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Seawater carbon and strontium isotope variations through the late Ediacaran to late Cambrian in the Tarim Basin

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Abstract

The radiation of early animals during the Cambrian explosion was accompanied by widespread perturbations in oceanic and atmospheric conditions. However, the cause-effect relationships between evolutionary innovations, carbon (C) cycling, and continental weathering are still a matter of debate. Further paleoenvironmental reconstructions of poorly studied paleo-continents covering the Precambrian/Cambrian (PC/C) transition may improve the correlation of the events and their impact on the Earth system on a global scale. The Tarim Basin was presumably located proximal to the well-studied South China craton during the Ediacaran-Cambrian boundary and is expected to contribute equally to our paleoenvironmental reconstructions. Here we apply carbon and strontium (Sr) isotope records ($\delta^{13}C$ and $^{87}Sr/^{86}Sr$) obtained from well-preserved marine carbonates spanning the late Ediacaran to the late Cambrian from the Keping region in the northwestern Tarim Basin, NW China. By comparison to other carbonate successions from various paleo-continents, our obtained seawater $^{87}Sr/^{86}Sr$ curve in this study shows a stepwise first-order increase to more radiogenic values superimposed by second-order presumably regional variations. Our obtained $\delta^{13}C$ variations exhibit four prominent negative and three positive excursions identical to the global $\delta^{13}C$ record. Based on existing paleontological evidence and our new Tarim $\delta^{13}C$ and $^{87}Sr/^{86}Sr$ curves, we propose a stratigraphic correlation from the late Ediacaran to the late Cambrian between the Tarim Basin and other paleo-continents. Besides, our paleoenvironmental reconstruction reveals both regional and global sea-level changes that may have controlled the influx of radiogenic Sr and presumably the influx of nutrients controlling primary production and the rate of organic carbon burial in the Tarim Basin at the time of deposition. These feedbacks may have ultimately influenced the rapid diversification of metazoans and the oxygenation of the Cambrian ocean and atmosphere approaching-present pO_2 levels.

Keywords: Carbon isotopes, Strontium isotopes, Ediacaran-Cambrian, Tarim Basin, Paleoenvironment

1. Introduction

The transitional interval between the terminal Proterozoic and the early Paleozoic is marked by the sudden appearance of abundant and diverse metazoans (Valentine et al., 1999; Conway-Morris, 2003; Maloof et al., 2010; Kouchinsky et al., 2011). This early animal evolution was accompanied by extensive perturbations to climate, carbon (C) cycling, continental weathering and changes to the composition of the oceans (Shields-Zhou and Zhu, 2013). However, the cause-effect relationships between this evolutionary innovation and these environmental perturbations are still a matter of broad scientific debate and paleoenvironmental reconstructions of less studied paleo-continents covering the Precambrian/Cambrian (PC/C) transition can help this situation. Most studies from China covering this interval focus on sedimentary successions in South China. For instance, Li et al. (2013) suggested that muted continental weathering during the early Cambrian was indicated by decreasing seawater strontium isotope ratios ($^{87}Sr/^{86}Sr$). Ediacaran-Cambrian marine carbonate successions outcropping in the Tarim Basin were deposited on a shallow-water continental shelf, which was presumably situated close to the South China craton and at a similar low-middle latitude position (Fig. 1A) (McKerrow et al., 1992; Merdith et al., 2017; Huang et al., 2000; Zhan et al., 2007). In the past, Ediacaran-Cambrian Tarim Basin sediment successions have only received minor attention (Zhou et al., 2018; Zhu et al., 2019) but a few recent chemo-stratigraphic studies provided some preliminary results on their stratigraphic subdivision and paleoenvironmental conditions. He et al. (2007) and Guo et al. (2017) provided the first C isotope stratigraphic correlations for the late-Ediacaran Cambrian strata in the Tarim Basin between the Sugetbrak, the Penglaiba, the Wushi phosphorite and the Dongergou sections, while Wang et al. (2011b) presented paired C and Sr isotope trends from the Cambrian Series 2 to Series 3 at the Penglaiba section. However, there is still a chemo-stratigraphic gap (especially in regard of the Sr isotope stratigraphy) spanning the late-Ediacaran to the late-Cambrian time interval, which limits the stratigraphic correlation of the Tarim Basin with other parts of the world.

C and Sr isotope compositions obtained from well-preserved (i.e. low diagenetic fluid flow overprinted) ancient carbonate rocks may retain primary compositions of the contemporaneous seawater and provide constraints on the isotope-based chemo-stratigraphic correlation (DePaolo and Ingram, 1985; Montañez et al., 1996; Montañez and Banner, 2000; Zhu et al., 2007; Wang et al., 2011b; McArthur et al., 2012; Li et al., 2013; Zhou et al., 2018; Zhu et al., 2019). For instance, the first appearance datum of small shelly fossils (SSFs) is globally accompanied by a negative $\delta^{13}C$ excursion and a declining trend in seawater $^{87}Sr/^{86}Sr$ ratios (Cowie and Glaessner, 1975; Landing, 1989; Zhu et al., 2006; Li et al., 2013; Zhu et al., 2019), although there are still debates about the exact boundary definition between the Ediacaran and the Cambrian (Geyer and Landing, 2017).

Secular variations in seawater $^{87}Sr/^{86}Sr$ ratios have been widely applied to reflect continental uplift, silicate weathering and variations in atmospheric carbon dioxide concentration (Burke et al., 1982; Keto and Jacobsen, 1987; Kaufman et al., 1993; Derry et al., 1994; Montañez et al., 1996; Denison et al., 1998; Montañez and Banner, 2000; Ebneth et al., 2001; Thomas et al., 2001; Melezhik et al., 2009; Sawaki et al., 2010). The $^{87}Sr/^{86}Sr$ ratio of seawater is controlled by a balance between continental weathering-derived Sr, hydrothermal Sr input and seafloor weathering (Palmer and Edmond, 1989). Continental source generally yields more radiogenic $^{87}Sr/^{86}Sr$ ratios due to the more radiogenic ^{87}Sr resulting from the excess decay of incompatible ^{87}Rb (rubidium) in evolved differentiated crustal rocks, while hydrothermal and seafloor weathering origin Sr deplete in radiogenic ^{87}Sr and therefore low $^{87}Sr/^{86}Sr$ ratios. Neglectable isotopic fractionation of the Rb/Sr system occurs during marine authigenic carbonate formation (Banner and Kaufman, 1994), together with its long residence time in the modern oceans, Sr is homogeneously distributed in the oceans and the seawater $^{87}Sr/^{86}Sr$ ratio ideally represents a global signature (DePaolo and Ingram, 1985). Thus, well-preserved marine carbonate minerals likely preserve the $^{87}Sr/^{86}Sr$ signature of the ambient seawater during the carbonate deposition.

In this study, we present C and Sr isotope compositions obtained in carbonates from the late Ediacaran to the late Cambrian at the Shiairik and Xiaoerbrak sections in the Keping region, Tarim Basin, NW China. The main objective of this study is to 1) document the variations of seawater C and Sr isotope compositions recorded in carbonates deposited in the Tarim Basin covering the late Ediacaran-late Cambrian interval; 2) refine the stratigraphic correlation of the sedimentary successions within the Tarim Basin; 3) interpret the data in the context of reconstruction of continental weathering and sea-level changes. Our new interpretations further aim to contribute to understanding of mechanisms linking environmental changes and animal evolution during this critical period.

2. Geological setting

The Tarim Basin was presumably paleogeographically situated in a low to middle latitude position between Australia and South China (see Fig. 1A) during the late Ediacaran to the late Cambrian. Both Tarim and its adjacent South China craton developed widespread shallow water carbonate platforms (McKerrow et al., 1992; Merdith et al., 2017; Huang et al., 2000; Zhan et al., 2007). Our studied Keping region is located at the north-western Tarim Basin, NW China (Fig. 1B) where the Ediacaran-Cambrian carbonate successions are outcropped at the Xiaoerbrak and Shiairik sections (GPS locations: N 40° 55′ 23″, E 79° 53′ 53″ and N 40° 59′ 12″, E 79° 59′ 35″ respectively).

2. 1 Shiairik section

The Shiairik section is located 30 km SW of Aksu city, NW Xinjiang Uygur Autonomous Region (Fig. 1B). This section records an Ediacaran-Cambrian succession of more than 200 m sedimentary rocks including the Ediacaran Sugetbrak and Chigebrak Formation (abbreviated throughout the manuscript Fm.) as well as the Cambrian Yurtus Fm. The Ediacaran Chigebrak and the overlying Cambrian Yurtus Fm. are separated by a disconformity between them (Fig. 2).

The Sugetbrak Fm. is mainly composed of reddish quartz sandstones interbedded with mudstone and dolostone layers and is in conformable contact with the overlying Chigebrak Fm. Rhythmic stratification, which reflects the influence of tidal currents, is widely developed in the middle and upper part of the Sugetbrak Fm., suggesting a tidal flat depositional environment (Deng et al., 2019). The Chigebrak Fm. contains thin-layered limestones interbedded with brown sandstones at the bottom, overlain by medium-thick (10-50 cm) layers of dolostones in the middle part and grey and yellowish medium-thick-layered dolostones in the upper part (He et al., 2007; Wang et al., 2010).

The Chigebrak Fm. is in contact with the lower Cambrian Yurtus Fm. by a widely distributed unconformity in the Tarim Block (Wu et al., 2018). In the strata just below the unconformity, various features of karstification including breccias, cave-sediment infills, solution vugs and pores, were observed in the uppermost layers of dolostones in the Chigebrak Fm., indicating a possible erosional surface (He et al., 2018).

The lower Cambrian Yurtus Fm. mainly comprises black shales and phosphatic siliceous rocks with barite concretions, syn-genetic paleo-weathering crusts and abundant benthic SSFs (Qian and Xiao, 1984; Qian et al., 2000; He et al., 2007; Peng, 2009; Qian et al., 2009; Chen et al., 2010; Yao et al., 2014).

In the Shiairik section, 49 carbonate samples were collected from the 156 m-thick Ediacaran Chigebrak Fm. (n = 41) and the top part of the underlying Sugetbrak Fm. (n = 8), respectively (Fig. 2).

2. 2 Xiaoerbrak section

The Xiaoerbrak section is located 50 km SW of Aksu city, NW Xinjiang Uygur Autonomous Region and comprises more than 800 m sedimentary rocks subdivided in ascending stratigraphic order into the Cambrian Yurtus, Xiaoerbrak, Wusonger, Shayilik, Awatagh and Xiaqiulitagh Fm. (Fig. 1 & 2), which are stratigraphically continuous without any depositional gaps.

The 21 m thick Yurtus Fm. is subdivided into two parts: the lower Yurtus Fm. consists of black shales, phosphatic siliceous rocks with the occurrence of sponge spicules (Peng, 2009; Chen et al., 2010) and rare SSFs (*Protohertzina Anabarica Missarzhevsky* and *Anabarites* (Zhou and Chen, 1990; Zhou, 2001), Fig. 2), while the upper part comprises thin-layered (1-10 cm) siliceous dolostones and muddy dolostones containing abundant SSFs (Qian and Xiao, 1984; Zhou and Chen, 1990; Qian et al., 2000; Zhou, 2001; Qian et al., 2009; Chen et al., 2010), such as *Paragloborilus Spinatus* (Qian and Xiao, 1984) and *Lapworthella* (Zhou and Chen, 1990). The sedimentary facies of the Yurtus Fm. are described in the literature as deposited in a restricted basin environment inferred from sedimentologic and geophysical studies as well as from molybdenum isotope compositions analyzed in the Yurtus black shales (Bai et al., 2019; Yao et al., 2014; Yao et al., 2017; Yu et al., 2009).

The conformably overlying Xiaoerbrak Fm. consists of 138 m-thick medium-thick layered micritic dolostone, which hosts three trilobite assemblages (in ascending stratigraphic order: *Shizhudiscus* Zone, *Metaredlichioides* or *Ushbaspis* Zone, *Kepingaspis-Tianshanacephalus* Zone) (Wang et al., 1985; Zhou and Chen, 1990; Zhou, 2001; Peng, 2009; Fig. 2).

The 95 m-thick Wusonger Fm. is mainly composed of fine-grained micritic dolomites and muddy dolostones (Sun et al., 2004; Yu et al., 2004; Wang et al., 2010). Fossils of *Paokannia* considered to survive from the Cambrian late Stage 3 to the end of Cambrian Series 2 were identified in this formation (Zhou, 2001; Peng, 2009, Fig. 2 orange arrow indicates fossil range). Semi-restricted carbonate platforms were indicated from thin-section studies of the muddy dolostone (Bai et al., 2019).

The overlying Shayilik Fm. contains the *Kunmingaspis-Chittidilla* biozone and comprises micritic and fine-grained dolomites in the lower part, dolomite-bearing limestones in the upper part, and is topped by algal microbialites (Zhou, 2001; Peng, 2009, Fig. 2 yellow arrow indicates fossil range).

The Awatagh Fm. spans about 200 m and comprises anhydrite and anhydrite-bearing dolostones at the bottom overlain by purple medium-thin (5-30 cm) bedded micritic dolomites and fine-grained dolomites in the upper part. Bai et al. (2019) studied the sedimentary characteristics of the Shayilik and Awatagh Fm. and proposed their sedimentary environments experienced a transgression and changed from gentle slope environments to restricted platforms. The overlying Xiaqiulitagh Fm. mainly consists of dolomites, intercalated with stromatolitic dolomites (Wang et al., 2011b) and siliceous dolomites exposed in the middle part of this Fm.

At the Xiaoerbrak section, 126 carbonate samples and three phosphorite samples were collected (Fig. 2). Based on regional lithostratigraphic and biostratigraphic correlation as well as sedimentary facies comparison, Shiairik and Xiaoerbrak sections are integrated into one single lithological succession in this study (Fig. 1B).

3. Analytical methods

Hand specimen free from visible alteration such as weathering crusts and visible calcite veins were collected in the field and subsequently crushed into small pieces of ~0.5 cm diameter. The best-preserved pieces (i.e. showing no microscopic calcite veins) were handpicked and ground into fine powder in an agate mill (200 mesh) for further geochemical analyses.

0.5 mg of sample powders were baked for ten hours at 70 °C and then reacted with orthophosphoric acid at 70 °C for 1.5 hours to extract CO_2 (McCrea, 1950; Craig, 1953) for subsequent measurement of C and oxygen (O) isotope compositions. C and O isotope compositions were determined using a Gas-isotope ratio mass spectrometer (IRMS) of type Finnigan Delta Plus XP at the State Key Laboratory for Mineral Deposits Research, Nanjing University, China. Chinese carbonate reference materials (TTB-1 and TTB-2) were analyzed along with our samples for isotopic calibration and to test for memory effects. C and O isotope compositions were measured relative to analyses of an in-house standard and the obtained data is here expressed relative to the international V-PDB (Vienna Pee Dee Belemnite) reference material against which our in-house standard was calibrated. We use the δ notation throughout the manuscript, which is the permille deviation of the C and O isotope compositions from the V-PDB reference material. The analytic precision of our method was determined on repeated analyses of the external reference materials and is in the order of ±0.15 ‰ for $\delta^{13}C$ and ±0.1 ‰ for $\delta^{18}O$ respectively (2SE) (Table 1).

