Paleo-environmental cyclicity in the Early Silurian Yangtze Sea (South China): Tectonic or glacio-eustatic control?

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ABSTRACT

The Yangtze Sea of the South China Craton experienced strong environmental perturbations during the first ~5-6 Myr of the Silurian. The Lower Silurian Lungmachi Formation in the Jiaoye-1 drillcore (Chongqing, China) records three sea-level cycles (stratigraphic sequences) characterized by rapid deepening at the base of each cycle followed by slow shallowing. Each deepening event was associated with shifts toward higher marine productivity, more intensely reducing bottomwater conditions, reduced watermass restriction, lower detrital input, and enhanced siliciclastic weathering intensity. We infer control by glacio-eustasy, in which deglacial sea-level rises and concurrent climatic warming triggered the observed environmental changes. Shallowing was associated with renewed glaciation and characterized by the opposite set of conditions. The results of this study thus provide evidence in support of continued continental glaciation during the Early Silurian, i.e., following the termination of the major end-Ordovician Hirnantian glaciation. The study core also exhibits a long-term shallowing trend through the entire Lower Silurian, the origin of which may have been tectono-epeirogenic uplift of the South China Craton during the Kwangsian Orogeny.

Keywords: eustasy; sea level; Hirnantian; redox; organic carbon; Kwangsian Orogeny

1. Introduction

The Ordovician-Silurian (O-S) transition was a time of global climatic and biotic perturbations. Cooling during the Middle to Late Ordovician culminated in a brief (~0.5-Myr)

but intense episode of continental glaciation during the Hirnantian, the final stage of the Ordovician (Brenchley et al., 2003; Chen et al., 2004; Zhang et al., 2009, 2010; Young et al., 2010; Gorjan et al., 2012; Dronov, 2013; Algeo et al., 2016), which was accompanied by a two-stage mass extinction eliminating 86 % of extant species (Jablonski, 1991; Rong et al., 2002, 2007; Chen et al., 2005; Fan et al., 2009; Melchin et al., 2013; Harper et al., 2014).

Global climate conditions ameliorated rapidly at the termination of the Hirnantian glaciation but remained somewhat disturbed during the Rhuddanian (early Early Silurian), which was characterized by higher-than-background rates of turnover among conodonts (Aldridge et al., 1993) and graptolites (Chen et al., 2000, 2005; Fan et al., 2013; Harper et al., 2014).

The climatic conditions of the Early Silurian remain under debate. Some workers have inferred global greenhouse conditions (Brenchley et al., 1994; Armstrong and Coe, 1997; Sheehan, 2001; Munnecke et al., 2010; Finlay et al., 2010) whereas others have suggested renewed continental glaciation, although less extensive than during the Hirnantian (Jeppsson and Calner, 2003; Melchin and Holmden, 2006; Cramer and Saltzman, 2007; Munnecke et al., 2010; Finnegan et al., 2011; Trotter et al., 2016). Many of the latter studies have provided evidence for three to four episodes of expansion and contraction of continental icesheets during the Early Silurian, yet there is little agreement among published sea-level curves for this interval (Fig. 1; Munnecke et al., 2010).

[Place Figure 1 here]

The Yangtze Platform of the South China Craton preserves a continuous record of

sedimentation through the O-S transition (Chen et al., 2004). A number of studies have reconstructed environmental conditions within the Yangtze Sea (Yan et al., 2009, 2010, 2012; Zhou et al., 2015; Li et al., in review; Liu et al., in review), although most have focused on the Upper Ordovician rather than the Lower Silurian. In the most thoroughly investigated sections, Wangjiawan and Nanbazi, only a few meters of the Lower Silurian Lungmachi Formation is exposed in outcrop, corresponding to the *Akidograptus ascensus* to *Coronograptus cyphus* graptolite zones of the Rhuddanian stage (Yan et al., 2009, 2010; Zhou et al., 2015). As a consequence, these studies have not recognized third-order (~1-3-Myr-long) cycles in this formation. In contrast, the Jiaoye-1 drillcore spans a much longer interval, through the *Stimulograptus sedgwichii* Zone, comprising the full Rhuddanian and Aeronian stages of the Lower Silurian. The Jiaoye-1 drillcore was studied recently by Chen-L et al. (2015), who analyzed conditions within the Early Silurian Yangtze Sea on the basis of petrographic data. However, the geochemistry of the Jiaoye-1 drillcore has not been investigated to date. Here, we report total organic carbon (TOC), major and trace element, $\delta^{13}\text{C}_{\text{org}}$, and pyrite framboid size data, as well as new graptolite biostratigraphic data, for a ~85-m interval of mostly Lower Silurian strata in the Jiaoye-1 drillcore. Our goals are (1) to reconstruct environmental changes within the Early Silurian Yangtze Sea, and (2) assess the tropical marine environmental record of possible climatic perturbations in the aftermath of the end-Ordovician Hirnantian glaciation.

2. Geological background

2.1. Location and paleogeographic setting

The South China Craton was located close to the paleo-equator, in a tectonically active region along the northwestern margin of Gondwana, during the Early Paleozoic (Fig. 2A) (Metcalf, 1994). During the Ediacaran to Early Ordovician, the Nanhua Basin (a failed intracratonic rift) opened between the Yangtze and Cathaysia blocks, producing a broad shelf on the northeastern (present-day southeastern) margin of the Yangtze Block (Fig. 2A; Liu and Xu, 1994; Macouin et al., 2004; Li et al., 2009; Zhang et al., 2013) (note: all further references to compass directions are paleo-orientations for the Late Ordovician, when South China was rotated $\sim 90^\circ$ counter-clockwise relative to its modern position). The Nanhua Basin began to close in the Middle Ordovician (~ 460 Ma) along the Wuyi-Yunkai orogenic belt, resulting in suturing of the Yangtze and Cathaysia blocks to form a unified South China Craton by the Early Silurian, an event known as the “Kwangsi Orogeny” (Wang, 1985; Wan and Zhu, 1990; Faure et al., 2009; Charvet et al., 2010; Li et al., 2010). As a consequence, the northeastern margin of the Yangtze Block was transformed from an extensional shelf margin to a compressional foreland basin, with continued tectonic activity until the Early Devonian (Li, 1998; Su et al., 2007; Huang et al., 2011).

[Place Figure 2 here]

The study area is located in the western part of the Yangtze Block, in the Jiaoshiba shale gas field, which is in Jiaoshi Town, Fuling District of Chongqing Municipality (Fig. 2B). The study material was collected from the Jiaoye-1 drillcore, which was acquired by the China Petroleum and Chemical Corporation in 2012 and is now curated by the Exploration Southern

Company of Sinopec. The Jiaoye-1 drillcore consists of 84.8 m of Upper Ordovician and Lower Silurian strata from 87.1 m of borehole (Fig. 3), representing a core recovery rate of ~97 %. This drillcore is from the crest of an anticline, and the intersected strata are nearly horizontal (Fig. 2B, inset). The thermal maturity of the study section is moderately high, as shown by *Ro* (vitrinite reflectance) values from 2.2 % to 3.1 % (mean 2.6 %), which correspond to the post-maturation stage of hydrocarbon generation (Table S1).

[Place Figure 3 here]

Semi-restriction of the shallow Yangtze Sea developed during the Late Ordovician to Early Silurian as a result of far-field stresses of the Kwangsian Orogeny, which caused relative uplift of the Chengdu and Hunan-Hubei submarine highs and Yunnan-Guizhou Oldland on the western Yangtze Block (Chen et al., 2004). The study site was located in the middle of the western Yangtze Sea, approximately equidistant from these uplifts (Fig. 2A). Beginning in the late Katian, the Yangtze Sea accumulated siliciclastics, primarily black shales of the Wufeng Formation, owing to a combination of weathering of uplifted orogens and restriction of the interior of the Yangtze Sea (Wang, 1985; Chen et al., 2004). A major eustatic fall during the Hirnantian glaciation (Brenchley, 1988; Chen et al., 2004) led to widespread deposition of calcareous mudstone and limestone of the Kuanyinchiao Bed (Zhang et al., 2016). A post-glacial transgression led to renewed deposition of black shales, now assigned to the Lungmachi (= Longmaxi in Pinyin) Formation of latest Ordovician to Early Silurian age.

2.2. Stratigraphy and sedimentology

The Wufeng Formation, which disconformably overlies limestones of the Upper Ordovician Linhsiang Formation, consists of 4.26 m of black carbonaceous shale and dark gray silty shale (Fig. 3). This unit is conformably overlain by the Kuanyinchiao Bed, which consists of ~0.1 m of light-colored argillaceous limestones. According to regional correlations, the Kuanyinchiao Bed is equivalent to the upper part of the *M. extraordinarius* Zone and the lower part of the *M. persculptus* Zone of the lower to middle Hirnantian Stage (Fan et al., 2009). It contains the Hirnantia Fauna, a cool-water shelly fauna that spread rapidly at the onset of the Hirnantian glaciation and disappeared abruptly at its termination (Rong et al., 2002; Zhan et al., 2010). The Lungmachi Formation, which conformably overlies the Kuanyinchiao Bed, consists of ≥ 79 m of dark gray to black silty shale (Fig. 3; note that the minimum thickness cited here represents the cored interval in the Jiaoye-1 drillcore). In the study core, we subdivided the Lungmachi Formation into three units (Unit 1, 2408.5-2390 m; Unit 2, 2390-2353 m; Unit 3, 2353-2330 m) that have sequence stratigraphic significance (see Sections 4.2 and 5.1).

