozoic (Fig. 1B) (Wang et al., 2010; Cawood et al., 2018).

In the Yangtze Block, the Guadalupian succession is dominated by marine carbonate rocks. The main domain of carbonate sedimentation is termed the YCP. It was surrounded by the South Qinling carbonate basin to the north, the Jiangnan chert basin to the east and the Dian – Qian – Gui carbonate basin to the south (Fig. 1B) (Liu and Xu, 1994; Ma et al., 2009). Within these basins, small, isolated carbonate platforms (ICPs) developed with dimensions of tens (occasionally more than one hundred) of kilometres. In the Cathaysia Block, the Guadalupian succession is mainly composed of clastic rocks (Fig. 1B). The clastic material, shedding from the Cathaysia oldland, accumulated in the Jiangnan Basin until the end Guadalupian (Fig. 1B) (Liu and Xu, 1994; Feng et al., 1996).

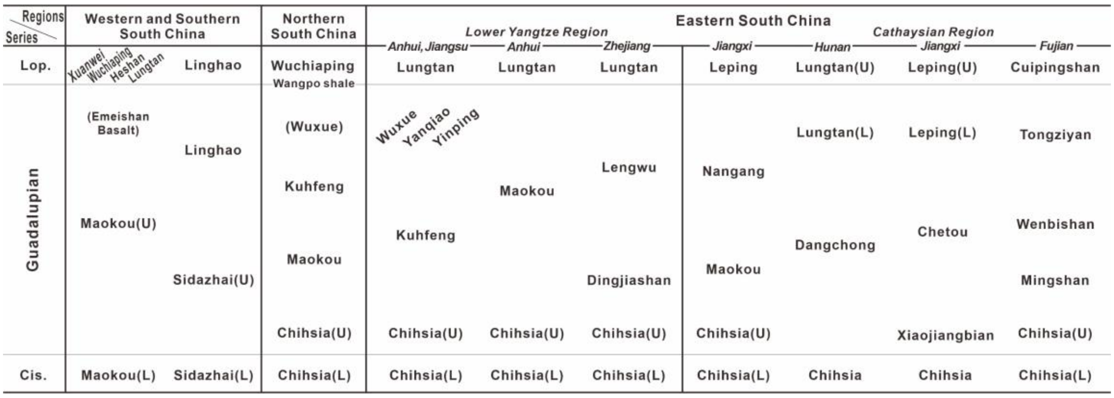

Fig. 2. Summary of lithostratigraphic units of the Guadalupian Series from different regions of South China. Units are arranged in chronological order, but not to actual scale. “Lungtan”, “Chihsia”, “Dingjiashan” and “Tongziyan” are also spelled as “Longtan”, “Qixia”, “Tingchiashan” and “Tungtseyan” respectively. Abbreviations: Cis. - Cisuralian; Lop. - Lopingian; L - Lower part of the unit; U - Upper part of the unit.

The Permian System is divided into three series, the Cisuralian, Guadalupian and Lopingian in ascending order. In South China, the Guadalupian Series is represented by the Maokou Formation and its equivalents in the YCP and ICPs (Fig. 2) (Shen et al., 2018). In the western YCP, the Maokou Formation commonly consists of dolomite and is interbedded with Emeishan basalts in the upper part (Figs. 1B, 2). In the middle-eastern part of the YCP, the Maokou Formation is dominated by bioclastic limestone (Fig. 1B). Outwards from the platform, limestone grades into siliceous limestone and marlstone of the Maokou Formation, or into siliceous, mud-rich deposits (SRDs) of the Kuhfeng Formation in northern South China, and the Sidazhai and Linghao formations in southern South China (the deep-water areas around the ICPs) (Figs. 1B, 2). The SRDs are dominated by thin-bedded chert, sometimes with interbeds of (siliceous) limestone, and/or mudstone, and/or shale.

In the Lower Yangtze region (LYR) and the Cathaysian region of eastern South China, the Guadalupian lithostratigraphic units are commonly named provincially (Fig. 2). In terms of lithology, these units comprise carbonate strata of the Maokou, Wuxue, Lengwu, Chihsia (also as "Qixia", upper part) and Xiaojiangbian formations, and SRDs of the Kuhfeng, Nangang, Dangchong, Chetou and Mingshan formations, and clastic strata of the Yanqiao, Yinping, Dingjiashan (also as "Tingchiashan"), Wenbishan, Tongziyan (also as "Tungtseyan"), Lungtan (also as "Longtan", lower part) and Leping (lower part) formations (Fig. 2). It should be noted that the carbonate Chihsia Formation, and the clastic Lungtan and Leping formations are diachronous, and straddle the Cisuralian – Guadalupian and Guadalupian – Lopingian boundaries respectively (see section 5.3).

Beneath the Guadalupian succession in South China, the upper Cisuralian Series is composed mainly of the carbonate Chihsia Formation, as well as parts of the Sidazhai and Maokou formations (Fig. 2). The Chihsian limestone is characterised by great thickness and persistence of lithofacies throughout South China. This contrasts with the variation among lithofacies (platform vs. basin) seen in the Maokou Formation during the Guadalupian.

The lower part of the Lopingian Series, the Xuanwei, Wuchiaping, Heshan, Lungtan, Leping and Cuipingshan formations, overlie the Guadalupian Series in South China (Fig. 2) (Hou et al., 2020). A shale unit containing coal and volcaniclastic materials, termed the Wangpo Shale, is developed at the base of the Wuchiaping Formation (Fig. 2). An unconformity is widely distributed on the top of the Guadalupian succession of the YCP and interpreted to record uplift related to the

ELIP and/or a Guadalupian regression (He et al., 2003; Sun et al., 2010). However, the time range and duration of this hiatus remain to be determined in most parts of the platform. Carbonate deposition still occurred in South China during the Wuchiapingian, but exhibited considerable differences to Guadalupian time (Liu and Yan, 2009; Yan et al., 2019).

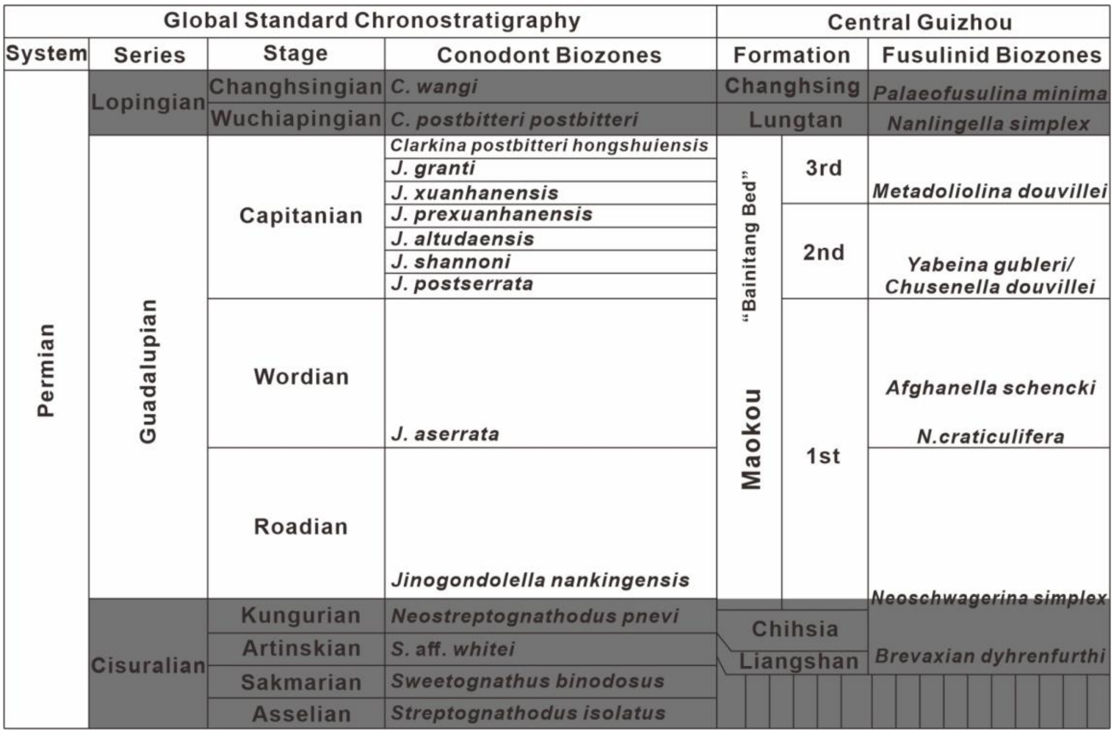

Fig. 3. Subdivision of the Permian sequence of the central part of Guizhou Province and its correlation with the global standard Permian System (combined with works in this research and in Chen et al., 1984; Bureau of Geology and Mineral Resources of Guizhou Province, 1987; Jin et al., 1997; Henderson, 2016; Shen et al., 2018). Conodont biozones and fusulinid biozones of the Cisuralian Series and Lopingian Series are only shown by their First Appearance Datum of species.

The global standard for the Guadalupian biostratigraphy was traditionally based on fusulinid biozones, but more recently a much higher resolution conodont zonation has been devised. Three stages of the Guadalupian Series, namely Roadian, Wordian and Capitanian, are defined by the nine species within the genus *Jinogondolella* lineage and *Clarkina* lineage (Fig. 3) (Henderson, 2016). In South China, the marine Guadalupian succession is highly fossiliferous, temporally well-constrained by its fossil content (*e.g.*, fusulinid, brachiopod, coral and ammonoid). Stage and series assignment in sections and correlations across the entire region have been mostly achieved by fusulinids and macrofossils (Sheng and Jin, 1994; Shen et al., 2018). However, many sections have conodont biostratigraphic constraints and provide useful reference sections (*e.g.*, Wang, 1995a; Henderson and Mei, 2003; Zhang et al., 2007; Cheng et al., 2017; Sun et al., 2017).

3. Materials and methods

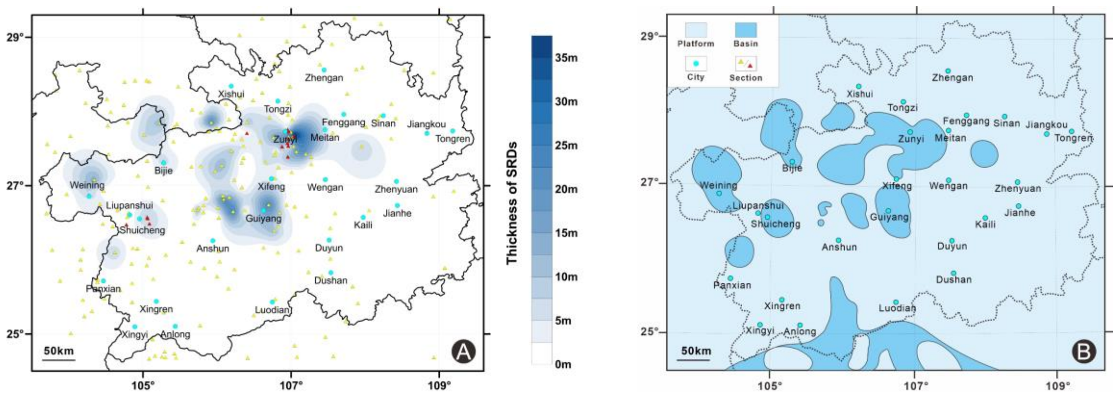

Fig. 4. Middle Permian palaeogeographic map of Guizhou Province. (A) Locations of the collected sections containing Maokou Formation and thickness contour lines of the thin-bedded SRDs. (B) Interpretation of Middle Permian lithofacies in Guizhou. The basin in the southern part of the map is after Liu and Xu (1994) and Ma et al. (2009).

To depict the spatial distribution of the Guadalupian SRDs, a Guadalupian lithofacies map of

Guizhou was compiled (Fig. 4). Stratal sections for the map are collected from the related geological survey sheets covering the region, combined with those undertaken in this study and others observed by the senior authors. Total thickness of the thin-bedded Guadalupian SRDs from the collected sections (data in Appendix A) were plotted on the palaeogeographic map, then contoured with the gridding method by the Surfer 13 software, to constrain the boundaries between the SRDs and the adjacent carbonate deposits (Fig. 4A). The lithofacies map was based on the contour line analysis, along with facies interpretations and biostratigraphic constraints detailed below.

In the Zunyi and Liupanshui – Anshun areas, the Maokou Formation and the lower part of the Lungtan Formation were seen in nine outcrops (Sangshuwan, Baiguowan, Xinpu, Sancha, Shangji, Yushe, Jiangjiazhai, Xujiazhai, Pianpozhai) and four underground drillcores (Xiangping, Shenxi, Fuxing, Xinglong), and were logged and sampled for microfacies analysis and conodont biostratigraphy (Fig. 1C, D).

A total of 272 samples were thin-sectioned for microfacies analysis. Additionally, 38 samples were collected from the limestone beds below, above or within the SRDs for conodont biostratigraphic study in the Pianpozhai, Sangshuwan, Baiguowan, Sancha and Shangji sections. Conodont samples, weighing >5 kg each, were collected from every bed. They were broken into fragments, dissolved using 10% diluted acetic acid, wet sieved through 20 # and 160 # meshes and air-dried. The heavy fraction of insoluble residues was separated using heavy liquid (solution of lithiumheteropolytungstates in water). Conodont specimens were handpicked under binocular microscopes. A total of 405 P1 elements were obtained and seven species were recognised in the genus *Jinogondolella* (Figs. 5, 6).