For the measurements of the concentrations of calcium (Ca), magnesium (Mg), manganese (Mn), Sr and Rb in the carbonate fraction of the studied samples, ~50 mg of sample powder was weighed into a 15 ml centrifuge tube. First, 3 ml 1 M sub-boiling double-distilled acetic acid (HAc) were added and the tubes placed in an ultrasonic bath for 30 minutes. Then the samples reacted at room temperature for 12 hours in order to sufficiently dissolve carbonate minerals. Finally, the solutions were centrifuged for 15 minutes at 4000 rpm, and the supernatant was placed into a 15 ml PFA beaker.

Subsequently, 2 ml ultrapure water was added into the 15 ml centrifuge tube to wash the remaining residue. After being placed in an ultrasonic bath for another 30 minutes and centrifugation for 15 minutes, the supernatant solutions were added into the 15 ml PFA beakers and this rinsing step was repeated. The combined 7 ml of supernatant (dissolved carbonate fraction) were then dried down at 120 °C and re-dissolved in 3% v/v double-distilled nitric acid with a dilution factor of 1:1000 for concentration measurements (the detailed dissolution procedure summarized in Fig. 3). Mn, Sr and Rb concentrations were determined using a Thermo Finnigan Element XR sector-field ICP MS at the State Key Laboratory for Mineral Deposits Research (Nanjing University, PR China) with a Scott type quartz spray chamber and a 100 µl/min nebulizer. For instrument drift correction, we doped 10 ppb rhodium into the sample solutions for elements analyzed in low-resolution mode. The Ca and Mg concentrations were determined using a Sky-ray ICP OES at the same laboratory. Along with the ICP OES analyses and the ICP MS analyses, a dolomite reference material (JDo-1) was repeated one time per 10 analyses, the repeated analyses of JDo-1 on ICP MS and ICP OES yielded < 10 % accuracy and < 5 % precision.

Samples with relatively high Sr concentrations and low Mn/Sr ratios were selected for Sr isotope measurements. Solutions containing around 20 ng Sr were dried at 120 °C and dissolved in 1 ml 3 M double-distilled nitric acid followed by standard Sr-spec resin ion-chromatography for Sr purification. Sr isotope compositions were obtained on an MC-ICP-MS (Thermo Neptune Plus) at the State Key Laboratory for Mineral Deposits Research in Nanjing University, China. Mass fractionation was corrected by applying an assumed $^{86}Sr/^{88}Sr$ ratio = 0.1194 and the exponential law (Nier, 1938). The NIST SRM 987 standard was analyzed along with the samples and yielded a mean value of 0.710271±30 (2SE, n= 30). The Sr isotope ratios and their corresponding standard error of the mean are listed in Table 1. Data mentioned in the text has been normalized to the recommended NIST SRM 987 standard value of 0.710248 (McArthur et al., 2012).

4. Results

4.1 C isotope feature at the Precambrian/Cambrian (PC/C)

All C isotope data obtained on carbonate leachates are reported in the delta (δ) notation relative to V-PDB and the $\delta^{13}C$ values of all analyzed carbonate samples are plotted in Fig. 4 and tabulated in Table 1, respectively. The $\delta^{13}C$ values fluctuate between -13.3 ‰ and +7.9 ‰.

The $\delta^{13}C$ curve obtained in carbonates from the Sugetbrak Fm. to the Xiaquilitagh Fm. shows four prominent negative excursions (“B”, “C”, “E”, “F”, Fig. 4, here we define “excursion” as a visible deviation, i.e. more than 10 % offset from their pre-excursion background values) and three positive excursions (“A”, “D”, “G”, Fig. 4).

The first $\delta^{13}C$ positive excursion appears at the boundary between the Chigebrak Fm. and Sugetbrak Fm. In this interval, the $\delta^{13}C$ values increase from -5.5 ‰ to +7.9 ‰ (Fig. 4, positive excursion “A”), and then plunge down to ~+1 ‰ resulting in a small negative excursion (Fig. 4, negative excursion “B”). Above the negative excursion “B”, the $\delta^{13}C$ values increase to a steady background value around +2.5 ‰ through the rest of the Chigebrak Fm.

The $\delta^{13}C$ values increase from -2.5 ‰ to +0.5 ‰ in the lower part of the Yurtus Fm., decrease rapidly to -5.1 ‰ in the middle part (“C-1”, Fig.4), and then increase again to +0.9 ‰ over a short interval in the upper part of the Yurtus Fm. (“C-2”, Fig. 4). At the top of the Yurtus Fm., the $\delta^{13}C$ values slowly decrease to -0.5 ‰ (“C-3”, Fig. 4).

The whole Xiaoerbrak Fm. demonstrates a prolonged positive C excursion (“D”, Fig. 4, $\delta^{13}C$ values vary from -0.5 ‰ to +3 ‰) with two negative swings. One occurs at the base of the Xiaoerbrak Fm., varying from +2.5 ‰ to +1.7 ‰, and then recovers to +3 ‰ (“D-1”, Fig. 4). The other negative excursion occurs in the middle part of the Xiaoerbrak Fm., decreasing from +3 ‰ to +1.8 ‰, and then recovering to +2.5 ‰ (“D-2”, Fig. 4).

At the boundary between the Wusonger and the Xiaoerbrak Fm., the $\delta^{13}C$ values gradually decrease to ~0 ‰ followed by a plunge to -2.0 ‰. The $\delta^{13}C$ values subsequently recover to ~0 ‰ at the top of the Wusonger Fm., completing the negative excursion “E” (Fig. 4). The $\delta^{13}C$ values decrease sharply to a prominent negative nadir of -13.3 ‰ forming the most negative excursion “F” in our dataset (Fig. 4) at the bottom of the Shayilik Fm. In the upper part of the Shayilik Fm., the $\delta^{13}C$ values recover to ~0 ‰.

A $\delta^{13}C$ positive excursion (Fig. 4, positive excursion “G”) varying from -1.5 ‰ to +1.5 ‰ was identified in the mid-upper part of the Awatagh Fm. The positive excursion “G” consists of positive excursion “G-1” (varying from -1.5 ‰ to +1.5 ‰) and “G-2” (varying from -1.5 ‰ to 0 ‰). Subsequently, almost the entire Xiaqiulitagh Fm. sustains a constant $\delta^{13}C$ value at ~-1.0 ‰ with a minor rise to ~0 ‰ at the top of this formation.

4.2 O isotope compositions at the PC/C

The here presented O isotope data are obtained on carbonate leachates and reported in the delta (δ) notation relative to V-PDB. The $\delta^{18}O$ values of the analyzed carbonates are plotted in Fig. 4 and tabulated in Table 1. The $\delta^{18}O$ values from the Ediacaran samples (from the Shiairik section) systematically differ from those obtained in the Cambrian samples (from the Xiaoerbrak section) (Fig. 4). The $\delta^{18}O$ values in the Ediacaran stratigraphic units range from -9 ‰ to 0 ‰ and have a mean value of -3 ‰. By contrast, almost all samples in the Cambrian units have $\delta^{18}O$ values ranging from -12 ‰ to -3 ‰ (mean= -7 ‰) except some samples in the Yurtus Fm. which show more negative values.

Sawaki et al. (2010) reported $\delta^{18}O$ values ranging from -10 ‰ to 0 ‰, averaging at -4 ‰ for the Ediacaran Three Gorges area, South China. Besides, Wang et al. (2011b) obtained $\delta^{18}O$ values ranging from -7.9 ‰ to +3.5 ‰ through the Cambrian Series 2 to Series 3 in the Penglaiba section of the Keping region. Thus, our $\delta^{18}O$ values of the two studied section are generally consistent with existing Ediacaran and Cambrian datasets.

4.3 Sr isotope compositions at the PC/C

Samples selected for Sr isotope measurements yield low $^{85}Rb/^{86}Sr$ ratios (all samples < 0.0016) (Table 1), suggesting that the chromatographic Sr separation from matrix elements using the specific Sr resin was effective. Further, samples with low Mn/Sr ratios (< 5) and relatively high Sr concentrations (generally > 50 ppm) were selected to perform Sr isotope analyses to minimize possible influence of diagenetic alteration (see discussion in section 5.1.2).

As shown in Fig. 4, the obtained $^{87}Sr/^{86}Sr$ ratios in our carbonate leachates vary between 0.7085 and 0.7102. From the uppermost part of the Sugetbrak Fm. to the middle Chigebrak Fm., the $^{87}Sr/^{86}Sr$ values decrease from ~0.7095 to ~0.7085. Then the $^{87}Sr/^{86}Sr$ values increase to ~0.7090 at the top of the Chigebrak Fm.

In the Yurtus Fm., the $^{87}Sr/^{86}Sr$ ratios change rapidly within the range of 0.7086 to 0.7102, followed by a plateau at ~0.7088 through most of the Xiaoerbrak Fm. Finally, a rise to ~0.7090 at the top of the Xiaoerbrak Fm. coincides with the positive $\delta^{13}C$ excursion “D” and a $\delta^{13}C$ platform at 0 ‰ above the excursion “D”. In the Wusonger Fm., the $^{87}Sr/^{86}Sr$ ratios rise slowly from ~0.7088 to ~0.7090 and sharply increase to 0.7092 at the bottom of the Shayilik Fm.

At the bottom of the Shayilik Fm., the recorded $^{87}Sr/^{86}Sr$ ratios drop to ~0.7087, coinciding with the prominent negative $\delta^{13}C$ excursion “F”. Subsequently, the $^{87}Sr/^{86}Sr$ ratios gradually increase to ~0.7092 in the mid-upper part of Awatagh Fm., coinciding with the rising limb of the positive $\delta^{13}C$ excursion “G”. Apart from only one outlier at 0.7086 in the lower part of Xiaqiulitagh Fm., the $^{87}Sr/^{86}Sr$ ratios recorded in carbonates from the rest of the Cambrian (i.e. Xiaqiulitagh Fm. and Upper Awatagh Fm.) is constant at ~0.7091.

5. Discussion

5.1 Assessing possible processes altering the C/Sr isotope compositions

5.1.1 The effect of silicate leaching and dolomitization on the $^{87}Sr/^{86}Sr$

To evaluate the influence of possible silicate leaching with our digestion method on the obtained $^{87}Sr/^{86}Sr$ ratios, a cross plot of $^{87}Sr/^{86}Sr$ versus [Rb] (i.e. Rb concentration normalized by CaO+MgO) is provided in the Fig. 6G and 6N. The absence of apparent correlations between Sr concentrations as well as $^{87}Sr/^{86}Sr$ ratios and [Rb] indicates our digestion method avoids substantial silicate mineral leaching as these contain much higher Rb contents and more radiogenic Sr than carbonate minerals. Moreover, there are no apparent correlations in all analyzed carbonates in the cross plots of $^{87}Sr/^{86}Sr$-Mg/Ca and $^{87}Sr/^{86}Sr$-carbonate content (%) (Fig. 6E, 6F, 6L, 6M), indicating $^{87}Sr/^{86}Sr$ ratios may not be influenced by dolomitization. Thus, we consider our sample digestion method as a suitable method to treat ancient marine carbonates, and our obtained $^{87}Sr/^{86}Sr$, $\delta^{18}O$ and $\delta^{13}C$ values as suitable proxies to record ambient seawater compositions (section 5.1.2 and 5.1.3).

5.1.2 The influence of diagenetic alteration

Any possible diagenetic overprint superimposed on the primary geochemical seawater signatures of marine carbonates (e.g. late-stage fluid alteration causing carbonate dissolution and reprecipitation) requires to be examined in detail, and any altered samples need to be rigorously screened out. In general, typical chemical characteristics of carbonates that experienced a high degree diagenetic alteration yield elevated Mn/Sr ratios, low $\delta^{18}O$ values and possibly low $\delta^{13}C$ values and elevated $^{87}Sr/^{86}Sr$ ratios (e.g. Marshall, 1992; Kaufman et al., 1993; Derry et al., 1994). Thus, cross-plots of Mn/Sr ratios, $\delta^{18}O$, $\delta^{13}C$ and $^{87}Sr/^{86}Sr$ ratios are useful and widely used tools to evaluate the degree of alteration. Moreover, Mn/Sr ratios and $\delta^{18}O$ values are considered useful diagnostic tools of late-stage alteration of water-rock interaction (Marshall, 1992; Derry et al., 1994; Kaufman and Knoll, 1995). Almost all the dolostone samples in this study, as well as those from other sections in the Keping area (e.g. Guo et al., 2017), have low Sr concentrations (< 300 ppm) which is possibly indicative for low initial seawater Sr concentrations at the time of deposition (cf. Li et al., 2011). Here we selected more generous cut-offs for our studied dolostone samples during our pre-screening: samples that clearly deviate from the main data group (Mn/Sr > 5 and $\delta^{18}O$ < -10 ‰, Fig. 5) are considered as not representing primary seawater $^{87}Sr/^{86}Sr$ signals (Derry et al., 1994; Kaufman and Knoll, 1995; Li et al., 2013; Zhou et al., 2020).

Nearly all Ediacaran samples show isotopically heavier $\delta^{18}O$ values (> -10 ‰) and lower Mn/Sr (< 5) (Fig. 5). However, carbonates from the Sugetbrak Fm. show a positive correlation between $\delta^{18}O$ and $\delta^{13}C$ ($R^2 = 0.95$; Fig. 5A). Besides, correlations between radiogenic $^{87}Sr/^{86}Sr$ ratios and low $\delta^{13}C$ and $\delta^{18}O$ values are present ($R^2 = 0.89$ and 0.95 respectively; Fig. 6H and 6I), suggesting that the Sugetbrak Fm. might have suffered substantial diagenetic alteration by basin-type fluids which enriched in "light" C and O and radiogenic Sr isotopes (see Hohl et al., 2015b).

Most Cambrian samples show no apparent correlations in the cross-plot of $\delta^{13}C$ versus $\delta^{18}O$ (Fig. 5B). However, carbonate leachates of the Yurtus Fm. show elevated Mn/Sr ratios (> 5; circles in Fig. 6D) and a positive correlation between $^{87}Sr/^{86}Sr$ ratios and $\delta^{18}O$ values (Fig. 6B), which indicates that the obtained Sr isotope compositions in the Yurtus Fm. have been influenced by the mixing of restricted basin Sr signals with less radiogenic fluids (see section 5.1.3). However, the obtained $\delta^{13}C$ values in the Yurtus Fm. carbonates show no correlation with obtained $^{87}Sr/^{86}Sr$ and $\delta^{18}O$ values (Fig. 6A and 5B), suggesting that the carbonates in the

Yurtus Fm. likely preserved the primary seawater C isotope composition.

In the Wusonger Fm., the $\delta^{18}O$ value of one sample yields -10 ‰ and its $\delta^{13}C$ value is similarly low (-6.7 ‰) (Fig. 5B). Therefore, the sample is excluded from the seawater C and Sr evolution curve (in sections 5.2 and 5.3). Likewise, the $\delta^{18}O$ value of another data point in the Shayilik Fm. is lower than -10 ‰. However, no paired decrease in the $\delta^{13}C$ excludes the possibility of significant post-depositional fluid alteration for this sample.

