Coupled sedimentary and $\delta^{13}\text{C}$ records of late Mississippian platform-to-slope successions from South China: Insight into $\delta^{13}\text{C}$ chemostratigraphy

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Abstract

Variability in stratigraphic accumulation rates and distribution of stratal hiatuses along with strong endemism of index fossils hinder regional to global stratigraphic correlation of the Visean-Serpukhovian (V–S) boundary interval (late Mississippian) and thus geological inferences regarding the onset of the late Paleozoic ice age. Here we integrate high-resolution $\delta^{13}\text{C}$ time series with detailed sedimentary facies analysis of late

Visean-early Serpukhovian (V-S) carbonate platform-to-slope successions from South China to evaluate the influences of local depositional and diagenetic processes on $\delta^{13}\text{C}_{\text{carb}}$ and to address the correlation issue.

Analysis of 11 sedimentary facies from 5 outcrop sections indicates a restricted platform setting (Yashui section) dominated by bioclastic wacke-packstone to lime mudstone in which paleokarst developed, and contemporaneous carbonate slope settings dominated by thin-bedded lime mudstones intercalated with slumps and calciturbidites. Based on vertical facies assemblages, three depositional units are recognized, recording a significant sea-level drawdown across the V-S boundary. Multiple negative $\delta^{13}\text{C}$ excursions ($>1\text{‰}$) can be correlated across the V-S boundary interval in several slope sections (Naqing, Luokun, and Narao sections). Variability in the V-S boundary $\delta^{13}\text{C}$ record in some sections is interpreted to record truncation physically by submarine erosion by slumps or chemically during karstification. A long-term decrease in $\delta^{13}\text{C}$ values through the Serpukhovian of the Yashui section likely records local influences on carbon cycling in the restricted platform setting. This negative $\delta^{13}\text{C}$ trend and associated depositional facies at the Yashui section can be correlated to the Arrow Canyon section, USA, which, together with other coeval global sedimentary and geochemical records, indicate a widespread eustatic drawdown in the late Visean with initial buildup of Gondwanan ice sheets. We conclude that integrated sedimentary facies analysis and $\delta^{13}\text{C}$ chemostratigraphy can be used for stratigraphic correlation when interpreted within a well-constrained sedimentary and carbon-isotope regional framework.

Keywords: carbon isotopes of carbonates, sedimentary facies, stratigraphic correlation,
Visean, Serpukhovian, late Paleozoic ice age, karstification

1. Introduction

In addition to being an important proxy for paleoclimate reconstruction and paleoceanographic perturbation (Kump and Arthur, 1999; Montañez et al., 2000, 2007; Jiang et al., 2007), carbon isotopic compositions ($\delta^{13}\text{C}$) has been widely used for marine stratigraphic correlation ranging from the Precambrian to Quaternary (Saltzman and Thomas, 2012). However, marine carbonate $\delta^{13}\text{C}$ ($\delta^{13}\text{C}_{\text{carb}}$) does not necessarily reflect the global $\delta^{13}\text{C}$ of seawater given the potential influence of local to regional factors including compositional/mineralogical variability, water mass differences, and diagenesis among other processes (Patterson and Walter, 1994; Holmden et al., 1998; Swart and Eberli, 2005; Batt et al., 2007; Swart, 2008, 2015; Grotzinger et al., 2011; Buggisch et al., 2011; Dyer and Maloof, 2015). It is for these reasons that careful petrographic screening and microdrilling are necessary to evaluate the reliability of measured $\delta^{13}\text{C}_{\text{carb}}$ (Batt et al., 2007; Bishop et al., 2009; Brand et al., 2012; Metzger and Fike, 2013; Swart, 2015) coupled with interpretation of the $\delta^{13}\text{C}_{\text{carb}}$ values within a facies and depositional environment context (Swart, 2008, 2015; Saltzman and Thomas, 2012).

The application of $\delta^{13}\text{C}_{\text{carb}}$ excursions for stratigraphic correlation has been questioned given the influence of non-global processes. The correlation of $\delta^{13}\text{C}_{\text{carb}}$ excursions among sections, which are poorly constrained geochronologically, by

alignment of lithologic units can be misleading given the diachronous nature of facies (Dyer and Maloof, 2015). On the other hand, where biostratigraphic constraints exist they can be used to variable levels of success for $\delta^{13}\text{C}$ chemostratigraphic correlation and (Glumac and Spivak-Birndorf, 2002; Chen et al., 2012; Qie et al., 2015). An integrated sedimentologic, stratigraphic, and geochemical approach has proven to provide robust constraints on paleoenvironmental conditions and the nature of C isotopic variation within and between stratigraphic successions (e.g., Jiang et al., 2007; Bishop et al., 2009; Qie et al., 2015).

The Visean-Serpukhovian (V-S) boundary interval (late Mississippian) has been proposed to record the onset of the late Paleozoic ice age (LPIA) on the basis of stratigraphic and isotopic evidence (Smith and Read, 2000; Fielding et al., 2008; Bishop et al., 2009; Barham et al., 2012; Montañez and Poulsen, 2013; Fielding and Frank, 2015). A better understanding of the nature and timing of the onset of the LPIA is largely hindered by the incomplete nature of stratigraphic successions that were deposited during glacioeustatic drawdown (Richards & Task Group, 2003; Kabanov et al., 2013). Global stratigraphic correlation of this interval is further challenged by the strong endemism of index fossils, including conodonts, foraminifers, and ammonoids (Bishop et al., 2009; Groves et al., 2012; Cózar and Somerville, 2015). To that end, carbonate slope successions, which are interpreted to record near continuous sedimentation, have become the main focus for the effort to define the V-S GSSP (Global Boundary Stratotype Section and Point). The Naqing section in South China (Fig. 1) has been selected as one of two

potential candidates (Groves et al., 2012; Qi et al., 2014b). Bed-by-bed, high-resolution conodont biostratigraphy of the Naqing and other slope sections (e.g., Dianzishang, Luokun, and Narao sections) from South China (Qi et al., 2014a, 2014b; Wang et al., 2014) provides a robust framework for global correlation. Correlation of the slope successions with the contemporaneous shallow-water platform carbonates in South China and elsewhere, however, remains poorly constrained due to the strong endemism of index fossils (Groves et al., 2012).

Here we present high-resolution $\delta^{13}\text{C}$ time-series of bulk rock samples and petrographically-screened, microdrilled samples coupled with detailed sedimentary facies analysis to address this stratigraphic-correlation dilemma for the V-S boundary interval. We document coupled changes in depositional facies and isotopic compositions across a carbonate platform-to-slope transect. The South China records are compared to the Arrow Canyon, USA succession to evaluate the potential for global stratigraphic correlation and the reconstruction of climatic and sea-level conditions across the V-S boundary interval.

2. Geological settings

During the Mississippian, the South China Block was located in the northeast Paleotethys Ocean near the paleoequator ($\sim 0\text{--}15^\circ\text{S}$) (Fig. 1). The block consists of the Yangtze and Cathaysia blocks that were merged towards the end of the early Paleozoic (Mei et al., 2005). The southwestern part of the South China Block was subdivided into several paleogeographic elements during the Mississippian, including the Yangtze Land,

Dian-Qian-Gui-Xiang Platform, and Qian-Gui Basin (Jiao et al., 2003). The Qian-Gui Basin, which was inherited from an active Devonian rifted basin, was tectonically quiescent by the late Mississippian (Li et al., 1997; Qie and Wang, 2012). Thick carbonate deposits accumulated on the relatively stable settings of the Qian-Gui Basin and Dian-Qian-Gui-Xiang Platform (Jiao et al., 2003). Outcrop sections (Naqing, Luokun, Narao, Dianshishang, and Yashui sections) used in this study are from the Qian-Gui Basin and Dian-Qian-Gui-Xiang Platform (Fig. 1), which are located in the middle and south of the Guizhou Province.