The Wufeng and Lungmachi black shales contain many well-preserved graptolite specimens (Fig. 4, Appendix 1) that permit correlations to international biozonation schemes. On this basis, the *Dicellograptus complexus*, *Paraorthograptus pacificus*, and *Metabolograptus extraordinarius* biozones of late Katian to early Hirnantian age were recognized in the Wufeng Formation (Fig. 3). The lowermost 3.24 m of the Lungmachi Formation is assigned to the *M. persculptus* Zone of the late Hirnantian Stage. The remainder

of the Lungmachi Formation is of Early Silurian age and assigned to the *Akidograptus ascensus*-*Parakidograptus acuminatus*, *Cystograptus vesiculosus*, *Coronograptus cyphus* zones of the Rhuddanian Stage, and the *Demirastrites triangulatus*, *Lituigraptus convolutus*, and *Stimulograptus sedgwickii* zones of the Aeronian Stage (Fig. 3).

[Place Figure 4 here]

The 79-m-thick Lungmachi Formation in the Jiaoye1 core was deposited over a ~6-Myr interval (Cooper and Sadler, 2012), yielding an average sedimentation rate of ~13 m Myr⁻¹. A depth-age analysis based on graptolite zones (Fig. 3) suggests that the late Hirnantian and earliest Silurian (early Rhuddanian) portions of this formation had a lower sedimentation rate (~3.0 m Myr⁻¹) than its younger Silurian (late Rhuddanian-Aeronian) portions (~18 m Myr⁻¹). Similar estimates have been reported from correlative sections in South China (e.g., Fan et al., 2013).

3. Materials and methods

3.1. Samples

Two sets of samples were collected from the Jiaoye-1 drillcore with an average sampling resolution of 0.5 m. The first set consisted of 173 samples, including 7 from the Wufeng Formation, 3 from the Kuanyinchiao Bed, and 163 from the Lungmachi Formation. These samples were analyzed for TOC (total organic carbon), and a subset of 87 of these samples was analyzed for mineralogic composition. The second set consisted of 156 samples,

including 6 from the Wufeng Formation, 2 from the Kuanyinchiao Bed, and 148 from the Lungmachi Formation. These samples were analyzed for major and trace element concentrations and organic carbon isotopes. The major focus of this study is the Lungmachi Formation; the smaller numbers of samples collected from the Wufeng Formation and Kuanyinchiao Bed were intended to provide baseline values against which to evaluate geochemical variation in the Lungmachi Formation.

3.2. Analytical methods

For each sample, about 100 g of fresh rock was crushed to gravel-sized particles using a steel jaw crusher, and then was powdered to finer than 200 mesh size in an agate mortar. After crushing, each powdered sample was split into several fractions for geochemical and mineralogic analyses.

Total organic carbon analyses were carried out in the Experimental Research Center (ERC) of Wuxi Petroleum Geology Institution (Sinopec Petroleum Exploration and Development Research Institute). About 200 mg of powder was weighed and treated with 10% hydrochloric acid (HCl) at 60 °C to remove carbonate, and then washed with distilled water to remove remaining HCl. Afterwards, the residue was dried and analyzed using a Leco CS-200 elemental analyzer. Instrumental accuracy is ± 0.5 % of reported TOC values, based on replicate analyses of the Chinese national standard GB/T 19145-2003 (Zeng et al., 2013).

Mineralogic analyses were also carried out at ERC via X-ray diffraction (XRD) using a Panalytical X'Pert PRO MPD. Powdered samples were placed in copper holders and scanned using a D/max-3B X-ray diffractometer. Samples were scanned from 5° to 90° for

whole-rock composition, and 3° to 30° for clay-fraction composition, both at 2° min^{-1} with a step width of 0.02° , using $\text{CuK}\alpha$ radiation and a graphite monochromator. The accuracy of goniometer measurement is better than 0.02° based on replicate analyses of the Chinese national standard SY/T 6201-1996 (Zeng et al., 2013; Zhao et al., 2016).

Major element concentrations were analyzed at the State Key Laboratory of Geological Processes and Mineral Resources of China University of Geosciences, Wuhan. The oxides of each major element were determined by chemical titration methods. Taking SiO_2 as an example, powdered samples were fused with the fluxing agent Na_2CO_3 , leached with HCl , evaporated to a small volume of coacervated silicate with polyethylene oxide, filtered, heated, and the solid part weighed as M_1 . Then SiO_2 was extracted and removed as SiF_4 through dissolution in hot HF and H_2SO_4 , and the remaining solid part was weighed as M_2 . Residues were fused again with the fluxing agent $\text{K}_2\text{S}_2\text{O}_7$ and extracted using distilled water. The solution was combined with the previous HCl filtrate. After depolymerization, the residual SiO_2 in the filtrate was measured as M_3 using molybdenum blue spectrophotometry. Finally, total SiO_2 concentration was calculated as $(M_1 - M_2 + M_3)$. The results have an accuracy of $\pm 0.1\%$ of reported SiO_2 contents, based on replicate analyses of the Chinese national standard GB/T 14506.3-2010 (Yuan et al., 2016). The concentrations of the oxides of Al, Fe, Ca, Mg, Ti, P, Mn, K, and Na were determined by similar methods with a corresponding degree of accuracy ($\pm 0.1\%$) based on the Chinese national standard GB/T 14506 (Yuan et al., 2016).

Trace elemental compositions were analyzed using an Optimass 9500 inductively coupled plasma time-of-flight mass spectrometer (ICP-TOF-MS) at the State Key Laboratory of Biogeology and Environmental Geology of China University of Geosciences, Wuhan.

About 50 mg of powdered sample was weighed into a Teflon bomb, moistened with a few drops of ultrapure water, followed by addition of 1 ml HNO₃ and 1 ml HF. The sealed bomb was heated at 190 °C in an oven for >48 h. After cooling, the bomb was opened and evaporated at 115 °C to incipient dryness, followed by addition of 1 ml HNO₃ and evaporation again to incipient dryness. The resultant salt was re-dissolved by adding 3 ml of 30 % HNO₃ and resealed and heated in the bomb at 190 °C for >12 h. Finally, the concentrated solution was transferred to a polyethylene bottle and diluted to ~100 g (corresponding to a dilution factor of ~1:2000) with mixture of 2 % HNO₃ prior to ICP-MS analysis. Analytical precision and accuracy were evaluated by duplicate analyses of samples and three international reference standards (BCR-2, BHVO-2, and GSR-6). Analytical precision for trace element concentrations is better than ±5 %.

The $\delta^{13}\text{C}_{\text{org}}$ analyses were performed using a Finnigan MAT 252 mass spectrometer at the State Key Laboratory of Geological Processes and Mineral Resources of China University of Geosciences, Wuhan. Isotopic measurements were calibrated with the U.S. National Bureau of Standards standard NBS-19 ($\delta^{13}\text{C} = +1.95 \text{ ‰}$). The results are expressed in the standard delta notation with respect to the VPDB standard, and with an analytical standard deviation of ±0.03 ‰.

Pyrite framboid diameters were measured on SEM images using the free software “ImageJ”. For each sample, 5 to 10 SEM images with pyrite framboids were selected and the diameters of 20-70 pyrite framboids were measured.

3.3. Proxy calculations

To determine the enrichment degree of redox-sensitive trace elements for redox analysis, we calculated elemental enrichment factors (EFs) (Tribovillard et al., 2006):

$$EF_X = (X/Al)_{\text{sample}} / (X/Al)_{\text{AUCC}} \quad (1)$$

where X is the element of interest. Values of EF_X that are >1.0 and <1.0 indicate enrichment and depletion, respectively, of element X relative to its concentration in average upper continental crust (AUCC; cf. McLennan, 2001, his table 5).

To estimate the non-detrital fraction of a given element, the detrital fraction was subtracted from the total concentration of that element. The detrital fraction (detr) was estimated based on the concentration of Al (i.e., an element of dominantly detrital origin) and a detrital X/Al ratio determined from a X-versus-Al crossplot:

$$X_{\text{non-detr}} = X_{\text{sample}} - (Al_{\text{sample}} \times (X/Al)_{\text{detr}}) \quad (2)$$

We used this equation to calculate the non-detrital fractions of Ba and Cu, based on $(Ba/Al)_{\text{detr}}$ and $(Cu/Al)_{\text{detr}}$ ratios of 100 and 55, respectively. We interpret all non-detrital Ba or Cu as biogenic (bio) in origin and therefore of use as paleoproductivity proxies.