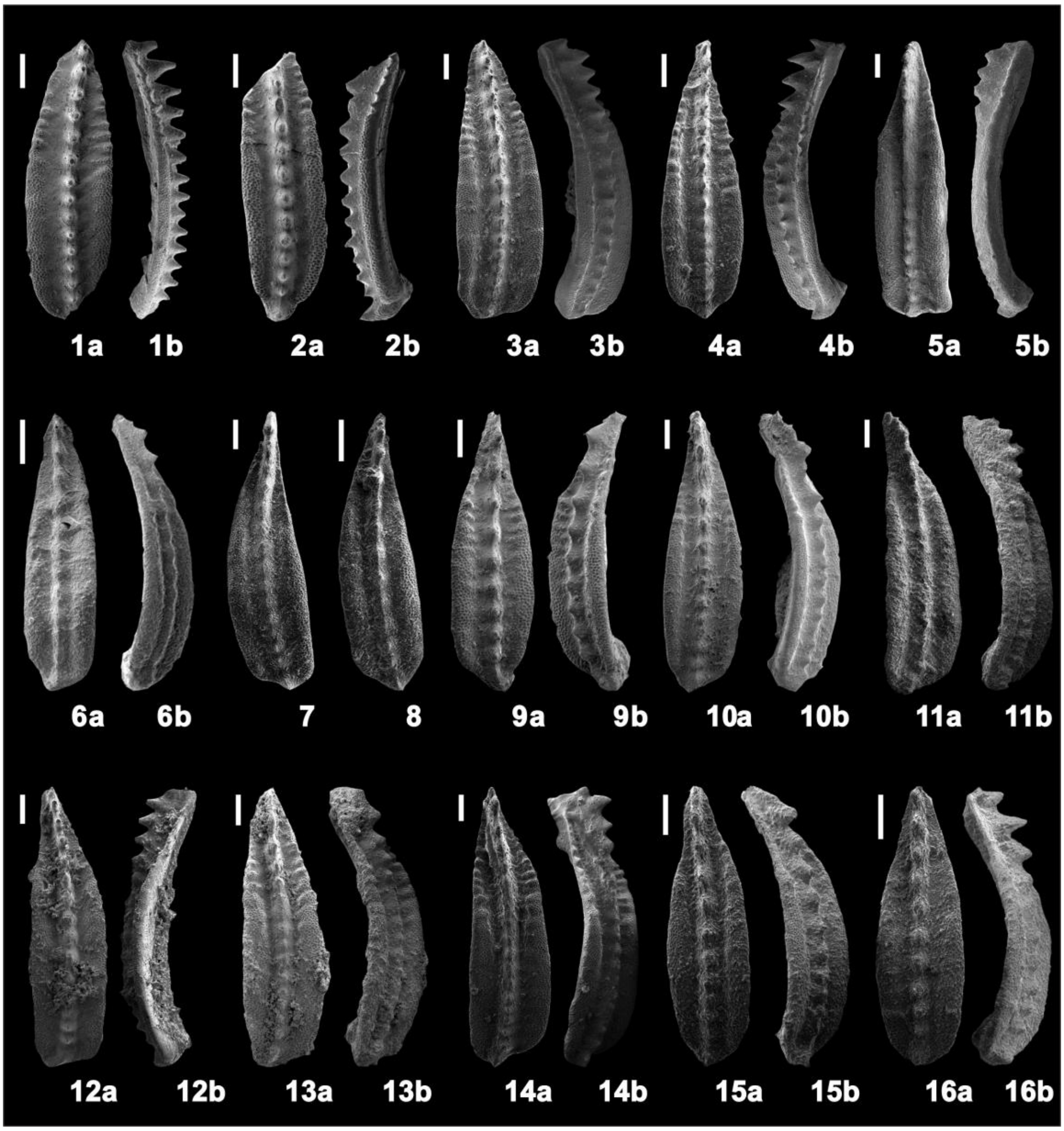

Fig. 5. P1 elements obtained from central Guizhou, scale bar for 100μm, “a” for upper view, “b” for lateral view. 1-2 from Pianpozhai, 3-4 from Shangji, 5-10 from Sancha, 11-16 from Baiguowan. 1, 2, 5: *J. nankingensis* (Wardlaw, 1998), 1. PPZ1, i001001; 2. PPZ1, i001003; 5. SC2, i003028. 3, 9-14: *J. postserrata* (Behnken, 1975), 3. SJ1, i002002; 9. SC3, i003014; 10. SC3, i003015; 11. BGW1, i004134; 12. BGW3, i004106; 13. BGW6, i004056; 14. BGW7, i004014. 4, 15-16: *J. shannoni* (Wardlaw, 1998), 4. SJ1, i002014; 15. BGW1, i004139; 16. BGW2, i004004. 6-8: *J. aserrata* (Clark and Behnken, 1979), 6. SC1, i003024; 7. SC2, i003030; 8. SC2, i003019.

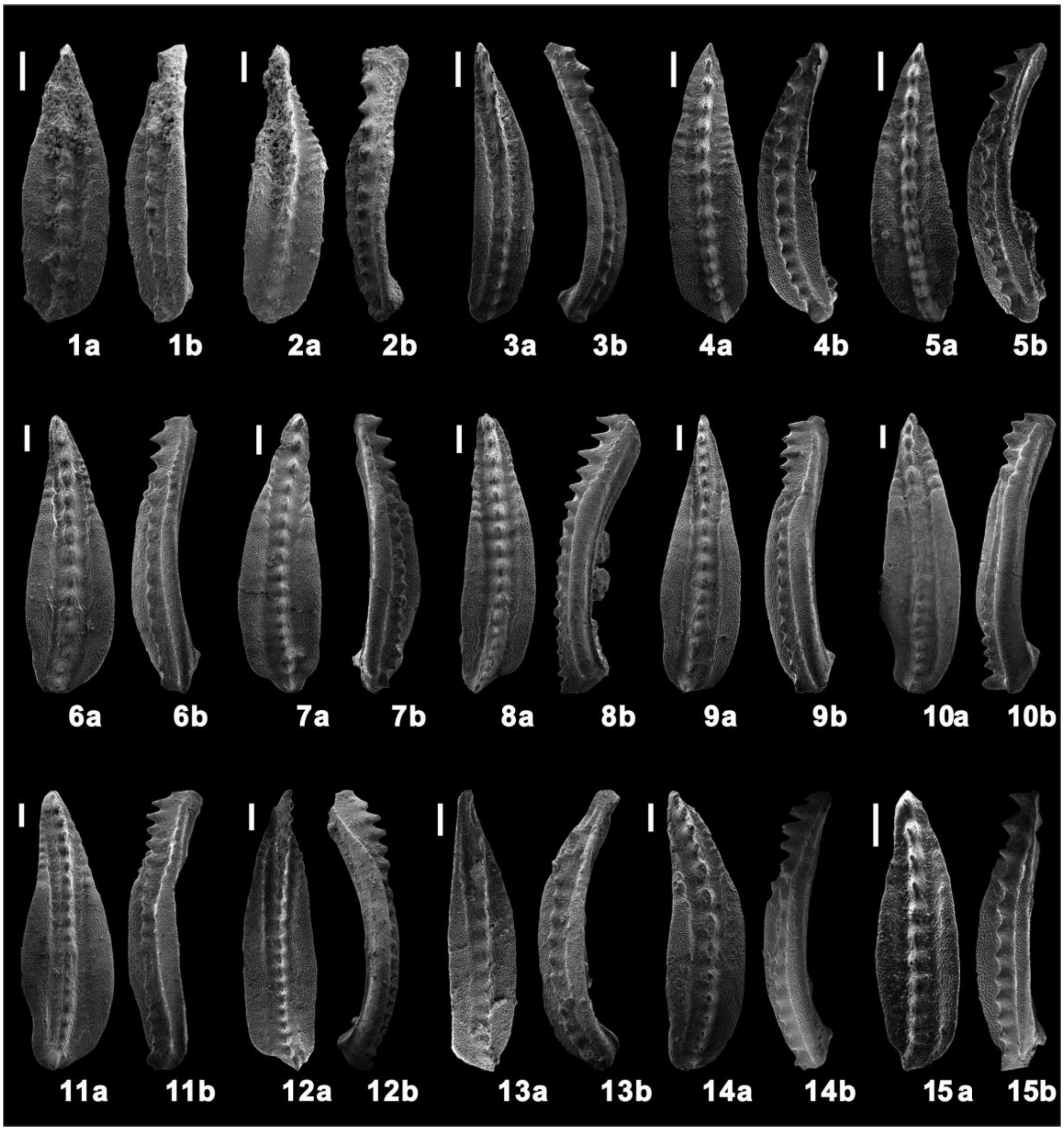

Fig. 6. P1 elements obtained from central Guizhou, scale bar for 100μm, "a" for upper view, "b" for lateral view. 1-11 from Baiguowan, 12-15 from Sangshuwan. 1-3, 13-15: *J. shannoni* (Wardlaw, 1998), 1. BGW3, i004102; 2. BGW6, i004052; 3. BGW7, i004037; 13. SSW1, i006036; 14. SSW2, i006029; 15. SSW3, i006012. 4-7: *J. altudaensis* (Kozur, 1992), 4. BGW8, i005009; 5. BGW9, i005017; 6. BGW10, i005052; 7. BGW11, i005027. 8, 9: *J. prexuanhanensis* (Mei and Wardlaw, 1994), 8. BGW10, i005056; 9. BGW11, i005032. 10, 11: *J. xuanhanensis* (Mei and Wardlaw, 1994), 10. BGW10, i005062; 11. BGW11, i005039. 12: *J. postserrata* (Behnken, 1975), 12. SSW1, i006035.

4. Guadalupian stratigraphy and facies changes in central Guizhou

Combined faunal and lithological features allowed the carbonate strata of the Maokou Formation in Guizhou Province to have been traditionally divided into three members (Fig. 3) (Bureau of Geology and Mineral Resources of Guizhou Province, 1987). However, macrofossils and fusulinid biozone schemes are difficult to apply in central Guizhou, *e.g.*, in the vicinity of Zunyi City (Fig. 1C), where the upper part of the Guadalupian succession above the thick-bedded carbonate rocks of the first Maokouan member consists of thin- to medium-bedded limestone with chert nodules, marlstone, SRDs and manganese deposits. The SRDs in the Zunyi area were assigned to the second Maokouan member by lithology and locally termed the "Bainitang Bed" (Fig. 3) (He and He, 1953). The third Maokouan member was ambiguously thought of either being absent or a manganese (carbonate) deposit (Bureau of Geology and Mineral Resources of Guizhou Province, 1987; Liu et al., 1989). It is still to be determined how many strata from the top of the Maokou Formation have been eroded in this area. A similar situation to the Zunyi area also occurs in the area extending to Liupanshui – Anshun (Fig. 1D). These areas have unique late Guadalupian successions where the "intra-platform trough of central Guizhou" developed (Chen et al., 1984). To understand the stratigraphy and deposition of those Guadalupian successions of the "trough", the Zunyi and Liupanshui – Anshun areas have been studied in detail.

4.1 *Zunyi area*

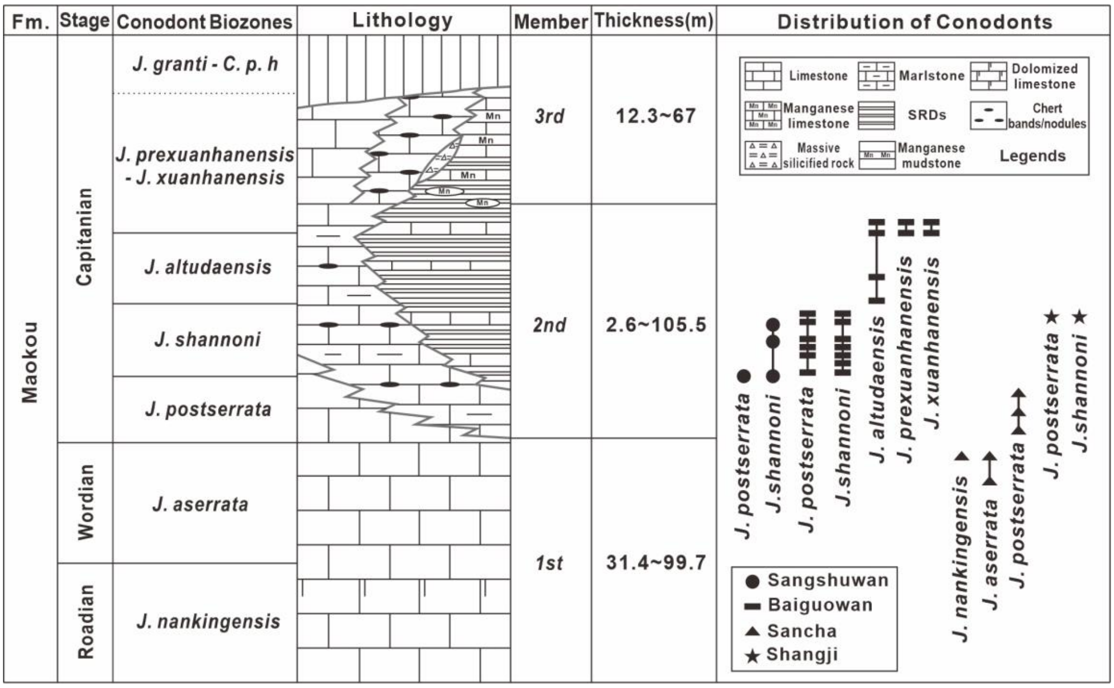

Fig. 7. Composite column of the Maokou Formation in the Zunyi area showing the lithology, thickness and conodont ranges.

The first member of the Maokou Formation is characterised by thick-bedded to massive wacke-packstone of pale grey colour, which is partly dolomitized in the lower part (Figs. 7, 8F, G). It comprises over 40% skeletal grains dominated by foraminifers (including fusulinids), calcareous algae and alatoconchid bivalves (Fig. 8A, B, J-L). Skeletal grains of brachiopods, sponges, gastropods and corals are also common (Fig. 8C-E). At Sancha, *J. nankingensis* (Fig. 5.5) and *J. aserrata* (Fig. 5.6-5.8) specimens were obtained from two samples from the top part of this member, suggesting a Wordian age (Figs. 1C, 7).

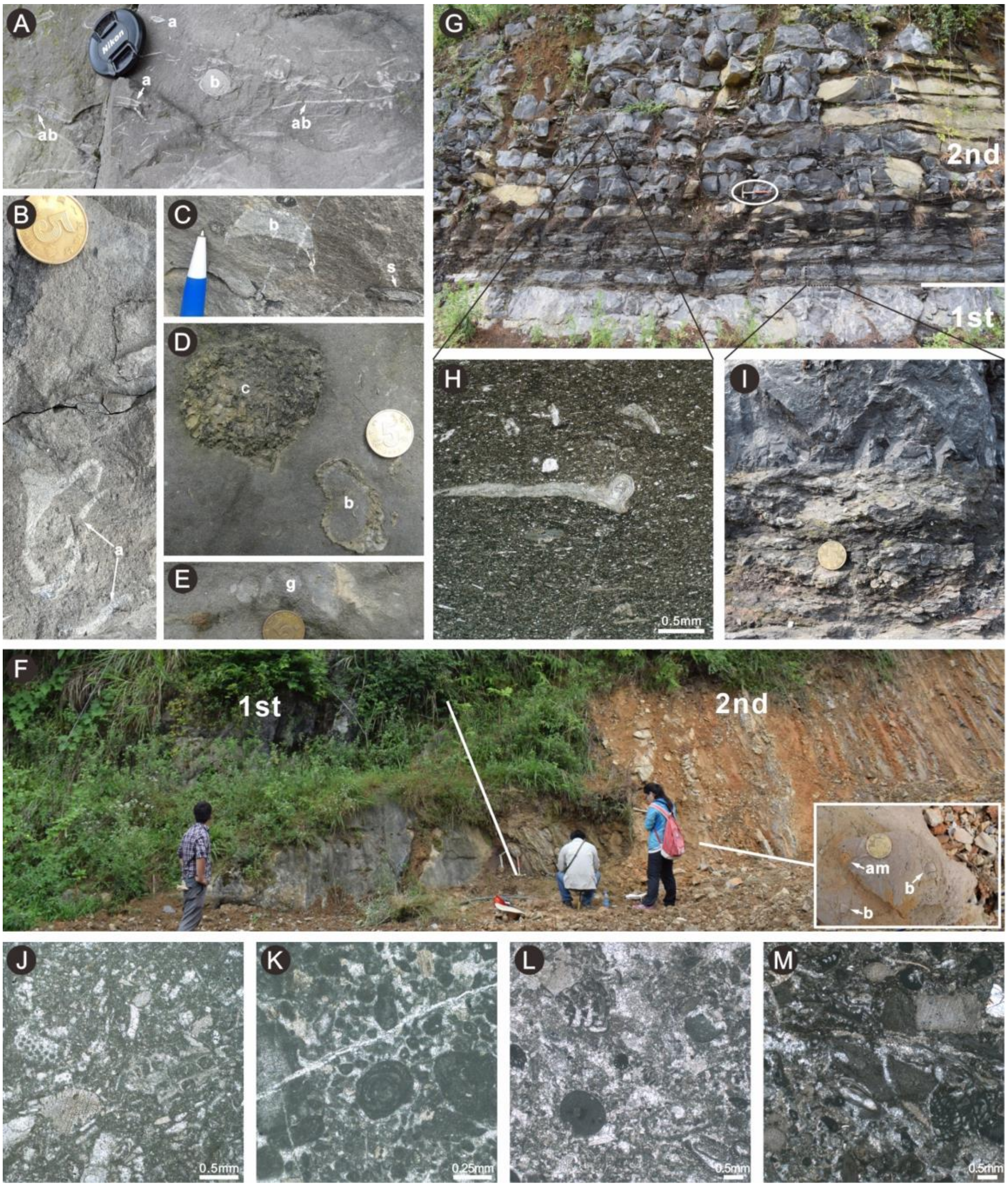

Fig. 8. Sedimentary features of the Maokou Formation in central Guizhou. (A-C) Packstone of the first member at Sangshuwan. (D, E) Packstone of the first member at Sancha. (F) Outcrop photos of the Sancha section. The close-up shows brachiopods and ammonoids found in the SRDs. (G) Outcrop photo of the Baiguowan section. (H, I) Photomicrograph and close-up of wackestone containing brachiopods at Baiguowan. (J) Packstone of the first member containing green algae, foraminifers, crinoids and bryozoans at Baiguowan. (K) Packstone of the first member containing

peloids, foraminifers and crinoids at Sancha. (L) Packstone of the first member containing green algae, foraminifers, fusulinids and crinoids at Shangji. (M) Packstone of the first member containing fusulinids, foraminifers, calcareous algae, crinoids and brachiopods at Pianpozhai. The coin is 2 cm in diameter. Abbreviations: a - calcareous algae; ab - alatoconchid bivalves; am - ammonoids; b - brachiopods; c - corals; g - gastropods; s - sponges.