3. Methods

Five outcrop sections that were previously measured by colleagues as part of the effort to define the GSSPs for the Serpukhovian, Moscovian, Kasimovian, and Gzhelian stages of the Carboniferous (Fig. 2) were utilized in this study to define a platform-to-slope transect. Bed-by-bed sedimentologic logging and description (Fig. 3) were carried out using a metric tape and the aluminum spikes with metric marks that were previously placed every meter by colleagues. This paper focuses on a ~20-m-thick interval that straddles the V-S boundary.

In total, 434 carbon isotope samples and 343 sedimentologic samples were collected from the five sections. Hand samples were slabbed, polished, scanned, and thin-sectioned in order to examine detailed sedimentary and diagenetic features. Fine-grained, micritic sediment (50–150 μg) was drilled from the fresh surfaces of samples using a dental drill

with stainless-steel bits for bulk carbonate analysis. 369 bulk carbonate samples from the slope sections were analyzed on a Kiel IV carbonate device and a MAT 253 mass spectrometer in the State Key Laboratory of Palaeobiology and Stratigraphy at Nanjing Institute of Geology and Palaeontology.

In order to evaluate the reliability of $\delta^{13}\text{C}$ values from bulk carbonates, 87 samples from slope sections and 65 from the platform Yashui section were thin-sectioned and stained with Alizarin Red-S and Potassium Ferricyanide (Dickson, 1965). Target areas for microsampling and analysis (e.g., well-preserved micrite, diagenetic cements, and brachiopods) were identified using transmitted and cathodoluminescent microscopy. Pink-staining (i.e., low $[\text{Fe}^{2+}]$), non/dully luminescent micrite was microdrilled from thick sections (70–100 μm thick) using a fully automated Merchantek microdrilling system with ~ 20 μm spatial resolution. Powders (50–100 μg) were collected and roasted for 30 min at 375 $^{\circ}\text{C}$ *in vacuo* in order to remove organic volatiles. In total, 172 (87 from slope sections and 85 from platform Yashui section) microsampled powders were analyzed using a Fisons Optima isotope ratio mass spectrometer with a 90 $^{\circ}\text{C}$ Isocarb common acid bath autocarbonate system in the Stan Margolis Stable Isotope Lab, University of California, Davis. All isotope values are reported relative to Pee Dee Belemnite (PDB) using standard delta notation; analytical precision for both $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ is better than $\pm 0.1\%$.

4. Platform-to-slope sedimentary facies

The Viséan-Serpukhovian boundary interval is well exposed in Guizhou Province, South China. The Naqing (NQ), Luokun (LK), Narao (NR), Dianzishang (DZS), and Yashui (YS) sections used in this study (Fig. 1) are well studied with respect to biostratigraphy (Groves et al., 2012; Qi et al., 2014a, 2014b) and define a platform-to-slope transect (Fig. 3). Based on detailed sedimentological study in the field, and on polished slabs and thin sections, 11 facies are defined, and described and interpreted below.

4.1. Carbonate platform succession

The carbonate platform succession is well exposed in the YS section (N 25° 59' 25", E 106° 44' 51") and crops out on a gentle slope along a paved road, near the town of Yashui, Huishui region, Guizhou Province (Fig. 1). The succession consists of peritidal to shallow-subtidal platform facies made up of carbonate mudstones, wacke-/packstones, calcimicrobe-rich facies, and a narrow interval (~0.8 m thick) of siliciclastic mudstone.

4.1.1. Mudstone (M)

Description: An interval of yellowish-gray mudstone with a homogeneous structure (Fig. 4A) occurs between 47.5 m and 48.25 m proximal to the V-S boundary (Fig. 4A). This mudstone contains (sub)vertically oriented caliche-coated columnar stromatolites in the lower and middle parts (Fig. 4A, B). Plant and metazoan fossils, and root-related structures (e.g., rootlets and rhizoliths) are absent. The stromatolites are a few cm in

width and up to 10 cm in height (Fig. 4B, C). Internally, the stromatolites consist of micritic coccoidal calcimicrobes and laminoid fenestrae filled with sparry calcite; blocky calcite spar is present in the center of some stromatolites (Fig. 4C, D, E, F). Alternating caliche and microbial laminae coat the outside of stromatolites (Fig. 4G, H). The $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ of the internal calcimicrobes and outer laminae overlap and are higher than those of the calcite cements (Fig. 4E). This mudstone interval overlies an irregular surface developed on the underlying lime mudstones with relief of up to 30 cm high (Fig. 4A); petrographically, this surface exhibits evidence of dissolution/corrosion.

Interpretation: Groves *et al.* (2012) interpreted the yellowish-gray mudstone as a paleosol and the stromatolites as “large, spherical, concentric oncolites or lithorelicts”. However, the homogeneous nature of the mudstone and lack of clear ped structures, horizonation, vertic features, and root-related redoximorphic mottling indicate it is not a well-developed *in situ* paleosol (cf. Tabor and Montañez, 2004). Furthermore, the lack of plant debris suggests the mudstone was not deposited in a fluvial or deltaic environment. We interpret the mudstone as an accumulation of insoluble residue created by limestone dissolution (karstification) during exposure as well as siliciclastic sediment washed into sub-surface solution caves. The columnar stromatolites, which show crusts of freshwater caliche, formed during deposition of the underlying, karsted limestone. Syndepositional marine cementation of the stromatolites led to selective dissolution of the less well cemented matrix during exposure to undersaturated meteoric fluids leaving the stromatolites as residual carbonate in the siliciclastic mudstone.

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190 *4.1.2. Lime mudstone (LM)*

191 *Description:* Lime mudstones are thin- to thick-bedded (a few to 90 cm in thickness),
192 homogenous to weakly laminated, and gray in color (Fig. 5A). Laminations are
193 horizontal to undulatory and irregular. Lime mudstone are composed of micrite matrix
194 with micro-peloids, intraclasts, and ostracod and gastropod shells showing geopetal
195 structures. The lime mudstone proximal to the V-S boundary (46.6–47.5 m; Fig. 4A)
196 consists of micro-peloids (<40 μm) in a micrite matrix and in some cases sporadic
197 microbial grains and subangular to rounded peloids within a pink-staining,
198 nonluminescent microspar matrix. A few caliche-coated microbial clots and ‘fingers’ are
199 embedded in the microspar (Fig. 5B, C); caliche crusts are corroded (Fig. 5D). The
200 mudstone in this interval contains siliciclastic filled fractures and cavities and is capped
201 by an irregular surface (Figs. 4A and 5A). Larger irregular cavities (a few cm in
202 dimension) are filled with brownish mudstone in which patches of physically abraded and
203 broken angular to subangular calcite crystals (40 μm to 5 mm long) occur (Fig. 5E, F).
204 *Microcodium* and alveolar structures are present (Fig. 5G, H).