Non-clay SiO_2 was calculated as total SiO_2 minus clay-fraction SiO_2 . Clay-fraction SiO_2 was calculated from total Si and Al concentrations and XRD data concerning the relative abundances of various types of clay minerals in the sample (cf. Algeo et al., 2007). This calculation is based on the assumptions that all Al is present in clay minerals, and that TOT clay minerals (principally illite and smectite) have a molar Si:Al ratio twice as large as TOTO clay minerals (principally chlorite; note that T and O indicate tetrahedral and octahedral layers, respectively). First, total Al was apportioned between TOT and TOTO clay minerals:

$$Al_{TOT} = Al_{total} \times \%TOT \quad (3)$$

$$Al_{TOTO} = Al_{total} \times \%TOTO \quad (4)$$

where %TOT and %TOTO are the relative abundances of TOT and TOTO clay minerals (summing to 100 %) as determined from XRD analysis. Second, the amount of Si in TOT and TOTO clay minerals was calculated assuming molar Si:Al ratios of 2:1 and 1:1, respectively:

$$Si_{TOT} = Al_{TOT} \times 2 \times 28.1 / 27.0 \quad (5)$$

$$Si_{TOTO} = Al_{TOTO} \times 28.1 / 27.0 \quad (6)$$

where 28.1 g and 27.0 g are the molar atomic masses of Si and Al, respectively. Third, non-clay SiO_2 was calculated as the difference between total SiO_2 and clay-fraction SiO_2 :

$$SiO_{2(non-clay)} = SiO_{2(total)} - (Si_{TOT} + Si_{TOTO}) \times 60.1 / 28.1 \quad (7)$$

where 60.1 g is the molar atomic mass of SiO_2 . $SiO_{2(non-clay)}$ gives a maximum estimate of SiO_2 of biotic origin, i.e., $SiO_{2(bio)}$, although it should be noted that $SiO_{2(non-clay)}$ also includes SiO_2 in the form of detrital quartz and diagenetic chert. Petrographic analysis revealed the presence of abundant radiolarian tests and sponge spicules in the study samples (Fig. 5A-C), so it is reasonable to infer that much of the $SiO_{2(non-clay)}$ fraction is of biogenic origin. The presence of detrital quartz was assessed using QEMSCAN image analysis of samples from an adjacent drillcore (Pengye-1), which showed that quartz grains were abundant in the lower part of the study section but decreased rapidly upsection (Chen-L et al., 2015).

To evaluate climate and weathering conditions, we calculated the chemical index of

alteration (CIA*) (Nesbitt and Young, 1982; Price and Velbel, 2003):

$$\text{CIA}^* = \text{Al}_2\text{O}_3 / (\text{Al}_2\text{O}_3 + \text{Na}_2\text{O} + \text{K}_2\text{O}) \quad (8)$$

which is a molar ratio of the major-element oxides Al_2O_3 , Na_2O , and K_2O . CIA* is a modified version of the original CIA equation of Nesbitt and Young (1982) from which CaO is omitted. CaO can be meaningfully included in analyses of weathered material derived from igneous parent rock, but it is difficult to accurately determine and subtract carbonate-derived CaO in calcareous sedimentary rocks, making it preferable to leave CaO out of the index entirely.

Owing to differences in number and spacing of samples between the two sample sets, direct comparisons of geochemical values (i.e., in crossplots) were based on resampled profiles, in which values were calculated at 0.5-m intervals from 2330 m to 2410 m based on linear interpolation between the closest two measured sample points. The original and resampled data are both given in Table S1.

[Place Figure 5 here]

4. Results

4.1. Petrographic observations

The black shales of the Wufeng and Lungmachi formations contain abundant well-preserved graptolites (Fig. 4) as well as frequent sponge spicules and radiolarian tests (Fig. 5A-C). The black carbonaceous mudstone is mostly homogenous and without visible fabric, although silty laminae are present locally (Fig. 5D). The silty shale intervals are characterized by laminations reflecting variation in grain size or organic matter content (Fig.

5E). Bioturbation structures are found sporadically in the upper part of the study section (~2326-2332 m), with an ichnofabric index mostly of 2 to 4 on the scale of Droser and Bottjer (1993) (Fig. 5F).

As seen using scanning electron microscopy (SEM), most of the silt-sized particles in the Wufeng and Lungmachi formations are composed of quartz or chert grains. Organic particles are commonly dispersed among quartz or pyrite grains, and there are some pores within the organic matter (Fig. 6A-C). These pores occurred due to the burial and maturation of the organic material (Jarvie et al., 2007; Roger and Neal, 2011). Blocky authigenic pyrite crystals are found in association with organic matter in some samples (Fig. 6D), and pyrite framboids are common in all samples that were examined petrographically (e.g., Fig. 6C, E-F).

[Place Figure 6 here]

4.2. Total organic carbon

The total organic carbon (TOC) content of the study samples ranges from 0.55 % to 5.9 % (mean 2.5%) (Table S1). The study section can be divided into three units based on the TOC profile: Unit 1 (2408.5-2390 m), Unit 2 (2390-2353 m), and Unit 3 (2353-2330 m) (Fig. 7A). Both the range and mean of TOC values decrease upsection for successive units: 1.3-5.5 % (mean 3.6 ± 0.9 %) in Unit 1, 0.9-4.0 % (mean 2.2 ± 0.8 %) in Unit 2, and 0.5-3.3 % (mean 1.7 ± 0.7 %) in Unit 3. Within each unit, TOC content shows a gradual decrease upsection, followed by a sharp increase at the base of the overlying unit. This cyclic variation has sequence stratigraphic significance (see Section 5.1).

[Place Figure 7 here]

4.3. Mineralogy

Total SiO₂ content, as determined by XRD, ranges from 18 % to 71 % with a mean of 37±10 % (Table S1). The SiO_{2(total)} profile exhibits a broadly similar pattern to the TOC profile, exhibiting both quasi-cyclic variation within units as well as a long-term trend toward lower concentrations upsection (Fig. 7C). The range and mean of SiO_{2(total)} content decrease upsection for successive units: 31-57 % (mean 44 %) for Unit 1, 24-52 % (mean 35 %) for Unit 2, and 18-37 % (mean 28 %) for Unit 3. SiO_{2(total)} content as determined by XRD corresponds closely to SiO_{2(non-clay)} concentration, as determined from Equation 7 (Fig. 7D).

The main clay minerals in the study samples are mixed-layer illite/smectite, illite, and chlorite (Table S1). Smectite and kaolinite were not detected in measurable amounts. Mixed-layer illite/smectite is the most abundant clay mineral, ranging from 25 % to 85 % (mean 54 %), with a decreasing trend upsection. Illite content ranges from 12 % to 68 % (mean 39 %), with an increasing trend upsection. Chlorite content ranges from 1 % to 20 % (mean 6 %) and shows a fluctuating pattern with peak values in the middle of the study section (Table S1).

Pyrite framboids occur in samples throughout the study section, with a size range from ~1 µm to 6 µm (mean 4.3±1.5 µm). However, both the mean diameter and standard deviation (SD) of framboids vary subtly between units: 4.2±1.2 µm in Unit 1, 4.5±1.3 µm in Unit 2, and 4.2±1.7 µm in Unit 3 (Fig. 8A). Pyrite framboids occur mainly as densely packed

spherical aggregates (Fig. 6C, F). In addition to framboids, euhedral and irregular pyrite crystals are found associated with the clay-mineral matrix or concentrated around organic particles (Fig. 6E).

[Place Figure 8 here]

4.4. Isotopic and elemental geochemistry

The study core yielded $\delta^{13}\text{C}_{\text{org}}$ values ranging from -31.8‰ to -26.5‰ (mean $-29.5 \pm 0.62\text{‰}$; Table S2). Units 1-2 exhibit a weak trend upsection toward heavier $\delta^{13}\text{C}_{\text{org}}$ values but with considerable sample-to-sample variability (Fig. 7G). Unit 3 exhibits obvious fluctuations, a subtle decrease followed by increased variation and significantly heavier values, followed by a decrease near the top. However, the ranges of $\delta^{13}\text{C}_{\text{org}}$ values between units overlap broadly, and the means are only modestly different: -30.4‰ to -29.5‰ (mean -30.0‰) for Unit 1, -31.8‰ to -28.8‰ (mean -29.3‰) for Unit 2, and -30.0‰ to -26.5‰ (mean -29.2‰) for Unit 3.

Aluminum oxide (Al_2O_3) concentrations, which serve as a proxy for total clay content, range from 4.1% to 22.0% (mean $13.1 \pm 3.4\%$) in the study section (Table S2). The Al_2O_3 profile exhibits a gradually increasing trend upsection (Fig. 7B), as shown by the range and mean values for each unit: $7.1\text{--}15.7\%$ (mean 10.2%) for Unit 1, $7.9\text{--}22.0\%$ (mean 13.2%) for Unit 2, and $4.1\text{--}19.3\%$ (mean 16.1%) for Unit 3.

$\text{SiO}_{2(\text{bio})}$, $\text{Ba}_{(\text{bio})}$, and $\text{Cu}_{(\text{bio})}$ were used as paleoproductivity proxies (Table S2), with $\text{SiO}_{2(\text{non-clay})}$ serving as an estimate of maximum $\text{SiO}_{2(\text{bio})}$ (see Section 3.3; cf. Schoepfer et al.,

2015). The $\text{SiO}_{2(\text{bio})}$ profile exhibits a decreasing trend upsection, with lower values in successive units: 19-59 % (mean 41 %) in Unit 1, 3.9-44 % (mean 30 %) in Unit 2, and 0-27 % (mean 17 %) in Unit 3. The $\text{SiO}_{2(\text{bio})}$ profile is similar to the chert profile (Fig. 7D), possibly because the clay-fraction SiO_2 comprises less than half of total SiO_2 (10-49 %, mean 29 %). In contrast, the $\text{Ba}_{(\text{bio})}$ and $\text{Cu}_{(\text{bio})}$ profiles exhibit no pronounced upsection trend (Fig. 7E, F). $\text{Ba}_{(\text{bio})}$ exhibits a pronounced enrichment in the lowest part of Unit 3, to maximum values of ~6,700 ppm, but fluctuates mostly between 0 and 200 ppm through the rest of the section. $\text{Cu}_{(\text{bio})}$ fluctuates mostly between 0 and 20 ppm through the section.