In the second member, thin-bedded marlstone and SRDs are common (Fig. 7). In the outlying area, the second member is represented by medium-bedded limestone with intercalations of marl and chert bands (Figs. 7, 8G, 9A, C). The medium-bedded wacke-packstone contains 15% – 70% of skeletal grains dominated by brachiopods and also includes foraminifers, crinoids and ostracods (Fig. 8H, I). Eastwards from Sangshuwan and southwards from Magou, marl and chert bands gradually decrease and SRDs become dominant (Figs. 1C, 4). Because of the exploration of carbonate manganese deposits in the Zunyi area, lots of drillcores are available through the Guadalupian succession. The drillcores and outcrops show that the SRDs are dark grey to black in colour and are occasionally interbedded with thin-bedded wackestone or siliceous limestone (Fig. 9D, E). The thickness of the SRDs reaches the maximum at Xiangping (Figs. 1C, 4A). The SRDs have only 5% – 10% of fine skeletal grains including sponge spicules, radiolarians, ostracods, ammonoids and brachiopods (Figs. 8F, 9G-J).

In the lower part of the second member, *J. postserrata* is present at Sancha (Fig. 5.9, 5.10), and assemblages consisting of *J. postserrata* (Figs. 5.11-5.14, 6.12) and *J. shannoni* (Figs. 5.15, 5.16, 6.1-6.3, 6.13-6.15) are recovered at Baiguowan and Sangshuwan. In limestone interbeds of the SRDs, an assemblage with *J. postserrata* (Fig. 5.3) and *J. shannoni* (Fig. 5.4) occurs at Shangji, and

J. altudaensis (Fig. 6.4, 6.5) at Baiguowan. These conodont fossils indicate that the second member ranges from the early to middle Capitanian age (Fig. 7).

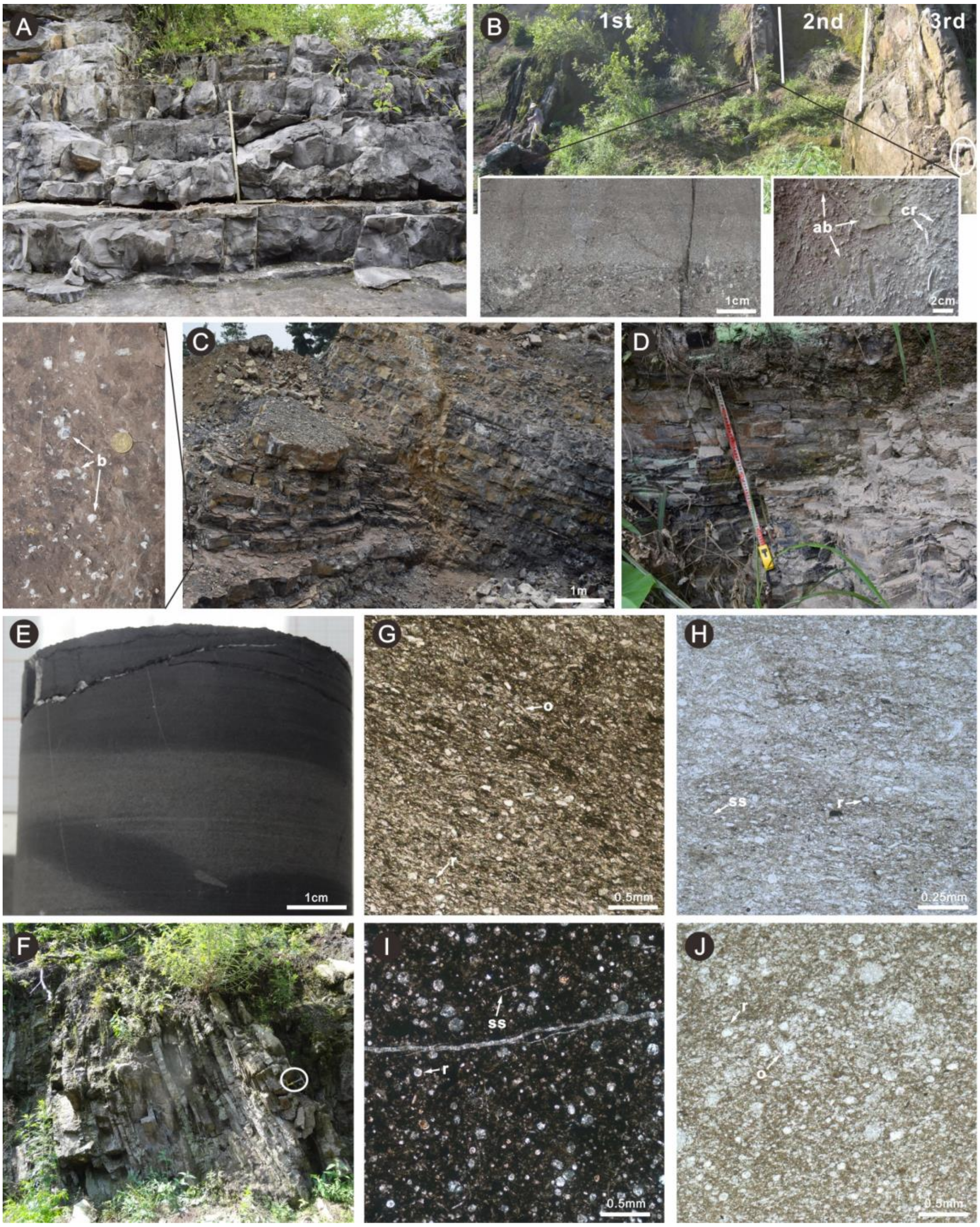

Fig. 9. Sedimentary features of the Maokou Formation in central Guizhou. (A) Wackestone interbedded with thin-bedded limestone and chert bands of the second member at Sangshuwan.

(B) Outcrop photos of the Pianpozhai section. The close-ups show packstone of the uppermost part of the first member containing alatoconchid bivalves and crinoids. (C) Wackestone interbedded with thin-bedded marlstone of the second member at Baiguowan. The close-up shows brachiopod fossils. (D-F) SRDs of the second member at Shangji (D), Shenxi (E) and Jiangjiazhai (F). (G-J) Photomicrographs of SRDs of the second member containing radiolarians, sponge spicules and ostracods at Baiguowan (G), Sancha (H), Xiangping (I) and Shangji (J). The coin is 2 cm in diameter. Each section of the folding ruler is 20 cm in length. Abbreviations: ab - alatoconchid bivalves; b - brachiopods; cr - crinoids; o - ostracods; r - radiolarians; ss - sponge spicules.

The development of the third member in the Zunyi area is indicated by conodonts of late Capitanian age. At Baiguowan, limestone interbeds of the uppermost SRDs yielded an assemblage consisting of *J. altudaensis* (Fig. 6.6, 6.7), *J. prexuanhanensis* (Fig. 6.8, 6.9) and *J. xuanhanensis* (Figs. 6.10, 6.11). Thus, a stratal record of the upper Capitanian Stage in the Zunyi area has been confirmed, at least as high as the *J. prexuanhanensis* – *J. xuanhanensis* Zone (Fig. 7).

The palaeogeography of the late Capitanian is generally similar to that of the early-middle Capitanian. In the siliceous deposition area (*e.g.*, in the Shangji section and the four drillcores), there occur manganese mudstone or manganese carbonates, intercalated with siliceous deposits (Figs. 7, 10J). Laterally, medium-bedded micrite and wackestone with chert nodules occur, and these deposits gradually transition into thick-bedded packstone, *e.g.*, at Magou and Sancha (Fig. 10I). Skeletal grains in packstone comprise mainly crinoids, foraminifers (including fusulinids), brachiopods, and minor ostracods and bryozoans (Fig. 10H).

At certain locations, an interval of approximately 10-m-thick massive silicified rocks is observed. This interval has informally been referred to as "city-wall chert" in the literature for its brecciated and weathering-resistant appearance (Regional Geological Survey Report, 1971). At Sangshuwan, there are many siliceous clasts supported by silicified matrix in the lower part of these massive rocks (Fig. 10C, D). The clasts are platy or cubic in shape, with diameters ranging from 3 to 35 cm. Most clasts exhibit dark grey banding (Fig. 10C). The orientations of the clasts, judging by their lengths or interior bands, range from chaotic to roughly parallel to the original bedding upwards (Fig. 10C). The upper part of the interval exhibits continuous beds of thin-bedded, banding silicified deposits (Fig. 10A, B) to massive, nodular sediments. At Baiguowan, the interval is massive (Fig. 10E). The clasts in the rocks are angular and range from 0.5 to 30 cm in diameter, whilst no banding is identified in these clasts (Fig. 10E). At both locations, skeletal grains in silicified clasts, including crinoids, foraminifers (including fusulinids) and calcareous algae, are observed (Fig. 10F, G).

The Guadalupian Maokou Formation in the Zunyi area is unconformably overlain by plant-bearing clay, siltstone or coal layers of the Lungtan Formation (Fig. 10I, K).

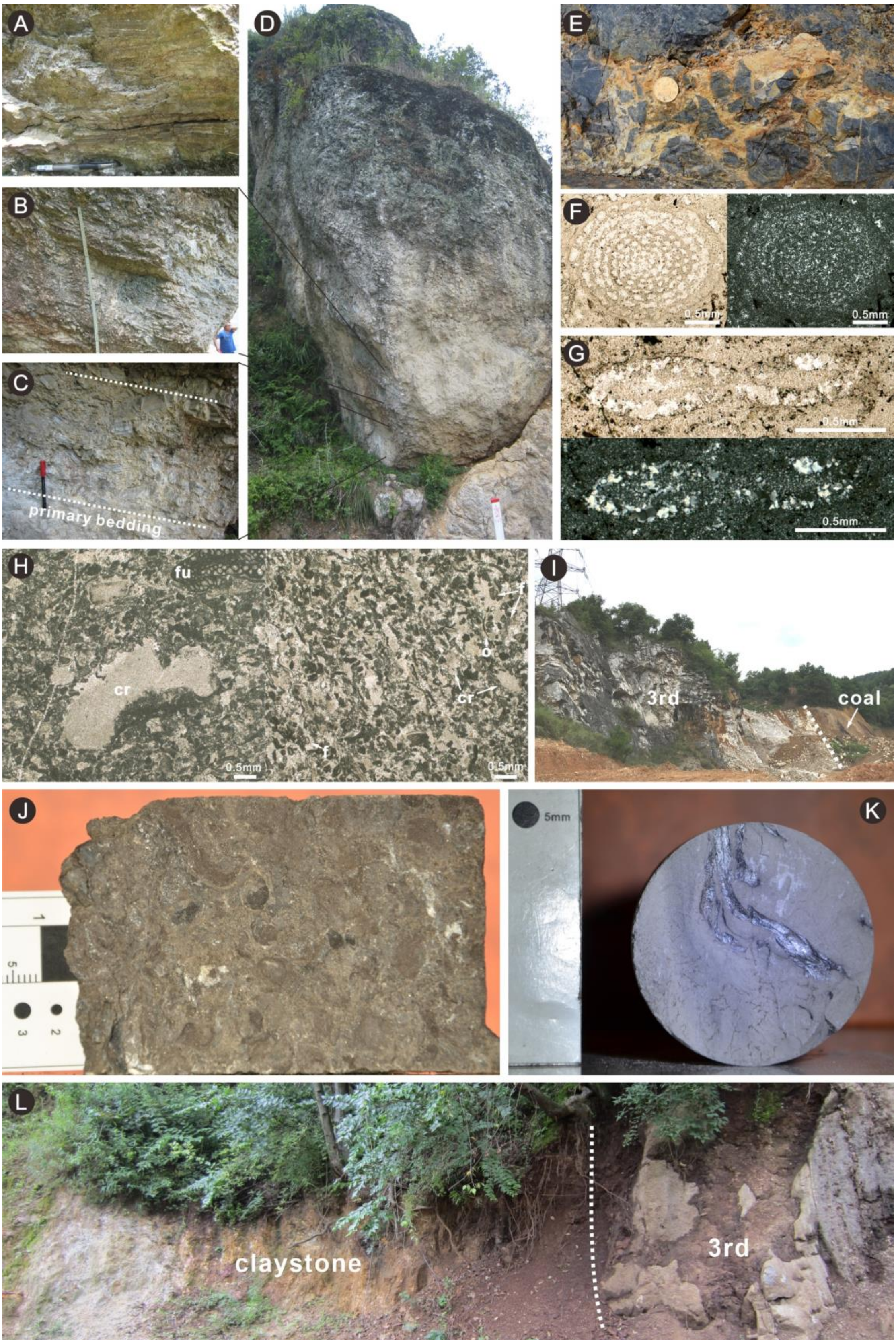

Fig. 10. Sedimentary features of the Maokou Formation in central Guizhou. (A-C) Close-ups show massive silicified rocks of the third member at Sangshuwan. Note the primary bedding and clasts

with dark grey bands in C. (D) Outcrop photo of massive silicified rocks of the third member at Sangshuwan. A is found behind D and could be traced to the upper part of D. (E) Outcrop photo of massive silicified rocks of the third member at Baiguowan. (F, G) Photomicrographs of silicified fusulinid (F), green algae (G) and matrix in the massive silicified rocks at Baiguowan. They are under the plane and perpendicular polarized light respectively in the same horizon. (H) Photomicrographs of packstone of the third member at Sancha. (I) Outcrop photo of the Sancha section. (J) Manganese carbonate of the third member at Shenxi. (K) Claystone of the Lungtan Formation at Shenxi. (L) Outcrop photo of the Xujiazhai section. Each section of the folding ruler is 20 cm in length. Abbreviations: cr - crinoids; f - foraminifers; fu - fusulinids; o - ostracods.