205 *Interpretation:* The dominance of micrite and few low diversity ostracod and
206 gastropod shells indicate a depositional setting either too deep or too shallow for
207 metazoans to flourish. The light gray, organic-poor lime mudstone, lack of sedimentary
208 structures, and abundance of microbialites argue for a restricted shallow-water
209 environment. The upper part of the lime mudstone was modified by karstification. Lack

of both other subaerial exposure structures (e.g., desiccation cracks, tepee structures, rhizoliths) and syndepositional vadose diagenetic features (e.g., pendent or meniscus cement) suggest that subaerial exposure and karstification postdate cementation of lime mud sediment. Brownish structureless mudstone in the solution fractures and cavities formed as a residue during karstification of the limestones and from downward infiltration of overlying alluvial sediment. Corroded and angular, truncated calcite crystals are interpreted as forming by hydraulic fracturing of meteoric cements that partially lined unsaturated fractures and cavities (Scholle and Ulmer-Scholle, 2003). *Microcodium* structures are interpreted as calcification of mycorrhizal associations (Klappa, 1978), whereas alveolar structures result from calcification and encrustation of rootlet or fungal structures (Goldstein, 1988). Both *Microcodium* and alveolar structures are indicative of subaerial exposure and meteoric diagenesis (Ueno et al., 2013). Iron-poor, nonluminescent microspar cements are interpreted as meteoric phreatic cements, whereas caliche crusts formed on relict stromatolites as vadose precipitates analogous to speleothems. Their corrosion attests to the repeated episodes of under- and over-saturation with respect to low-Mg calcite as is characteristic of cave settings.

4.1.3. Thin-bedded wackestone (Wtb)

Description: This facies is dominated by thin-bedded (few cm to ~10 cm, and rarely up to 20 cm) bioclastic wackestones and less so packstones that are intercalated with argillaceous limestone beds or calcareous shales (a few cm) or seams (a few mm) of

irregular, undulatory character (Fig. 6A). Wacke-packstones contain rugose and tabulate coral fragments, foraminifers, brachiopods, echinoderms, and trilobites floating in a homogeneous micrite matrix (Fig. 6B). Solitary rugose corals, tabulate corals, and flattened *Giganoproductus* shells occur in some argillaceous limestone or shale layers.

Interpretation: An abundance of micrite matrix and lack of wave- or current-induced sedimentary structures suggest that this facies was deposited in a relatively low-energy environment (e.g., a restricted platform setting), most likely below fair-weather wave base and above storm wave base. A diversity of marine fossils indicate normal seawater salinity. Thin calcareous shale layers or shaly seams are indicative of intermittent influx of siliciclastic fines and/or temporary shutdown of the carbonate factory due to fluctuating environmental conditions.

4.1.4. Massive wackestone and packstone (W-Pm)

Description: Thick-bedded, massive or weakly laminated wackestones and packstones (Fig. 3) contain heterozoan fossils, including solitary, fasciculate, and massive rugose corals, tabulate corals, brachiopods, echinoderms, foraminifers and dasyclad algae (Fig. 6). Calcimicrobes commonly encrust fossil fragments such as corals and brachiopods (Fig. 6D). Calcispheres (Fig. 6E) and botryoidal and tubular calcimicrobes occur throughout the matrix (Fig. 6F). An interval rich in *Giganoproductus* brachiopod shells occurs at ~40.7 m, with most shells flat-lying and convex-up.

Interpretation: The occurrence of metazoan fossil fragments and abundant

calcimicrobes and calcispheres indicates deposition in the photic zone with sufficient light and nutrients. Microbes, most likely cyanobacteria, colonized the interspace among fossil fragments, and sometimes encrusted fossil fragments under suitable conditions. Abundant micrite and lack of wave- or current-induced structures indicate relatively low-energy conditions. Concentrations of brachiopod shells are interpreted as storm deposits.

4.1.5. Gastropod-ostracod micro-peloid packstone (Pgo)

Description: These packstones to grainstones, which are predominant between 41.60 and 42.10 m, consist of densely packed micro-peloids in a micritic matrix or calcite spar cement (Fig. 7A); matrix-dominated regions are homogeneous micrite. Peloids are subrounded to rounded, ~40–400 µm in diameter, and densely packed and exhibit normal grading within a bed. Calcispheres, brachiopods, ostracods, and gastropods with geopetal structures, as well as cement-filled, irregular fenestral structures are common (Fig. 7B).

Interpretation: The presence of calcispheres and metazoans, characteristic of restricted environments (i.e., ostracods and gastropods), indicates deposition in a peritidal or restricted platform environment (e.g., Reading, 1996; Pracht and Somerville, 2015). Micro-peloids may be calcified coccoidal cyanobacteria (Flügel, 2010). Irregular fenestral structures likely formed by decay of microbial mats or due to tidal pumping in the lower intertidal region (Groves and Read, 1978), corroborating a peritidal origin

4.1.6. *Stromatolitic-microbial grainstone (Gsm)*

Description: This facies, predominant between 43.95 and 44.70 m, is characterized by columnar/domal stromatolites and microbial grainstones (Fig. 7C). Stromatolites are composed mainly of coccoidal and hairy calcimicrobes (Fig. 7C), contain gastropod shells, some with geopetal structures (Fig. 7D), and some exhibit mm-scale microbialite columns or domes nucleating from their tops (Fig. 7E). Calcite cement-filled vugs are common (Fig. 7C). Some stromatolites occur within a microbial-peloidal grainstone matrix (Fig. 7C). Large (~0.5 mm) microbial grains and clots exhibit a clotted internal core and a dense micritic outer cortex, whereas smaller microbial grains are slightly clotted to homogeneous (Fig. 7E).

Interpretation: The predominance of calcimicrobes, associated calcite cement-filled vugs, and abundant gastropods suggests low-energy, shallow-water conditions. Alternation of laminated microbialites and randomly oriented microbial clots indicates fluctuating hydrodynamic conditions with laminated micritic stromatolites recording warm quiet water conditions and grainstone-hosted stromatolites with microbial clots/clasts indicative of higher-energy conditions

4.1.7. *Coral rudstone (Rc)*

Description: Coral rudstone beds, a few cm up to 15 cm thick (Fig. 3), consist of >30% debris of primarily fasciculate rugose corals (*Siphonodendron*) and sparse solitary rugose corals and brachiopod shells (Fig. 7F). Fasciculate coral debris are mostly

rod-shaped and rarely exhibit bifurcation. The matrix consists of either bioclastic wacke-
to packstone or yellowish calcareous shale (Fig. 5I).

Interpretation: Coral rudstones record high-energy conditions and are interpreted to
be storm deposits. Calcareous shale as the matrix of the rudstone most likely deposited
from suspension after storms.

4.2. Deep-water slope successions

Carbonate slope successions are continuously exposed in road-cut sections in the
Luodian and Anshun regions, Guizhou Province (Fig. 1). We studied four of these slope
successions: (1) NQ (N 25° 14' 40", E106° 29' 26"), (2) NR (N 25° 24' 39", E106° 36' 25"),
and (3) LK (N 25° 18' 30", E106° 34' 9") sections from the Luodian region, and (4) the
DZS section (N 25° 49' 18", E105° 52' 35") from the Anshun region (Fig. 1). These four
successions have been extensively studied for conodont biostratigraphy (Qi et al., 2014a,
2014b; Wang et al., 2014). Overall these successions are characterized by thin-bedded
lime mudstone intercalated with normal-graded intraclast-bearing wackestones to
packstones (Fig. 3).

4.2.1. Thin-bedded lime mudstones (LMtb)

Description: Lime mudstones are thin-bedded and mostly homogeneous (Fig. 8A),
or in some cases, weakly laminated, nodular, or conglomeratic. Scarce horizontal or
vertical burrows disrupt laminae (Fig. 8B). Lime mudstones are dominantly micritic with

dispersed fossil fragments such as foraminifers and crinoids. Thin-bedded lime mudstones can be intercalated with thin (a few mm to cm thick) black shales and wacke-packstone layers.

Interpretation: Scarce fossils and burrows and lack of current/wave-induced structures suggests that thin-bedded lime mudstones were deposited in low-energy, relatively deep-water conditions most likely below storm-wave base. Nodular and conglomeratic features formed during early burial compaction and diagenesis. Intercalated black shales are aeolian- or fluvial-transported, whereas intercalated wacke-packstones are interpreted as sediment gravity flow deposits.