Apart from these obviously altered samples, there is no clear correlation in the cross-plots of $\delta^{18}O$-$\delta^{13}C$ (Fig. 5), $^{87}Sr/^{86}Sr$-$\delta^{18}O$, $^{87}Sr/^{86}Sr$-$\delta^{13}C$, $^{87}Sr/^{86}Sr$-[Sr], $^{87}Sr/^{86}Sr$-Mn/Sr, $^{87}Sr/^{86}Sr$-Mg/Ca and $^{87}Sr/^{86}Sr$- Carbonate content (%) (Fig. 6). Thus, we consider these samples are well-preserved marine carbonates, of which their $^{87}Sr/^{86}Sr$ and $\delta^{13}C$ values may record primary seawater compositions.

5.1.3 Fluid mixing model for Sr isotope compositions in the Yurtus Fm.

Samples of the Yurtus Fm. show elevated Mn/Sr ratios (> 5; circles in Fig. 6D), arguing for excess in Mn possibly due to pore water-redox controlled remobilization of MnO colloids (Cowie et al., 1995; Thomson et al., 1998; Hohl et al., 2017). Besides, carbonate leachates of the Yurtus formation show a positive correlation between $^{87}Sr/^{86}Sr$ ratios and $\delta^{18}O$ values (Fig. 6B) with two endmembers, one with the least-radiogenic $^{87}Sr/^{86}Sr$ ratios and light $\delta^{18}O$ values, while the other endmember has more radiogenic Sr and heavier O isotopic compositions. However, diagenetic basin-type fluid-alteration usually results in a negative correlation between $^{87}Sr/^{86}Sr$ ratios and $\delta^{18}O$ values (Hohl et al., 2015b).

In order to explore the mechanisms that may generate the full range in O and Sr isotope compositions observed in the Yurtus Fm., a simple open-system fluid mixing model (Jacobsen and Kaufman, 1999) is applied. As inferred from the paired Sr and O isotope results, two endmembers have been identified: one is a radiogenic localized basin-type seawater ($^{87}Sr/^{86}Sr$ = 0.7102, $\delta^{13}C$ = 0.5 ‰, $\delta^{18}O$ = 0.5 ‰ and Sr = 50 ppm), while the other endmember is a meteoric fluid with $^{87}Sr/^{86}Sr$ of 0.7085, $\delta^{13}C$ of -5 ‰, $\delta^{18}O$ of -14 ‰ and 600 ppm Sr (Fig. 6A, 6B and 6C). Data of the Yurtus carbonates falls on a mixing trend line between the localized seawater and a lighter $\delta^{18}O$ bearing meteoric fluid, exhibiting mixing ratios between 0.01 to 5 (Fig. 6A, 6B, 6C).

In general, meteoric fluids comprise more radiogenic Sr and lower Sr concentrations than contemporaneous seawater (e.g. Xu and Han, 2009). However, the hypothetic global regression at the Precambrian/Cambrian boundary presumably causing the 'great unconformity' (Brasier, 1996; Ishikawa et al., 2008; Sawaki et al., 2008; He et al., 2017) might have exposed extensive amounts of Precambrian marine carbonates on the shelf, and thus, meteoric fluids leaching these lithologies, might have inherited their Ediacaran seawater-like Sr isotopic compositions and Sr concentrations. He et al. (2018) suggested a wide geographical distribution of paleo-karsts at the top of the Chigebrak Fm. and speculated that there is an Ediacaran carbonate paleo-upland adjacent to the Xiaoerbrak and Shiairik sections during this time interval. Furthermore, similar Sr isotope composition and Sr concentration of the Chigebrak Fm. carbonates with those of our modelled meteoric fluid ($^{87}Sr/^{86}Sr$ ratio ~0.7085; Sr concentration ~600 ppm) argue for a contribution to the carbonate $^{87}Sr/^{86}Sr$ in the Yurtus Fm. from such an endmember.

On the other hand, the Sr isotope composition of the seawater endmember (~0.7102) in this study is more radiogenic when compared to the reported $^{87}Sr/^{86}Sr$ seawater ratio from the correlative strata in South China (~0.7085, Li et al., 2013), which may be explained by a local rather than global signature, arguing for a restricted basin depositional environment on the shelf of the Tarim craton during this period (cf. Bai et al., 2019; Yao et al., 2017).

Several syn-genetic ancient weathering crusts in the Yurtus Fm. imply subaerial exposure environments and the occurrence of black shale in the lower part of the Yurtus Fm. indicate facies change to deep-water deposition. Hence the two depositional settings probably represent sea-level fluctuations. Because the early Cambrian Tarim Basin was presumably restricted and underwent several second-order sea-level shifts (Yao et al., 2017), we argue that meteoric fluids leaching from aerial exposed Ediacaran shelf carbonates are mixed with confined basin waters. The basin may be localized at the late-Ediacaran and the basin water may be influenced by enhanced silicate weathering of hinterland during the global regression at the late-Ediacaran (Haq and Schutter, 2008).

5.2 C isotope stratigraphic and bio-stratigraphic correlation

Before performing a stratigraphic correlation between the Tarim block and other time-equivalent paleo-continents, it is essential to note that our obtained C isotope variation curve can be correlated to the existing datasets from other sections in the Tarim Basin. For example, similar rapid increasing $\delta^{13}C$ trends in the top part of the Sugetbrak Fm. have been reported at a composite profile from seven sections in the Quruqtagh area and at the Penglaiba section in the Keping area by Xu et al. (2003) and He et al. (2007), respectively. Further, Guo et al. (2017) reported extreme negative $\delta^{13}C$ values (down to -14.3 ‰) at the boundary of the Wusonger/Shayilik Fm. in the Penglaiba section in the Keping area, similar to our negative $\delta^{13}C$ excursion “F” with respect to the total magnitude of the excursion (down to -13.3 ‰).

During the PC/C, the Tarim Basin was situated in a low latitude location on the continental margin shelf (Yu et al., 2009), leading to the deposition of shallow water carbonate sediments. This makes the Tarim Basin sections lithologically and paleogeographical similar to other shallow marine PC/C sections on the Yangtze Platform (McKerrow et al., 1992; Merdith et al., 2017; Huang et al., 2000; Zhan et al., 2007; Hohl et al., 2015a). Therefore, we here apply the Ediacaran/Cambrian bio-chemostratigraphic division of South China as the primary reference frame for the correlation with our studied bio-chemostratigraphic data from the two sections. Fig. 7 shows the carbonate C isotope evolution curve from the Keping region and our proposed C isotope stratigraphic correlation with the reported composite carbonate C isotope evolution curves of South China from the late Ediacaran (Jiang et al., 2007), across the Ediacaran-Cambrian transition (Li et al., 2013), and through the Cambrian Series 2 to Series 3 (Guo et al., 2010). On a global scale, we also compare our Tarim Basin carbonate C isotope evolution curve with records from North India during late Ediacaran (Kaufman et al., 2006) and the Cambrian global seawater standard $\delta^{13}C$ variations proposed by Zhu et al. (2006), which is mainly derived from South China shallow-water carbonates. Similarly, carbonate C isotope datasets from Laurentia (Montañez and Banner, 2000) and Siberia (Wotte et al., 2011) spanning the Cambrian Series 2 – Series 3 strata and carbonate C isotope data from Southwest Mongolia spanning the Ediacaran-Cambrian transition (Brasier et al., 1996) are used for correlation.

According to the current Ediacaran chronostratigraphy proposed by Zhu et al. (2007), the upper Doushantuo negative $\delta^{13}C$ excursion (DOUNCE, equivalent to the Shuram/Wonoka excursion) ranges from ca. 560 Ma to 551 Ma and the following positive $\delta^{13}C$ excursion that ended at around 548 Ma, which provide a general baseline reference for C isotope stratigraphy in the mid-late Ediacaran. In South China, a sharp $\delta^{13}C$ rise predates the positive $\delta^{13}C$ excursion at the boundary between the Doushantuo and the Dengying Fm. (e.g. Jiang et al., 2007). Similar $\delta^{13}C$ variations were observed at the boundary between the Sugetbrak and Chigebrak Fm. in the Tarim Basin which lead to the hypothesis that the Chigebrak Fm. is stratigraphically equivalent to the Dengying Fm. in South China (He et al., 2007). The studied Shiairik section is merely 40 km away from the site investigated by He et al. (2007) and similarly obtained carbonate $\delta^{13}C$ curves at both sites suggest that the Chigebrak carbonates have only experienced minimal diagenetic alteration and thus, may archive the $\delta^{13}C$ signals of coeval Ediacaran paleo-seawater. As mentioned above, the positive excursion “A” occurring at the boundary between the Chigebrak and Sugetbrak Fm. is consistent with the Ediacaran division plan proposed by Zhu et al. (2007) and He et al. (2007). Since, late Ediacaran South China strata bears no other obvious negative $\delta^{13}C$ excursion above the ca. 548 Ma old positive $\delta^{13}C$ excursion (Fig. 7I) while North Indian strata of the same time reveals two negative $\delta^{13}C$ excursions (Fig. 7J), the negative excursion “B” obtained in the Tarim Basin carbonates cannot be considered as a global $\delta^{13}C$ excursion.

A slight $\delta^{13}C$ decline at the top of Chigebrak Fm. is similar to the $\delta^{13}C$ excursion at the top of the Dengying Fm. (Fig. 7I) and a subsequent sharp decrease (negative Excursion “C”) across the top of the Chigebrak Fm. and the lower Yurtus Fm. is suggested as equivalent to the globally identified BACE (BAsal Cambrian C isotope Excursion) (Fig. 7B, 7G, 7H). This interpretation is consistent with previous studies in which the parallel unconformity between the Chigebrak Fm. and the Yurtus Fm. is considered to represent the boundary between the Ediacaran and Cambrian in the Tarim Basin (Yao et al., 2014; He et al., 2018).

The appearance of SSFs, that are widely distributed during the Terrenenvian Series (Zhu et al., 2006), can further be used for robust biostratigraphic correlation. In South China, the *Anabarites-Protohertzina-Arthochites* assemblage, the first Terreneuvian SSFs assemblage in the Southwest Chinese Province of Yunnan, agrees well with the earliest skeletal faunas on the Siberian Platform and the Canadian Cordillera (Qian and Bengtson, 1989; Steiner et al., 2004;

Yang et al., 2014a; Yang et al., 2014b). The lower Yurtus Fm. in the Tarim Basin, which contains an *Anabarites-Protohertzina* assemblage (Qian and Xiao, 1984), can be correlated to the lower Terreneuvian. Moreover, because the second assemblage (*Siphogonuchites-Paragloborilus*) and the last assemblage (*Lapworthella-Tannuolina-Sinosachites*) found in Yunnan (Qian and Xiao, 1984; Qian and Bengtson, 1989; Yang et al., 2014a) also appear in the upper Yurtus Fm. (i.e. *Paragloborilus-Lapworthella* assemblage), we suggest that all or most of the Yurtus Fm., conformably underlying the trilobite-bearing Xiaoerbrak Fm. can be best assigned to the Terrenenvian Series.

The Terrenenvian Series (the first series of the Cambrian) is characterized by more than three carbonate δ^{13}C excursions recorded in South China sections, i.e. BACE, ZHUCE (ZHUjiaqing C isotope Excursion) and SHICE (SHIyantou C isotope Excursion) (Fig. 7B). One negative δ^{13}C excursion (Fig. 4, “C-2”) and two positive δ^{13}C excursions (Fig. 4, “C-1” and “C-3”) are recorded in the Yurtus Fm. carbonates, not perfectly matching the reported excursions’ amounts and magnitudes during the Terreneuvian epoch in South China. As syn-genetic ancient weathering crusts were recognized in field observation, several depositional hiatuses in the Yurtus Fm. may explain the differences in the early Cambrian δ^{13}C profiles between South China and the Tarim Basin.

The Cambrian Stage 3 is marked by the first appearance of trilobites and two positive δ^{13}C excursions (Cambrian Arthropod Radiation isotope Excursion (CARE) and MIngxinsi C Isotope Excursion (MICE), Zhu et al., 2006). Due to the appearance of *Shizhudiscus* at the bottom of the Xiaoerbrak Fm., it may be assigned to the base of the Cambrian Series 2. Besides, due to the consistency in δ^{13}C variation trends (a major positive excursion with a sub-negative excursion) and magnitudes (increase from ~0 ‰ to +3 ‰) between the upper part of the positive δ^{13}C excursion “D” (with the sub-negative excursion “D-2”) and the MICE (Fig. 7), this study correlates the upper part of the excursion “D” as equivalent to the MICE in South China. Although the lower part of the excursion “D” is not perfectly matching in shape to the CARE, the similar magnitude of the excursion “D-1” (decreasing from +2.5 ‰ to +1.7 ‰) and the appearance of *Shizhudiscus* broadly confirms that the negative δ^{13}C excursion “D-1” likely corresponds to the falling limb of the CARE. Therefore, combining the evidence that the appearance of *Shizhudiscus* occurs slightly later than the start of the Cambrian Series 2 (Fig.

7B, the green arrow represents the strata with *Shizhudiscus* occurrence), the boundary between the Yurtus Fm. and the Xiaoerbrak Fm. is considered to be stratigraphically higher than the base of the Cambrian Stage 3 in the Tarim Basin.

The ROECE (Redlichiid–Olenellid Extinction C isotope Excursion; Zhu et al., 2006), the most negative δ^{13}C excursion throughout the Cambrian Series 2 to Furongian (Fig. 7C, 7D, 7E, 7F, e.g. Fan et al., 2011; Guo et al., 2017), is considered to represent the boundary between the Cambrian Stage 4 and the Wuliuan Stage (Montañez and Banner, 2000; Zhu et al., 2006; Guo et al., 2010; Wotte et al., 2011; Guo et al., 2017). Accordingly, we consider that the most negative δ^{13}C excursion “F” in our studied carbonates is equivalent to the ROECE and the bottom of the Shayilik Fm. probably represents the base of the Cambrian Series 3 (Fig. 7). This correlation is consistent with the stratigraphic correlations by Guo et al. (2017) and Wang et al. (2011b) and the appearance of the *Kunmingaspis-Chittdilla* (Fig. 2) in the Shayilik Fm. (Zhou, 2001; Peng, 2009), the biozone range of which overlaps with the Wuliuan Stage (Wang et al., 2011a, Fig. 7, the yellow arrow represent the strata with *Kunmingaspis-Chittdilla* occurrence). Below the negative δ^{13}C excursion “F”, by comparing the δ^{13}C variation magnitude, the negative δ^{13}C excursion “E” in the Wusonger Fm. may be equivalent to the Archaeocyathid Extinction C isotope Excursion (AECE, Zhu et al. (2006)). This hypothesis is further supported as *Paokannia* is considered to survive from mid-Cambrian Series 2 (late Stage 3) to the end of Cambrian Series 2 (Zhou, 2001; Peng, 2009) (Fig. 7, the orange arrow represents the *Paokannia*-bearing strata).

The positive δ^{13}C excursion “G” is the only apparent positive excursion overlying the negative δ^{13}C excursion "F", and this temporal distribution is similar to the chronological order between SPICE (Steptoean C-Isotope Positive Excursion) and ROECE. However, the positive δ^{13}C excursion “G” is not as distinct as the SPICE reported from other paleo-continents and its magnitude of 3‰ is smaller than the magnitude of the SPICE recorded elsewhere (~4-6‰) (Li et al., 2018), which may suggest depositional hiatuses existing at the base of the Furongian in the Keping area. Considering that no depositional hiatus is reported in the Awatagh Fm., we tentatively speculate that the positive excursion “G” is identical to SPICE but suggest further verification by independent sedimentological and geochemical proxies. Because the SPICE is the typical characteristic of the start of the Paibian stage of the Furongian Series (Saltzman et al., 1998; Peng et al., 2004; Zhu et al., 2006; Fan et al., 2011), a correlation of the positive excursion “G” with SPICE would mean that the upper Awatagh Fm. in the NW Tarim Basin may be preliminarily assigned to the Furongian Series.