4.2 *Liupanshui – Anshun area*

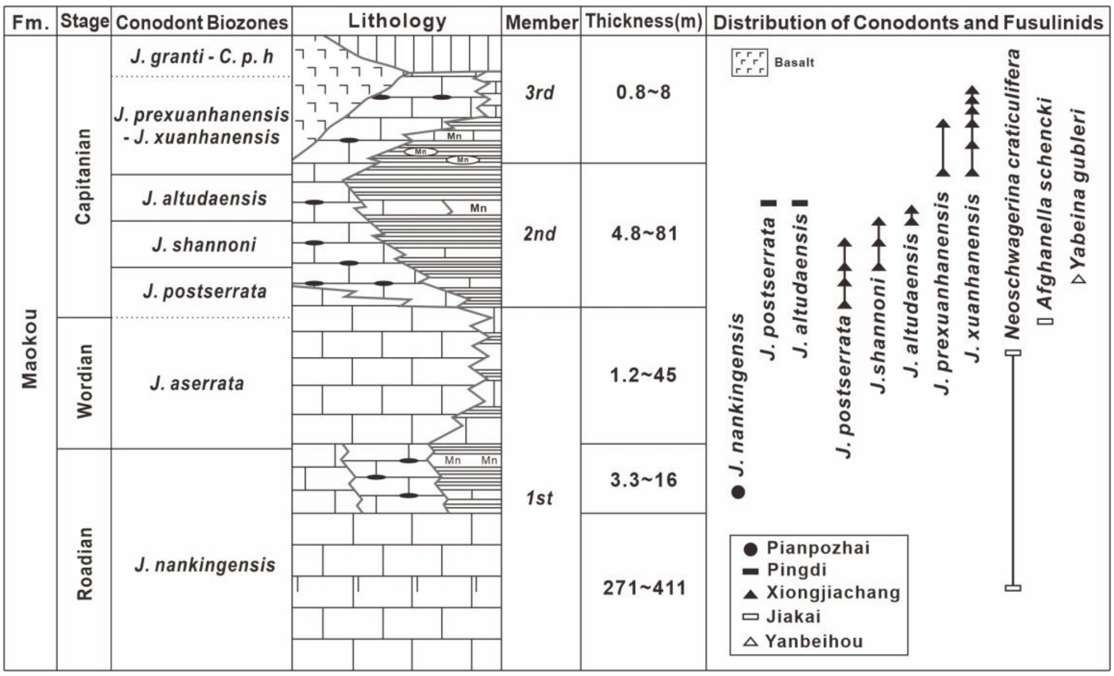

Fig. 11. Composite column of the Maokou Formation in the Liupanshui – Anshun area showing the lithology, thickness, conodont and fusulinid ranges. See Fig. 7 for legends.

The first member of the Maokou Formation is mainly characterised by grey, thick-bedded to massive packstone, with a bed of dolomitized limestone in the lower part (Fig. 11). The packstone comprises ~70% skeletal grains including foraminifers (including fusulinids), calcareous algae, crinoids and brachiopods (Fig. 8M). Fusulinids *Neoschwagerina craticulifera* (Schwager, 1883) and *Afghanella schencki* (Thompson, 1946) collected from this member at Jiakai indicate a Roadian-Wordian age (Fig. 11) (Regional Geological Survey Report, 1971). However, in the middle or upper part of the first member, thin- to medium-bedded wackestone alternates with SRDs and manganese mudstone at Jiangjiazhai, Xujiazhai and Pianpozhai (Fig. 1D), and upwards the limestone beds become thicker packstone with fragmented shells of alatoconchid bivalves and crinoids (Figs. 9B, 11) (Chen et al., 2018). Laterally, the wackestone is interbedded with chert bands at Yushe, Maluojing and Yanbeihou (Figs. 1D, 11). Skeletal grains comprise only ~15% of the wackestone and include brachiopods, foraminifers, ostracods, whilst 10% of the SRDs consist of sponge spicules. *Jinogondolella nankingensis* specimens (Fig. 5.1, 5.2) are identified from the thin-bedded wackestone at Pianpozhai. Thus, these siliceous beds in the first member accumulated during the Roadian (Fig. 11).

In the second member, thin-bedded SRDs occur at Jiangjiazhai, Xujiazhai, Pianpozhai, Pingdi and Xiongjiachang, and they are occasionally interbedded with manganese mudstone (Figs. 1D, 9F, 11). The SRDs have ~10% skeletal grains including sponge spicules, ostracods, radiolarians and foraminifers. Laterally, the second member passes from SRDs to medium- to thick-bedded wackestone with chert bands (Fig. 11). According to the conodont biostratigraphic investigation of Sun et al. (2010) at Xiongjiachang and Pingdi, this member of SRDs and wackestone is of early-middle Capitanian age (Fig. 11).

The third member is characterised by alternations of medium- to thick-bedded micrite with chert nodules and packstone (Figs. 10L, 11). The packstone comprises ~55% crinoids, foraminifers (including fusulinids) and brachiopods. The carbonates are commonly covered by basalt in the Liupanshui – Anshun area whilst in the Xujiazhai and Yanbeihou sections, the carbonates are overlain by the claystone of the Lungtan Formation (Fig. 10L). At Xiongjiachang, Sun et al. (2010) identified the *J. prexuanhanensis* – *J. xuanhanensis* Zone of the late Capitanian in this member, implying that the late Guadalupian erosion has removed less than two conodont zonations of strata, not as much as almost the entire Capitanian suggested by He et al. (2003).

4.3 *Facies interpretations and changes in central Guizhou*

Based on field observations and microfacies analysis, three facies types have been recognised in the Maokou Formation from central Guizhou.

Type 1: Light grey, thick-bedded wacke-packstone with abundant skeletal grains or peloids. Skeletal grains consist of foraminifers (including fusulinids), calcareous algae, brachiopods, crinoids and alatoconchid bivalves, suggesting deposition in warm, sunlit waters of shallow platform environments (Schlager, 2003; Flügel, 2010).

Type 2: Grey, thin- to medium-bedded wacke-packstone sometimes alternating with marl or siliceous beds. Skeletal grains are lower in diversity than *Type 1* facies and are dominated by brachiopods and gastropods. Other grains, such as foraminifers, ostracods and crinoids are only present occasionally. This assemblage indicates that this facies type develops in a somewhat deeper platform environment as evidenced by the absence of calcareous algae.

Type 3: Dark grey to black SRDs composed mainly of thin-bedded siliceous and argillaceous strata, and occasionally interbedded with thin-bedded limestone or lenticular, manganese mudstone. Skeletal grains are dominated by sponge spicules and radiolarians. Brachiopods, ammonoids, ostracods and foraminifers occur occasionally, and are smaller than the fossils seen in facies *Type 2*. The fossil assemblage suggests a deep-water basinal setting. The total thickness contour lines of SRDs in this type indicate that the faices are scattered on the platform (Fig. 4), rather than forming a linear trough or a series of rifts in the region (Chen et al., 1984; Luo et al., 1988). Thus, this facies is interpreted to have formed in intra-platform depressions in central Guizhou.

Based on the facies interpretations and the biostratigraphy above mentioned, the spatial-temporal change in depositional facies has been constrained. During the early Roadian, central Guizhou was a shallow-water carbonate platform across the region. By the late Roadian, deeper-water facies commenced in the Liupanshui – Anshun area (*e.g.*, *Type 2* at Xujiazhai, Yushe, Maluojing and Yanbeihou; *Type 3* at Jiangjiazhai and Pianpozhai), whilst shallow platform facies persisted elsewhere (Fig. 11). Later during the Wordian, the basinal deposits at Jiangjiazhai and Pianpozhai alternated with packstone containing abundant fragmented alatoconchid bivalves, and deeper platform deposits at the other locations turned to shallow platform wacke-packstone (*Type 1*) with well-preserved fusulinid shells (Fig. 11).

During the early Capitanian, deeper platform and local basinal facies recurred in the Liupanshui – Anshun area and initiated in the Zunyi area (Figs. 7, 11). In the *J. prexuanhanensis* – *J. xuanhanensis* Zone of the late Capitanian, limestone with chert nodules or manganese carbonates commonly filled depressions in both the Zunyi and Liupanshui – Anshun areas (Figs. 7, 11).

The depositional interpretation of the massive silicified rocks during the late Capitanian requires further investigation. These rocks only occur in specific sections in the Zunyi area and differ from the SRDs of facies *Type* 3, *e.g.*, these massive silicified rocks contain shallow-water organisms and low organic matter and argillaceous contents. All of the clasts and the corresponding matrix were silicified, so these rocks appear massive. In addition, the silicified rocks differ from place to place in regard to the clast types, sizes, shapes and depositional successions. Xu et al. (2020) recently measured the silicon isotopes and REE values of these massive silicified rocks and suggested that the observed brecciation and silicification resulted from hydrothermal migration along potential syndepositional faults during the stage of the ELIP.

5. Guadalupian sequence stratigraphy around the YCP

In order to evaluate the spatial-temporal relationship between the Guadalupian carbonate rocks and the SRDs, and the origin of deep-water deposits in central Guizhou, we have investigated the Guadalupian successions around the YCP. The lithostratigraphic units from different regions are summarized in Fig. 2 and detailed information on sections is in Appendix B.

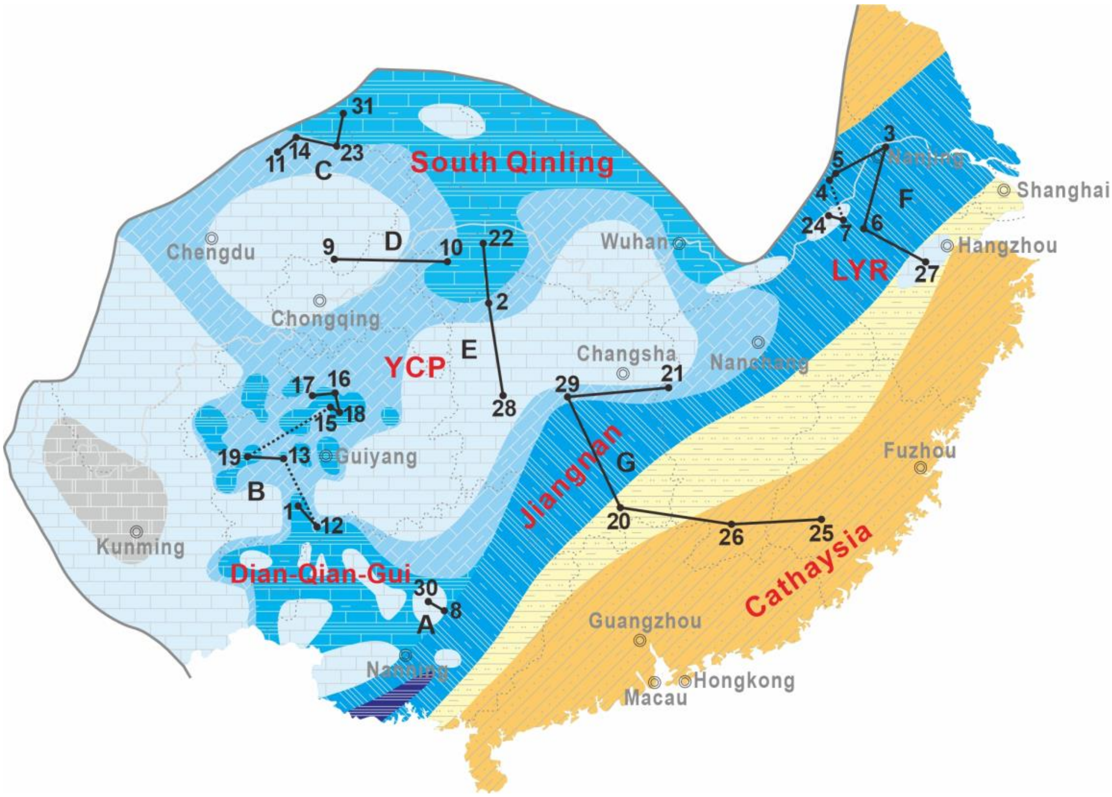

Fig. 12. Index map showing localities of transects for correlations of the Guadalupian sequence stratigraphy in South China.

5.1 *Southern South China*

In the southern region of South China, the YCP passes into the Dian – Qian – Gui Basin (Fig. 1B). SRDs interfinger with shallow-water carbonates on the margins of the YCP as well as the ICPs found within the basin. Transect A is along depositional dip and intersects the Laibin – Heshan ICP (Fig. 12). Transect B involves a correlation from the exterior to the interior of the southern part of the YCP (Fig. 12).

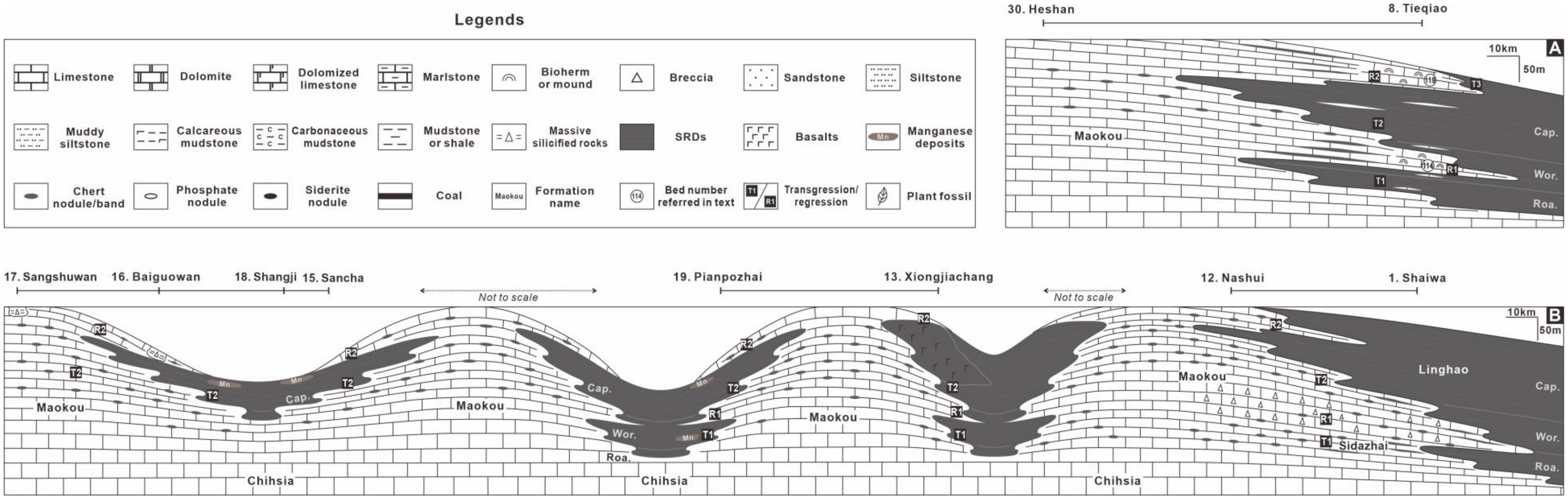

Fig. 13. Correlation of the Guadalupian sequence stratigraphy in the southern region of South China (transects A, B). See Fig. 12 for localities of transects and Appendix B for detailed information on sections. Conodont biostratigraphic version follows that in Fig. 3. Abbreviations: Roa. - Roadian; Wor. - Wordian; Cap. - Capitanian.