4.2.2. Laminated wacke- to packstones (W-Pl)

Description: Thin-bedded bioclastic wacke- to packstones are weakly planar to cross laminated or massive (Fig. 8C). Laminated wacke- to packstones consist of alternating dark-gray micrite and light-gray bioclastic calcisilts. Fossil fragments are elongated parallel to laminations. Laminae show either distinct or gradational boundaries.

Interpretation: Bioclastic wacke- to packstones were deposited under moderate-energy conditions, partly reworked by currents. Bioclastic laminae overlain by micrite laminae are interpreted as distal storm/turbidite deposits.

4.2.3. Normal-graded packstone (Png)

Description: Normal-graded packstones (few cm to 70 cm thick) overlie lime

mudstones with sharp, irregular lower boundaries (Fig. 8D). Individual beds exhibit normal grading from intraclast-bearing, coarse packstones to fine-grained wacke-packstones that are weakly laminated (Fig. 8E). A few packstones exhibit reverse grading. Subangular to subrounded intraclasts (up to several cm long) of either lime mudstone or wacke-packstone occur within the lower part of some beds (Fig. 8D, E). Elongate intraclasts are aligned parallel to bedding. Packstones contain fossil fragments of crinoids, foraminifers, and brachiopods as well as peloids (Fig. 8D, F, G). Load cast structures occur locally at the base of packstone beds.

Interpretation: Normal-graded, intraclast-bearing coarse packstones and overlying parallel-laminated packstones are interpreted as the lower portions (A–C) of turbidite sequences, which form by decelerating sediment gravity flows carrying intraclasts and fossil fragments, as well as lime mud (Korn, 2008; Reijmer et al., 2012). Intraclasts were derived by erosion of underlying strata, whereas fossil fragments were derived from the carbonate platform. Rare reverse-graded packstones may be debris flow deposits.

4.2.4. *Slump deposits (SL)*

Description: An interval (47.5–51 m) of slumped thin-bedded carbonates at the DZS section (Figs. 3 and 9A) exhibits recumbent, meter-scale folds (Fig. 9B) and has a sharp and undulatory lower boundary and irregular upper boundary with lime mudstones (Fig. 9A). The slumped unit consists of coarse-grained carbonate and carbonate breccias. Breccia clasts are poorly sorted and polymictic, varying in lithology from lime mudstone

to packstone (Fig. 9C). Ammonoids shells and crinoids fragments are present.

Interpretation: Slump beds typically form on depositional slopes, where gravitational instability leads to sediment failure (e.g., C3zar et al., 2006; Oliveira et al., 2009). Slumping of coherent carbonates can be triggered by rapid sea-level changes or seismic activity (e.g., Hilbrecht, 1989; Spence and Tucker, 1997; Alsop and Marco, 2012; Mastrogiacomo et al., 2012). Occurrence of both breccias and folds indicate that the slumped sedimentary strata were partially lithified, experiencing plastic to brittle deformation (cf. Chen et al., 2011).

5. Depositional processes and sea-level changes

Facies analysis indicates that the V-S boundary interval at the YS section was deposited in a restricted carbonate platform setting. The composition of metazoan fossil fragments, abundant calcimicrobes, domal to columnar stromatolites, and fenestrae and paleokarst-related features collectively indicate relatively shallow-marine, photic-zone environments. Lack of wave- or current-induced sedimentary structures is indicative of low-energy conditions, although brachiopod shell concentrations and coral rudstones likely record short-lived storm reworking (e.g., Jin et al., 2013). Conversely, lithofacies from the NQ, LK, NR, and DZS sections, which are characterized by thin-bedded, homogeneous lime mudstone and rare occurrences of burrows and fossil fragments, suggest a relatively deep-water setting. Episodic occurrence of normal-graded, intraclast-bearing packstone and slump beds further indicate a carbonate slope setting,

which in turn suggest depositional water depths of up to several hundred meters (Bahamonde et al., 2004; C3zar et al., 2006; Korn, 2008; Reijmer et al., 2012).

Based on the facies stacking patterns within each succession and their correlation between the five sections, the V-S boundary interval is divided into three sequence-stratigraphic units, which are separated by a subaerial unconformity and its correlative conformity, and by a transgressive surface (Figs. 10 and 11), all of which record relative sea-level changes

5.1. Highstand systems tract (HST)

The highstand systems tract deposits of the carbonate platform (YS section) consist primarily of the massive wackestone to packstone facies in the lower part and lime mudstones in the upper part. The composition of the metazoan fossil fragments, the domal/columnar stromatolites, and the lime mudstones indicate overall shallow-marine conditions during sea-level highstand and a shallowing-upward trend through the unit (i.e., normal regression). The abundance of calcimicrobes in the late highstand deposits records the strong keep-up ability of the carbonate factory leading to shallow, and perhaps restricted water conditions, where microbes flourish (Fig. 11A). At the platform YS section, the highstand deposits are separated from the overlying unit by lime mudstones and domal/columnar stromatolites that exhibit extensive evidence of karstification and associated meteoric diagenesis (e.g., *Microcodium* and alveolar structures) formed during sea-level fall and lowstand (Fig. 10). The HST in the slope

sections is dominated by thin-bedded lime mudstones and wackestones, intercalated with subordinate amounts of fine- to coarse-grained packstones (Fig. 10). In slope sections, highstand deposits are separated from the overlying unit by a shift in sedimentologic facies to intraclast-bearing, normal-graded coarse-grained packstones (NQ, LK, and DZS sections), or to thickening-upward cycles of lime mudstones (NR section) (Fig. 10).

5.2. Lowstand systems tract (LST)

During sea-level lowstand, deposits (lime mudstones and domal/columnar stromatolites) of the previous high systems tract in the YS section were modified by meteoric diagenetic alteration, leading to extensive development of paleokarst features (breccias, corroded margins and clasts of mudstone, solution-enlarged fractures, and sparite-filled cavities and veins), and localized and weak soil development (*Microcodium*, and alveolar structures). Redeposited and broken calcite crystals, of meteoric origin (i.e., nonluminescent, non-ferroan) and caliche coatings indicate the development of vadose caverns (Fig. 11B) into which greenish-gray siliciclastic mudstones were deposited by freshwater flow. These features collectively indicate a significant hiatus (10^3 to 10^5 yr). In the slope sections, the LST is characterized by the prevalence of intraclast-bearing, normal-graded packstones. At the NR section, a few thickening-upward cycles of lime mudstone occur in the interval of packstones in the other slope sections. We interpret these cycles as recording local variation in sediment supply by sediment gravity flows (e.g., So et al., 2013). The discrepancy in depositional style between the NR and other

slope sections implies that autocyclic processes influenced local sediment supply from the carbonate platforms to the depositional slope regions, given that the carbonate sediment source to the slope during the Carboniferous was from the platform as calcareous plankton had not yet evolved (Riding, 1993). The thick interval (~47.5–51 m) of slumped lime mudstones in the DZS section is interpreted to have been triggered by a sea-level fall rather than by regional tectonic activity given the relatively stable tectonic setting of the Qian-Gui Basin. A rapid shift in pore-water pressure within the carbonate mudstones in response to sea-level fall could have induced sediment failure on the slope (cf. Hilbrecht, 1989; Spence and Tucker, 1997; Chen et al., 2011).

5.3. Transgressive systems tract (TST)

The transgressive systems tract deposits of the platform (YS section) consist of thin-bedded lime mudstones with micro-peloids and ostracod and gastropod shells, passing upward into wacke-packstones with abundant metazoan fossils, suggesting a return to normal salinity conditions upward through the TST (Fig. 11C). The predominance of thin-bedded lime mudstones and infrequent influx of sediment gravity flow deposits in the slope sections (Fig. 10) record the increasing water depths and decrease in platform carbonate source. A volcanic ash, proximal to the TST deposits can be confidently correlated between the NQ, LK, and NR sections, and is currently being dated. The maximum flooding surface between the TST and the overlying HST is not readily recognized, which we attribute to high accumulation rates of platform carbonate

and near-continuous deposition on the slope.