5.3 Sr isotope stratigraphic correlation

Based on the carbonate C isotope stratigraphic correlation, a carbonate-based seawater $^{87}Sr/^{86}Sr$ curve (Fig. 8 red dashed line) is constructed using a “lower limit” Sr isotope data approach (Jones et al., 1994). Fig. 8 shows the obtained least-diagenetically altered $^{87}Sr/^{86}Sr$ ratios from carbonate leachates in the Tarim Basin of this study (Fig. 8D), a proposed standard Cambrian global seawater $^{87}Sr/^{86}Sr$ curve (Fig. 8A, Peng et al., 2012; Zhu et al., 2019) constructed after the work of Derry et al. (1994); Montañez and Banner (2000); Ebneth et al. (2001); Kouchinsky et al. (2008) as well as the $^{87}Sr/^{86}Sr$ values recorded in carbonates from the Yangtze Platform, South China during the Ediacaran-Cambrian transition period (Fig. 8B, C, Sawaki et al., 2008; Sawaki et al., 2010).

In this study, our inferred seawater $^{87}Sr/^{86}Sr$ ratios from the Keping region decrease from ~0.7090 to ~0.7085 in the lower part of the Chigebrak Fm. with a slight rebound to more radiogenic values (~0.7091) at the top of this formation. This trend mimics the variation recorded in the terminal Proterozoic Dengying and Doushantuo Fm. on the South China Yangtze Platform (Sawaki et al., 2010). Sawaki et al. (2008) and Hohl et al. (2015a) reported increasing $^{87}Sr/^{86}Sr$ ratios in the upper Dengying Fm., coinciding the similar increasing trends recorded in the contemporarily deposited upper Chigebrak Fm. carbonates (Fig. 8). However, the magnitude of the total rise in $^{87}Sr/^{86}Sr$ ratios is smaller than those recorded by Sawaki et al. (2008) but comparable to that reported by Hohl et al. (2015a). This may be explained with erosion and denudation of the uppermost Chigebrak carbonate platform (He et al., 2018) and the incomplete outcrop situation at the Xiaofenghe section in South China (Hohl et al., 2015a). As a summary from the above discussion, we consider the Chigebrak Fm. to be equivalent to the Dengying Fm., revealing consistent Sr and C isotope stratigraphy (Fig. 7).

As discussed above, the scattering $^{87}Sr/^{86}Sr$ ratios obtained in carbonates from the Yurtus Fm. are likely results of restricted-basin seawater mixed with carbonate weathering fluid (section 5.1.3, Fig. 6A, 6B and 6C) and thus may not be compared to the global seawater $^{87}Sr/^{86}Sr$ curve. We, therefore, exclude these data points from our Sr isotope composition curve as they may represent a restricted-basin signal (Bai et al., 2019; Yao et al., 2017) that is unsuitable for global correlations.

From the bottom of the Xiaoerbrak Fm. to the bottom of Shayilik Fm., the obtained $^{87}Sr/^{86}Sr$ ratios in carbonate leachates rise from ~0.7087 to ~0.7090, consistent with the reported seawater $^{87}Sr/^{86}Sr$ evolutionary curve during the Cambrian Series 2 (Montañez and Banner, 2000; Ebneth et al., 2001; Peng et al., 2012; Zhu et al., 2019) (Fig. 8). At the bottom of the Shayilik Fm., the $^{87}Sr/^{87}Sr$ ratios in the carbonates leachates decrease from ~0.7090 to ~0.7087, analogous to the global seawater $^{87}Sr/^{86}Sr$ curve at the Stage 4/Wuliuan Stage boundary (Peng et al., 2012; Zhu et al., 2019) (Fig. 8). Subsequently, $^{87}Sr/^{87}Sr$ ratios increase again at the middle part of the Shayilik Fm. and reach a peak of ~0.7092 in the middle part of the Awatagh Fm., coinciding the starting point of the $\delta^{13}C$ positive excursion “G” (Fig. 7). The SPICE and the maximum $^{87}Sr/^{87}Sr$ peak at around ~0.7092 characterize the boundary between the Guzhuangian and Paibian Stage (Montañez and Banner, 2000; Ebneth et al., 2001; Kouchinsky et al., 2008; Fan et al., 2011). As we inferred hereinbefore, based on the coincidence between the $^{87}Sr/^{87}Sr$ peak and the $\delta^{13}C$ positive excursion “G”, this study correlates the middle-upper part of the Awatagh Fm. approximately to the boundary between the Guzhuangian and Paibian Stage. Apart from only one outlier with a much lower $^{87}Sr/^{86}Sr$ value (~0.7086) in the lower part of Xiaqiulitagh Fm., the $^{87}Sr/^{86}Sr$ ratios slowly decrease from ~0.7092 to ~0.7091 from the middle part of the Awatagh Fm. to the Xiaqiulitagh Fm., which are broadly similar to those ratios reported by Ebneth et al. (2001) and Montañez and Banner (2000).

In this study, as a whole, the $^{87}Sr/^{86}Sr$ ratios and the C isotope variations obtained in carbonates from two sections of the Keping region spanning the late-Ediacaran to the late-Cambrian show global consistency (Fig. 7 and Fig. 8), except those analyses performed in presumably diagenetically altered Sugetbrak Fm. and the Yurtus Fm. carbonates. Our results from the stratigraphic correlation between the Keping area and other parts of the world from the late-Ediacaran to the late-Cambrian are as follows:

i. The Chigebrak Fm. is presumably equivalent to the Dengying Fm. of the South China

Yangtze Platform.

ii. The base of the Xiaoerbrak Fm., which contains *Shizhudiscus* fossils, is slightly stratigraphically higher than the base of the Cambrian Stage 3.

iii. The most negative $\delta^{13}C$ excursion “F” together with a decrease in $^{87}Sr/^{86}Sr$ characterize the bottom of the Shayilik Fm., which can be regarded as the boundary between the Cambrian Stage 4 and Cambrian Wuliuan in the Tarim Basin. This stratigraphic correlation is consistent with the existing paleontological constraints in the Tarim Basin (Peng, 2009).

iv. The carbonate samples of the upper-middle part of the Awatagh Fm. archive a positive $\delta^{13}C$ excursion “G” (likely equivalent to SPICE) and a peak value of $^{87}Sr/^{86}Sr$ ratios at ~0.7092, presumably marking the base of the Cambrian Paibian Stage.

5.4 Paleoenvironmental implications for the Tarim Basin at the PC/C

Overall, the obtained seawater $^{87}Sr/^{86}Sr$ ratios in carbonates from this study and those recorded in numerous carbonate successions from different paleo-continents show a stepwise increase from late Ediacaran to late Cambrian (Furongian) which is superimposed by several minor declines (Fig. 8 and Fig. 9). Meanwhile, the obtained $\delta^{13}C$ values in our study show four prominent negative and three positive excursions that are identical to trends found in various carbonate-based C isotope records at a global scale (Fig. 9).

5.4.1 Late Ediacaran paleoenvironmental reconstruction of the Tarim Basin

The integrated global carbonate $\delta^{13}C$ trend of late Ediacaran sections (Fig. 7) suggests that the carbonate $\delta^{13}C$ record from the Keping region probably archives a primary global seawater $\delta^{13}C$ signal during this interval. According to our C and Sr isotope stratigraphic correlation (Fig. 7 and Fig. 8), the lower part of the positive $\delta^{13}C$ excursion “A” in this study probably corresponds to the uppermost Doushantuo negative $\delta^{13}C$ excursion which ended at around 551 Ma (Zhu et al., 2007). It is controversial whether DOUNCE/Shuram-Wonoka is representative of the primary coeval seawater chemistry or not. Some researchers explained the extreme negative $\delta^{13}C$ values of the DOUNCE/Shuram-Wonoka event by global changes in the C isotopic composition of Ediacaran seawater as a result of the remineralization of a vast dissolved organic carbon reservoir (cf. Rothman et al., 2003; Fike et al., 2006). Others suggested a large flux of methane released from the seafloor (Bjerrum and Canfield, 2011) as the cause, while Derry (2010) argued for a large-scale post-depositional diagenetic alteration.

In this study, samples from the Sugetbrak Fm., which constitute for the most of the lower part of the positive excursion "A", exhibit a strong positive correlation in the $\delta^{13}C$-$\delta^{18}O$ plot (Fig. 5A) and negative correlations in the $^{87}Sr/^{86}Sr$-$\delta^{13}C$, $^{87}Sr/^{86}Sr$-$\delta^{18}O$ plots (Fig. 6H, 6I) in the Sugetbrak Fm. and the correlations support a severe diagenetic alteration imprint. $^{87}Sr/^{86}Sr$ data of the diagenetically altered samples of this study is also comparable to the altered samples in the Doushantuo Fm. from South China (Sawaki et al., 2010) (Fig. 8). However, although this diagenetic alteration may relate to Derry's large-scale or even global post-depositional diagenetic $\delta^{13}C$ variation hypothesis (Derry, 2010), it requires further diagenesis evaluation by other proxies at a global scale (e.g. lithological evidence of widespread deposition of authigenic carbonate).

Immediately above the altered samples from the Sugetbrak Fm., samples from the base of the Chigebrak Fm. delineate the peak of the positive $\delta^{13}C$ excursion "A" and its recovery towards lower values. Wood et al. (2015) also reported a positive $\delta^{13}C$ excursion (ca. 548 Ma, equivalent to the excursion "A") in the carbonates of the Nama group from Namibia postdating the appearance of calcified metazoans. The steep increase of calcified metazoans in the total primary production could have amplified organic carbon burial and led to a net drawdown of atmospheric CO_2, ultimately shifting the seawater $\delta^{13}C$ to more positive values (Knoll et al., 1996). In this study, the positive $\delta^{13}C$ excursion "A" is accompanied by an $^{87}Sr/^{86}Sr$ decrease, coinciding a lithological transition from more siliciclastic carbonates to micritic carbonates (Fig. 9) that may relate to a transgression event (Deng et al., 2019). Therefore, the recorded $^{87}Sr/^{86}Sr$ decrease in carbonates may imply suppressed continental silicate weathering due to reduced continental silicates being exposed to physical-chemical weathering on land (Jiang et al., 2007) and possible associated atmospheric CO_2 drawdown (Knoll et al., 1996).

Carbonates deposited in the Keping region above the positive excursion "A" show a negative $\delta^{13}C$ excursion "B" accompanied by a decrease in seawater $^{87}Sr/^{86}Sr$ ratios (Fig. 9), which was also reported in North India (Fig. 7, Jiang et al., 2007; Kaufman et al., 2006). The further decrease in seawater $^{87}Sr/^{86}Sr$ might result from continued continental flooding and suppressed continental weathering. At the same time, the $\delta^{13}C$ negative excursion "B" could be explained by global ocean stratification by upwelling of deep-sea ^{13}C depleted suboxic to euxinic waters in the terminal Proterozoic (cf. Jiang et al., 2007). Finally, the upper Chigebrak Fm. carbonates record a rising seawater $^{87}Sr/^{86}Sr$ trend, consistent with the global seawater record in the late Ediacaran pristine carbonate (Brasier et al., 1996; Sawaki et al., 2008), possibly indicating enhanced silicate weathering as a result of increased silicates exposure during large-scale regression (Brasier, 1996; Haq and Schutter, 2008).

5.4.2 Terrenenvian Series and Cambrian Series 2 paleoenvironmental reconstruction of the Tarim Basin

As discussed in section 5.1.3, $^{87}Sr/^{86}Sr$ variations recorded in the Yurtus Fm. carbonates are possibly caused by restricted basin seawater mixing with meteoric waters (Fig. 6). Further, the widespread karst structures may represent a widely distributed hiatus between the Chigebrak and Yurtus Fm. (Wu et al., 2018), which indicates a large scale sea-level drop during the Ediacaran-Cambrian transition and is consistent with a proposed global sea-level shift during this interval (Haq and Schutter, 2008; He et al., 2017). He et al. (2018) studied the geographical distribution of the paleo-karst at the top of the Chigebrak Fm. and speculated that there was an Ediacaran carbonate paleo-upland near to the Xiaoerbrak and Shiairik sections during this time. Additionally, the paleo-karst structures in the uppermost of the Chigebrak Fm. may indicate that meteoric fluids (with light O isotopic compositions) and dissolution of Ediacaran carbonates (and their Sr isotope compositions) might have influenced the Yurtus Fm. carbonate Sr isotope values.

The obtained $\delta^{13}C$ values in the Yurtus Fm. carbonates show no variation with $^{87}Sr/^{86}Sr$, however, given the existence of a paleo-upland which might contribute to the "great unconformity" at the PC/C boundary (He et al., 2017; He et al., 2018), the negative $\delta^{13}C$ excursions in the Yurtus Fm. may be a result of enhanced dissolved organic carbon (with low $\delta^{13}C$ signals) remineralization as a result of oceanic ventilation during regression (He et al., 2019).

During the Cambrian Series 2, the global seawater $^{87}Sr/^{86}Sr$ trend shows a slow rise from ~0.7087 to ~0.7090 (Kaufman et al., 1993; Derry et al., 1994; Montañez et al., 1996; Montañez and Banner, 2000; Ebneth et al., 2001, Fig. 8). This rise may record an increased radiogenic ^{87}Sr input from enhanced continental weathering. Further, increased weathering enhanced organic matter burial, and eventually caused the distinct positive $\delta^{13}C$ excursion "D". Finally, organic matter burial could have induced a net increase in oxygen in the deep waters (Guo et al., 2017) which potentially triggered the appearance of *Shizhudiscus* (Fig. 9, Qian and Xiao, 1984), that has higher oxygen demand for its cellular respiration and motile activities.

5.4.3 Causes of the ROECE event

The prominent global decrease in $^{87}Sr/^{86}Sr$ ratios at the boundary between the Cambrian Stage 4 and the Cambrian Wuliuan Stage is commonly explained by one of the following hypotheses:

i. Large-scale sea-level rise (Montañez and Banner, 2000; Hough et al., 2006). The flooding of continental areas decreases the input of radiogenic Sr to the oceans and therefore lower the $^{87}Sr/^{86}Sr$ ratios of seawater.

ii. Widespread rifting along with the Weddell Sea-South African sector of the paleo-Pacific margin of Laurentia (Curtis et al., 1999) exposes mafic rocks (enriched in ^{86}Sr) resulting in a net decrease in radiogenic Sr input.