V/Cr , Mo_{EF} , U_{EF} and $\text{C}_{\text{org}}/\text{P}$ were used as proxies for watermass redox conditions (Lewan and Maynard, 1982; Algeo and Ingall, 2007; Algeo and Tribovillard, 2009; Ross and Bustin, 2009; Tribovillard et al., 2012). The V/Cr , U_{EF} , Mo_{EF} , and $\text{C}_{\text{org}}/\text{P}$ profiles exhibit similar patterns of decreasing values upsection (Fig. 8B, D, E, F), with each successive unit showing lower values. For V/Cr , the range and mean are 0.9-4.9 (mean 3.6) in Unit 1, 1.3-4.0 (mean 2.3) in Unit 2, and 1.3-5.3 (mean 2.5) in Unit 3. For Mo_{EF} , the range and mean are 0.4-77 (mean 36) in Unit 1, 3.1-54 (mean 11) in Unit 2, and 0.2-23 (mean 8.8) in Unit 3. For U_{EF} , the range and mean are 0.8-13 (mean 8.1) in Unit 1, 1.7-8.6 (mean 3.6) in Unit 2, and 1.0-9.9 (mean 2.9) in Unit 3. For $\text{C}_{\text{org}}/\text{P}$, the range is 84-316 (mean 188) in Unit 1, 52-232 (mean 116) in Unit 2, and 38-150 (mean 87) in Unit 3.

Mo/TOC ratios were used as a paleohydrographic proxy (Algeo and Lyons, 2006; Algeo and Rowe, 2012). Mo/TOC values decline upsection into Unit 2, as reflected by the range and mean of ratios: 3.3-21 (mean 10) for Unit 1, and 3.8-13 (mean 6.6) for Unit 2 (Fig. 8F). A sharp increase in Mo/TOC is observed at the Unit 2-3 contact, and Unit 3 exhibits modestly

higher, but significantly more variable, ratios 1.2-13 (mean 8.0).

CIA* and Zr/Al ratios were used as terrestrial input proxies (Nesbitt and Young, 1982; Schnetger, 1992; Bahlburg, 2009). Their values show also the three units (Fig. 8G, H), although the boundary for Unit 1-2 is obscure. The Zr/Al ratios (all values are $\times 10^{-3}$) are 1.8-4.7 (mean 3.0) in Unit 1, 1.8-6.2 (mean 2.7) in Unit 2, and 1.2-5.4 (mean 2.2) in Unit 3. The CIA* values are 0.72-0.74 (mean 0.73) in Unit 1, 0.72-0.78 (mean 0.73) in Unit 2, and 0.73-0.78 (mean 0.76) in Unit 3. Within Units 2 and 3, Zr/Al and CIA* exhibit a distinct pattern of negative covariation (Fig. 8G, H).

5. Discussion

5.1. Sequence stratigraphy

The Jiaoye-1 study section exhibits long-term trends upsection in most geochemical profiles. This is evident in the TOC profile, in which both the range and mean of values for each successive unit decrease upsection (Fig. 7A), as well as in the profiles for Al_2O_3 , chert, biogenic SiO_2 , and redox proxies (V/Cr, Mo_{EF} , and U_{EF}) (Fig. 7C, D; Fig. 8B, D, E). This pattern suggests a progressive, long-term reduction in the sinking flux of biogenic material and improved oxygenation/ventilation of the seafloor, due to one or more of the following factors: (1) sea-level shallowing (leading to lower productivity or higher detrital sediment inputs in nearshore areas), (2) a decline in surface-water productivity (at stable sea levels), or (3) an increase of detrital inputs related to the suturing process of the Yangtze and Cathaysia blocks (Faure et al., 2009; Charvet et al., 2010; Li et al., 2010).

The study section can be subdivided into three intervals (Units 1-3) on the basis of abrupt

shifts in a number of geochemical profiles. For example, the profiles of TOC, V/Cr, and U_{EF} exhibit a gradual decrease upsection within each unit, followed by a sharp increase at the base of the overlying unit (Fig. 7A, Fig. 8B, D, E). Other profiles show similar patterns although with minor variations, e.g., the chert and $SiO_{2(bio)}$ profiles exhibit sharp increases at the base of Units 1 and 2, but a decrease at the base of Unit 3 (Fig. 7D). The observation that multiple geochemical profiles show breaks (i.e., abrupt increases or decreases) at the contacts between the defined units demonstrates the robustness of the tripartite subdivision of the Lower Silurian part of our study section. These patterns suggest that each unit is a stratigraphic sequence delimited by sequence boundaries.

The durations of these three sequences can be estimated from our age-depth analysis of the study core: ~2.3, 0.7, and 2.5 Myr for Units 1, 2, and 3, respectively (Fig. 3; see Section 2). The durations of the sequences are thus longer than the well-established components of Milankovitch-band orbital variation (~0.02-0.41 Myr; Hinnov, 2000; Berger, 2013), falling instead in the range of third-order sequences that have typical durations of 1 to 3 Myr and poorly-understood origins (Haq, 1991; Posamentier and Allen, 1999). One possible origin is climate variation linked to the ~2-Myr orbital eccentricity cycle, which is an ultralong modulation of the shorter ~100- and 413-kyr eccentricity cycles caused by planetary orbital resonances (Abels et al., 2010; Laskar et al., 2011; Boulila et al., 2012). A second possibility is control by episodic tectonic subsidence and relative sea-level changes within the Yangtze Sea linked to the Kwangsi Orogeny (Wang, 1985; Wan and Zhu, 1990; Faure et al., 2009; Charvet et al., 2010; Li et al., 2010). Before considering these possible origins further (see Section 5.6), we will analyze variation in redox conditions, watermass restriction, surface

productivity, and detrital fluxes in the Early Silurian Yangtze Sea and their relationships to Units 1-3 in order to elucidate controls on the formation of these stratigraphic sequences.

5.2. Marine redox conditions

Paleomarine redox conditions can be reconstructed using trace metal and mineral proxies.

In this study, we chose V/Cr, Mo_{EF}, U_{EF}, C_{org}/P, and pyrite framboids to examine the redox conditions of the Jiaoye-1 section. V/Cr is a useful redox proxy (Lewan and Maynard, 1982; Ross and Bustin, 2009) for which thresholds of <2.0 for oxic conditions, 2.0-4.25 for dysoxic conditions, and >4.25 for anoxic conditions were estimated based on Jurassic shales (Jones and Manning, 1994). Both Mo and U typically are present at low concentrations (<5 ppm) in oxic environments (Zheng et al., 2000; Morford et al., 2009a, b), but at much higher concentrations in moderately (~25-100 ppm Mo) and strongly (>100 ppm Mo) reducing environments (Scott and Lyons, 2012). In the latter systems, Mo_{EF} and U_{EF} values can exceed 100 (Algeo and Tribovillard, 2009; Tribovillard et al., 2012). C_{org}/P ratios show a strong relationship with benthic redox conditions: modern marine sediments yield median C_{org}/P ratios of <50 in oxic facies, ~50-100 in suboxic facies, and >100 in anoxic facies (Algeo and Ingall, 2007). In addition to these elemental proxies, the abundance of pyrite framboids and their size distribution can be used to identify water-column euxinia (Wilkin et al., 1996; Wignall et al., 1998; 2005). If the chemocline lies above the sediment-water interface, pyrite can form syngenetically as fine-grained framboids, characteristically <~6 μm in diameter and having a narrow size distribution (Wilkin and Barnes, 1997; Tribovillard et al., 2006). On the other hand, if the chemocline lies at or below the sediment-water interface, pyrite forms

diagenetically as larger framboids having a wider size distribution.

All elemental redox proxies (i.e., V/Cr, Mo_{EF}, U_{EF}, C_{org}/P) document a long-term shift toward less intensely reducing conditions upsection as well as pronounced cyclic variation, with sharp shifts toward more reducing conditions at the base of each unit followed by a gradual lessening of intensity upward within the unit (Fig. 8B, C, D, E). Based on C_{org}/P ratios, redox conditions in Unit 1 were strongly anoxic at the unit base, declining to weakly anoxic at the unit top. In Unit 2, redox conditions were moderately anoxic at the unit base, ameliorating to suboxic by mid-unit, after which little change occurred. In Unit 3, redox conditions were weakly anoxic at the unit base, shifting to suboxic or oxic at the unit top. Although the absolute redox interpretations yielded by different proxies vary somewhat (e.g., V/Cr suggests less intensely reducing conditions than C_{org}/P at all stratigraphic levels), all elemental redox proxies exhibit the same patterns of relative variation upsection (Fig. 8).

On a U_{EF}-Mo_{EF} crossplot, the trend toward lower enrichment factors for both Mo and U from Unit 1 to Unit 3 is consistent with a general amelioration of redox conditions upsection within the Lungmachi Formation (Fig. 9). About one third of the study samples plot in the “unrestricted marine” field with the others plotting between the “unrestricted marine” and “particulate shuttle” fields. This distribution implies that the accumulation of Mo was partially influenced by operation of a particulate shuttle, which suggests that the chemocline was generally located close to or above the sediment-water interface (Algeo and Tribovillard, 2009).

[Place Figure 9 here]

The mean size of pyrite framboids in all of the examined samples ($n = 16$) is $<6\ \mu\text{m}$, and a mean size-versus-SD crossplot indicates a syngenetic origin (Fig. 10). Although sampling for pyrite framboid size analysis was not continuous (Fig. 8A), the observation that all 16 samples analyzed contain syngenetic framboids suggests that water-column euxinia existed throughout the interval of deposition of the study section.