6. Carbon and Oxygen isotopic compositions

The $\delta^{13}\text{C}$ values of the slope carbonates (NQ, NR, and DZS sections) range from 1.2‰ to 4.0‰, with an average value of 3.1‰ (Figs. 10 and 12A). Slope carbonate $\delta^{18}\text{O}$ values range from -0.2‰ to -9.9‰, with an average value of -2.8‰. Notably, a population of matrix carbonate from one slope section (LK; Fig. 12A) have $\delta^{13}\text{C}$ values that span a much larger range (-4.2‰ to 3.6‰, average of 1.9‰), including anomalously low $\delta^{13}\text{C}$ values relative to the other slope successions. Average $\delta^{13}\text{C}_{\text{carb}}$ (3.3‰, n=204) for the NQ reported here are similar to values previously published (average of 3.2‰, n=24; Buggisch et al., 2011), but of much higher sampling resolution. Platform (YS) carbonates have the $\delta^{13}\text{C}$ values ranging from -1.2‰ to 3.6‰, with average of 1.7‰, but have $\delta^{18}\text{O}$ values that define a narrower range (-8.9‰ to -3.0‰, average of -4.7‰) (Fig. 12B). Microdrilled cements (including calcite spar filling fenestrae and solution cavities, and from fasciculate and massive corals) have overlapping to lower $\delta^{18}\text{O}$ values (ranging from -2.9 ‰ to -13.2‰) than the depositional carbonates from all sections, but similar $\delta^{13}\text{C}$ values (Fig. 12A), reflecting their carbonate rock-buffered compositions (Wynn and Read, 2007).

Overall, the upper Viséan interval is characterized by four negative excursions (of ~1–4‰) that can be correlated between the NQ, LK, and NR sections within the constraints of the conodont biostratigraphic framework (Fig. 10). Only three, small

magnitude negative C isotope excursions ($<1\text{‰}$) occur in the upper Viséan carbonates of the DZS section and might correlate to the lower three excursions in the other slope sections. A distinct negative $\delta^{13}\text{C}$ excursion (of 1.9‰) is recorded by carbonates immediately above the V-S boundary (17.92 m by Al-spike marks or 60.1 m by paint marks) in the NQ section. This $\delta^{13}\text{C}$ excursion, based on high-resolution conodont biostratigraphy (Qi et al. 2014b; Wang et al., 2014), can be correlated to negative $\delta^{13}\text{C}$ excursions recorded by carbonates immediately above the V-S boundary in the LK and NR sections (Fig. 10). A comparable $\delta^{13}\text{C}$ excursion is missing in the DZS carbonates proximal to the V-S boundary interval at 47.3 m.

Carbon isotopic compositions show more variable trends in the lower Serpukhovian interval among sections. In the NQ slope section, carbonate $\delta^{13}\text{C}$ values are near monotonic. Although the NR and DZS slope sections also show long-term invariance in $\delta^{13}\text{C}_{\text{carb}}$ through the lower Serpukhovian interval, two negative $\delta^{13}\text{C}$ excursions ($>1\text{‰}$ at 43.4 m and 46 m) occur in the NR section. In contrast, $\delta^{13}\text{C}_{\text{carb}}$ defines a long-term decreasing trend through the lower Serpukhovian interval of the LK section upon which are superimposed multiple and increasingly upward greater magnitude $\delta^{13}\text{C}$ excursions (Fig. 10).

For the shallow-water platform section (YS), negative excursions in the $\delta^{13}\text{C}_{\text{carb}}$ that occur within the upper Viséan interval might correlate to the aforementioned $\delta^{13}\text{C}_{\text{carb}}$ excursions of the slope sections. An abrupt shift at 48 m of $\sim 1.6\text{‰}$ from $\delta^{13}\text{C}$ values of $\sim 3\text{‰}$ to 1‰ occurs across the LST interval, which is ascribed to the stratigraphic discontinuity

due to erosion and karstification (cf. Chen et al., 2011; Myrow et al., 2015). This $\delta^{13}\text{C}$ negative excursion may correlate to the negative $\delta^{13}\text{C}_{\text{carb}}$ excursion proximal to the boundary between LST and TST in slope sections (Fig. 10), but not the one immediately above the V-S boundary given the sedimentary hiatus. The $\delta^{13}\text{C}_{\text{carb}}$ values define a gradual decrease through the lower Serpukhovian interval of the YS section with two superimposed small-scale negative shifts. Unlike the uppermost $\delta^{13}\text{C}_{\text{carb}}$ values of the slope sections, values in the uppermost portion of the studied YS interval define a large-scale positive excursion.

7. Discussion

7.1. Evaluation of the origin of the carbon isotope compositions

Carbon isotope excursions are typically interpreted as perturbations to global carbon cycling and in turn argued as a global stratigraphic correlation tool (Saltzman and Thomas, 2012), although the ‘whole ocean’ origin of some of these $\delta^{13}\text{C}$ excursions (e.g., the Ediacaran Shuram excursion) is still debated (Derry, 2010; Grotzinger et al., 2011; Swart and Kennedy, 2012).

Recent studies (e.g., Swart and Eberli, 2005; Swart, 2008; Swart and Kennedy, 2012; Oehlert and Swart, 2014) of the Quaternary to Neogene sediments reveal the potential for depositional and diagenetic processes, which are unrelated to global ocean-atmosphere C perturbation, to be the primary control on $\delta^{13}\text{C}_{\text{carb}}$ values. This potential influence is amplified in carbonates deposited in epicontinental seas (Swart, 2015) and in Neritan

Oceans, which lacked the carbonate buffering effect of open-ocean carbonate-shelled primary producers (Ridgwell and Zeebe, 2005). For example, $\delta^{13}\text{C}_{\text{carb}}$ can be facies dependent, in particular across a platform-to-basin transect, and influenced by mineral composition (calcite/aragonite), water-mass residence time (Algeo et al., 1992; Immenhauser et al., 2003; Panchuk et al., 2005; Swart and Eberli, 2005). Fluid-rock interaction during meteoric and burial diagenesis will overprint any depositional isotopic signatures (Allan and Matthews, 1977, 1982; Derry, 2010; Swart, 2015). Unraveling the influence of each of these factors on $\delta^{13}\text{C}_{\text{carb}}$ values of ancient deposits requires systematic petrographic and sedimentologic study coupled with a diagenetic evaluation of the carbonate textures and preservation potential (Montañez et al., 1996; 2000; Brand et al., 2012).

We evaluated the $\delta^{13}\text{C}_{\text{carb}}$ values of 87 pairs of bulk carbonate and petrographically screened microdrilled samples from the slope sections (NQ, LK, and NR sections) in order to address to what degree bulk carbonate $\delta^{13}\text{C}$ values are representative of the depositional conditions. Our assumption is that the $\delta^{13}\text{C}$ values of best-preserved micrites will record the local depositional water $\delta^{13}\text{C}$. This bulk carbonate-microdrilled carbonate comparison documents overlapping $\delta^{13}\text{C}$ values overall although microdrilled carbonate from the LK and NR slope successions are slightly (by $\sim 0.3\text{‰}$) lower than those of the corresponding bulk carbonate samples (Fig. 10, Table 1 and Supplementary Table). The lack of correlation between sedimentary facies and their average $\delta^{13}\text{C}_{\text{carb}}$ values (Fig. 10, Table 1, and Supplementary table) indicates a lack of lithology specific $\delta^{13}\text{C}$ values and

thus facies independence of measured $\delta^{13}\text{C}$ values in all studied sections.