Transect A

The Tieqiao section in the transect is a supplementary reference section of the GSSP for the GLB. Intensive biostratigraphic investigations on this section make its Guadalupian Series a key reference for intra- and inter-regional correlation (Sha et al., 1990; Mei et al., 1999; Wang and Sugiyama, 2001; Jin et al., 2006). The basal part of the Guadalupian succession consists of medium-bedded bioclastic limestone, yielding *J. nankingensis* of Roadian age. The main body of the succession comprises SRDs with two massive carbonate intervals (Bed 114 and 119 of Sha et al., 1990) (Fig. 13A). The latter two intervals are mound-shaped bodies, individually ranging from a few metres to tens of metres in thickness, and from tens of metres to over one hundred metres in extent (Chen et al., 2009; Yao et al., 2012). The carbonates include sponge-*Tubiphytes* bafflestone and bryozoan-*Tubiphytes* framestone, as well as packstone, grainstone and floatstone containing crinoids, bryozoans, calcareous algae and alatoconchid bivalves (Chen et al., 2009; Yao et al., 2012; Huang et al., 2018). Laterally, the mound flanks become thin-bedded SRDs and limestone. Bed 114 spans within the *J. aserrata* Zone of Wordian age (Sun et al., 2017), whilst Bed 119 spans the *J. xuanhanensis* Zone to the *C. postbitteri hongshuiensis* Zone of late Capitanian age (Sun et al., 2017). The transition northwest from Tieqiao into the interior of the Laibin – Heshan ICP is a transition into bioclastic limestone (Fig. 13A). Here, a late Guadalupian erosion surface is draped by a thin bauxite layer between the Maokou and the overlying Heshan formations (Yu et al., 2016).

Transect B

Transect B encompasses the “trough” located in the southern margin of the YCP where the Guadalupian succession comprises the Sidazhai and the overlying Linghao (or Shaiwa) formations in the Shaiwa section, and the main part of the Maokou Formation in the shallower Nashui section (Fig. 13B).

The Sidazhai Formation at Shaiwa consists mainly of thin- to medium-bedded micrite, with sparse clasts showing graded bedding (Wang et al., 2016). Starting in the Roadian Stage, thin-bedded SRDs also occur alternating with the micrite beds, and from the late Roadian to early Wordian, thick-bedded breccia beds are present (Fig. 13B) (Wang et al., 2016). The succeeding Linghao Formation, characterised by SRDs, started in the early Capitanian *J. postserrata* Zone (Wang et al., 2016).

In the Nashui section, the Guadalupian succession begins with Roadian strata: thick-bedded bioclastic limestone and is overlain by medium-bedded limestone with chert nodules (Fig. 13B) (Shi et al., 2000; Henderson and Mei, 2003). These are succeeded by massive coarse carbonate megabreccias, containing abundant fusulinids and calcareous algae, that span the late Roadian to the Wordian (Shi et al., 2000; Henderson and Mei, 2003). This interval correlates with the breccia beds found in the Shaiwa section (Fig. 13B). Starting at the late Wordian, the megabreccias were overlain by thin-bedded siliceous limestone and SRDs in ascending order (Shi et al., 2000). Subsequently, limestone with chert nodules reappeared in the late Capitanian *J. prexuanhanensis* – *J. xuanhanensis* Zone (Fig. 13B) (Henderson and Mei, 2003). The limestone beds are unconformably covered by strata of the Wuchiaping Formation.

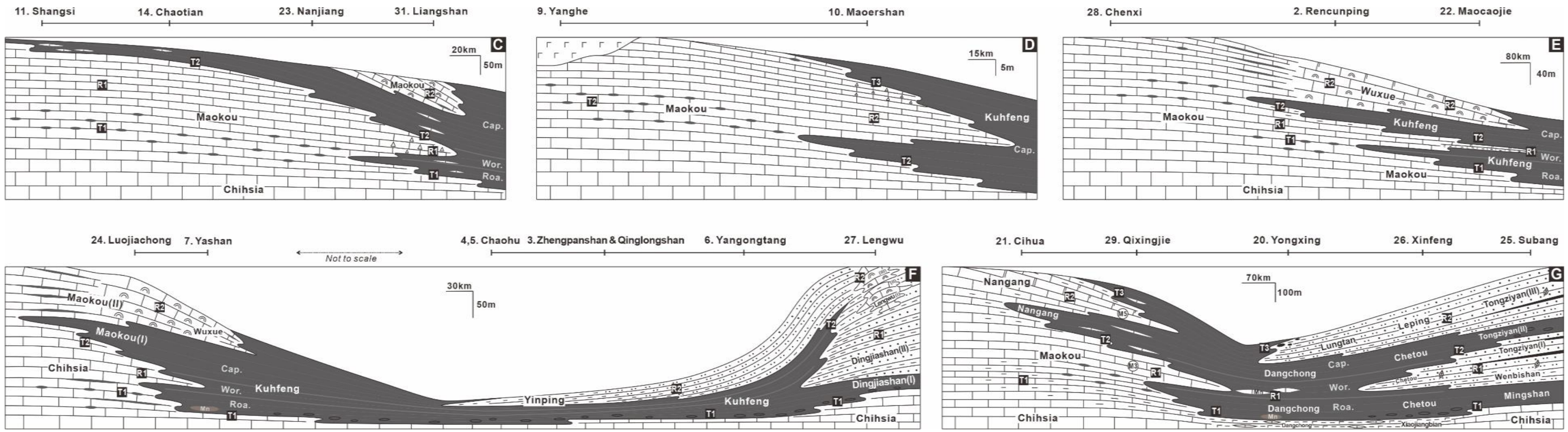

Fig. 14. Correlation of the Guadalupian sequence stratigraphy in the northern (transects C, D, E) and eastern (transects F, G) regions of South China. See Fig. 12 for localities of transects, Fig. 13 for legends and Appendix B for detailed information on sections. Conodont biostratigraphic version follows that in Fig. 3. Abbreviations: Roa. - Roadian; Wor. - Wordian; Cap. - Capitanian.

5.2 *Northern South China*

In the northern part of South China, the YCP passes northwards into the South Qinling Basin and an embayment developed in western Hubei (Fig. 1B). The bay formed during the early Guadalupian (Liu and Xu, 1994; Ma et al., 2009). Three transects (C, D, E) from basin to platform facies are correlated here (Fig. 12).

Transect C

This transect includes the Liangshan, Nanjiang, Chaotian and Shangsi sections (Fig. 14C). At Liangshan, the Guadalupian succession includes the uppermost part of the Chihsia Formation and the Maokou Formation (Kong, 2011). Three carbonate intervals are separated by two SRD intervals (Fig. 14C). The lower and the upper carbonate intervals are thick-bedded and massive bioclastic limestone respectively (Fig. 14C). Fossils in the former interval are dominated by crinoids and brachiopods, and in the latter by calcareous algae, crinoids and sponges (Kong, 2011; Cao et al., 2018). The middle carbonate interval is breccia consisting of micrite intraclasts. The lower limestone yielded the conodont *J. nankingensis* of the Roadian (Wang, 1978).

The Nanjiang, Chaotian and Shangsi platform sections are similar: the Maokou Formation consists of a lower limestone member and an overlying SRD member (Fig. 14C). The former is mainly packstone rich in calcareous algae and brachiopods, with chert nodules present in the lower part (Fig. 14C) (Lai et al., 2008). The limestone member continues to the early Capitanian *J. postserrata* Zone (Mei et al., 1994; Sun et al., 2008). The layers with chert nodules are dated to the lower *J. aserrata* Zone of the early Wordian (Sun et al., 2008). The SRD member contains intercalated limestone beds that increase upwards and also southwestwards from Chaotian to

Shangsi (Lai et al., 2008). However, the youngest conodont zone recorded in the upper SRD members varies: it is the *J. altudaensis* Zone at Nanjiang, the *J. shannoni* Zone at Chaotian and the upper *J. postserrata* Zone at Shangsi respectively (Mei et al., 1994; Sun et al., 2008). The top contact thus gets gradually older towards the platform interior (Fig. 14C).

The upper massive bioclastic limestone noted from the Liangshan section is obviously a distinct limestone horizon in the transect for it occurs above the SRDs (Fig. 14C).

Transects D and E

Transects D and E both trend from the basinal embayment in western Hubei to the northern part of the YCP (Fig. 12). The Maoershan section in transect D, as well as the Maocaojie and Rencunping sections in transect E, record a deeper-water environment, with the Rencunping section being closer to the platform (Fig. 12).

In transect D, the Capitanian Stage of the Maoershan section (Tianfengping of Wei et al., 2016) includes two intervals of limestone and two intervals of SRDs (Fig. 14D). The lower limestone is a bioclastic grainstone, with *J. altudaensis* yielded from the topmost horizons and the overlying SRDs also developed in the *J. altudaensis* Zone of the middle Capitanian (Zhang et al., 2007). The upper limestone is composed of packstone and breccia upwards. Conodont *J. granti* obtained in the top part of the breccia bed indicates a late Capitanian age for the following SRDs (Zhang et al., 2007).

In transect E, the Guadalupian succession includes, in ascending order, the Maokou, Kuhfeng and Wuxue formations. The Maokou Formation is a thick-bedded bioclastic limestone containing brachiopods, corals and fusulinids (Fig. 14E) (Tian et al., 2007; Cao et al., 2018). In the middle part of the Maokou Formation at Rencunping, there occurs an interval with abundant chert nodules (Fig.

14E) (Cao et al., 2018). The Kuhfeng Formation is mainly SRDs and contains ammonoids, sponge spicules and radiolarians (Shi et al., 2016; Cao et al., 2018). In the Maocaojie section, an interval of carbonaceous mudstone with dolomitized limestone interbeds occurs in the middle part of the SRDs (Fig. 14E) (Tian et al., 2007; Shi et al., 2016). Overlying the Kuhfeng Formation is the thick-bedded to massive limestone of the Wuxue Formation that contains abundant carbonate mud and calcareous algae and fusulinids, and is partly dolomitized in the upper part (Fig. 14E) (Tian et al., 2007; Cao et al., 2018).

According to the biostratigraphic data from the Maocaojie section and the nearby Luojiaba section, the Maokou Formation underlying the Kuhfeng SRDs remains within the Roadian Stage (Fig. 14E), supported by fossils of the *J. nankingensis* conodont Zone and *Pseudoalbaillella globosa* radiolarian Zone (Ma et al., 2016; Shi et al., 2016). In the Kuhfeng SRDs, the intercalated dolomitized limestone and carbonaceous mudstone are constrained to the *Follicucullus monacanthus* radiolarian Zone of the Wordian Stage (Shi et al., 2016). Upwards, the Kuhfeng SRDs continue into the middle Capitanian *F. scholasticus* radiolarian Zone, suggesting a late Capitanian age of the Wuxue Formation in the top part of the Maocaojie section (Fig. 14E) (Shi et al., 2016).

In the carbonate-dominated Rencunping section, the Maokouan limestone exhibits a longer range than that at Maocaojie and spans from the Roadian to the early Capitanian *J. postserrata* Zone (Fig. 14E) (Cao et al., 2018). Horizons bearing chert nodules in the middle part of the limestone yielded *J. aserrata* of the Wordian (Cao et al., 2018), coeval with those nodules in the Shangsi section from transect C (Fig. 14C, E). Specimens of *J. prexuanhanensis* identified in the uppermost part of the Kuhfeng Formation and *J. xuanhanensis* identified in the Wuxue Formation indicate that the *J. prexuanhanensis* – *J. xuanhanensis* Zone is the highest Capitanian biozone in the area (Cao et al., 2018).

The interior YCP successions in the Yanghe and Chenxi sections are mainly bioclastic limestone (Fig. 14D, E) (Mu et al., 1997; Liu et al., 2016a). In the Yanghe section, chert nodules are observed in the upper part of the Maokou Formation within the middle Capitanian *J. altudaensis* Zone. The Maokou Formation there is overlain by Emeishan basalts (Fig. 14D) (Liu et al., 2016a). The Guadalupian successions in both sections are truncated at an unconformity surface.

5.3 *Eastern South China*

During the Guadalupian, there were three depositional regimes in the eastern part of South China, the carbonate deposits of the YCP's margin and the ICPs in the west, the siliceous deposits of the Jiangnan Basin in the middle, and the terrestrial clastic deposits of the Cathaysian region in the east (Fig. 1B). Transects F and G stretch across the three regimes allowing lateral correlations of the carbonate, siliceous and terrigenous depositional systems to be seen (Fig. 12).

Transect F

Transect F is mainly in the LYR where the Chaohu, Zhengpanshan, Qinglongshan, Yangongtang and Lengwu sections array along a depositional trend approaching and intersecting with the Cathaysian region (Figs. 12, 14F). The Guadalupian biostratigraphy of the LYR has been intensively studied. The Chaohu, Zhengpanshan and Yangongtang sections are basinal locations that yield abundant deep-water fossils such as radiolarians, ammonoids and conodonts. The Zhengpanshan section is the locality where the holotype of the conodont species *J. nankingensis* was found (Jin, 1960). The standard radiolarian biozones of the Middle Permian, as well as the widely recognised Kuhfeng Formation, have been established in this area (Wang and Qi, 1995).

The Guadalupian successions of the Chaohu, Zhengpanshan, Qinglongshan and Yangongtang sections are composed of thick-bedded limestone in the basal part, thin-bedded SRDs in the lower part, and thin- to thick-bedded clastic deposits in the upper part (Fig. 14F). The Chihsia Formation in the basal part contains abundant fusulinids, corals and brachiopods (Kametaka et al., 2005). The SRDs of the Kuhfeng Formation are subdivided into a lower mudstone member with phosphate nodules, and an upper chert member yielding radiolarians, sponge spicules and bivalves (Kametaka et al., 2005; Kametaka et al., 2009; Zhang et al., 2019b). The overlying Yinping (or Yanqiao) Formation consists of siltstone at Yangongtang, which passes basinwards (northwestwards) into shale with trace fossils at Zhengpanshan, Qinglongshan and Chaohu (Figs. 12, 14F) (Kametaka et al., 2005; Du et al., 2010; Zhang et al., 2020).

At Lengwu, close to the Cathaysia oldland, the Guadalupian succession is composed of the Chihsia (top part), Dingjiashan and Lengwu formations in ascending order (Figs. 2, 14F) (Wang, 1993; Liu et al., 2016b). The Chihsia Formation is mainly a thick-bedded bioclastic limestone (Feng, 1991). The Dingjiashan Formation comprises a lower member of SRDs with phosphate nodules and fossils of ammonoids and sponge spicules, and an upper member of mudstone-sandstone alternations that yielded ammonoids, brachiopods and foraminifers (Fig. 14F) (Feng, 1991). The carbonate Lengwu Formation consists of massive sponge framestone and bafflestone in the lower and upper members, and medium-bedded crinoid-bearing marlstone with siltstone interbeds in the middle member (Fig. 14F) (Wang, 1993; Liu et al., 2016b).