Considering that the negative $\delta^{13}C$ excursion "F" obtained in the carbonates from the Keping region (down to ~-13 ‰) is much lower than -6 ‰, we rule out explanation "ii" as hydrothermal fluid CO_2 exhalation of mid-ocean ridge-type (with C isotope compositions usually > -6 ‰; Ishibashi et al., 1995) shows no such negative C isotope compositions. By contrast, the sea-level rise hypothesis represents a more plausible mechanism for the $^{87}Sr/^{86}Sr$ decrease in the contemporaneous seawater and is supported by the transgression event revealed in the sedimentary characteristics (Bai et al., 2019). Moreover, the extreme negative $\delta^{13}C$ recorded in the carbonates, which exceeds the magnitude of other ROECE-equivalent negative excursions in South China (Guo et al., 2010), can be hardly explained alone by a reduced dissolved inorganic carbon input into the ocean (Kump and Arthur, 1999). The extreme negative $\delta^{13}C$ values observed in our studied carbonates are therefore probably superimposed by an upwelling of ^{13}C-depleted deep-waters alongside the transgression, leading to extreme low $\delta^{13}C$ values in the regional water-mass and the deposited carbonates (Deng et al., 2019; Guo et al., 2017).

5.4.4 Cambrian Series 3 (Miaolingian) and Furongian Series paleoenvironmental reconstruction of the Tarim Basin

During the Cambrian Series 3 (Miaolingian), the ambient seawater $^{87}Sr/^{86}Sr$ ratio increases steadily to ~0.7092, while $\delta^{13}C$ values likewise rise to zero per mil succeeding the excursion "F" (Fig. 9). Following Montañez and Banner (2000), this increase to more radiogenic seawater Sr ratios may have resulted from orogenic events in Antarctica and Australia which is presumably proximal to the Tarim Basin during the Miaolingian (Huang et al., 2000) as they exposed more silicates on the continents prone to physicochemical weathering. This process may have further enhanced the influx of key-nutrients (e.g. phosphorus) into the ocean (Montañez et al., 1996), and ultimately leading to an amplification in the primary productivity and organic carbon burial. As a result, the $\delta^{13}C$ values of co-precipitating carbonates from surface waters could have increased towards higher values.

The beginning of the Furongian is characterized by a peak in both $\delta^{13}C$ and $^{87}Sr/^{86}Sr$ in well-preserved carbonates (e.g. Montañez and Banner, 2000; Ebneth et al., 2001, Fig. 8). In this study, contemporarily precipitated carbonates from the Tarim Basin yield $^{87}Sr/^{86}Sr$ ratios between 0.7092 to 0.7091, which are close to modern seawater Sr isotope compositions (i.e. present seawater $^{87}Sr/^{86}Sr$: 0.70916 (Ebneth et al., 2001)) and may imply a contribution from a more radiogenic Sr-bearing rocks during the denudation of Damara Belt as part of the Pan-African orogeny (Montañez and Banner, 2000; Ebneth et al., 2001). Moreover, a slight decrease in the $^{87}Sr/^{86}Sr$ towards the end of the Furongian in the Keping region as well as in the global record may imply the waning of the Pan-African orogeny (Fig. 9).

6. Conclusions

This study presents an integrated C and Sr isotope stratigraphic framework from late Ediacaran (ca. 551 Ma) to late Cambrian (< ca. 497 Ma) fully marine strata from the Keping region in the Tarim Basin, North-West China. The C and Sr isotope curves obtained from well-preserved carbonates are similar in their overall trends to those obtained in sections from other carbonate platforms worldwide. A chemo-stratigraphic correlation of our C and Sr isotope curves with globally reported isotope curves from various PC/C sections provides a stratigraphic framework for the late Ediacaran-late Cambrian strata in the Keping area, NW Tarim Basin. Furthermore, the obtained curves may contribute to the global seawater $\delta^{13}C$ and $^{87}Sr/^{86}Sr$ evolution curves and provide a reference for future stratigraphic correlations in the Tarim Basin and elsewhere.

Finally, our paleoenvironmental reconstruction shows bio-geochemical variations of the Tarim Basin seawater that were presumably governed by fluctuations in the hinterland weathering regime and temporal changes in the connection to the open ocean. The studied carbonates record both local as well as global seawater level changes, which may have controlled the influx of radiogenic Sr and key nutrients, restricting primary production, into the Tarim Basin. These cause-effect relationships provided in this study may have ultimately impacted the extent of organic C burial and the rapid diversification of metazoans.

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Figure captions:

Fig. 1 A) Paleogeographic maps of the Tarim Basin (yellow) and South China (orange) during the Ediacaran (A-1) and the middle Cambrian (A-2) (modified from Merdith et al., 2017), abbreviations: T for the Tarim Basin and SC for South China. B) Proterozoic and Phanerozoic geological units of the Keping region (red square on the inlet map B-1) with locations of the sections indicated by red stars (modified from Zhou et al., 2015).

Fig. 2 Comprehensive stratigraphic log of sections from the Keping region (Modified after Zhou and Chen, 1990; Zhou, 2001; Peng, 2009; Wang et al., 2011b). The markers in the "Sampling Location" column represent the heights/distances of samples (the height scale of 50 m can be found in the upper-right corner), the Ediacaran samples are from the Shiairik section (red dashes), and the Cambrian samples are from the Xiaoerbrak section (black dashes). The dotted lines represent less defined formation boundaries; the colored arrows mark the biozone and fossil occurrence range.

Fig. 3 Schematic flow diagram of the carbonate dissolution procedure.

Fig. 4 C, O and Sr isotopic isotope compositions obtained in carbonate leachates from the late Ediacaran to Cambrian at the studied sections in the Keping region. The green/orange fields represent positive/negative C isotope excursions, respectively. The inserted column shows the $\delta^{13}C$ variation in the Yurtus Fm. in 10-fold height magnification for better visibility.

Fig. 5 C, O isotope compositions and Mn/Sr ratios for the late-Ediacaran samples (A) and Cambrian samples (B). The yellow layer represents $\delta^{18}O < -10$ ‰.

Fig. 6 Sr isotopic composition vs. $\delta^{13}C$, $\delta^{18}O$, Sr concentration, Mn/Sr, Mg/Ca, carbonate content (%, = CaO (%) + MgO (%)) and [Rb] for the Cambrian samples (A-G) and late Ediacaran samples (H-N). For the fluid-carbonate mixing model in Yurtus Fm., a meteoric fluid composition of $^{87}Sr/^{86}Sr = 0.7085$, $\delta^{18}O = -14$, $\delta^{13}C = -5$ and Sr concentration of 600 ppm was chosen together with a presumed localized seawater of $^{87}Sr/^{86}Sr = 0.7102$, $\delta^{18}O = 0.5$, $\delta^{13}C = 0.5$, and Sr concentration of 50 ppm was used in the model.

Fig. 7 Comparison of $\delta^{13}C$ signatures recorded in carbonate rocks spanning late-Ediacaran to late-Cambrian strata in A) the Tarim Basin (this study), B) the presumable global Cambrian seawater standard $\delta^{13}C$ variations, mainly derived from South China (Zhu et al., 2019), C-D-G-I) South China (Jiang et al., 2007; Guo et al., 2010; Li et al., 2013), H) South-western Mongolia

(Brasier et al., 1996) E) Laurentia (Montañez and Banner, 2000), F) Siberia (Wotte et al., 2011) and J) North India (Kaufman et al., 2006). Red dots represent screened samples with severe diagenetic alteration; the dotted line indicates uncertain boundaries; the red line indicates the first appearance of trilobites. Paleontological data are derived from Qian and Xiao (1984); Qian and Bengtson (1989); Peng (2009); Wang et al. (2011b). The age of the beginning of the Paibian stage is set to 497 Ma (Ogg et al., 2016).

Fig. 8 Comparison of $^{87}Sr/^{86}Sr$ ratios from the late-Ediacaran to the late-Cambrian in A) Cambrian global seawater $^{87}Sr/^{86}Sr$ standard proposed by Peng et al. (2012); Zhu et al. (2019); of which data is derived from Derry et al. (1994); Montañez and Banner (2000); Ebneth et al. (2001); Kouchinsky et al. (2008); dotted lines represent less defined formation boundaries, B) & C) Sr isotope data from the Yangtze Platform, South China (Sawaki et al., 2008; Sawaki et al., 2010) and D) Keping Area, NW Tarim Basin (this study) where the data from the altered carbonates from the Yurtus and Sugetbrak Fm. is excluded.

Fig. 9 C and Sr isotope evolution of seawater in the Keping region, NW Tarim. This figure includes the stratigraphic correlation and supposed age constraints. Colored (green/orange) rectangular boxes represent the durations of the $\delta^{13}C$ excursions. Colored arrows pointed out the occurrences of the fossil-bearing strata; red dots indicate diagenetically altered $^{87}Sr/^{86}Sr$ values and the diagenetically altered $^{87}Sr/^{86}Sr$ values from the Yurtus and Sugetbrak carbonates are excluded in this figure.

Table 1. Analytical results of C, O, Sr isotope compositions and Ca, Mg, Mn, Sr, Rb concentrations of samples from the Keping area, NW Tarim Basin, NW China.

Sample	Formation	Lithology	Height (m)	δ^{13}C (V-PDB, ‰)	δ^{18}O (V-PDB, ‰)	^{87}Sr/^{86}Sr	±σ	Ca (%)	Mg (%)	Mn (ppm)	Rb (ppm)	Sr (ppm)	Mn/Sr	Mg/Ca	^{85}Rb/^{86}Sr
SARK-1	Sugetbrak	limestone	-217.9	-5.49	-9.88	0.709471	0.000005	32.4	0.52	2180	1.56	700	3.11	0.02	0.000341
SARK-2	Sugetbrak	limestone	-215.0	-4.68	-9.44					2340	1.94	590	3.96		
SARK-3	Sugetbrak	dolostone	-156.2	1.00	-5.90	0.709365	0.000005	19.7	6.42	781	1.29	171	4.57	0.33	0.001164
SARK-4	Sugetbrak	dolostone	-153.0	3.67	-4.20					436	0.83	160	2.73		
SARK-5	Sugetbrak	dolostone	-151.4	3.90	-3.96					700	4.17	148	4.72		
SARK-6	Sugetbrak	dolostone	-150.5	4.14	-1.83					642	2.13	62.1	10.3		
SARK-7	Sugetbrak	dolostone	-148.5	4.67	-2.53	0.709158	0.000004	13.4	7.39	669	1.60	147	4.55	0.55	0.000280
SARK-8	Sugetbrak	dolostone	-146.0	5.60	-0.85					488	1.06	175	2.78		
SARK-9	Chigebrak	dolostone	-143.0	6.37	-3.02	0.708849	0.000005	15.3	8.64	219	0.69	235	0.93	0.57	0.000458
SARK-10	Chigebrak	dolostone	-142.1	6.61	-3.56					277	0.65	171	1.62		
SARK-11	Chigebrak	dolostone	-141.0	6.56	-3.49	0.709069	0.000005	12.2	6.71	369	1.46	253	1.46	0.55	0.000726
SARK-12	Chigebrak	dolostone	-140.0	6.96	-0.98					205	0.69	201	1.02		
SARK-13	Chigebrak	dolostone	-138.0	7.94	-2.50					182	0.81	227	0.80		
SARK-14	Chigebrak	dolostone	-135.0	5.59	-5.32	0.708934	0.000005	16.3	9.12	499	1.31	141	3.54	0.56	0.001235
SARK-15	Chigebrak	dolostone	-133.5	5.01	-2.41					312	1.23	180	1.73		
SARK-16	Chigebrak	dolostone	-132.0	5.68	-2.78					241	0.57	174	1.38		
SARK-17	Chigebrak	dolostone	-130.5	5.54	-2.60	0.708962	0.000006	15.7	9.16	259	0.74	171	1.51	0.58	0.000738
SARK-18	Chigebrak	dolostone	-129.6	5.93	-2.63					363	1.16	207	1.76		
SARK-19	Chigebrak	dolostone	-129.0	5.69	-2.36					243	0.62	203	1.20		
SARK-20	Chigebrak	dolostone	-127.0	4.87	-3.65	0.708994	0.000005	13.4	7.83	402	1.67	226	1.78	0.58	0.000579
SARK-21	Chigebrak	dolostone	-125.2	5.88	-2.16					279	0.65	180	1.55		
SARK-22	Chigebrak	dolostone	-124.2	5.93	-2.58					218	0.78	176	1.24		
SARK-23	Chigebrak	dolostone	-122.0	4.41	-0.72	0.708806	0.000005	12.6	7.27	410	1.23	191	2.15	0.58	0.000467
SARK-24	Chigebrak	dolostone	-119.0	4.40	-2.30					297	1.42	115	2.58		
SARK-25	Chigebrak	dolostone	-117.2	4.12	0.15					222	0.77	142	1.56		

Sample	Formation	Lithology	Height (m)	$\delta^{13}C$	$\delta^{18}O$	$^{87}Sr/^{86}Sr$	±σ	Ca (%)	Mg (%)	Mn (ppm)	Rb (ppm)	Sr (ppm)	Mn/Sr	Mg/Ca	$^{85}Rb/^{86}Sr$
				(V-PDB, ‰)											
SARK-26	Chigebrak	dolostone	-114.0	3.50	-2.80	0.708816	0.000004	17.5	10.1	190	0.69	158	1.20	0.58	0.000260
SARK-27	Chigebrak	dolostone	-111.0	3.63	-2.47					156	0.72	220	0.71		
SARK-28	Chigebrak	dolostone	-109.0	2.06	-3.56					152	0.54	170	0.89		
SARK-29	Chigebrak	dolostone	-107.0	0.97	-3.10	0.708750	0.000005	16.8	9.80	149	0.54	169	0.88	0.58	0.000474
SARK-30	Chigebrak	dolostone	-100.3	1.70	-2.87	0.708611	0.000005	18.9	11.1	207	0.51	104	1.99	0.59	0.000455
SARK-31	Chigebrak	dolostone	-99.6	1.69	-2.98					18.7	0.45				
SARK-32	Chigebrak	dolostone	-98.0	2.35	-3.11					199	0.50	161	1.23		
SARK-33	Chigebrak	dolostone	-95.0	2.57	-3.12					108	0.96	73.5	1.47		
SARK-34	Chigebrak	dolostone	-90.0	2.05	-5.08	0.708746	0.000005	17.9	10.1	396	0.68	127	3.11	0.57	0.000641
SARK-35	Chigebrak	dolostone	-86.0	2.13	-5.16					464	0.47	144	3.23		
SARK-36	Chigebrak	dolostone	-83.0	2.52	-1.52					256	0.44	130	1.96		
SARK-37	Chigebrak	dolostone	-80.0	3.17	-2.45					165	0.44	123	1.34		
SARK-38	Chigebrak	dolostone	-75.0	3.05	-1.04	0.708464	0.000005	19.8	11.5	173	0.44	138	1.25	0.58	0.000689
SARK-39	Chigebrak	dolostone	-66.0	2.70	-2.10					179	0.43	163	1.10		
SARK-40	Chigebrak	dolostone	-60.0	2.75	-0.10					177	0.47	440	0.40		
SARK-41	Chigebrak	dolostone	-56.0	2.86	-0.05	0.708544	0.000005	19	11.0	229	0.50	125	1.83	0.58	0.000469
SARK-42	Chigebrak	dolostone	-53.0	3.22	0.42					110	0.45	107	1.03		
SARK-43	Chigebrak	dolostone	-48.2	2.88	-1.91					136	0.61	78.8	1.73		
SARK-44	Chigebrak	dolostone	-45.0	2.98	-0.91	0.708775	0.000005	18.2	10.5	155	0.61	116	1.33	0.58	0.001286
SARK-45	Chigebrak	dolostone	-40.0	2.75	-1.78					185	0.59	104	1.78		
SARK-46	Chigebrak	dolostone	-34.5	2.33	-3.92					336	0.55	94.2	3.57		
SARK-47	Chigebrak	dolostone	-6.0	1.97	-5.54	0.708880	0.000005	13.3	7.70	557	0.94	143	3.90	0.58	0.000579
SARK-48	Chigebrak	dolostone	-3.0	1.65	-6.49	0.709089	0.000004	14.6	8.20	604	1.98	208	2.91	0.56	0.000610
SARK-49	Chigebrak	dolostone	-2.0	1.98	-5.97					546	0.92	167	3.26		
XBK-1	Yurtus	phosphorite	0.5							1988	1.31	386	5.15		
XBK-2	Yurtus	phosphorite	2.5			0.709950	0.000008	1.04	0.19	0.36	0.25	58.4	0.01	0.18	0.000483