[Place Figure 10 here]

There are several possible explanations for the apparent disagreement between the elemental redox proxies, which document fluctuations between anoxic and suboxic conditions, and the pyrite framboid size proxy, which indicates continuous euxinia. For example, euxinia can develop in oxygen-minimum zones at intermediate water depths, delivering sinking syngenetic framboids to the sediment, even though bottom-waters are less reducing (e.g., suboxic) (see Algeo et al., 2011, for a detailed discussion). However, this would have required water depths great enough to allow formation of a deep watermass below a thermocline, i.e., much deeper than the estimates of 100-200 m for most of the Yangtze Sea (Li et al., in review; Liu et al., in review). More likely is that the ubiquitous presence of syngenetic pyrite reflects high-frequency redox variations, with intervals of euxinia sufficient to allow formation of framboids in the water column but not long enough to cause strong enrichments of Mo, U and V. Our data therefore suggest that short intervals of euxinia continued to occur even as water-column redox conditions became, on average, less

reducing upsection.

Several earlier studies concluded that long-term stable anoxia existed within the Yangtze Sea following the Hirnantian glaciation (Chen et al., 2004; Yan et al., 2010, 2012; Zhou et al., 2015). However, the multiple redox proxies of the present study reveal that the Yangtze Sea experienced unstable redox conditions during the Early Silurian. Our redox data demonstrate a pronounced cyclicity, with rapid shifts toward more intensely reducing conditions at the base of each unit followed by slow trends toward less reducing conditions upsection within each unit. This pattern reflects an overall sequence stratigraphic control on environmental redox conditions.

5.3. Basin hydrography

The study area was located in the interior of the Yangtze Sea, surrounded by uplifts or oldlands and in communication with the open ocean through relatively narrow straits (Fig. 2B; Zhang et al., 2016). For this reason, the Yangtze Sea was potentially prone to watermass restriction (Chen et al., 2004). Hydrographic restriction of paleomarine basins can be tested using Mo/TOC ratios (Algeo and Lyons, 2006; Algeo and Rowe, 2012). Given anoxic deepwaters, increasing watermass restriction correlates with lower aqueous Mo concentrations (owing to limited deepwater Mo resupply) and lower sediment Mo/TOC ratios. In modern anoxic marine systems, Mo/TOC ratios range from ~45 for weakly restricted basins such as Saanich Inlet to ~4.5 for strongly restricted basins such as the Black Sea (Algeo and Lyons, 2006). The study samples yield Mo/TOC ratios mostly of 5-15, with an overall average of ~9 (Fig. 8F, Fig. 11), which is equivalent to moderately intense restriction,

as exhibited by the modern Framvaren Fjord (Algeo and Lyons, 2006; Algeo and Rowe, 2012). Although a long-term decline in Mo/TOC ratios upsection (Fig. 8F) was probably redox-controlled, i.e., related to progressively ameliorating deep-water redox conditions, a large increase in Mo/TOC at the Unit 2/3 contact is consistent with a sharp rise in sea level leading to more unrestricted watermass conditions.

[Place Figure 11 here]

One point of uncertainty is whether Early Silurian global seawater had the same Mo concentration as modern seawater. If global seawater Mo concentrations were lower at that time, then sediment Mo/TOC ratios would have been reduced without any need to invoke basinal restriction. However, the ranges of Mo concentrations and Mo/TOC ratios in anoxic marine facies have been relatively uniform throughout the Phanerozoic (Scott et al., 2008), suggesting no major changes in global seawater Mo concentrations. For this reason, we infer that the low Mo/TOC ratios of the study section imply significant basinal restriction.

In addition to watermass restriction, a regionally humid climate probably also played a role in generating deepwater anoxia. The Yangtze Block was located near the paleo-equator during the O-S transition (Scotese and McKerrow, 1991; Young et al., 2008), where humid climatic conditions prevailed and precipitation and runoff into the Yangtze Sea are likely to have exceeded evaporation. Given these conditions, large-scale estuarine circulation is likely to have existed within the Yangtze Sea, which would have promoted strong water-column stratification and deepwater anoxia, as in the modern Black Sea or Baltic Sea (Algeo et al.,

2008). These types of marine systems, in which water-column stratification depends on hydrologic/hydrographic factors in addition to basinal restriction, tend to exhibit particularly dynamic watermass redox conditions (Algeo and Lyons, 2006; Algeo and Heckel, 2008), which is consistent with our inferences regarding redox conditions in the Early Silurian Yangtze Sea (see Section 5.2).

5.4. Marine productivity

Marine productivity can be assessed using biogenic SiO_2 , Ba, and Cu proxies (Dymond et al., 1992; Morse and Luther, 1999; Piper and Perkins, 2004; Griffith and Paytan, 2012). Biogenic SiO_2 is abundantly present in the study samples, as shown by the remains of radiolarians and sponge spicules (Fig. 5A-C; see Section 4.1). However, biogenic SiO_2 cannot be readily distinguished from detrital quartz analytically, and both are included in the total SiO_2 (Fig. 7C) and non-clay SiO_2 profiles (Fig. 7D). Given that the location of the study section was far removed from sources of detrital quartz, and that non-clay SiO_2 shows a strong positive correlation to TOC (Fig. 12), we infer that the largest share of SiO_2 is of biogenic origin, and henceforth we will refer to the non-clay SiO_2 curve as “biogenic SiO_2 ”. Ba accumulates authigenically in sinking organic matter to form intraparticulate barite crystals (Dymond et al., 1992; Griffith and Paytan, 2012). Cu is one of the micronutrients that show a nutrient-type vertical concentration profile in the ocean (Huerta-Diaz and Morse, 1992; Morse and Luther, 1999; Piper and Perkins, 2004).

[Place Figure 12 here]

570

571 In the Jiaoye-1 study core, the biogenic SiO₂ profile exhibits a similar pattern to the
572 elemental redox proxies (see [Section 5.2](#)), i.e., both unit-scale cyclicity and a long-term trend
573 toward lower values upsection ([Fig. 7](#)). Each of units 1-3 is characterized by high values of
574 biogenic SiO₂ at its base, followed by a gradual decline upsection ([Fig. 7D](#)). This similarity
575 of pattern suggests a causal link between productivity and redox conditions, e.g., that
576 increases in productivity and organic carbon sinking fluxes triggered deepwater oxygen
577 depletion (cf. [Demailson and Moore, 1980](#); [Parrish, 1982](#); [Pedersen and Calvert, 1990](#)).

578 The biogenic Ba profile exhibits mostly low values through the study section, although
579 with pronounced enrichment at the base of Unit 3 ([Fig. 7E](#)). Its relative absence in Units 1
580 and 2 may reflect a redox control on the preservation of barite ([Dymond et al., 1992](#); [Falkner](#)
581 [et al., 1993](#); [Paytan and Kastner, 1996](#); [Schenau et al., 2001](#)), or possibly the relative
582 shallowness of the Yangtze Sea (barite formation in sinking organic floccules correlates with
583 water depth). The biogenic Cu profile also shows mostly low values through the study section,
584 although with increases of ~5-10× background levels, to 20-30 ppm, at the bases of Units 2
585 and 3 ([Fig. 7F](#)). The Ba and Cu profiles thus support the inference of elevated marine
586 productivity at the base of each unit.

587 $\delta^{13}\text{C}_{\text{org}}$ records may also reflect variations in marine productivity. As photosynthetic
588 algae preferentially make use of ¹²C-enriched CO₂ in the ocean-surface layer, the remaining
589 DIC becomes isotopically heavier, a signature that is enhanced by intense productivity ([Rau](#)
590 [et al., 1992, 1997](#); [Kump and Arthur, 1999](#)). However, if $\delta^{13}\text{C}_{\text{org}}$ tracks the $\delta^{13}\text{C}$ of
591 contemporaneous seawater dissolved inorganic carbon (DIC), then it serves as a global rather

than a local productivity proxy. In the Jiaoye-1 study core, the $\delta^{13}\text{C}_{\text{org}}$ profile shows a near-continuous rise of +2 ‰ upsection through Units 1 and 2, followed by more variable but generally declining $\delta^{13}\text{C}_{\text{org}}$ in Unit 3 (Fig. 7G). The slow changes in $\delta^{13}\text{C}_{\text{org}}$ through most of the study section are consistent with control by $\delta^{13}\text{C}_{\text{DIC}}$ and, thus, a global signal. However, small but abrupt changes in $\delta^{13}\text{C}_{\text{org}}$ at the Unit 1/2 and Unit 2/3 contacts (Fig. 7G) are indicative of stratigraphic condensation and/or a change in carbon cycling at the sequence boundaries represented by these contacts.

5.5. Terrigenous fluxes

The $\%\text{Al}_2\text{O}_3$ profile provides the most direct proxy for the siliciclastic content of the study section (Fig. 7B). The $\%\text{Al}_2\text{O}_3$ profile exhibits smooth changes upsection within each unit, but sharp shifts across the unit contacts (i.e., lower at the Unit 1/2 contact, and higher at the Unit 2/3 contact). This pattern is consistent with continuous sedimentation within each unit, followed by breaks or slowdowns at unit contacts (i.e., sequence boundaries). Higher $\%\text{Al}_2\text{O}_3$ values upsection within each unit, as well as for the section as whole (Fig. 7B), suggests a long-term increase in siliciclastic supply to the study site.

The chemical index of alteration (CIA) can provide information about the intensity of subaerial weathering in the continental sourcelands of marine siliciclastic sediments. As an approximate guide to their interpretation, sediment CIA values of 0.8-1.0 correspond to hot and humid tropical climates, 0.7-0.8 to temperate climates, and <0.7 to cold and arid climates (Nesbitt and Young, 1982; Bahlburg, 2009). However, the chemical composition of the source rock can influence CIA values as well, especially for relatively immature sediments

with low CIA values (Bahlburg, 2009). Because of the potential influence of sediment provenance, it is generally best to assess positive and negative shifts in CIA profiles as evidence of increasing and decreasing humidity during weathering, rather than attempting to interpret absolute climate conditions.