Lack of co-variation between the $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values of the studied carbonates (cf. Cramer and Bostic, 2015; Qie et al., 2015), with the exception of the LK section (Fig. 12), and overlapping $\delta^{13}\text{C}$ values of this study with those of Mississippian slope successions from the Appalachian Basin, USA (Wynn and Read, 2007) and contemporaneous platform successions of the Antler foreland basin, USA (Batt et al., 2007) indicate that a large majority of $\delta^{13}\text{C}_{\text{carb}}$ values of the southern Chinese successions record depositional seawater compositions. Further corroboration is that a large subset of the slope carbonate $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values from the NQ, NR, and DZS successions overlap those of well-preserved time-equivalent brachiopods (Mii et al., 1999, 2001) (Fig. 12A). The exceptions to this interpretation that the $\delta^{13}\text{C}_{\text{carb}}$ values from the late Mississippian carbonates in South China faithfully record the depositional seawater values are a subset of samples from the LK slope and the YS platform successions (Fig. 12).

The C and O stable isotopic compositions of the LK carbonate samples defines an ‘inverted J’ distribution (Fig. 12A) that has long been interpreted to be indicative of alteration of marine carbonates during meteoric diagenesis (Lohmann, 1982, 1988). An ‘inverted J’ isotopic distribution reflects the initial and typically wholesale alteration of the $\delta^{18}\text{O}_{\text{carb}}$ values followed by increasing degrees of resetting of $\delta^{13}\text{C}_{\text{carb}}$ values by soil-C and ^{12}C -enriched vadose and phreatic waters. This interpretive model is not viable for the LK slope carbonates given that they were deposited in a slope environment comparable to that of the proximal NQ succession. Moreover, there is no sedimentologic or

petrographic evidence in the LK succession for subaerial exposure (i.e., rooting or development of karst fractures, caverns and intraformational caverns) or meteoric diagenesis (i.e., nonluminescent, nonferroan cements, aragonite molds). The origin of the subset of most negative $\delta^{13}\text{C}_{\text{carb}}$ and $\delta^{18}\text{O}_{\text{carb}}$ values is presently equivocal and could primarily record shifts in seawater $\delta^{13}\text{C}$ values. It is as feasible that the more negative isotopic values record burial diagenetic alteration of more permeable and/or diagenetically reactive carbonates by advecting organic-rich, basinal (hot) brines (cf. Derry, 2010).

Similarly, the large range in $\delta^{13}\text{C}_{\text{carb}}$ values ($\sim 4.5\text{‰}$) relative to the smaller spread in $\delta^{18}\text{O}_{\text{carb}}$ values ($\sim 2.7\text{‰}$) for the carbonate platform YS succession (Fig. 12B) could be interpreted as recording extensive meteoric diagenesis of these generally shallow-water carbonates. Evidence of meteoric diagenesis, however is lacking from the YS succession with the exception of the karstified interval immediately below the V-S boundary (Fig. 10). We cannot, however, completely discount a meteoric diagenetic origin for some of the negative $\delta^{13}\text{C}$ excursions in this succession given its shallow-water depositional environment and known short-term eustatic fluctuations at this time (Rygel et al., 2008; Bishop et al., 2009, 2010).

The $\delta^{13}\text{C}_{\text{carb}}$ values of the platform YS succession further define a long-term negative shift across the late Viséan up to the early Serpukhovian. The negative $\delta^{13}\text{C}$ excursion unlikely records a long-term shift in the global seawater DIC reservoir given that the slope successions in the study region (NQ, NR, and DZS), which would have

567 been bathed in more open ocean waters than the restricted carbonate platform, are
568 relatively invariant in $\delta^{13}\text{C}_{\text{carb}}$ throughout this interval. Rather, we suggest that the
569 long-term negative shift in $\delta^{13}\text{C}_{\text{carb}}$ through the Serpukhovian interval of the YS
570 succession records changes in local C cycling within the restricted platform setting.
571 Seawater $\delta^{13}\text{C}$ in these environments is influenced by the distribution and residence time
572 of water masses, contribution of ^{12}C -enriched inorganic C from microbial processes, and
573 influx of ^{12}C -enriched organic carbon from terrestrial and estuarine sources (e.g.,
574 Patterson and Walter, 1994; Immenhauser et al., 2003; Panchuk et al., 2005; Saltzman
575 and Thomas, 2012; Swart, 2015; Qie et al., 2015). Moreover, variation across a slope
576 transect in the degree of sediment contribution by shedding of carbonate sediment from
577 the platform could explain the differences in magnitude of the short-term excursions in
578 $\delta^{13}\text{C}_{\text{carb}}$ observed in all slope successions and the dampening of these excursions in the
579 most offshore section (NQ; Fig. 10) (Swart and Eberli, 2005).

580 The long-term $\delta^{13}\text{C}$ negative shift in the YS succession is similar to that recorded by
581 $\delta^{13}\text{C}$ values of the contemporaneous interval in the Arrow Canyon succession, USA (Fig.
582 13A; Bishop et al., 2009). The negative excursion recorded by bulk carbonates in the
583 Arrow Canyon succession has been interpreted as evidence of extensive meteoric
584 alteration (Bishop et al., 2009). Given that the trend begins below the first evidence of
585 exposure (i.e., the first paleosol) it is possible that the meteoric diagenetic trend is
586 superimposed on a long-term trend in seawater $\delta^{13}\text{C}$ to more negative values. The
587 relatively invariant trend in $\delta^{13}\text{C}_{\text{carb}}$ through the time-equivalent interval in the slope

successions argues against a seawater shift. Although the negative $\delta^{13}\text{C}$ trends can be correlated in the YS and Arrow Canyon sections we interpret the similarity in the long-term negative trends to record similar perturbations to processes that impact local carbon cycling (e.g., terrestrial influx of ^{12}C -enriched organic carbon). Both records record subsequent diagenetic alteration by freshwater driven by exposure with eustatic fall immediately prior to the V-S boundary.

7.2. Regional stratigraphic correlation and the utility of $\delta^{13}\text{C}$ chemostratigraphy in the V-S boundary interval of South China

Endemism of index fossils hampers regional- to global-scale stratigraphic correlation in Carboniferous-Permian successions of the South China region. For the V-S boundary interval of the South China Block, carbonate platform successions yield abundant and diverse foraminifers but are barren of conodonts. In contrast, carbonate slope successions contain abundant conodonts but only sparse foraminifers, which are present in intercalated calciturbidite beds (Groves et al., 2012). Both the foraminifers and conodonts exhibit patterns of endemism that limit confidence in intra- and inter-basin correlations. Ammonoids are absent from these South China successions with the exception of one report from a slump deposit in the DZS section. Therefore, it is not plausible to correlate successions across a platform to slope transect using biostratigraphy.

Carbon isotope chemostratigraphy as a correlation tool has been widely applied in

Carboniferous successions worldwide (e.g., Saltzman et al., 2003a, 2003b; Katz et al., 2007; Buggisch et al., 2011; Saltzman and Thomas, 2012). High-resolution $\delta^{13}\text{C}$ records of the V-S boundary interval, however, are lacking precluding their use as a correlation tool intra- or inter-basinally. The moderate negative $\delta^{13}\text{C}$ excursion defined immediately above the conodont-constrained V-S boundary in the studied South China successions (Fig. 10) may offer an independent marker for correlation. This negative $\delta^{13}\text{C}$ excursion, however, is missing from the DZS succession, the reason of which is discussed below.