Sample	Formation	Lithology	Height	$\delta^{13}C$	$\delta^{18}O$	$^{87}Sr/^{86}Sr$	±σ	Ca	Mg	Mn	Rb	Sr	Mn/Sr	Mg/Ca	$^{85}Rb/^{86}Sr$
			(m)	(V-PDB, ‰)				(%)	(%)	(ppm)	(ppm)	(ppm)			
XBK-3	Yurtus	phosphorite	4.5			0.708852	0.000008	1.33	0.55	1.50	0.13	578		0.41	0.000917
XBK-4	Yurtus	dolostone	6.5	-2.63	-6.32	0.709560	0.000007	34.6	17.7	415	2.30	362	1.15	0.51	0.000787
XBK-5	Yurtus	limestone	8.5	-0.58	-12.1					344	0.34	206	1.67		
XBK-6	Yurtus	limestone	9.0	0.46	-13.2	0.708695	0.000011	36.2	0.20	213	0.43	123	1.74	0.01	0.000400
XBK-7	Yurtus	dolostone	9.3	-5.05	-9.20					496	0.35	103	4.84		
XBK-8	Yurtus	dolostone	10.2	-1.90	-7.77	0.709264	0.000009	2.74	0.75	44.9	0.56	63.6	0.70	0.27	0.000506
XBK-9	Yurtus	dolostone	11.0	-0.77	-6.45	0.710168	0.000009	3.27	1.63	121	0.70	24.2	4.98	0.50	0.001347
XBK-10	Yurtus	dolostone	12.0	-1.12	-9.94					950	0.45	68.9	13.8		
XBK-11	Yurtus	dolostone	13.0	-0.05	-8.33					935	0.40	46.9	19.9		
XBK-12	Yurtus	dolostone	13.5	0.30	-8.22					881	0.32	50.2	17.6		
XBK-13	Yurtus	dolostone	14.2	0.73	-7.67	0.709472	0.000009	18.3	10.2	1062	0.37	73.8	14.4	0.56	0.000990
XBK-14	Yurtus	dolostone	15.6	0.61	-6.85					0.10	0.00				
XBK-15	Yurtus	dolostone	17.8	0.30	-7.06					1017	0.34	50.0	20.3		
XBK-16	Yurtus	dolostone	18.7	0.22	-6.99					902	0.33	44.9	20.1		
XBK-17	Yurtus	dolostone	19.5	-0.12	-7.18					966	0.13	43.6	22.1		
XBK-18	Yurtus	dolostone	21.5	-0.53	-7.11					1428	0.34	41.9	34.1		
XBK-19	Yurtus	dolostone	21.5	0.21	-7.53	0.709451	0.000009	18.4	10.8	472	0.43	58.5	8.07	0.59	0.000861
XBK-20	Yurtus	dolostone	21.6	-0.04	-7.73					361	0.08	44.8	8.07		
XBK-21	Xiaoerbrak	dolostone	21.7	0.08	-7.34					277	0.15	53.6	5.16		
XBK-22	Xiaoerbrak	dolostone	24.9	1.21	-7.7	0.708971	0.000009	21.8	12.5	250	0.11	71.7	3.49	0.57	0.000793
XBK-23	Xiaoerbrak	dolostone	25.9	1.26	-7.52					172	0.03	75.9	2.27		
XBK-24	Xiaoerbrak	dolostone	31.6	2.20	-6.79	0.708817	0.000011	23.8	13.9	136		91.4	1.49	0.58	0.000392
XBK-25	Xiaoerbrak	dolostone	33.4	2.54	-7.05					147		89.2	1.65		
XBK-26	Xiaoerbrak	dolostone	36.1	1.73	-7.39					129		67.4	1.92		
XBK-27	Xiaoerbrak	dolostone	41.4	2.78	-7.26	0.708790	0.000009	19.0	11.3	142		81.7	1.73	0.59	0.001545
XBK-28	Xiaoerbrak	dolostone	45.9	2.59	-6.71					139		92.2	1.51		

Sample	Formation	Lithology	Height (m)	$\delta^{13}C$	$\delta^{18}O$	$^{87}Sr/^{86}Sr$	±σ	Ca (%)	Mg (%)	Mn (ppm)	Rb (ppm)	Sr (ppm)	Mn/Sr	Mg/Ca	$^{85}Rb/^{86}Sr$
				(V-PDB, ‰)											
XBK-29	Xiaoerbrak	dolostone	52.4	2.55	-6.97	0.708765	0.000009	21.9	12.9	117		73.7	1.58	0.59	0.000309
XBK-30	Xiaoerbrak	dolostone	59.5	2.69	-7.32					124		67.3	1.84		
XBK-31	Xiaoerbrak	dolostone	63.1	2.99	-7.33	0.708731	0.000009	21.2	12.5	132	0.09	87.7	1.50	0.59	0.000555
XBK-32	Xiaoerbrak	dolostone	63.3	2.89	-7.23					111		73.2	1.51		
XBK-33	Xiaoerbrak	dolostone	65.4	3.18	-7.59					96.0		83.9	1.14		
XBK-34	Xiaoerbrak	dolostone	71.6	3.12	-6.94	0.708733	0.000009	20.8	12.2	122		112	1.08	0.58	0.000373
XBK-35	Xiaoerbrak	dolostone	75.9	3.17	-7.24					101		74.5	1.36		
XBK-36	Xiaoerbrak	dolostone	80.3	2.87	-8.49					112		55.5	2.02		
XBK-37	Xiaoerbrak	dolostone	85.5	2.96	-7.96	0.708803	0.000008	21.4	12.8	124		65.2	1.91	0.59	0.000649
XBK-38	Xiaoerbrak	dolostone	90.3	2.66	-8.78					141		64.4	2.18		
XBK-39	Xiaoerbrak	dolostone	103.7	2.73	-8.27					134		53.9	2.49		
XBK-40	Xiaoerbrak	dolostone	113.7	3.15	-7.12	0.708817	0.000015	20.0	11.8	82.8		45.3	1.83	0.59	0.000562
XBK-41	Xiaoerbrak	dolostone	118.0	2.70	-8.24					109	0.03	43.8	2.48		
XBK-42	Xiaoerbrak	dolostone	123.5	2.47	-7.32	0.708888	0.00001	20.6	11.6	91.1		47.7	1.91	0.56	0.000990
XBK-43	Xiaoerbrak	dolostone	126.5	1.73	-7.21					95.8		36.5	2.62		
XBK-44	Xiaoerbrak	dolostone	132.6	2.07	-6.90	0.708810	0.000009	19.5	11.5	92.5		43	2.15	0.59	0.000599
XBK-45	Xiaoerbrak	dolostone	135.6	2.66	-6.44					72.6		30.9	2.35		
XBK-46	Xiaoerbrak	dolostone	142.6	2.85	-6.41					95.4		35.2	2.71		
XBK-47	Xiaoerbrak	dolostone	145.0	2.86	-6.45	0.708788	0.000009	17.8	10.6	86.7		36.0	2.41	0.60	0.000567
XBK-48	Xiaoerbrak	dolostone	149.4	2.33	-6.81					120		39.8	3.02		
XBK-49	Xiaoerbrak	dolostone	151.6	2.43	-6.44					112		39.5	2.83		
XBK-50	Xiaoerbrak	dolostone	155.5	2.45	-6.49	0.708795	0.000009	20.3	12.0	114		57.5	1.98	0.59	0.000615
XBK-51	Xiaoerbrak	dolostone	159.8	2.46	-6.74					104	0.10	51.9	2.00		
XBK-52	Xiaoerbrak	dolostone	165.3	2.51	-6.45					116	0.06	64.1	1.82		
XBK-53	Xiaoerbrak	dolostone	166.0	2.54	-5.50	0.708831	0.000008	19.7	11.5	98.9	0.11	67.0	1.48	0.59	0.00033
XBK-54	Xiaoerbrak	dolostone	167.5	2.42	-5.60					88.4	0.02	64.6	1.37		

Sample	Formation	Lithology	Height	$\delta^{13}C$	$\delta^{18}O$	$^{87}Sr/^{86}Sr$	±σ	Ca	Mg	Mn	Rb	Sr	Mn/Sr	Mg/Ca	$^{85}Rb/^{86}Sr$
			(m)	(V-PDB, ‰)				(%)	(%)	(ppm)	(ppm)	(ppm)			
XBK-55	Xiaoerbrak	dolostone	168.9	2.50	-5.61	0.708827	0.00001	5.53	3.01	14.8		24.7	0.60	0.54	0.000830
XBK-56	Xiaoerbrak	dolostone	170.1	2.25	-3.53	0.708796	0.000008	17.3	10.1	79.3	0.12	86.1	0.92	0.58	0.000534
XBK-57	Xiaoerbrak	dolostone	171.7	1.22	-4.56					111	0.09	91.5	1.21		
XBK-58	Xiaoerbrak	dolostone	174.1	0.71	-6.7	0.708962	0.000009	18.8	10.7	135	0.01	62.2	2.17	0.57	0.000761
XBK-59	Xiaoerbrak	dolostone	174.0	1.08	-6.12					139	0.00	60.4	2.31		
XBK-60	Xiaoerbrak	dolostone	176.8	-0.69	-6.33	0.708989	0.000008	17.9	10.3	137	0.14	69.0	1.99	0.58	0.000492
XBK-61	Xiaoerbrak	dolostone	178.2	0.19	-8.09					157	0.32	112	1.40		
XBK-62	Xiaoerbrak	dolostone	179.0	-0.19	-6.05	0.708990	0.000008	19.9	11.3	120	0.04	66.1	1.81	0.57	0.000509
XBK-63	Xiaoerbrak	dolostone	182.3	0.33	-6.08					151	0.19	81.9	1.84		
XBK-64	Xiaoerbrak	dolostone	186.7	0.34	-6.12	0.709057	0.00001	19.3	11.1	110	0.14	77.8	1.41	0.58	0.000517
XBK-65	Xiaoerbrak	dolostone	188.9	-0.14	-6.36					143	0.14	68.7	2.09		
XBK-66	Xiaoerbrak	dolostone	192.5	0.62	-6.10	0.708863	0.00001	18.5	10.4	139	0.03	89.6	1.55	0.56	0.000506
XBK-67	Wusonger	dolostone	196.1	-1.05	-6.47					241	0.19	72.3	3.34		
XBK-68	Wusonger	dolostone	202.5	-1.77	-5.80					323	1.73	106	3.04		
XBK-69	Wusonger	dolostone	210.1	-1.21	-6.44	0.708902	0.000007	16.8	9.45	260	0.21	81.9	3.17	0.56	0.000587
XBK-70	Wusonger	dolostone	213.2	-1.89	-6.33					214	0.94	77.9	2.75		
XBK-71	Wusonger	dolostone	244.7	-1.71	-5.14	0.708934	0.000008	13.5	7.88	207	1.17	97.9	2.11	0.58	0.000747
XBK-72	Wusonger	dolostone	254.7	-1.11	-5.47	0.708980	0.000007	19.1	10.9	273	1.36	209	1.30	0.57	0.000628
XBK-73	Wusonger	dolostone	256.2	-0.21	-5.54	0.708941	0.000007	19.8	10.8	298	0.95	219	1.36	0.55	0.000425
XBK-74	Wusonger	dolostone	257.7	-0.27	-5.76	0.708974	0.000007	20.4	11.8	243	0.92	257	0.94	0.58	0.000378
XBK-75	Wusonger	dolostone	259.2	-0.54	-5.84	0.709179	0.000008	18.1	10.2	335	1.26	203	1.65	0.56	0.000242
XBK-76	Wusonger	dolostone	260.7	-0.31	-5.67	0.709054	0.000008	19.3	10.4	295	1.19	220	1.34	0.54	0.000378
XBK-77	Wusonger	dolostone	262.2	0.98	-5.27	0.709169	0.000009	18.5	10.8	276	0.99	213	1.29	0.58	0.000508
XBK-78	Wusonger	dolostone	263.7	-6.66	-12.9	0.709150	0.000006	2.32	0.15	57.9	3.50	1213	0.05	0.06	0.000209
XBK-79	Wusonger	dolostone	265.2	-0.67	-5.59					314	1.13	343	0.92		
XBK-80	Wusonger	dolostone	266.7	-1.18	-5.80					276	0.96	240	1.15		

Sample	Formation	Lithology	Height (m)	$\delta^{13}C$	$\delta^{18}O$	$^{87}Sr/^{86}Sr$	±σ	Ca (%)	Mg (%)	Mn (ppm)	Rb (ppm)	Sr (ppm)	Mn/Sr	Mg/Ca	$^{85}Rb/^{86}Sr$
				(V-PDB, ‰)											
XBK-81	Wusonger	dolostone	268.2	-3.65	-5.48	0.708749	0.000008	20.0	11.4	265	0.74	209	1.27	0.57	0.000483
XBK-82	Wusonger	dolostone	269.7	-2.23	-5.77					281	1.16	214	1.31		
XBK-83	Wusonger	dolostone	271.2	-11.7	-5.48	0.708771	0.000008	18.8	10.6	352	0.89	248	1.42	0.56	0.000609
XBK-84	Shayilik	dolostone	273.0	-13.3	-7.75					471	1.48	184	2.56		
XBK-85	Shayilik	dolostone	277.9	-2.60	-6.17	0.708923	0.000009	20.3	9.50	320	0.96	213	1.50	0.47	0.001105
XBK-86	Shayilik	dolostone	285.6	-2.45	-4.81			13.6	7.90	196	2.70	153	1.28	0.58	
XBK-87	Shayilik	dolostone	301.7	-2.44	-6.37	0.708723	0.000008	13.6	7.80	287	1.80	128	2.24	0.57	0.000928
XBK-88	Shayilik	dolostone	308.7	-1.35	-6.98					614	2.52	131	4.68		
XBK-89	Shayilik	limestone	326.2	-0.35	-10.5	0.709021	0.000009	35.3	0.40	330	0.99	615	0.54	0.01	0.000329
XBK-90	Shayilik	limestone	336.4	-0.65	-8.05					217	0.64	201	1.08		
XBK-91	Shayilik	limestone	340.0	-0.08	-8.13	0.708853	0.000010	30.2	3.40	192	0.52	570	0.34	0.11	0.000260
XBK-92	Shayilik	limestone	350.4	-0.61	-8.47					369	0.81	425	0.87		
XBK-93	Awatagh	limestone	369.1	0.67	-5.94					598	3.16	130	4.60		
XBK-94	Awatagh	limestone	371.1	0.97	-8.59	0.708880	0.000007	36.3	0.50	135	0.61	5906	0.02	0.01	0.000746
XBK-95	Awatagh	dolostone	386.0	-0.66	-5.76	0.708892	0.000007	12.0	6.80	484	2.14	144	3.37	0.56	0.000542
XBK-96	Awatagh	dolostone	421.5	0.16	-8.81	0.708988	0.000007	19.2	6.50	1085	0.69	183	5.93	0.34	0.000628
XBK-97	Awatagh	dolostone	449.4	-1.50	-7.03					1109	1.21	180	6.17		
XBK-98	Awatagh	dolostone	449.6	-0.25	-6.47	0.709189	0.000008	15.2	7.10	545	0.70	287	1.90	0.47	0.000849
XBK-99	Awatagh	dolostone	468.1	-0.54	-7.47					82.1	0.55	38.9	2.11		
XBK-100	Awatagh	dolostone	483.0	0.70	-6.31	0.709083	0.000009	11.6	6.20	493	4.76	222	2.22	0.53	0.000347
XBK-101	Awatagh	dolostone	483.0	1.11	-6.82					482	1.15	182	2.65		
XBK-102	Awatagh	dolostone	485.0	1.34	-6.45					284	1.05	268	1.06		
XBK-103	Awatagh	dolostone	498.9	-1.38	-7.07	0.709053	0.000007	14.8	8.20	169	1.31	157	1.07	0.56	0.000289
XBK-104	Awatagh	dolostone	500.2	0.06	-6.94					223	1.56	95.5	2.33		
XBK-105	Awatagh	dolostone	517.6	0.09	-6.63	0.709134	0.000008	17.2	9.60	239	1.30	112	2.14	0.56	0.000610
XBK-106	Awatagh	dolostone	519.0	-1.56	-8.69					159	0.80	81.9	1.94		