The study samples yield CIA* values mostly of 0.70-0.80, with an overall average of $\sim 0.74 \pm 0.02$ (Fig. 8G), which is indicative of moderate weathering intensity. Even though the range of CIA* variability in Units 1-2 is small (~ 0.72 -0.75, except for isolated outliers), sample-to-sample variability is quite limited, so small upsection changes in CIA* values are potentially meaningful. Closely examined, the CIA* profile shows evidence of cyclic variation within each unit, with maximum values at the unit base and top and minimum values in mid-unit, although this pattern is subtle in Unit 1 (Fig. 8G). Within Unit 1, CIA* declines from 0.73 at the base to 0.72 in mid-unit and then rises to 0.74 at its top. Unit 2 exhibits a similar pattern, with a minimum of 0.72 in mid-unit followed by a rise to 0.75 at its top. The base of Unit 3 is marked by a sharp increase to 0.77 followed by a decline to 0.75 in mid-unit (excluding some outliers with even lower values) and finally a rise to 0.78 at the unit top. The CIA* profile thus exhibits both a long-term trend toward higher values as well as cyclic variation within each unit, similar to the patterns of variation seen in many other geochemical proxies in the study section.

Zr/Al ratios can be used as a proxy for climate change, specifically eolian inputs and, thus, wind shear (Straub et al., 2013). The element Zr is present mainly in the heavy mineral zircon, which is commonly concentrated in the silt fraction, and large quantities of silt are

transported to oceanic regions downwind of major deserts, e.g., the modern Sahara or Gobi (Schnetger, 1992; Wehausen and Brumsack, 1999; Hao et al., 2010). The Zr/Al profile for the Jiaoye-1 core shows an almost mirror-image relationship to the CIA* profile, with a long-term trend toward lower values as well as cyclic variation within Units 2 and 3 (but not Unit 1), in which lower Zr/Al ratios are observed at the unit base and top and higher ratios in mid-unit (Fig. 8H). This indicates that eolian inputs and/or wind shear were least at the base and top of units, where weathering intensity was strongest (based on CIA*), and greatest in the middle of units, where weathering intensity was weakest.

5.6. Tectono-eustatic or glacio-eustatic controls on cyclicity?

The three sequences present in the Lower Silurian Lungmachi Formation of the Jiaoye-1 core have durations ranging from ~0.7 to 2.5 Myr and, thus, are equivalent to third-order sequences (Haq, 1991; Posamentier and Allen, 1999). The origins of such sequences can be either climato-eustatic, e.g., related to ultralong orbital periodicities (Abels et al., 2010; Laskar et al., 2011; Boulila et al., 2012), or tectonic, e.g., related to regional basin subsidence (Galloway, 1989). There is a basis for inferring either origin for the sequences of the Lower Silurian Lungmachi Formation. On the one hand, the Early Silurian world is thought to have been an interval of continental glaciation (Cramer and Saltzman, 2007; Finnegan et al., 2011; Trotter et al., 2016), and on the other hand, the Early Silurian Yangtze Sea was subject to the regional Kwangsian Orogeny, a compressional event related to collision of the Yangtze and Cathaysia blocks (Wang, 1985; Wan and Zhu, 1990; Faure et al., 2009; Charvet et al., 2010; Li et al., 2010). Evaluating these two hypotheses may be possible based on our detailed

analysis of redox, hydrographic, productivity, and terrigenous proxies in the Jiaoye-1 study core.

The key to evaluating controls on sequence development based on multiple geochemical proxies is examining covariation within individual sequences. Sequence boundaries are commonly associated with rising sea level, which can limit detrital siliciclastic inputs and lead to stratigraphic condensation (Galloway, 1989; Posamentier and Allen, 1999). These processes tend to result in higher TOC concentrations, lower %Al₂O₃ concentrations, and a shift toward more reducing conditions associated with greater water depths. Deepening may also lead to more unrestricted watermass circulation (see Cariaco Basin discussion in Algeo and Lyons, 2006) as well as to enhanced nutrient upwelling, stimulating marine productivity and organic carbon sinking fluxes, thus intensifying pre-existing patterns. All of these processes are consistent with observed changes in redox, hydrographic, productivity, and detrital proxies just above the sequence boundaries in the study core (Fig. 13A; see Sections 5.2-5.4). Conversely, the highstand to falling stages of sequences tend to be associated with increased siliciclastic inputs (owing to greater proximity to shorelines), reduced productivity (owing to weaker nutrient upwelling), less reducing conditions (owing to water-column shallowing), and greater restriction (if a basin-margin sill is present). Again, all of these processes are consistent with observed changes in the multiple proxies analyzed within individual units (Fig. 13B; see Sections 5.2-5.4).

[Place Figure 13 here]

The development of sequences in marine successions can be controlled by either glacio-eustatic fluctuations, e.g., related to the growth and decay of continental icesheets, for which there is substantial evidence in the Early Silurian (Cramer and Saltzman, 2007; Finnegan et al., 2011; Trotter et al., 2016), or tectonically-driven basin subsidence, e.g., via thrust loading, which has been inferred for the regional Kwangsian Orogeny (Li et al., 2010; Liu et al., in review). The suite of marine environmental proxies discussed above show causally interrelated changes that are more closely linked to sequence development rather than to the specific mechanism that generated the sequences. Thus, these proxies are generally not sufficiently diagnostic to allow one mechanism or the other to be proven, and additional, independent proxies are needed for this assessment.

Terrigenous proxies (e.g., CIA* and Zr/Al), which are largely independent of the marine proxies discussed above, provide information about coeval changes in climate or terrestrial environments (Bahlburg, 2009; Straub et al., 2013), although possible influences of water depth and facies proximity on marine clay-mineral assemblages and, thus, CIA should be noted (Khim, 2003; Bahlburg, 2009). In the study section, terrigenous proxies show that weathering intensity was strongest and eolian inputs and/or wind shear least at the base and top of sequences (Fig. 13A), whereas weathering intensity was reduced and eolian inputs and/or wind shear intensified in the middle of each sequence (Fig. 13B). This pattern is perfectly consistent with a climato-eustatic origin for the Lungmachi Formation sequences. In this scenario, the base of each sequence represents a rapid sea-level rise triggered by deglaciation (Fig. 13A; cf. Lambeck and Chappell, 2001). Deglaciations were associated with climatic warming, leading to intensified chemical weathering (hence higher CIA*) (cf. Vance

et al., 2009) and reduced wind shear (hence lower Zr/Al) (cf. Straub et al., 2013). As glacial conditions were renewed, cooling led to reduced chemical weathering intensity (hence lower CIA*) and increased wind shear (hence higher Zr/Al), as inferred for the middle of each sequence (Fig. 13B).

A tectonic origin of the Lungmachi Formation sequences is more difficult to reconcile with the observed changes in terrigenous proxies. Thrust loading, i.e., the putative trigger for development of a sequence boundary, is generally associated with orogenic uplift (cf. Heller et al., 1988), which would be likely to induce rapid erosion and/or cause development of a rain shadow (cf. Clark et al., 2005; Sepulchre et al., 2006), in either case limiting chemical weathering and resulting in lower CIA* values. Siliciclastic influx should also peak early, owing to the enhanced physical weathering associated with uplift (cf. Clark et al., 2005), leading to a spike in %Al₂O₃ at the bases of sequences. Neither of these patterns is observed in the study core--rather, exactly the opposite patterns are present. For these reasons, we favor a glacio-eustatic mechanism over a regional tectonic mechanism for the origin of Lungmachi Formation sequences.

Several additional considerations favor a glacio-eustatic mechanism. First, the pattern of rapid sea-level rises and slow sea-level falls is highly characteristic of glacio-eustatic fluctuations (Imbrie et al., 1993; Augustin et al., 2004). Second, Lower Silurian marine successions globally have provided evidence of pronounced third-order eustatic fluctuations that have been widely interpreted as of glacio-eustatic origin (Bennacef et al., 1971; Berry and Boucot, 1973; Grahn and Caputo, 1992; Caputo, 1998; Melchin and Holmden, 2006; Johnson, 2006, 2010; Díaz-Martínez and Grahn, 2007; Calner, 2008; Lehnert et al., 2010;

Munnecke et al., 2010; Clayer and Desrochers, 2014; Davies et al., 2016). The three Early Silurian sea-level cycles recorded in the Jiaoye-1 drillcore are consistent with the short-term global sea-level changes shown in the short-term eustatic curve of Haq and Schutter (2008) (Fig. 1).

The origin of the long-term trends in our geochemical proxy records may be tectonic rather than eustatic. There is widespread evidence of a long-term fall in sea level on the Yangtze Platform during the Early Silurian (Chen et al., 2004; Fan et al., 2009; Chen-L et al., 2015). This pattern differs from that of published global sea-level curves, all of which show a gradual rise during this interval (Loydell, 1998; Johnson, 2006, 2010; Haq and Schutter, 2008). This difference suggests control of water depths on the Yangtze Platform by epeirogenic uplift. During the Early Silurian, the South China Craton experienced tectono-epeirogenic movements related to final welding of the Yangtze and Cathaysia blocks (“Kwangsi Orogeny”). The suture zone, which was active from the Middle Ordovician to the Early Devonian, is the Wuyi-Yunkai orogenic belt (Wang, 1985; Wan and Zhu, 1990; Faure et al., 2009; Charvet et al., 2010; Li et al., 2010) (see Section 2.1). The compressional stresses associated with this tectonic event induced slow uplift of significant areas of the South China Craton, potentially accounting for the pattern of long-term sea-level fall during the Early Silurian observed in the Jiaoye-1 drillcore.