Bed-by-bed, cm-scale conodont study has placed the V-S boundary in the studied slope sections (NQ, LK, and NR sections) based on the first appearance datum (FAD) of *Lochriea ziegleri* (Qi et al., 2014a, 2014b; Wang et al., 2014). The first occurrence of *L. ziegleri* in the DZS slope succession occurs in the lowermost portion of a slump deposit at 48 m (previously marked as 49 m in paint marks in Qi et al., 2014a) (Fig. 3). Notably, *Lochriea nodosa*, the ancestor of *L. ziegleri* that is found in the NQ, LK, and NR successions is absent in the DZS section (Qi et al., 2014a, 2014b). This suggests that the lower to middle part of the lineage has either not yet been discovered or is missing due to submarine erosion during slumping. Therefore, the occurrence of *L. ziegleri* at 48 m might record reworking of the contemporaneously older strata during slumping. Preliminary study of foraminifers indicates that the V-S boundary occurs below the interval of slumped carbonates at 47.3 m or lower, which could explain the lack of a negative $\delta^{13}\text{C}$ shift above the boundary.

Detailed foraminiferal study of the platform succession (YS) provisionally places

the V-S boundary at 49 m based on the appearance of “*tortula*-like” foraminifers (Groves et al., 2012), which first occurred with the first appearance of conodont *Lochriea Ziegleri* in the type section of Serpukhovian at Moscow Basin. However, the absolute placement of the boundary at the YS section is uncertain given that the interval (41.6–48.9 m) is mostly barren of foraminifers (Groves et al., 2012). Moreover, the paleokarst development and associated siliciclastic mudstone at 47.20–48.25 m indicate a major depositional hiatus, which further complicates placement of an accurate stage boundary at the YS section. In fact, the V-S stage boundary in the YS succession could be lower than at 49 m. Sedimentary facies analysis and correlation indicate that the onset of sea-level fall is recorded by carbonates below the V-S boundary in the slope successions (Fig. 10). This implies that carbonates, which were deposited across the V-S boundary are missing at the YS section, with the boundary now preserved in the fossil-barren siliciclastic mudstone that is interpreted as a paleocave filling deposit. In summary, the coupled facies analysis and $\delta^{13}\text{C}$ record for the YS succession suggest that the V-S boundary is at best placed at the base of the mudstone (i.e., the top surface of the karsted lime mudstone bed) at 47.5 m.

7.3. Implications for mid-Carboniferous glaciation

The late Paleozoic ice age was the longest lived and most widespread of the Phanerozoic and has been considered as an important analogue to the modern ice age (Raymond and Metz 2004; Isbell et al., 2008; Montañez and Poulsen, 2013). Its

651 glaciation history was characterized by repeated discrete glaciations interrupted by
652 interglacial, greenhouse conditions (Fielding et al., 2008; Isbell et al., 2012; Montañez
653 and Poulsen, 2013; Limarino et al., 2014). Although short-lived glacial events may have
654 occurred in the latest Devonian and early Mississippian of western Gondwana (Isaacson
655 et al., 2008) and southeastern Euramerica (Brezinski et al. 2008), the mid-Carboniferous
656 is conventionally considered to be the period of onset of major glaciation in the late
657 Paleozoic (Isbell et al., 2003; Fielding et al., 2008). This paradigm has been questioned
658 over the past decade.

659 The V-S boundary interval in ‘near-field’ successions of southern Gondwana has
660 been widely recognized as a major unconformity (e.g., Isbell et al., 2003, 2012; Fielding
661 et al., 2008; Gulbranson et al., 2010), most likely indicating extensive glaciation and
662 eustatic drawdown (e.g., “C1” event of Fielding et al., 2008; Fig. 13B). It has been
663 argued that the far-field stratigraphic proxies may be more complete and stratigraphically
664 continuous records of sea-level change, from which glacial events and their magnitude
665 can be inferred (e.g., Ross and Ross, 1985; Heckel et al., 2007; Belt et al., 2011; Eros et
666 al., 2012a, 2012b). Some far-field stratigraphic records suggest a late Viséan onset of
667 mid-Carboniferous glaciation (Smith and Read, 2000; Wright and Vanstone, 2001;
668 Bishop et al., 2009; Barham et al., 2012; Fielding and Frank, 2015), whereas others
669 suggest glacial onset at or above the V-S boundary interval (Mii et al., 2001; Buggisch et
670 al., 2008). Comparison of stratigraphic stacking patterns and $\delta^{13}\text{C}$ isotope stratigraphy of
671 the carbonate platform succession in South China (YS section) with the

mid-Carboniferous interval of the Arrow Canyon succession, which hosts the mid-Carboniferous GSSP (Lane et al., 1999), reveals strong similarity. It also indicates that the karstified interval in the YS section likely correlates with the oldest observed interval of paleosol development in the Arrow Canyon succession (Fig. 13A). If correct, then this global correlation corroborates previous evidence for onset of ice buildup across the V-S boundary. Integrated sedimentologic, stratigraphic, and $\delta^{13}\text{C}$ stratigraphic studies in contemporaneous successions in other basins are needed to further test this hypothesis.

8. Conclusions

The Viséan-Serpukhovian boundary succession is well exposed in Guizhou Province, South China, including well biostratigraphically constrained and exposed slope successions. Detailed sedimentary facies analysis of the study area indicates a restricted platform to slope transect, which defines three sequence-stratigraphic units that record significant sea-level drawdown across the V-S boundary, interpreted to record onset of ice buildup in the latest Viséan. High-resolution $\delta^{13}\text{C}$ chemostratigraphy indicates that the slope sections (NQ, LK, and NR) can be well correlated using short-lived $\delta^{13}\text{C}$ negative excursions proximal to the V-S boundary. Absence of these excursions in the DZS slope succession likely indicates missing succession due to submarine erosion by slumping and further explains the missing lower lineage of the conodont *Lochriea ziegleri*. Conversely, the nature of the $\delta^{13}\text{C}$ excursions across the V-S boundary interval in the restricted platform succession (YS), together with paleokarst development, indicates a significant

hiatus in the V-S boundary interval. A long-term negative $\delta^{13}\text{C}$ trend through the lower Serpukhovian (48.55–58.5 m) of the platform succession (YS), as well as in one slope succession (LK), is interpreted to record local carbon cycling rather than a whole-ocean carbon isotopic shift despite similarity in trend with the time-equivalent $\delta^{13}\text{C}$ record of Arrow Canyon, USA. This suggests correlation of $\delta^{13}\text{C}$ records within and between basins is best carried out when done with consideration of local, regional, and global factors that can influence $\delta^{13}\text{C}_{\text{carb}}$ values.

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Figure caption

Fig. 1. Paleogeographic map of the study area in Guizhou Province, South China (modified after Jiao et al., 2003). The red box in the inset Mississippian global paleogeographic map (after <http://jan.ucc.nau.edu/~rcb7/RCB.html>, courtesy of Ron Blakey) indicates the study area in South China Block. YS=Yashui section; DZS=Dianzishang section; NR=Narao section; LK=Luokun section; NQ=Naqing

section.

Fig. 2. Carboniferous regional and international chronostratigraphy (modified from Wang and Jin, 2005; International Chronostratigraphic Chart, 2015). Shaded area is the study interval. Dashed lines indicate Carboniferous stages lacking GSSPs.

Fig. 3. Correlation of stratigraphic successions and detailed facies distributions (scale of 1:50) for the NQ, LK, NR, DZS, and YS sections. The datum is the currently defined Viséan-Serpukhovian boundary based on conodont and foraminifera biostratigraphy (Groves et al., 2012; Qi et al., 2014a, 2014b; Wang et al., 2014). For facies code, see text.