Sample	Formation	Lithology	Height	$\delta^{13}C$	$\delta^{18}O$	$^{87}Sr/^{86}Sr$	±σ	Ca	Mg	Mn	Rb	Sr	Mn/Sr	Mg/Ca	$^{85}Rb/^{86}Sr$
			(m)	(V-PDB, ‰)				(%)	(%)	(ppm)	(ppm)	(ppm)			
XBK-107	Awatagh	dolostone	521.2	-0.58	-6.61					249	1.16	183	1.36		
XBK-108	Xiaquilitagh	dolostone	557.6	-0.41	-6.37	0.709156	0.000011	19.2	11.0	106	0.82	119	0.88	0.57	0.000381
XBK-109	Xiaquilitagh	dolostone	563.9	-0.05	-7.22					149	0.76	106	1.40		
XBK-110	Xiaquilitagh	dolostone	570.8	-0.08	-7.38					105	0.69	76.2	1.38		
XBK-111	Xiaquilitagh	dolostone	594.8	-1.30	-5.04	0.708608	0.000008	19.8	11.2	96.7	0.76	91.3	1.06	0.57	0.000679
XBK-112	Xiaquilitagh	dolostone	603.7	-1.37	-4.43					104	1.12	90.7	1.14		
XBK-113	Xiaquilitagh	dolostone	612.7	-0.96	-6.23	0.709018	0.000008	19.1	10.9	68.2	0.67	116	0.59	0.57	0.000277
XBK-114	Xiaquilitagh	dolostone	615.1	-1.22	-6.50					83	0.64	141	0.59		
XBK-115	Xiaquilitagh	dolostone	623.7	-1.12	-6.32	0.709107	0.000009	20.4	9.80	84.3	1.08	126	0.67	0.48	0.000181
XBK-116	Xiaquilitagh	dolostone	631.9	-1.08	-6.25					64.5	0.74	140	0.46		
XBK-117	Xiaquilitagh	dolostone	643.7	-1.16	-6.10	0.709114	0.000008	19.2	10.7	57.5	0.69	118	0.49	0.56	0.000546
XBK-118	Xiaquilitagh	dolostone	669.2	-0.99	-6.16					65.1	0.79	115	0.57		
XBK-119	Xiaquilitagh	dolostone	694.3	-0.87	-6.55	0.709087	0.000008	15.1	8.10	73.5	0.55	100	0.73	0.54	0.000312
XBK-120	Xiaquilitagh	dolostone	700.8	-1.04	-6.16					88.9	0.72	130	0.69		
XBK-121	Xiaquilitagh	dolostone	712.9	-0.91	-6.37	0.709127	0.000008	16.5	9.00	55.1	0.71	151	0.36	0.54	0.000257
XBK-122	Xiaquilitagh	dolostone	722.6	-1.18	-6.70					59.3	1.10	146	0.41		
XBK-123	Xiaquilitagh	dolostone	738.2	-0.96	-6.37	0.709138	0.000008	18.1	10.0	104	0.80	204	0.51	0.56	0.000450
XBK-124	Xiaquilitagh	dolostone	773.2	-1.17	-6.48					118	0.74	227	0.52		
XBK-125	Xiaquilitagh	dolostone	784.9	-0.97	-6.58	0.709143	0.000009	20.3	11.4	120	0.81	245	0.49	0.56	0.000230
XBK-126	Xiaquilitagh	dolostone	797.9	-0.86	-6.49					69.5	0.77	150	0.46		
XBK-127	Xiaquilitagh	dolostone	816.8	-0.82	-7.17	0.709195	0.000008	21.1	10.4	62.8	0.81	126	0.50	0.49	0.000360
XBK-128	Xiaquilitagh	dolostone	837.0	-0.52	-5.68					73.6	0.77	129	0.57		
XBK-129	Xiaquilitagh	dolostone	870.7	-0.29	-6.03	0.709107	0.000009	18.4	10.1	88.3	0.58	136	0.65	0.55	0.000317

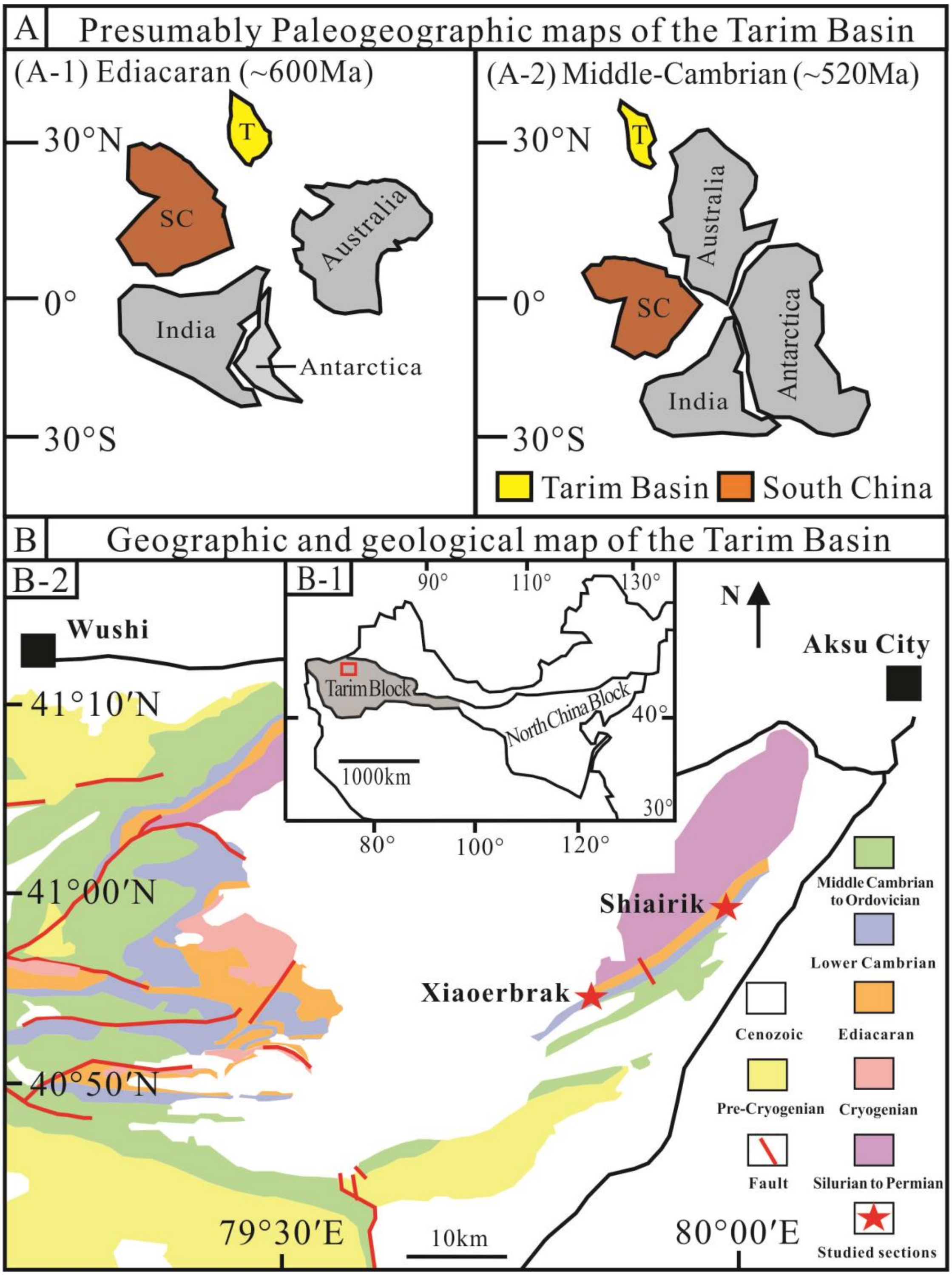

A
Presumably Paleogeographic maps of the Tarim Basin
(A-1) Ediacaran (~600Ma)
(A-2) Middle-Cambrian (~520Ma)
30°N
0°
30°S
T
SC
Australia
India
Antarctica
Tarim Basin
South China
B
Geographic and geological map of the Tarim Basin
B-2
B-1
Wushi
Aksu City
N
41°10′N
41°00′N
40°50′N
79°30′E
80°00′E
10km
90°
110°
130°
40°
30°
80°
100°
120°
Tarim Block
North China Block
1000km
Shiairik
Xiaoerbrak
Middle Cambrian to Ordovician
Lower Cambrian
Cenozoic
Ediacaran
Pre-Cryogenian
Cryogenian
Fault
Silurian to Permian
Studied sections

Figure 1

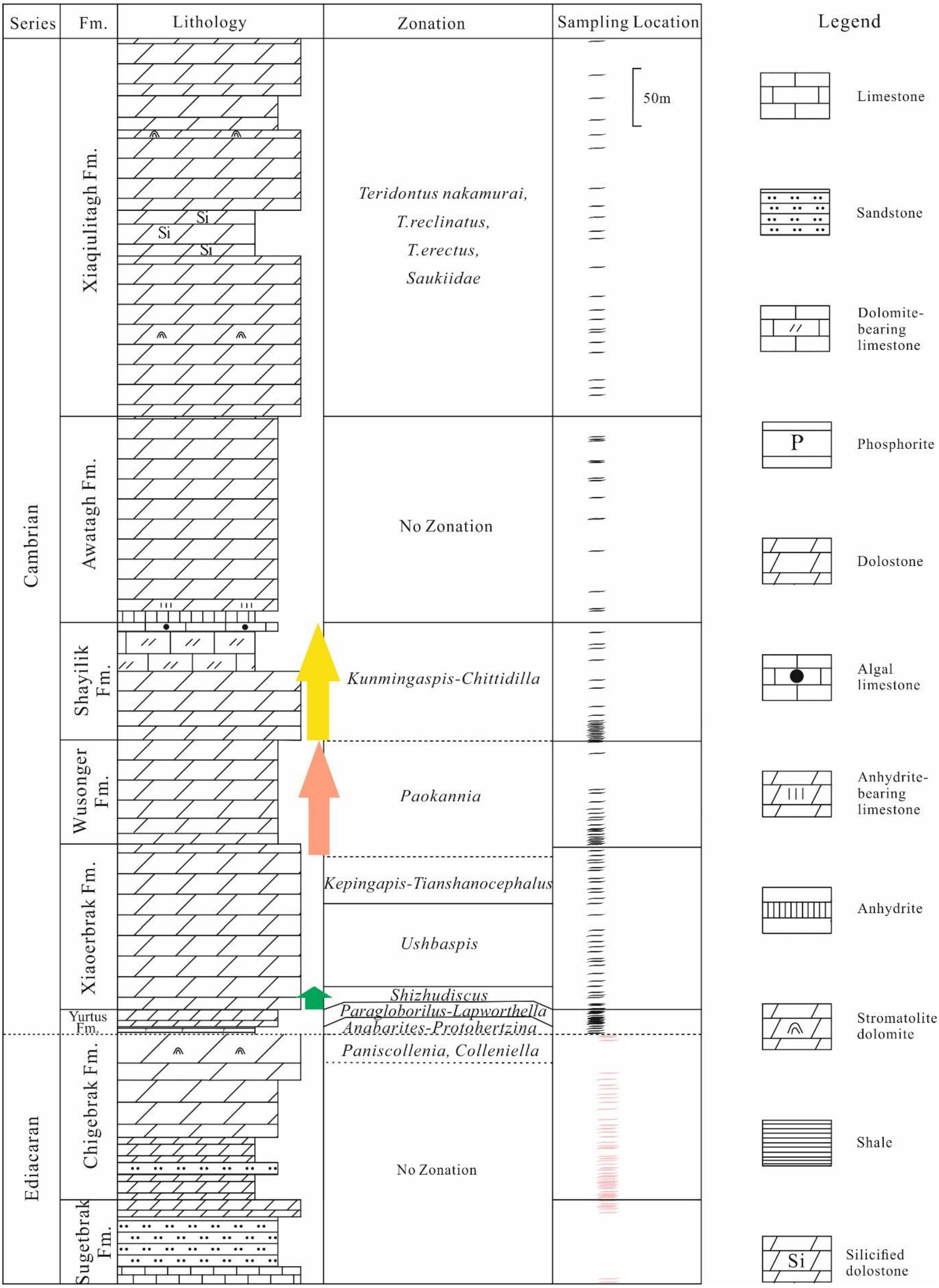
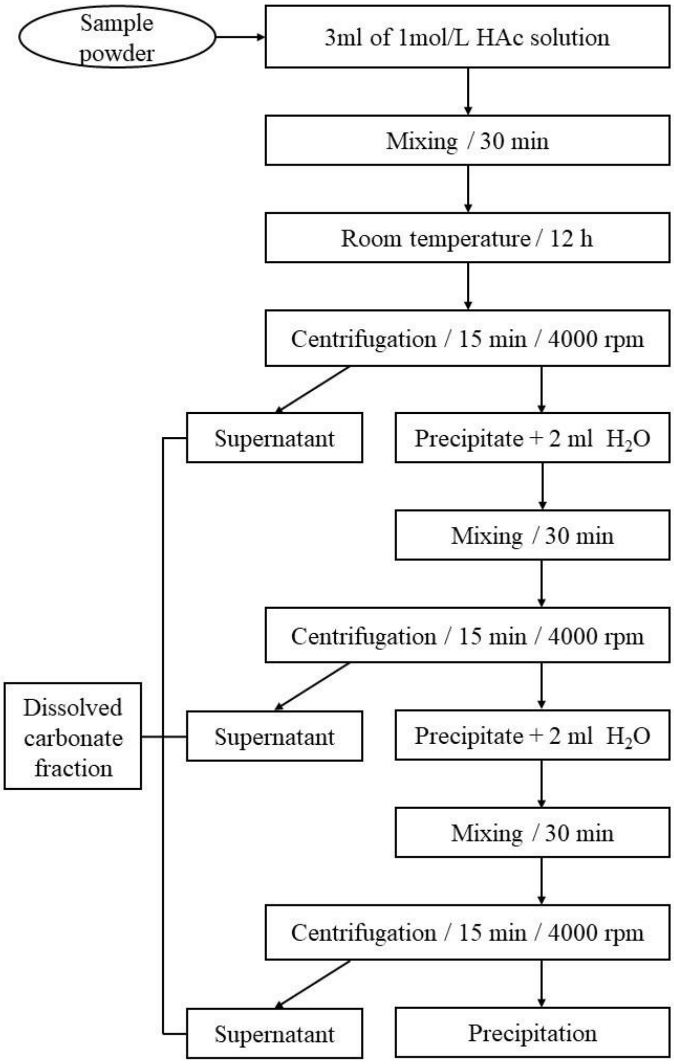