6. Conclusions

Following the end-Ordovician Hirnantian glaciation, the Early Silurian was characterized by major changes in environmental conditions within the Yangtze Sea of the South China

Craton. The post-Hirnantian sea-level rise triggered development of euxinic conditions within the inner Yangtze Sea. Subsequently, three glacio-eustatic cycles (stratigraphic sequences) were characterized by rapid rises and slow falls of sea level. During sea-level rises, which corresponded to intervals of reduced continental glaciation, the study area experienced elevated primary productivity, more reducing conditions, diminished watermass restriction, lower detrital input, and increased siliciclastic weathering intensity. During sea-level falls, which corresponded to intervals of increased continental glaciation, the opposite set of conditions prevailed. Besides these cyclic fluctuations, the study area underwent long-term shallowing during the Early Silurian, an event possibly related to tectono-epeirogenic uplift associated with the Kwangsi Orogeny that marked collision of the Yangtze and Cathaysia blocks of the South China Craton.

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Figure captions

Fig. 1. Stratigraphy and sea-level variation for the Late Ordovician to Late Silurian. (A) High-frequency (“short term”) and low-frequency (“long term”) eustatic (global sea-level) curves of [Haq and Schutter \(2008\)](#). (B) High-frequency sea-level curves of [Ross and Ross \(1995\)](#) for the Ordovician (of North America) and [Loydell \(1998\)](#) for the Silurian (of Europe). (C) Low-frequency sea-level curves of [Kangyin et al. \(2010\)](#) for the Ordovician (of Russia) and [Johnson \(2006, 2010\)](#) for the Silurian (of North America). (D) High-frequency sea-level variation during the latest Ordovician-Early Silurian in South China ([this study](#)). Units 1-3 are defined in [Sections 4.2 and 5.1](#).

Fig. 2. (A) Global paleogeography of the Late Ordovician (after [Witzke, 1990](#); [Scotese and McKerrow, 1991](#); [Young et al., 2008](#)). Note that South China Craton was rotated ~90° counter-clockwise relative to its modern orientation. (B) Geologic map showing Late

1184 Ordovician depositional facies of the Yangtze Block and the location of the Jiaoye-1 drillcore
1185 (after [Chen et al., 2004](#); [Yan et al., 2010](#); [Zhou et al., 2015](#)). The cross-section at lower left
1186 shows the present subsurface distribution of Paleozoic units in the Szechuan Basin (after [Hu](#)
1187 [et al., 2014](#)).

1188

1189 **Fig. 3.** Depth-age relationships in the Jiaoye-1 drillcore based on graptolite zones ([this study](#)).

1190 Time scale from [Cooper and Sadler \(2012\)](#). Abbreviations: Aero., Aeronian; Rhud.,
1191 Rhuddanian; Hirn., Hirnantian; WF, Wufeng Formation; KYC, Kuanyinchiao Bed. Graptolite
1192 zones after [Chen et al. \(2000\)](#) and [Zhan et al. \(2010\)](#). Hirn., Hirnantian; WF, Wufeng
1193 Formation; KYC, Kuanyinchiao Bed; WF2, *Dicellograptus complexus* Zone; WF3,
1194 *Paraorthograptus pacificus* Zone; WF4, *Metabolograptus extraordinarius* Zone; LM1,
1195 *Metabolograptus persculptus* Zone; LM2, *Akidograptus ascensus* Zone; LM3,
1196 *Parakidograptus acuminatus* Zone. Graptolite zone labels after [Chen-X et al. \(2015\)](#).

1197

1198 **Fig. 4.** Graptolites from the Jiaoye-1 drillcore. (A) *Coronograptus* cf. *cyphus* (Lapworth),
1199 from 2378.57 m, *Coronograptus cyphus* Zone, Lungmachi Fm.; (B) *Paraorthograptus*
1200 *pacificus* (Ruedemann), from 2411.88 m, *Paraorthograptus pacificus* Zone-*Metabolograptus*
1201 *extraordinarius* Zone, Wufeng Fm.; (C) *Campograptus communis* (Lapworth), from 2350.36
1202 m, *Demirastrites triangulatus* Zone, Lungmachi Fm.; (D) *Cephalograptus cometa* (Geinitz),
1203 from 2348.22 m, *Lituigraptus convolutus* Zone, Lungmachi Fm.; (E) *Demirastrites*
1204 *guizhouensis* (Mu et al.), from 2367.23 m, *Demirastrites triangulatus* Zone, Lungmachi Fm.;
1205 (F) *Stimulograptus* cf. *halli* (Barrande), from 2340.87 m, *Stimulograptus sedgwickii* Zone,

1206 Lungmachi Fm.; (G) *Cystograptus vesiculosus* (Nicholson), from 2403.82, *Cystograptus*
 1207 *vesiculosus* Zone, Lungmachi Fm. Note the less blunt rounded proximal part due to
 1208 preservation; (H) *Akidograptus cuneatus* Chaletzkaya, from 2407.42 m, *Akidograptus*
 1209 *ascensus* Zone-*Parakidograptus acuminatus* Zone, Lungmachi Fm.; (I) *Lituigraptus*
 1210 *convolutus* (Hisinger), from 2348.22 m, *Lituigraptus convolutus* Zone, Lungmachi Fm.; and
 1211 (J) *Dicellograptus ornatus* (Elles & Wood), from 2412.69 m, *Dicellograptus complexus* Zone,
 1212 Wufeng Fm. The scale is the same in all photos.

1213
 1214

1215 **Fig. 5.** Thin-section features of the study section. (A) Sponge spicules altered to pyrite
 1216 (carbonaceous mudstone, 2392.74 m). (B) Radiolarians (carbonaceous mudstone, 2389.31 m).
 1217 (C) Sponge spicules (carbonaceous mudstone, 2388.52 m). (D) Silty mudstone (carbonaceous
 1218 mudstone, 2381.91 m). (E) Laminated silty shale (silty mudstone, 2369.63 m). (F) Laminated
 1219 to moderately bioturbated silty mudstone (2332.63 m).

1220

1221 **Fig. 6.** SEM features of the study section. (A) Clay minerals and interparticle organic
 1222 material (carbonaceous mudstone, 2411.05 m). (B) Pores formed due to organic material
 1223 maturation (carbonaceous mudstone, 2397.13 m). (C) Pyrite framboids (silty mudstone,
 1224 2376.05 m). (D) Authigenic pyrite crystals and organics (silty mudstone, 2366.74 m). (E)
 1225 Irregular pyrite crystals (silty mudstone, 2335.30 m). (F) Pyrite framboids (silty mudstone,
 1226 2330.46 m).

1227

1228 **Fig. 7.** Stratigraphic profiles of paleoproductivity proxies: total organic carbon (TOC), Al₂O₃,
 1229 chert, biogenic SiO₂, biogenic Ba, biogenic Cu, and $\delta^{13}\text{C}_{\text{org}}$. Chert content was determined by

XRD, and biogenic SiO₂ content by chemical analysis. The horizontal dashed lines show the division of the section into Units 1-3, which was based mainly on the TOC profile.

Fig. 8. Stratigraphic profiles of paleoredox and climatic proxies: pyrite framboid size distributions, V/Cr, C_{org}:P ratios, Mo_{EF}, U_{EF}, Mo/TOC, chemical index of alteration (CIA*), and Zr/Al. Pyrite framboid size distributions are shown as bar-and-whisker plots where ‘Min’, ‘Med’, and ‘Max’ indicate minimum, median, and maximum diameters, respectively. The dotted vertical line in column A shows euxinic conditions for pyrite framboid size < 6 µm. The dotted vertical lines in column B show redox conditions for V/Cr (oxic < 2.0, dysoxic 2.0-4.25, and anoxic > 4.25) (after [Lewan and Maynard, 1982](#); [Ross and Bustin, 2009](#)), in column C show redox conditions for Corg: P (oxic < 50, dysoxic 50-100, and anoxic > 100) (after [Algeo and Ingall, 2007](#)), and in column D show redox conditions for Mo-EF (oxic < 25, dysoxic 25-100, and anoxic > 100) (after [Algeo and Tribovillard, 2009](#); [Tribovillard et al., 2012](#)).

Fig. 9. Enrichment factors (EFs) of Mo versus U. SW, the elements enrichment factor in sea water (after [Algeo and Tribovillard, 2009](#)).

Fig. 10. Mean size versus standard deviation of pyrite framboids (after [Wignall and Newton, 1998](#)). All of the Jiaoye-1 samples fall in the syngenetic pyrite field.

Fig. 11. Mo concentration versus TOC. Dashed lines show the regression trends for four modern anoxic marine systems (Saanich Inlet, Cariaco Basin, Framvaren Fjord, and Black Sea), reflecting variable degrees of hydrographic restriction ([Algeo and Lyons, 2006](#)). The

1254 regression line (red) indicates the mainly moderate watermass restriction of the Jiaoye-1
1255 section.

1256

1257 **Fig. 12.** $\text{SiO}_{2(\text{bio})}$ versus TOC. The diagonal line shows the positive correlation between these
1258 proxies.

1259

1260 **Fig. 13.** Model showing oceanic environmental changes of Yangtze Sea before and after the
1261 end-Ordovician crisis (adapted from Jeppsson, 1990; Bickert et al., 1997; Chen et al., 2004;
1262 Algeo et al., 2008; Yan et al., 2012).