Fig. 4. Sedimentologic features and context of stromatolites of the paleokarst interval at the Yashui section. A. Outcrop relationships of the karsted lime mudstone (irregular contact shown by dashed yellow line), overlying yellowish-gray siliciclastic mudstone, and capping thin-bedded lime mudstones. The Viséan-Serpukhovian boundary was previously defined as the first occurrence of “*tortula-like*” foraminifers at 49 m. B. Close-up of (sub)vertically orientated, caliche-coated columnar stromatolites (arrows), showing concentric structures in an oblique/transverse surface. Longitudinal (C) and transverse (D) views of a stromatolite. Both C and D show fenestrae structures and blocky calcite cement in the stromatolite core, and laminated overgrowth structures on the outer part. E. Scanned thin section of a longitudinal slab, showing $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values of different components (back Sharpie circles/squares 1 & 2: outer microbial laminae;

3, 4 & 5: caliche laminae; 6: coccoidal calcimicrobes; 7: blocky calcite cement; yellow rectangular boxes indicate the microdrilling raster area). F. Coccoidal calcimicrobes floating in blocky calcite cement. G. Caliche laminae between the inner region of coccoidal calcimicrobes and an outer layer of filamentous calcimicrobes. H. Close-up of outer layer of filamentous calcimicrobes.

Fig. 5. Paleokarst-related features at the Yashui section. A. Karsted lime mudstone below the siliciclastic mudstone. B. Slab from the basal part of this lime mudstone (~46.8 m), containing two caliche-coated microbial clots. C. Scanned thin section from part of (B), showing $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values of different components (back Sharpie circles 1: intraclast/peloid-bearing micro-sparite; 2, 3 & 4: filamentous calcimicrobes; 5: blocky calcite cement). D. Corroded caliche coating on microbial clot and matrix of peloid/intraclast-bearing micro-sparite. E. Scanned thin section of the karsted lime mudstone at ~47 m, showing $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values of different components (back Sharpie circles 1 & 2: homogeneous micrite; 3: transported calcite crystals; 4: *Microcodium*). F. Angular to sub-rounded calcite crystals in mudstone matrix. G. *Microcodium* structures (arrows). H. Alveolar structures (arrows).

Fig. 6. Bioclastic wacke-packstone facies at the Yashui section. A. Thin-bedded wackestone intercalated with shale (lower half). B. Foraminifer-bearing bioclastic wackestone with homogeneous micrite matrix. C. “Spiny” brachiopod, possible belonging to productids, encrusted by calcimicrobes. D. Rugose coral encrusted by calcimicrobes. E. Bioclastic wacke-packstone with abundant calcispheres. F.

Abundant calcimicrobes in the matrix of a bioclastic wacke-packstone.

Fig. 7. Micro-peloid packstone, stromatolite, and coral rudstone facies at the Yashui section. A. Micropeloids within a micrite matrix and regions of calcite microspar cement in the micro-peloid packstone facies. B. Micro-peloid packstone containing gastropod shells with associated calcite spar. C. Polished slab of stromatolites with calcite-filled vugs, surrounded by grainstone matrix. D. Gastropod shells in stromatolites showing geopetal structures. E. Micro-stromatolites encrusting on the stromatolite laminae, overlain by microbial-peloidal grainstone. F. Fasciculate rugose coral rudstone with a few convex-up brachiopod shells (arrow)

Fig. 8. Slope facies. A. Thin-bedded, homogeneous lime mudstone, with light-color nodular chert in some beds. Hammer head for scale. B. Lime mudstone with discrete burrows disrupting laminae. C. Laminated wackestone to packstone. D. Intraclast-bearing, coarse-grained bioclastic packstone, overlying a brecciated lime mudstone via a sharp boundary (arrow). E. Intraclast-bearing, normal-graded packstone passing upward into parallel-laminated wackestone, interpreted as basal units of a Bouma turbidite sequence. F. Bioclastic packstone containing intraclasts, peloids, and fossil fragments (a foraminifer in the center). G. A fossil fragment in packstone showing geopetal structure.

Fig. 9. Slump deposit overlying the Viséan-Serpukhovian boundary at the Dianzishang (DZS) section. A. Outcrop photo of slump beds between thin-bedded deposits. Hammer and notebook for scale. B. Recumbent fold in slump deposit. C. Polished

slab of a limestone breccia with polymictic clasts and crinoid oscicles.

Fig. 10. Integrated sedimentary facies and $\delta^{13}\text{C}$ time-series across the Visean-Serpukhovian boundary interval. The $\delta^{13}\text{C}$ time series for the slope sections consist of bulk carbonate values (blue dots) and microdrilled micrite values (red squares) where present, whereas the $\delta^{13}\text{C}$ time series for Yashui section is plotted using only microdrilled micrite.

Fig. 11. Depositional model for the Visean-Serpukhovian (V-S) boundary interval in the study area. A. Sea-level highstand during the late Visean. HST: highstand systems tract. B. Sea-level lowstand during the V-S boundary interval. LST: lowstand systems tract. C. Sea-level rise during early Serpukhovian. TST: transgressive systems tract.

Fig. 12. Cross-plots of carbonate $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values separated by material/method of sampling for all of the data (A) and for the Yashui section (B). Values from the Luokun section and some of the Yashui data define an “inverted J” curve (Lohmann, 1982), whereas the remaining Yashui values define a steeper “inverted J” curve.

Fig. 13. Lithofacies and $\delta^{13}\text{C}$ time-series for the Yashui section (this study) and the Arrow Canyon succession, USA (Bishop et al., 2009), and global distribution of glaciation for the late Visean to early Serpukhovian (V-S). A. Comparison of the V-S boundary interval between the two regions, showing similarity in depositional sequences and paleosol development and a long-term negative trend in $\delta^{13}\text{C}$. M: mudstone, LM: lime mudstone, W: wackestone, P: packstone, G: grainstone, R: rudstone. B.

1113 Literature compilation of inferred timing of glaciation across the V-S boundary
1114 interval (modified from Bishop et al., 2009). Thin colored bands indicate minor
1115 glaciation and thick bands indicate major glaciation.

Table 1. Average $\delta^{13}\text{C}$ values of the various lithofacies, and micro-drilled micrite and the corresponding bulk rock samples from the NQ, LK, NR, and DZS sections, Guizhou Province, South China.

Average $\delta^{13}\text{C}$ (‰ PDB)	NQ	LK	NR	DZS
Lime mudstone	3.25 (n=64)	2.26 (n=26)	2.94 (n=51)	2.87 (n=51)
Wackestone	3.33 (n=33)	2.47 (n=9)	2.65 (n=2)	3.10 (n=16)
Packstone	3.28 (n=76)	2.10 (n=18)	2.88 (n=4)	3.16 (n=9)
Slump				2.70 (n=11)
Micro-drilled micrite	3.08 (n=31)	1.42 (n=32)	2.43 (n=24)	
Corresponding bulk rock	2.98 (n=31)	1.74 (n=32)	2.72 (n=24)	

Fig. 1

Fig. 2

International chronostratigraphy				Regional chronostratigraphy (stage)		
System	Sub-system	Series	Stage (East Europe)	China	North America	West Europe
Carboniferous	Pennsylvanian	Upper	Gzhelian 298.9	Xiaodushanian	Virgilian	Autunian
			303.7		Missourian	Stephanian
		Middle	Kasimovian 307.0	Dalaan	Desmoinesian	Westphalian
			Moscovian		Atokan	
			315.2	Huashibanian	Morrowan	Namurian (upper)
		Lower	Bashkirian	Luosuan		
	Mississippian	Upper	Serpukhovian 323.2	Dewuan	Chesterian	Namurian (lower)
		Middle	330.9	Shangsian	Meramecian	Visean
			Visean	Jiusian	Osagean	
			346.7	Tangbagouan	Kinderhookian	Tournaisian
		Lower	Tournaisian			
			358.9

Fig. 4

Fig. 5

Fig. 6

Fig. 7

Fig. 8

Fig. 9

Fig. 10