As indicated by conodont biostratigraphy and zircon U-Pb dating, SRDs of the lower part of the Kuhfeng Formation and the lower part of the Dingjiashan Formation are of early Roadian age

(Wang, 1993; Wu et al., 2017; Zhang et al., 2019b). *Jinogondolella aserrata* of the Wordian has been reported from the lower carbonate member of the Lengwu Formation in the Lengwu section (Wang, 1993). These data constrain the mudstone-sandstone alternations of the upper part of the Dingjiashan Formation of the late Roadian, and the marlstone of the middle part of the Lengwu Formation of the early Capitanian (Liu et al., 2016b).

The distribution of siliciclastic deposits in the transect F demonstrates that there was a terrigenous supply derived from the southeastern Cathaysia oldland. From Lengwu to Yangongtang, and to Zhengpanshan and Chaohu, the siliciclastic successions gradually decrease in thickness and grain size as the terrigenous source becomes more distal (Fig. 14F). Consistent with the decrease, the base of siliciclastic deposits becomes younger to the northwest. Terrigenous deposits did not occur at Chaohu until the late Capitanian (Zhang et al., 2019b).

The Yashan and Luojiachong sections were located around the margin of an ICP, and were distant from the Cathaysian terrigenous input (Fig. 12). The Kuhfeng Formation at Yashan is similar to that at Chaohu, Zhengpanshan, Qinglongshan and Yangongtang, but the upper succession is the carbonate Wuxue Formation, pale grey, thick-bedded to massive bafflestone and bindstone built by calcareous algae, *Tubiphytes* and sponges, and wackestone to grainstone with abundant crinoids, calcareous algae and fusulinids (Figs. 2, 14F) (Luo et al., 1997; Du et al., 2010). These carbonates are mound-shaped with their dimensions ranging from tens to about one hundred metres, and laterally they pass rapidly into thin-bedded SRDs (Chen et al., 1995). The fusulinid genus *Metadoliolina* indicates a late Capitanian age for the Wuxue Formation (Luo et al., 1997; Kametaka et al., 2009). Considering the age assignment of the similar carbonate strata from the northern part of the YCP (*e.g.*, Rencunping and Maocaojie sections of Transect E), the Wuxue Formation of the

Yashan section probably belongs to the late Capitanian *J. prexuanhanensis* – *J. xuanhanensis* Zone.

The Luojiachong section was closer to the platform than the Yashan section. The Guadalupian succession is composed of two carbonate units separated by an SRD unit (Fig. 14F). The lower unit belongs to the upper part of the Chihsia Formation and the upper two units are grouped into the Maokou Formation (Wang, 1993). The upper part of the Chihsia Formation is mainly a dark grey, thick-bedded bioclastic limestone that yielded Roadian to Wordian conodonts of the *J. nankingensis* and *J. aserrata* zones (Wang, 1993). Similar to the Shangsi and Rencunping sections in the northern part of the YCP, horizons rich in chert nodules and chert interbeds are present in the lower part of the thick-bedded limestone (Fig. 14F). The succeeding SRD unit yielded *J. postserrata* of early Capitanian age (Wang, 1993). The upper carbonate unit is a pale grey, massive limestone with calcareous algae, crinoids, fusulinids and sponges (Luo et al., 1997). This unit is roughly dated as the late Capitanian by Chen et al. (1995), so it is correlatable with the above mentioned Wuxue Formation in the LYR and northern part of the YCP.

Transect G

The Subang, Xinfeng and Yongxing sections are a transition from the Cathaysia Block to the Jiangnan Basin, and are mainly the record of terrigenous and siliceous depositional systems (Fig. 12). Further to the northwest, the Qixingjie and Cihua sections progress from the Jiangnan Basin into the carbonates of the YCP (Fig. 12).

In the Subang section, the Guadalupian succession includes, in ascending order, limestone from the top part of the Chihsia Formation, the siliceous Mingshan Formation, and the siliciclastic Wenbishan and Tongziyan formations (Figs. 2, 14G). The Mingshan Formation (equivalent to the widespread Kuhfeng Formation in the LYR) consists of black SRDs (Du and Zhang, 1998; Li, 2008). The overlying Wenbishan Formation is mainly thin-bedded siltstone and silty mudstone with brachiopods, bivalves and ammonoids. The Tongziyan Formation includes three members (Fig. 14G). The lower and upper members are medium- to thick-bedded, fine-grained sandstone and siltstone with plant fossils and interbeds of coal. The middle member is mudstone with phosphate nodules (Li, 2008). The brachiopods *Neoplicatifera huangi* (Jin et al., 1974) and *Monticulifera sinensis* (Frech, 1911), and the ammonoid *Shouchangoceras shouchangensis* (Zhao and Zheng, 1977) present in the Wenbishan and Tongziyan formations indicate a Wordian-Capitanian age (Li, 2008). In the Wuping section, about 80 km from Subang, the fusulinid *Metadoliolina* obtained from the limestone interbeds in the upper part of the Tongziyan Formation indicates a late Capitanian age (Li, 2008).

In the Xinfeng and Yongxing sections from Jiangxi and Hunan provinces respectively, the Guadalupian successions comprise three lithological units. The first unit is characterised by calcareous mudstone with abundant limestone lenses, the second unit mainly includes SRDs, and the third unit is dominated by siliciclastics (Fig. 14G). Differences in the division scheme and naming of these units occur because geological surveys were conducted independently in these two provinces (Fig. 2). At Xinfeng, the first unit is called the Xiaojiangbian Formation, which is equivalent to the top part of the Chihsia Formation (Figs. 2, 14G) (Zhang, 1997; Liu, 2008). The second unit of SRDs is termed the Chetou Formation and contains phosphate nodules in the basal part and horizons of siltstone-sandstone with plant fossils in the middle part (Fig. 14G) (Liu, 2008). In the Yongxing section, the first two units are assigned to the Dangchong Formation (Figs. 2, 14G). The SRD unit consists of manganese mudstone interbeds in the basal part and an interval of manganese limestone in the middle part (Fig. 14G) (Wang, 1995b). The Chetou and Dangchong formations are commonly considered synonymous with the Kuhfeng Formation (Liu, 2008). In both sections, the lower part of the Chetou and Dangchong formations yielded ammonite genera (*Altudoceras*, *Waagenoceras* and *Paraceltites*) typical of the Roadian to Wordian interval (Bureau of Geology and Mineral Resources of Jiangxi Province, 1984; Zhang, 1997; Liu, 2008). The regional stratigraphic correlation by Wang (1995b) suggested that the horizons of siltstone-sandstone and the manganese limestone intercalated in the middle parts of the two formations were likely Wordian (Fig. 14G). Thus, the upper parts of these two formations probably belong to the Capitanian.

The third siliciclastic unit is locally termed the Leping Formation at Xinfeng and the Lungtan Formation at Yongxing (Figs. 2, 14G). These successions mainly consist of sandstone and siltstone fining upwards into mudstone and siltstone interbedded with siderite nodules at the latter locality (Fig. 14G) (Wang, 1995b; Liu, 2008). Limestone interbeds in the lower parts of the two formations yielded the fusulinid *Metadoliolina multivoluta* (Sheng, 1963) and ammonoids such as *Paratongluceras* (Zhang, 1997; Liu, 2008). The Leping and Lungtan formations are common Late Permian units in South China. The strata discussed here only belong to the lower part of the succession and contain late Guadalupian fossils.

Dominated by the terrigenous supply originating from the Cathaysia oldland, the distribution of siliciclastics of the Tongziyan, Leping and Lungtan formations in transect G is similar to that in transect F. Both show trends of getting finer and younger towards the west (Fig. 14G) (Zhou, 1985; Liu, 2008).

In the Qixingjie section from Hunan Province, the basal Guadalupian succession is a bioclastic limestone of the Chihsia Formation and the main body of the succession is the Maokou Formation

(Fig. 14G). The lower part of the Maokou Formation mainly comprises thin-bedded marlstone and siliceous limestone (Fig. 14G). The upper part includes two intervals of thick-bedded limestone (Bed M3 and M5 of Wang, 1995b), both of which are overlain by an interval of SRDs (Fig. 14G). Wang (1995b) assigned the basal part of the Maokou Formation to the Roadian and correlated Bed M3 with the manganese limestone horizons of the Wordian, and the dolomized Bed M5 with the sandstone and siltstone of the late Capitanian in the Yongxing section.

The Cihua section from Jiangxi Province was located at the eastern margin of the YCP. Its Guadalupian succession includes medium-bedded limestone of the top part of the Chihsia Formation, and the Maokou and Nangang formations in ascending order (Figs. 2, 14G). The Maokou Formation consists of black, thin-bedded marlstone and micrite yielding bivalves in the lower part, and of dark grey, thick-bedded to massive bioclastic limestone bearing fusulinids in the upper part (Fig. 14G) (Zeng et al., 2010). The Nangang Formation starts with SRDs and passes upwards into medium- to thick-bedded bioclastic limestone and dolomite yielding fusulinids and corals (Fig. 14G) (Zeng et al., 2010).

The occurrence of fusulinid *Neoschwagerina craticulifera* (Schwager, 1883) indicates that the top part of the Chihsia Formation and the Maokou Formation are of Roadian-Wordian age (Zeng et al., 2010). The overlying Nangang Formation started in the early Capitanian and the late Capitanian fusulinid *Metadoliolina* was found in its upper part (Zeng et al., 2010).

5.4 *Guadalupian sequence correlation in South China*

Correlations of the strata successions with the updated biostratigraphic constraints discussed above help establish a sequence stratigraphic framework for the Guadalupian succession of South

China. The sequences recorded here mainly reflect transgressive (T) - regressive (R) cycles (Catuneanu et al., 2009), namely T1 to T3, and R1 to R2 respectively. Although some gravity-flow deposits, such as massive breccia developed adjacent to some platform margins, are incorporated in the regressive phase, it is not possible to distinguish lowstand or shelf margin system tracts. During transgressive phases, SRDs of slope and basin facies expanded their extent and during regression shallow-water limestone was better developed.

Southern South China

In the Tieqiao reference section, the Maokou Formation is subdivided into three third-order sequences (Mei et al., 1999; Hu et al., 2012; Yao et al., 2012). Transgression T1 is indicated by the transition from the limestone of open platform facies to the SRDs of basin and slope facies during the Roadian (Fig. 13A) (Yao et al., 2012; Sun et al., 2017). Upwards, the two massive carbonate intervals (Bed 114 and 119 of Sha et al., 1990) are interpreted as bioherms based on their mound-like forms, smaller sizes (commonly tens of metres in dimension) and prominent components of sedentary organisms (such as calcareous algae and sponges) (Chen et al., 2009; Yao et al., 2012). Intercalated in siliceous basinal and slope deposits, and featured by phototrophic organisms (calcareous algae and alatoconchid bivalves), the massive mounds accumulated in relative shallow-water environments to the underlying and overlying siliceous deposits, correlating to the R1 and R2 regressions respectively (Fig. 13A). Bed 114 is constrained to the Wordian Stage, indicating that R1 belongs to the early Wordian, and the T2 transgression, above the Bed 114, started in the late Wordian (Sun et al., 2017). Wignall et al. (2009b) detailed the succession in the Bed 119 at Tieqiao and identified a transgressive surface within the Capitanian *J. granti* Zone. The thin-bedded limestone above the surface of Bed 119 coupled with the succeeding SRDs indicates the third transgression T3, which persisted into the Wuchiapingian (Fig. 13A). Thus, there were two and a half T-R cycles in the Guadalupian sequence. This scheme from Tieqiao is applicable throughout South China although the sequence boundaries require further study.

Guadalupian sequences on the southern slope of the YCP are recorded in the Shaiwa and Nashui sections (Fig. 13B). During the Roadian, the occurrence of chert beds and nodules within the limestone in both sections correlates well with the T1 transgression of Tieqiao (Fig. 13B). Above this, the breccia beds in the two places are traceable, and at Nashui the carbonate lithoclasts (megabreccias) were mainly derived from the shoal margin and had sizes up to tens of centimetres (Meng et al., 2018). The breccia beds are therefore interpreted as the products of R1 regression.

Along the slope, the T2 transgression is distinct (Fig. 13B). In the Shaiwa section, this transgression is evidenced by the substitution of the siliceous Linghao Formation for the carbonate-dominated Sidazhai Formation (Wang et al., 2016). In the Nashui section, the transgression is indicated by the decrease of bed thickness, from massive breccia to well-bedded siliceous limestone and SRDs in ascending order. The succeeding limestone with chert nodules represents the R2 regression (Fig. 13B) (Mei et al., 1999; Henderson and Mei, 2003).

In central Guizhou, the Guadalupian sequence in the intra-platform depressions is also correlatable with that of Tieqiao. Two transgressions, T1 and T2, are developed: the Roadian basinal SRDs in the Liupanshui – Anshun area, and the early Capitanian basinal SRDs and equivalent deeper platform siliceous limestone throughout central Guizhou (Fig. 13B). The intervening R1 regression could be represented by the local recovery of platform deposits in the basins of T1 in the Liupanshui – Anshun area (Figs. 11, 13B). Overlying the T2 deposits, the R2 regression could be related to the shallowing seen in the late Capitanian such as transitions from SRDs to manganese carbonate or limestone (Figs. 7, 11, 13B).

Northern South China

In northern South China, the Guadalupian T-R cycles recognised in the Tieqiao section are also traceable. The T1 transgression is seen as an onlap of the early Roadian platform carbonates by basinal SRDs in the Liangshan and Maocaojie sections (Fig. 14C, E). Traced towards the interior of the YCP, the T1 is manifest as horizons of deeper platform limestone with chert nodules in the Nanjiang, Chaotian, Shangsi and Rencunping sections (Fig. 14C, E). Subsequently, in basinal areas, the R1 regression is developed as breccia at Liangshan, and as carbonaceous mudstone and dolomitized limestone interbeds at Maocaojie. In the deeper platform area, the R1 is represented by the re-establishment of shallow, platform carbonate facies (Fig. 14C, E).

The succeeding T2 transgression of the early-middle Capitanian is widely present and saw the onlap of basinal SRDs onto platform limestone at Nanjiang, Chaotian, Shangsi, Maoershan and Rencunping (Fig. 14C-E). In the intra-platform area at Yanghe, the transgression is manifest as chert nodules within limestone (Fig. 14D). Above the basinal SRDs of the middle Capitanian, both of the massive limestone of bioherm facies at Liangshan, Maocaojie and Rencunping (Cao et al., 2018), and the packstone of shallow platform facies at Maoershan (Zhang et al., 2007) indicate a shallowing during the R2 regression (Fig. 14C-E). This regression resulted in an exposure of the platform and a major truncation of the preceding Capitanian succession in the area. Conodont biostratigraphic work reveals that at least two biozones of the Capitanian Stage are absent around the platform margin (*e.g.*, two absent at Rencunping) and more are absent towards the platform interior (*e.g.*, four absent at Nanjiang vs. six at Shangsi) (Figs. 3, 14C, D) (Mei et al., 1994; Sun et al., 2008).

The latest Capitanian deep-water SRDs dated by conodont *J. granti* at Maoershan confirm there was a transgression synchronous with the T3 transgression in the area (Fig. 14D) (Zhang et al., 2007).

Eastern South China

In eastern South China, a similar Guadalupian sequence framework is developed in the carbonate and terrigenous settings in the region.