1263

1264 **Appendix 1.** Graptolite specimens from Wufeng and Lungmachi formations of Jiaoye-1
1265 drillcore

1266

1267 Most graptolite biozones were recognized based on the first appearance (FA) of the
1268 zonal fossil (marked by an asterisk) or index fossils (marked by double asterisks). The
1269 *Dicellograptus complanatus* Zone (WF1) is known from only a few localities in the western
1270 Yangtze Sea (Chen et al., 2000). Although we did not find zonal or index fossils of the
1271 *Dicellograptus complexus* Zone (WF2), the base of the Wufeng Formation is considered
1272 equivalent to the base of the *Dicellograptus complexus* Zone (WF2) based on regional
1273 correlations. The base of the *Paraorthograptus pacificus* Zone (WF3) is marked only by the
1274 FA of *Rectograptus songtaoensis*, which sometimes occurs in the upper *Dicellograptus*
1275 *complexus* Zone but is most abundant in the *Paraorthograptus pacificus* Zone (WF3). We did
1276 not find any index fossils of the *Metabolograptus extraordinarius* Zone (WF4) from the
1277 uppermost Wufeng Formation. However, the recognition of the Kuanyinchiao Bed limestone
1278 indicates the existence of the underlying *M. extraordinarius* Zone at the Jiaoye-1 drillcore.

1279

1280 **Lungmachi Formation (above 2411.00 m):**

1281 *Stimulograptus sedgwickii* Zone (LM8):

1282 JY70 (2333.63 m): *Oktavites circularis* (Elles and Wood), *Pristiograptus* cf. *xiushanensis*

1283 Mu et al.;

1284 JY69 (2340.87 m): *Stimulograptus* cf. *halli* (Barrande);

- 1285 JY68 (2341.31 m): *Stimulograptus sedgwickii* (Portlock);
- 1286 JY67 (2341.43 m): *Stimulograptus sedgwickii* (Portlock);
- 1287 JY66 (2343.75 m): *Stimulograptus sedgwickii* (Portlock);
- 1288 JY65 (2344.19 m): *Stimulograptus sedgwickii* (Portlock)*;
- 1289
- 1290 *Lituigraptus convolutus* Zone (LM7):
- 1291 JY64 (2344.97 m): *Petalolithus minor* Elles, *Demirastrites* sp., *Monograptus* sp.;
- 1292 JY63 (2348.18 m): *Petalolithus folium* (Hisinger);
- 1293 JY62 (2348.22 m): *Lituigraptus convolutus* (Hisinger)*, *Petalolithus folium* (Hisinger);
- 1294
- 1295 *Demirastrites triangulatus* Zone (LM6):
- 1296 JY61 (2350.34 m): *Metaclimacograptus hughesi* (Nicholson), *Pseudoplegmatoraptus*
- 1297 *perlatus* (Nicholson);
- 1298 JY60 (2350.36 m): *Campograptus communis* (Lapworth);
- 1299 JY59 (2550.39 m): *Pseudoplegmatoraptus perlatus* (Nicholson);
- 1300 JY58 (2351.72 m): *Petalolithus folium* (Hisinger), *Campograptus communis* (Lapworth),
- 1301 *Monograptus arciformis* Chen and Lin, *Pseudorthograptus* sp., *Demirastrites* sp.;
- 1302 JY56 (2352.44 m): *Monograptus arciformis* Chen and Lin, *Campograptus communis*
- 1303 (Lapworth), *Demirastrites raitzhainiensis* (Eisel), *Diversograptus* sp.;
- 1304 JY55 (2356.82 m): *Demirastrites raitzhainiensis* (Eisel);
- 1305 JY54 (2358.67 m): *Glyptograptus* cf. *tamariscus* (Nicholson);
- 1306 JY53 (2358.84 m): *Monograptus involutus* Lapworth;

- 1307 JY52 (2363.64 m): *Demirastrites triangulatus* (Harkness), *Rastrites guizhouensis* Mu et al.;
- 1308 JY51 (2364.75 m): *Demirastrites triangulatus* (Harkness), *Rastrites guizhouensis* Mu et al.;
- 1309 JY50 (2367.23 m): *Demirastrites triangulatus* (Harkness);
- 1310 JY49 (2367.43 m): *Demirastrites triangulatus* (Harkness)*;
- 1311 JY47 (2368.97 m): *Rastrites longispinus* Perner;
- 1312
- 1313 *Coronograptus cyphus* Zone (LM5):
- 1314 JY44 (2378.57 m): *Coronograptus cyphus* (Lapworth);
- 1315 JY43 (2380.13 m): *Coronograptus* cf. *cyphus* (Lapworth);
- 1316 JY41 (2383.20 m): *Dimorphograptus confertus* (Nicholson);
- 1317 JY40 (2383.44 m): *Hamatograptus tenellus* (Chen and Lin);
- 1318 JY39 (2384.45 m): *Hamatograptus tenellus* (Chen and Lin);
- 1319 JY38 (2386.89 m): *Coronograptus* cf. *cyphus* (Lapworth), *Atavograptus atavus* (Jones);
- 1320 JY37 (2390.58 m): *Coronograptus* cf. *cyphus* (Lapworth)*, *Cystograptus penna* (Hopkison),
- 1321 *Atavograptus atavus* (Jones);
- 1322
- 1323 *Cystograptus vesiculosus* Zone (LM4):
- 1324 JY35 (2393.39 m): *Pseudorthograptus* sp.;
- 1325 JY29 (2400.23 m): *Atavograptus gracilis* Hutt;
- 1326 JY27 (2401.47 m): *Cystograptus vesiculosus* (Nicholson);
- 1327 JY26 (2403.82 m): *Cystograptus vesiculosus* (Nicholson);
- 1328 JY25 (2404.27 m): *Cystograptus vesiculosus* (Nicholson)*;

- 1329
- 1330 *Akidograptus ascensus* Zone – *Parakidograptus acuminatus* Zone (LM2-3):
- 1331 JY24 (2404.40 m): *Korenograptus jerini* (Koren' and Melchin)**;
- 1332 JY23 (2405.88 m): *Korenograptus lacinosus* (Churkin and Carter);
- 1333 JY22 (2405.89 m): *Normalograptus normalis* (Lapworth);
- 1334 JY21 (2406.08 m): *Korenograptus robustus* (Chen)**, *Neodiplograptus modestus*
- 1335 (Lapworth);
- 1336 JY20 (2407.11 m): *Normalograptus ajjeri* (Legrand)**;
- 1337 JY19 (2407.15 m): *Korenograptus Lungmaensis* (Sun);
- 1338 JY18 (2407.42 m): *Korenograptus lacinosus* (Churkin and Carter), *Akidograptus ascensus*
- 1339 Davies*;
- 1340 JY17 (2407.76 m): *Akidograptus?* sp.***, *Normalograptus wangiawanensis* (Mu and Lin);
- 1341
- 1342 *Metabolograptus persculptus* Zone (LM1):
- 1343 JY16 (2407.81 m): *Avitograptus avitus* (Davies)**, *Korenograptus lacinosus* (Churkin and
- 1344 Carter);
- 1345 JY15 (2407.84 m): *Normalograptus mirnyensis* (Obut and Sobolevskaya), *Normalograptus*
- 1346 sp.;
- 1347
- 1348 **Kuanyinchiao Bed (2411.00-2411.10 m):** fragments of brachiopods
- 1349
- 1350 **Wufeng Formation (2415.36-2411.00 m):**

- 1351 *Paraorthograptus pacificus* Zone - *Metabolograptus extraordinarius* Zone (WF3-WF4):
- 1352 JY14 (2411.74 m): *Dicellograptus turgidus* Mu, *Dicellograptus ornatus* Elles and Wood,
- 1353 *Appendispinograptus supernus* (Elles and Wood), *Paraorthograptus pacificus* (Ruedemann),
- 1354 *Climacograptus hastatus* T. S. Hall, *Paraplegmatograptus* sp.;
- 1355 JY13 (2411.88 m): *Paraorthograptus pacificus* (Ruedemann)*, *Appendispinograptus*
- 1356 *supernus* (Elles and Wood), *Pleurograptus* sp.;
- 1357 JY12 (2412.01 m): *Dicellograptus ornatus* Elles and Wood, *Anticostia lata* (Elles and
- 1358 Wood);
- 1359 JY11 (2412.06 m): *Dicellograptus minor* Toghill, *Anticostia lata* (Elles and Wood),
- 1360 *Climacograptus hastatus* T. S. Hall;
- 1361 JY10 (2412.52 m): *Rectograptus songtaoensis* Mu and Li**;
- 1362
- 1363 *Dicellograptus complexus* Zone (WF2):
- 1364 JY9 (2412.69 m): *Dicellograptus ornatus* Elles and Wood;
- 1365 JY8 (2412.84 m): *Rectograptus abbreviatus* (Elles and Wood);
- 1366 JY6 (2413.90 m): *Dicellograptus ornatus* Elles and Wood, *Anticostia lata* (Elles and Wood);
- 1367 JY4 (2414.13 m): *Rectograptus abbreviatus* (Elles and Wood);
- 1368 JY3 (2414.15 m): *Dicellograptus complexus* Davies*;
- 1369 JY2 (2415.19 m): *Rectograptus* sp.;
- 1370 JY1 (2415.20 m): *Appendispinograptus longispinus* (T. S. Hall), *Appendispinograptus*
- 1371 *supernus* (Elles and Wood).

Fig. 1

Fig. 2

Fig. 3

Fig. 4

Fig. 5

Fig. 6

Fig. 7

Fig. 8

Fig. 9

Fig. 10

Fig. 11

Fig. 12

Fig. 13