Overlying the Chihsian carbonates, the widespread deposition of basinal SRDs, including the Kuhfeng, Dingjiashan, Dangchong, Chetou and Mingshan formations, as well as the common phosphate nodules in the lower part of the SRDs, characterise the T1 transgression (Fig. 14F, G). Similar to that in the northern part of the YCP, this transgression in the platform interior is marked by the occurrence of chert-bearing layers or marlstone at Luojiachong of an ICP from the LYR (Fig. 14F), as well as at Qixingjie and Cihua from the eastern part of the YCP (Fig. 14G). The return of shallow platform limestone to these sections during the Wordian corresponds to the R1 regression (Fig. 14F, G).

The R1 regression is clearly seen in the Cathaysian region as the progradation of siliciclastic deposits: delta-front sandstone at Lengwu (Fig. 14F) (Feng, 1991), and coal-bearing coastal siltstone and sandstone at Subang and Xinfeng (Fig. 14G). Basinwards, this regression is evidenced by SRDs being replaced by manganese carbonates in the Yongxing section (Fig. 14G).

The subsequent T2 transgression during the early-middle Capitanian is clearly developed in both platformal and paralic areas. At Luojiachong, Qixingjie and Cihua, the transition from shallow platform limestone to SRDs characterises the T2 transgression in these carbonate platform settings (Fig. 14F, G) (Wang, 1993; Wang, 1995b; Zeng et al., 2010). At Lengwu, Xinfeng and Subang, the occurrences of marlstone and siltstone in the middle part of the Lengwu Formation, and SRDs of the upper part of the Chetou and the middle part of the Tongziyan formations mark a fining trend during T2 in these siliciclastic successions (Fig. 14F, G).

Above the T2 transgression, in carbonate systems, the R2 regression is marked by facies changes to the bioherms (the upper part of the Maokou Formation at Luojiachong and the Wuxue Formation at Yashan) surrounding an ICP from the LYR (Fig. 14F) (Luo et al., 1997) and to the shallow platform carbonates at Qixingjie and Cihua on the eastern part of the YCP (Fig. 14G) (Wang, 1995b; Zeng et al., 2010). Meanwhile, in terrigenous settings, the R2 regression is the progradation of terrestrial facies. The coarser-grained siliciclastics reached further Yangongtang, Zhengpanshan and Chaohu in the LYR (Fig. 14F), and Yongxing in the Jiangnan Basin (Fig. 14G), indicating a more substantial siliciclastic progradation than that of R1.

In eastern South China, the uppermost Guadalupian succession is either completely developed or slightly eroded because of the late Guadalupian emergence (Hu, 1994; Zhang et al., 2019b). Thus, it is hard to identify the spatial distribution of the T3 transgression. Some sections show potential evidence: at Yongxing there is a change from sandstone to mudstone and at Qixingjie from platform limestone to SRDs (Fig. 14G) (Wang, 1995b).

The discussions above illustrate that the Guadalupian sequence stratigraphy, identified at the GSSP section of Tieqiao, is recognisable across all carbonate and siliciclastic depositional systems in South China. Two regressions, R1 and R2, are recognised between three transgressions, T1, T2 and T3, constituting two and a half T-R sedimentary cycles. The first cycle T1-R1 spanned the early Roadian to the late Wordian. The second cycle T2-R2 occurred from the late Wordian to late Capitanian. The T3 transgression was present in the latest Capitanian.

6. Discussion

Having established a chronostratigraphic framework for the evolution of Guadalupian carbonate platforms in South China, we here discuss the interrelated factors that controlled the development of these platforms.

6.1 *Developing pattern of the YCP*

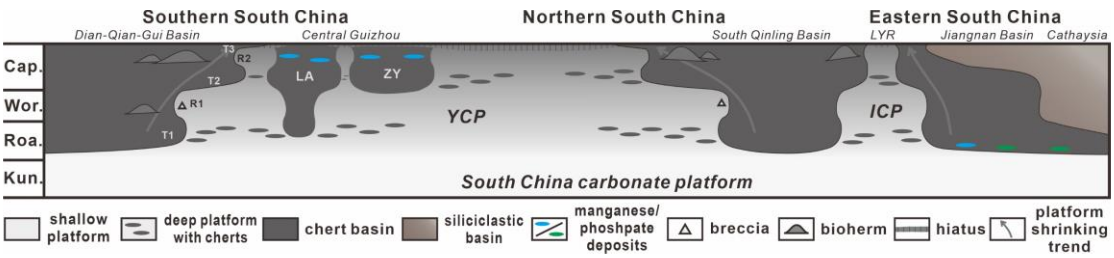

Fig. 15. Sketch showing carbonate platform evolution during the Guadalupian in different regions of South China. Abbreviations: Kun. - Kungurian; Roa. - Roadian; Wor. - Wordian; Cap. - Capitanian; LA - Liupanshui – Anshun area; ZY - Zunyi area.

Inherited from the late Kungurian, a shallow-water carbonate platform was still widespread during the earliest Guadalupian and termed the South China carbonate platform because it covered almost the entire region (Fig. 15) (Liu and Xu, 1994; Wang and Jin, 2000; Ma et al., 2009). However, a dramatic demise of the carbonate platform took place during the T1 transgression. As a result, the initial YCP as well as some ICPs developed in the basins of the South Qinling region and the LYR, replacing the South China carbonate platform (Figs. 1B, 15). Meanwhile, local intra-platform depressions formed in the Liupanshui – Anshun area of central Guizhou (Fig. 15).

The YCP and ICPs contracted during the T2 transgression, when siliceous basins onlapped the margins of the platforms, causing further area loss (Fig. 15). Within these platforms, the intra-platform depressions expanded markedly in the Liupanshui – Anshun and Zunyi areas during T2. Moreover, the deeper platform facies surrounding these basinal areas in the interior of the YCP also developed at the expense of shallow platform facies (Figs. 1B, 15).

Following the above two transgressive phases respectively, the two expanding phases of carbonate extent in our study interval saw the development of massive breccia and bioherms along the peripheries of the YCP and ICPs (Fig. 15). The breccia, observed during the R1 regression, indicates a shedding of the lithified YCP’s margin. The bioherms mainly coincided with the R2 sequence which terminated at a regional unconformity. Notably, these bioherms developed enclosed in slope or basinal environments and were probably independent of the broad platform area (Fig. 15). Thus, their occurrences in the preceding platform areas during the R1 regression, such as at Rencunping and Luojiachong (Fig. 14E, F), demonstrate that the extents of the YCP and ICPs did not recover to match those during R1 (Fig. 15).

After the R2 regression, a younger backstepping of platform carbonates occurred during T3, although much of its distribution remains to be verified biostratigraphically in the platform margin areas.

In summary, the YCP, which commenced after the drowning of the South China carbonate platform, shrank during the Guadalupian as the platform margin contracted whereas deep-water depressions developed within the platform interior (Fig. 15).

6.2 *Comparison to other Guadalupian carbonate platforms*

Marginal retreat accompanied with onlaps of the SRDs recorded in the YCP broadly occurs in other Guadalupian carbonate platforms worldwide.

The Permian Chert Event, corresponding to quantities of chert-bearing accumulation, formed a belt of glass ramps that substitute for carbonate production in many well-documented carbonate platform areas along the coast of boreal to tropical Pangea (Fig. 1A, locations 2-4) (Murchey and Jones, 1992; Ritterbush, 2019; Matheson and Frank, 2020a). The above substitution occurred simultaneously in the Phosphoria Sea and the Barents Sea during the Roadian (Blomeier et al., 2013; Davydov et al., 2018a; Matheson and Frank, 2020b) and emerged during the early Capitanian in the Sverdrup Basin (Beauchamp and Grasby, 2012). These two periods of substitution support the contraction of carbonate platforms associated with the T1 and T2 transgressions.

The subtropical to tropical paleo-Tethys is another region with thriving Guadalupian carbonate platforms (Fig. 1A). In the Arabian isolated platform and southern Primorye in Far East Russia (Fig. 1A, locations 5 and 9), the retreat of carbonate platform margins is evidenced by a facies shift from bioclastic limestone to mudstone or cherts in the upper part of the Guadalupian succession (Weidlich and Bernecker, 2007; Kani et al., 2018). Similar changes occur in Transcaucasia, the Abadeh area and the Pha Nok Khao platform (Fig. 1A, locations 6-8), where light grey, thick-bedded limestone becomes increasingly blackish and thinner, and contains a higher chert proportion upwards both in the middle and upper parts of the Capitanian succession, reflecting platform contraction during the T2 and T3 transgressions, respectively (Leven, 1998; Hada et al., 2015, fig. 4; Arefifard and Payne, 2020, fig. 10b).

In the Panthalassan paleo-atoll of southwest Japan (Fig. 1A, location 10), the marginal retreat has not been clearly observed on the seamount-capping carbonate mass, but the multiple subaerial exposures present in the carbonate interval during the Wordian and the late Capitanian could be attributed to the R1 and R2 regressions, respectively (Nakazawa and Ueno, 2004; Nakazawa et al., 2012; Kofukuda et al., 2014).

The measurement of the global lithofacies distribution by Kiessling et al. (2003) suggested that nearly 3×10^5 km^2 of shallow carbonate platform areas disappeared during the Guadalupian period. Comprehensive research of the process of Guadalupian platform contraction from platform to platform worldwide could undoubtedly be significant in revealing the mechanism of platform evolution. Nevertheless, the YCP could represent an irreplaceable case in this regard to be scrutinized for its complete Guadalupian succession and sound biostratigraphy basis.

6.3 *Regional tectonics on the carbonate platform evolution*

Demise of carbonate platforms is commonly attributed to increasing accommodation that can outpace platform growth, causing platform margins to retreat and give way to deeper-water facies (Godet, 2013; Menier et al., 2014). Rapid tectonic subsidence is one mechanism that may create accommodation and has generally been held responsible for SRDs development (*e.g.*, Kong and Gong, 1987; Feng, 1991; Jiang et al., 1994; Cheng et al., 2015). The ELIP was a major tectono-volcanic event in the western YCP region in the Middle to Late Permian. An "Emei taphrogenesis" was proposed to explain the occurrence of Emeishan basalts in narrow but thick (up to ~2000m) outcrops (Luo et al., 1988). The SRDs and manganese deposits in central Guizhou have long been attributed to similar tectonic activities occurring in the western YCP region (Chen et al., 1984; Luo et al., 1988; Jin and Feng, 1995). However, other factors may be responsible for SRDs formation.

In this study, we have shown that, in central Guizhou, the SRDs were developed in depressions of various shapes and their distributions were scattered on the top of the YCP (Fig. 4B). These depressions do not show linear trends predicted by active, fault-controlled rifting (*e.g.*, Zhu et al., 1989; Chen et al., 2003). Such features imply that these depocentres were not the product of taphrogenesis. In addition, the gradational transition of lithofacies between shallow-water carbonates and deeper-water SRDs, with a belt of siliceous limestone transitional facies in-between (Fig. 1B), further argues against sharply delineated syn-rift basins (*e.g.*, Woodfine et al., 2010; Silva-Casal et al., 2019). The synchronous change in depositional facies of the platform margins, including areas surrounding the YCP and ICPs, and in the YCP's interior, also does not support a local, syndepositional faulting control on the occurrence of SRDs.

6.4 *Guadalupian environmental conditions for carbonate platform developing*

Anoxia and upwelling are two factors commonly discussed in the interpretation of Guadalupian marine environments in South China, and are interrelated to some extent. Their possible impacts on the YCP's evolution are discussed below.

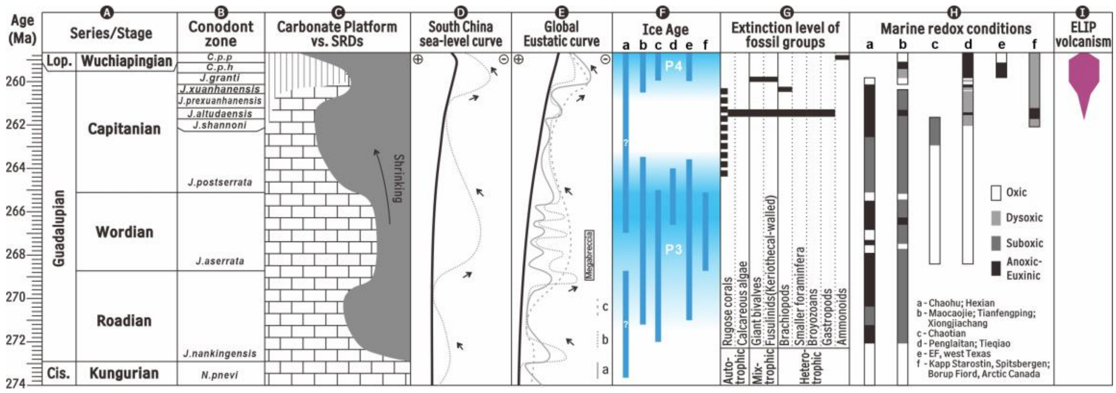

Fig. 16. Summary on the evolution of carbonate platforms in South China and possible influencing factors during Middle-Late Permian. (A, B) Timescale and biostratigraphic framework from Henderson (2016) and Shen et al. (2018). (C) Generalized platform margin onlaps based on this study. The interval filled by vertical lines shows missing strata caused by regression. (D) South China relative sea-level change modified from Wang et al. (1999). (E) Global eustatic sea-level changes modified from: a - Haq and Schutter (2008); b - Ross and Ross (1987); c - Haq and Al-Qahtani (2005). Megabreccia event from Haq and Schutter (2008). (F) Durations of glacial events. Data of a-d are from eastern Australia and recalibrated in this timescale based on: a - Fielding et al. (2008); b - Metcalfe et al. (2015); c - Frank et al. (2015); d - Garbelli et al. (2019). Data of e and f are from: e - NE Russia (Davydov et al., 2016; Davydov et al., 2018b); f - Arctic Canada (Beauchamp and Grasby, 2012). (G) Extinction levels of different fossil groups summarized from Wang and Sugiyama (2001), Weidlich (2002a), Ota and Isozaki (2006), Isozaki and Aljinović (2009), Shen and Shi (2009), Bond et al. (2010a) and Huang et al. (2018). (H) Comparison of marine redox conditions in: a - central Anhui (Wei et al., 2019; Zhang et al., 2019a); b - western Hubei (Shi et al., 2016; Wei et al., 2016) and central Guizhou (Wignall et al., 2009a); c - northern Sichuan (Saitoh et al., 2013b); d - central Guangxi (Zhang et al., 2015; Wei et al., 2016); e - West Texas (Zhang et al., 2015); f - Spitsbergen and Arctic Canada (Bond et al., 2015; Bond et al.,

2019). (I) Volcanism of the ELIP from Sun et al. (2010) and Wu et al. (2020).

Low oxygen (dysoxic) concentrations are commonly cited as a cause of impaired carbonate platform growth and may even cause platform shutdown, as documented by many Mesozoic examples (*e.g.*, Han et al., 2018; Jin et al., 2018; Bonvallet et al., 2019). In South China, dysoxic facies are common in the Guadalupian geological records. Thus, oxygen-deficient conditions were present in the Roadian recorded by the deposition of the Kuhfeng Formation, an organic rich shale with phosphate and manganese deposits (Tong and Zhou, 1985; Wu et al., 1994). Diverse indicators (*e.g.*, radiolarian palaeoecology, pyrite petrography and trace metal enrichment) indicate the Kuhfeng Formation was deposited in suboxic-euxinic conditions from the Roadian to early Capitanian, in basins surrounding the northern to eastern part of the YCP (Fig. 16H-a, b, c) (Saitoh et al., 2013b; Shi et al., 2016; Cao et al., 2018; Wei et al., 2019). In contrast, shallow-water areas remained oxygenated during this time (Wei et al., 2019).

Severe oxygen-deficiency occurred in the middle-late Capitanian, when the redox variation has been extensively investigated because of its potential role in the Capitanian extinction. Anoxic-euxinic conditions in the deep-water facies are evidenced by C-S-Fe systematics combined with enrichment of redox-sensitive trace element (Mo, U, V) contents in the LYR (Fig. 16H-a) (Zhang et al., 2019a), and by the small size of pyrite framboids in central Guizhou (Wignall et al., 2009a) and western Hubei (Fig. 16H-b) (Wei et al., 2016). At time these oxygen-poor conditions impinged in the northeastern margin of the YCP (Wei et al., 2019), and the southeastern margin of the Laibin – Heshan ICP (Fig. 16H-d) (Zhang et al., 2015).

Marine anoxia during the middle-late Capitanian was widespread and potentially a near-global phenomenon, seen in tropical to temperate latitudes, including the Delaware Basin of western Pangea (Fig. 16H-e) (Zhang et al., 2015; Smith et al., 2019), the Sverdrup Basin (Arctic Canada) and the cool, temperate shelf seas (Spitsbergen) of northern Pangea (Fig. 16H-f) (Bond et al., 2015; Bond et al., 2019).

Therefore, recurrent dysoxic conditions in the basins surrounding carbonate platforms in South China may have affected the carbonate productivity of these areas, with upwelling exacerbating the effect (Saitoh et al., 2013a; Zhang et al., 2021). However, an upwelling mechanism is unlikely to have been important in the formation of the intra-platform depressions far removed from oceanic circulation regimes. Therefore, caution should be taken when invoking upwelling as a control on the entirety of carbonate platform evolution of the YCP in the Guadalupian.

6.5 *Glaciation*

Other than the regional factors discussed above, eustasy also plays a critical role in the construction of carbonate platforms (Pomar and Kendall, 2008). Glacio-eustasy is likely to have played a role in the development of Guadalupian carbonate systems of South China because the interval marked the final stages of the Late Paleozoic Ice Age (LPIA), the most prolonged glacial interval of the Phanerozoic.

The LPIA spanned the Carboniferous to Permian and saw ice caps develop, especially in southern hemisphere Gondwanan locations. Eight discrete glacial intervals (with interglacial intervals in-between) have been recognised (Fielding et al., 2008). Among the eight glaciations, the last two intervals, P3 and P4, occurred during the Middle-Late Permian and were initially identified in high-latitude glacial sedimentary records retrieved from eastern Australia (Fig. 16F-a) (Fielding et al., 2008). In the Sydney and Bowen basins of eastern Australia, U-Pb CA-TIMS calibration of the glacial sedimentary formations suggests that P3 is late Roadian to early Capitanian in age, and P4 is latest Capitanian to Wuchiapingian in age (Fig. 16F-b) (Metcalfe et al., 2015). Based on the palaeotemperature proxies and sedimentological records in the same region, Frank et al. (2015) calibrated a slightly earlier P3 duration than that suggested by Metcalfe et al. (2015) (Fig. 16F-c). The recent $^{87}Sr/^{86}Sr$ ratio measured on brachiopod shells obtained from the Sydney Basin implied that the P3 duration is shorter than previously reported, extending only from the late Wordian to the early Capitanian (Fig. 16F-d) (Garbelli et al., 2019).

Sedimentary records from the northern hemisphere also confirm two glacial events in the Guadalupian. In the high-latitude NE Russia, the results of U-Pb dating on glacial sediments obtained by Davydov et al. (2016) and Davydov et al. (2018b) indicate that P3 spans the Wordian, and P4 is latest Capitanian to Wuchiapingian in age (Fig. 16F-e). In temperate regions, both glacial intervals coincide with the decrease of warm-water fusulinid diversity in North America (Davydov, 2014). Cooling in the Wordian was supported by the occurrence of a cool-water fauna of bryozoans and brachiopods, and dropstones in the Sverdrup Basin (Fig. 16F-f) (Beauchamp and Baud, 2002; Beauchamp and Grasby, 2012). In tropical South China, the P4 cooling interval has been conjectured from paleotemperature proxies during the earliest Wuchiapingian (Chen et al., 2013; Yang et al., 2018; Wang et al., 2020).

6.6 *Eustatic and relative sea-level changes*

The first order eustatic trend, derived from the study of depositional sequences of major cratons such as North America and Russian platform, consists of a regressive trend from the Roadian to the

Wuchiapingian with amalgamation of the Pangea (Fig. 16E) (Ross and Ross, 1987; Haq and Al-Qahtani, 2005; Haq and Schutter, 2008). Second order changes, superimposed on this trend, consist of three intervals, all three dominated by sea-level rise, with two phases of regression. The former regression began during the late Roadian, after which the sea-level experienced a change to high frequency (third-order) fluctuations, with a magnitude up to approximately 70 m during the Wordian (Fig. 16E). This interval coincides with the P3 glaciation discussed above (Rygel et al., 2008; Chen et al., 2016) and well-developed lowstand features such as widespread carbonate megabreccias (Fig. 16E) (Haq and Schutter, 2008). The latter regression during the late Capitanian marks one of the lowest points of sea-level in the Phanerozoic (Fig. 16E).

The relative sea-level changes recorded in the Guadalupian succession of South China closely follow the eustatic trend. The Guadalupian succession in South China occurs beneath a major unconformity (Figs. 15, 16C) (Hou et al., 2020). This unconformity, first recognised in the 1930s, has been attributed to the “Dongwu tectonic movement” or “Dongwu uplift” (Hu, 1994) and marks the conventional lithostratigraphic boundary between the Guadalupian and Lopingian in China. Below the Guadalupian strata are the Chihsian carbonates and underlying thin coal-bearing littoral deposits (locally termed the Liangshan Formation and equivalents) (Fig. 3). There exists another major unconformity below the Liangshan Formation (Shen et al., 2018). The succession from the Liangshan Formation to the Chihsian carbonates constitutes a major transgressive sequence based on the succession from coal-bearing littoral deposits to shallow shelf carbonates with a major basal unconformity, as well as the widespread Chihsia Formation in the region. Thus, the entire Guadalupian succession in South China comprises the falling phase of a major sea-level change cycle, consistent with the regressive trend of the first order eustatic fluctuation (Fig. 16D, E) (Chen et al., 1998; Mei et al., 1999; Shen et al., 2018). The minor cyclicity of the Guadalupian succession in South China, namely, the identified two and a half T-R cycles, also correlates well with the three intervals of the second order eustatic changes mentioned above (Wang et al., 1999; Hu et al., 2012) (Fig. 16D, E).

6.7 *Guadalupian mass extinction and carbonate factories*

The Guadalupian mass extinction has been demonstrated to be a global event (Shen and Shi, 2002; Clapham et al., 2009; Wignall et al., 2012; Bond et al., 2015), and recently ranked as the sixth major mass extinction (Rampino and Shen, 2019). It has been estimated that about 60% of species and 34% of genera of marine animals were eliminated during this biocrisis (Stanley, 2016). Photosynthetic and photosymbiont-bearing taxa, such as calcareous algae, keriothecal-walled fusulinids, alatoconchid bivalves and corals, were especially badly affected (Fig. 16G) (Isozaki and Aljinović, 2009; Wignall et al., 2009a). All of these taxa were important carbonate producers in the shallow platform environments of low-latitudes (Kiessling et al., 2003). During the crisis, reefs in the Guadalupian experienced a heavy loss in terms of both reef abundance and carbonate production, of 47% and 89% respectively (Flügel and Kiessling, 2002; Kiessling, 2002).

However, analysis of the geological literature suggests that reefs and carbonate platforms were generally two systems that were decoupled in quantity and quality (Kiessling et al., 2003). Reefs that flourished during the Guadalupian were mainly built by microbes, bryozoans and *Tubiphytes*, which had a minimal contribution to carbonate platforms (Kiessling, 2002; Weidlich, 2002b). If the shrinking process of the YCP shown above resulted from a less productive carbonate factory, then we must consider factors impacting the carbonate producers on the platform. As noted above, the surrounding basinal areas were oxygen poor raising the possibility that such conditions may have reduced carbonate production. The loss of photosynthetic taxa during the Guadalupian mass extinction could also have played a role although the contraction in the area of platforms was underway long before the crisis (Fig. 16C, G). However, the timing of this crisis is debated and some have suggested there was a prolonged gradual pattern of diversity decline but no crisis event (Clapham et al., 2009; Groves and Wang, 2013). Cooling during glacial intervals is another possibility but it is noteworthy that the earlier glaciations of the Permian, before P3 and P4, did not impact carbonate productivity. Further work is clearly needed but it is noteworthy that neither scenario – abrupt crisis versus prolonged diversity decline – correlates closely with the waxing and waning of carbonate platform extent in the Guadalupian of South China (Fig 16).

7. Conclusion

1. A unique Guadalupian succession of deep-water SRDs, as well as manganese carbonates developed on the YCP in central Guizhou. Spatially, these SRDs developed in isolated depressions scattered in the region, surrounded by limestone of deeper platform facies. Conodont identification of this succession indicates that these deep-water deposits commenced in the Liupanshui – Anshun area during the Roadian and then spread northeastwards to the Zunyi area during the Capitanian.

2. Sequence correlations across the carbonate, siliceous and terrigenous depositional areas in South China reveal that there are three transgressions (T1, T2 and T3) and two regressions (R1 and R2) in the Guadalupian succession. The transgressive-regressive cycles match the Guadalupian global eustacy well.

3. The Guadalupian evolutionary history of the YCP in South China is recovered. The YCP

commenced during the early Guadalupian and thereafter experienced gradual contraction. The contraction process was demonstrated by the commencement and expansion of depositional depressions within the platform interior (partial drowning) and the increasing onlaps of the SRDs along the platform margin (marginal retreat).

4. The partial drowning and marginal retreat during the course of platform shrinking reminded us of less productive shallow-water carbonate factories on the platform top. In addition to the global eustatic fluctuations, attention should be given to the Guadalupian mass extinction and the environmental deterioration also potentially impacting the evolution of carbonate platforms during the Guadalupian.

Acknowledgements

We thank Die Chen for helpful field work, and Xia Wang and Bowen Zan for useful discussions to improve the manuscript. This work was supported by the National Natural Science Foundation of China (Grant No. 41872117 and No. 41472087). Zhichen Liu and Deng Chen's work was supported by the Young Talent Funding of Guizhou (No. 2019-5654). We gratefully acknowledge the constructive review by Merlynd Nestell and an anonymous reviewer for their thoughtful comments and suggestions that improved this manuscript.

Declaration of Competing Interest

The authors declared that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

Appendix A. Supplementary data

Data of locations and total thickness of thin-bedded Guadalupian SRDs collected from the stratal sections in Guizhou Province.

Appendix B. Section information

Detailed information on sections for Guadalupian sequence correlation in South China

No.	Section	Location	Name of SRDs	Reference	Types of biostratigraphy
1	Shaiwa	Sidazhai, Ziyun, Guizhou	Linghao Fm.	Wang et al. (2016)	Conodont
2	Rencunping	Sangzhi, Hunan	Maokou Fm.	Cao et al. (2013); Cao et al. (2018)	Conodont
3	Zhengpanshan & Qinglongshan	Lungtan, Nanjing, Jiangsu	Kuhfeng Fm.	Jin (1960); Wang (1995a); Zhang et al. (2020)	Conodont
4	Chaohu (Anmenkou)	Chaohu, Anhui	Kuhfeng Fm.	Kametaka et al. (2005); Kametaka et al. (2009)	Radiolarian
5	Chaohu (Pingdingshan)	Chaohu, Anhui	Kuhfeng Fm.	Wang and Qi (1995); Wu et al. (2017); Zhang et al. (2019a); Zhang et al. (2019b)	Conodont; Radiolarian
6	Yangongtang	Jingxian, Anhui	Kuhfeng Fm.	Wang and Qi (1995); Du et al. (2010)	Radiolarian
7	Yashan	Nanling, Anhui	Kuhfeng Fm.	Luo et al. (1997); Du et al. (2010)	Radiolarian; Fusulinid
8	Tieqiao	Laibin, Guangxi	Maokou Fm.	Sha et al. (1990); Mei et al. (1999); Wang and Sugiyama (2001); Chen et al. (2009); Yao et al. (2012); Sun et al. (2017)	Conodont
9	Yanghe	Huaying, Sichuan	Maokou Fm.	Liu et al. (2016a)	Conodont
10	Maoershan/ Tianfengping	Enshi, Hubei	Maokou Fm.	Zhang et al. (2007)	Conodont

11	Shangsi	Guangyuan, Sichuan	Maokou Fm.	Lai et al. (2008); Sun et al. (2008)	Conodont
12	Nashui	Luodian, Guizhou	Maokou Fm.	Shi et al. (2000); Henderson and Mei (2003); Meng et al. (2018)	Conodont
13	Xiongjiachang	Zhijin, Guizhou	Maokou Fm.	Wignall et al. (2009a); Sun et al. (2010)	Conodont
14	Chaotian	Guangyuan, Sichuan	Maokou Fm.	Lai et al. (2008); Sun et al. (2008); Saitoh et al. (2013a); Saitoh et al. (2013b)	Conodont
15	Sancha	Zunyi, Guizhou	Maokou Fm.	This study	Conodont
16	Baiguowan	Zunyi, Guizhou	Maokou Fm.	This study	Conodont
17	Sangshuwan	Renhuai, Guizhou	Maokou Fm.	This study	Conodont
18	Shangji	Zunyi, Guizhou	Maokou Fm.	This study	Conodont
19	Pianpozhai	Shuicheng, Guizhou	Maokou Fm.	This study	Conodont
20	Yongxing	Chenzhou, Hunan	Maokou Fm.	Huang and Li (1992)	Conodont
21	Cihua	Yichun, Jiangxi	Nangang Fm.	Zeng et al. (2010)	Conodont
22	Maocaojie	Jianshi, Hubei	Kuhfeng Fm.	Shi et al. (2016)	Conodont
23	Nanjiang	Bazhong, Sichuan	Maokou Fm.	Mei et al. (1994)	Conodont
24	Luojiachong	Zhongming, Anhui	Kuhfeng Fm.	Wang (1993)	Conodont
25	Subang	Longyan, Fujian	Wenbishan Fm.	Li (2008)	Fusulinid
26	Xinfeng	Ganzhou, Jiangxi	Wenbishan Fm.	Liu (2008)	Fusulinid
27	Lengwu	Tonglu, Zhejiang	Dingjiashan Fm.	Wang (1993); Liu et al. (2016b)	Conodont

28	Chenxi	Huaihua, Hunan	Maokou Fm.	Mu et al. (1997)	No fossil record
29	Qixingjie	Lianyuan, Hunan	Maokou Fm.	Wang (1995b)	No fossil record
30	Heshan	Laibin, Guangxi	Maokou Fm.	Yu et al. (2016)	No fossil record
31	Liangshan	Hanzhong, Shaanxi	Maokou Fm.	Wang (1978); Kong (2011); Cao et al. (2018)	Conodont

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