Series
Fm.
Lithology
Zonation
Sampling Location
Legend
Cambrian
Ediacaran
Xiaqiulitagh Fm.
Awatagh Fm.
Shayilik Fm.
Wusonger Fm.
Xiaoerbrak Fm.
Yurtus Fm.
Chigebrak Fm.
Sugetbrak Fm.
Si
Teridontus nakamurai,
T.reclinatus,
T.erectus,
Saukiidae
No Zonation
Kunmingaspis-Chittidilla
Paokannia
Kepingapis-Tianshanocephalus
Ushbaspis
Shizhudiscus
Paragloborilus-Lapworthella
Anabarites-Protohertzina
Paniscollenia, Colleniella
No Zonation
50m
Limestone
Sandstone
Dolomite-bearing limestone
P
Phosphorite
Dolostone
Algal limestone
Anhydrite-bearing limestone
Anhydrite
Stromatolite dolomite
Shale
Si
Silicified dolostone

Figure 2



Sample powder
3ml of 1mol/L HAc solution
Mixing / 30 min
Room temperature / 12 h
Centrifugation / 15 min / 4000 rpm
Supernatant
Precipitate + 2 ml H_2O
Mixing / 30 min
Centrifugation / 15 min / 4000 rpm
Dissolved carbonate fraction
Supernatant
Precipitate + 2 ml H_2O
Mixing / 30 min
Centrifugation / 15 min / 4000 rpm
Supernatant
Precipitation

Figure 3

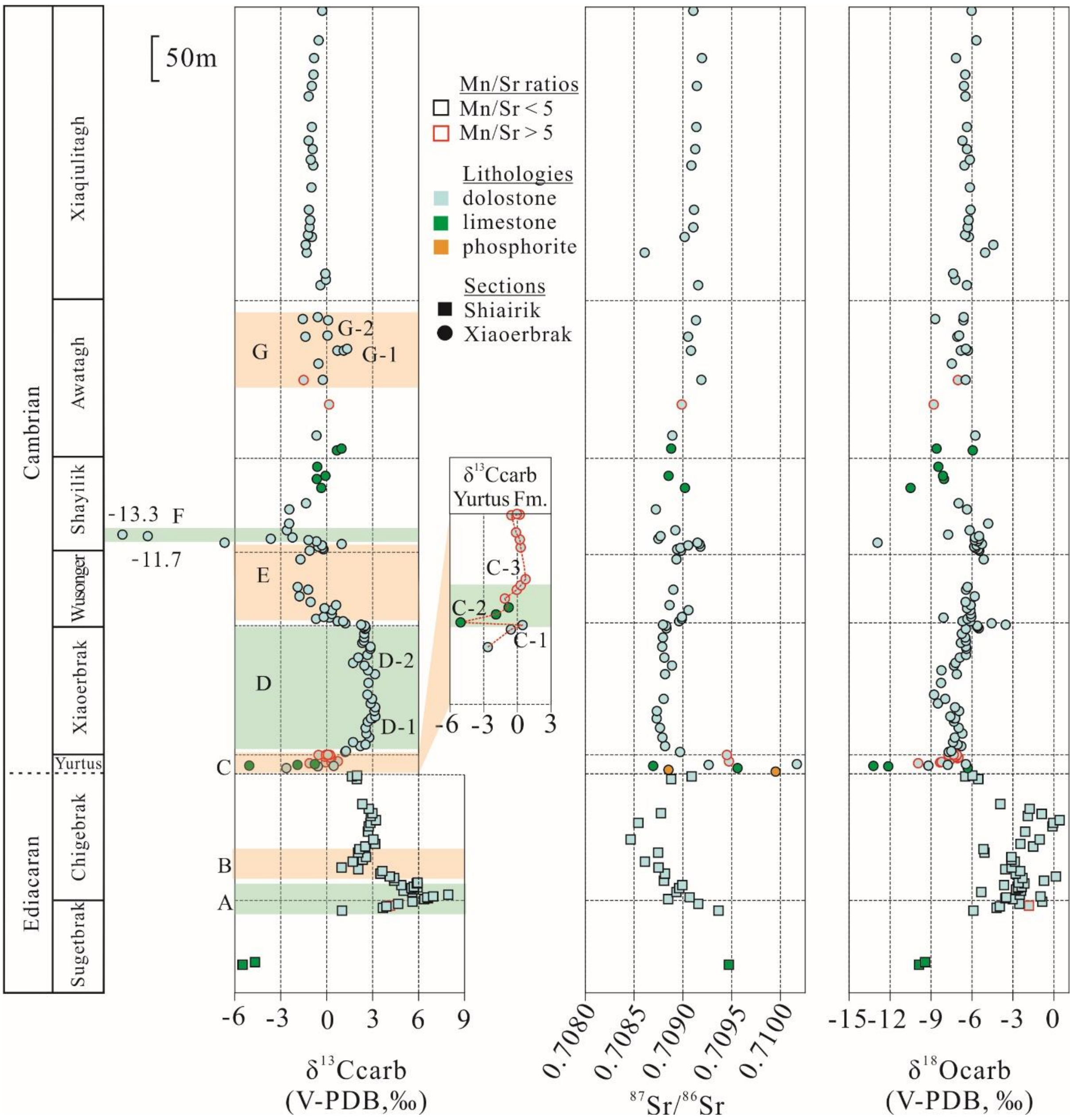

Cambrian
Ediacaran
Xiaqiulitagh
Awatagh
Shayilik
Wusonger
Xiaoerbrak
Yurtus
Chigebrak
Sugetbrak
50m
-13.3
F
-11.7
G
G-2
G-1
E
D
D-2
D-1
C
B
A
Mn/Sr ratios
☐ Mn/Sr < 5
☐ Mn/Sr > 5
Lithologies
■ dolostone
■ limestone
■ phosphorite
Sections
■ Shiairik
● Xiaoerbrak
δ^{13}Ccarb Yurtus Fm.
C-3
C-2
C-1
-6 -3 0 3
-6 -3 0 3 6 9
δ^{13}Ccarb (V-PDB,‰)
0.7080 0.7085 0.7090 0.7095 0.7100
^{87}Sr/^{86}Sr
-15 -12 -9 -6 -3 0
δ^{18}Ocarb (V-PDB, ‰)

Figure 4

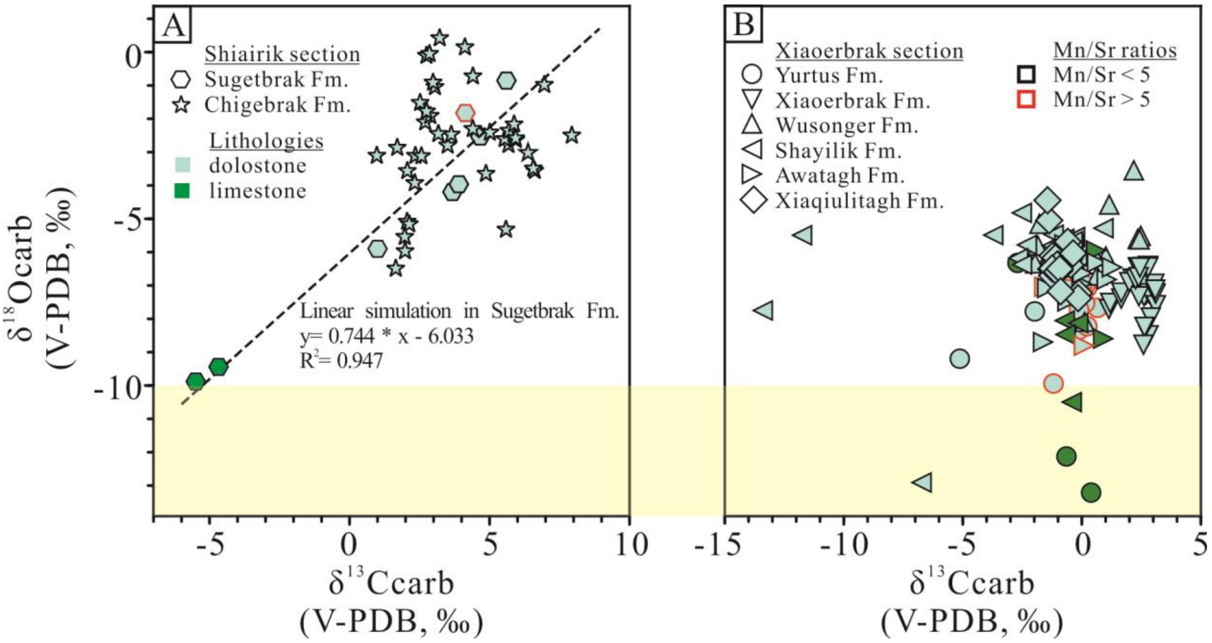

A
Shiairik section
Sugetbrak Fm.
Chigebrak Fm.
Lithologies
dolostone
limestone
Linear simulation in Sugetbrak Fm.
y= 0.744 * x - 6.033
R²= 0.947
δ18Ocarb (V-PDB, ‰)
0
-5
-10
-5
0
5
10
δ13Ccarb (V-PDB, ‰)
B
Xiaoerbrak section
Yurtus Fm.
Xiaoerbrak Fm.
Wusonger Fm.
Shayilik Fm.
Awatagh Fm.
Xiaqiulitagh Fm.
Mn/Sr ratios
Mn/Sr < 5
Mn/Sr > 5
-15
-10
-5
0
5
δ13Ccarb (V-PDB, ‰)

Figure 5

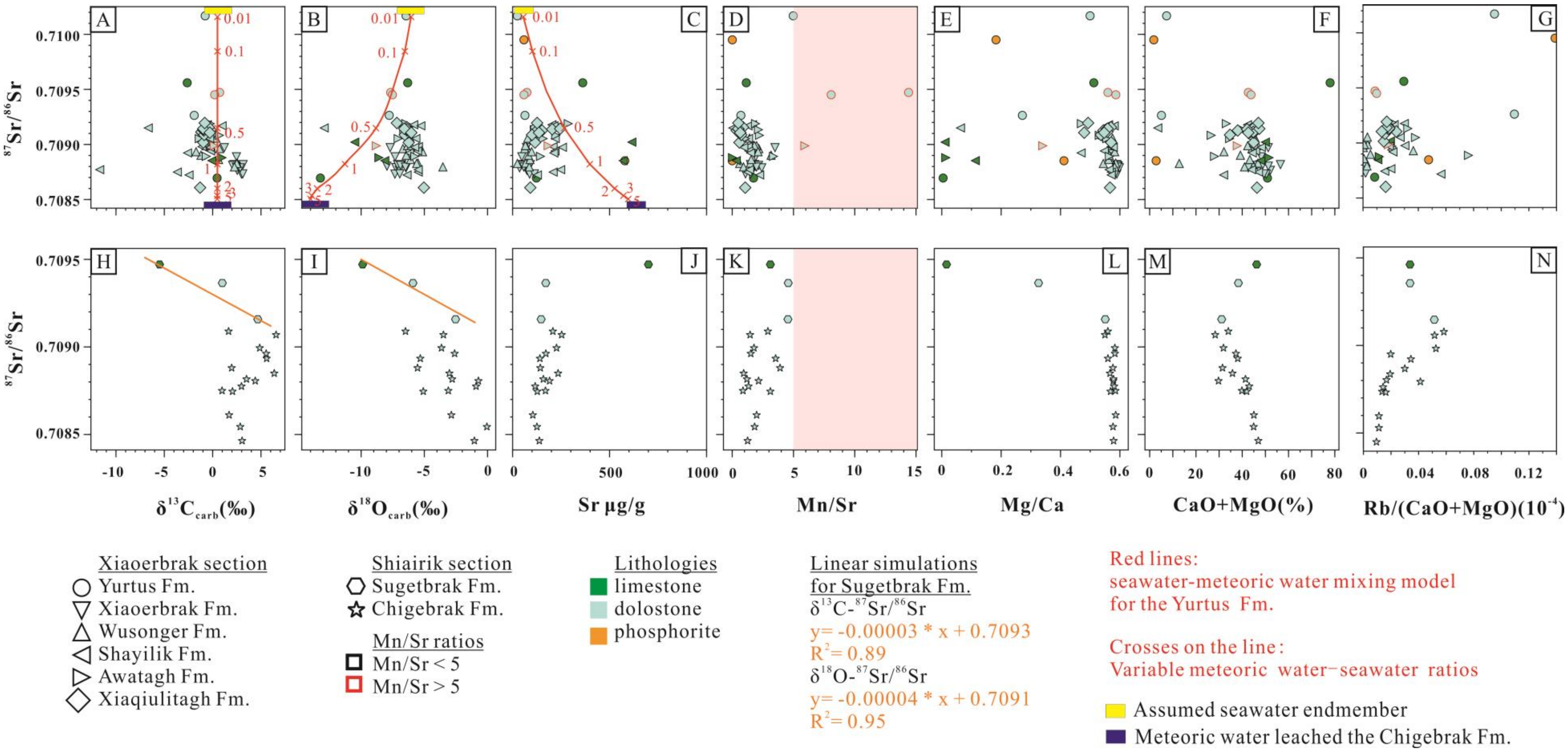

A
B
C
D
E
F
G
H
I
J
K
L
M
N
^{87}Sr/^{86}Sr
0.7085
0.7090
0.7095
0.7100
×0.01
×0.1
0.5
1
2
3
5
-10
-5
0
5
0
500
1000
10
15
0.2
0.4
0.6
20
40
60
80
0.04
0.08
0.12
δ^{13}C$_{carb}$(‰)
δ^{18}O$_{carb}$(‰)
Sr μg/g
Mn/Sr
Mg/Ca
CaO+MgO(%)
Rb/(CaO+MgO)(10^{-4})
Xiaoerbrak section
Yurtus Fm.
Xiaoerbrak Fm.
Wusonger Fm.
Shayilik Fm.
Awatagh Fm.
Xiaqiulitagh Fm.
Shiairik section
Sugetbrak Fm.
Chigebrak Fm.
Mn/Sr ratios
Mn/Sr < 5
Mn/Sr > 5
Lithologies
limestone
dolostone
phosphorite
Linear simulations for Sugetbrak Fm.
δ^{13}C-^{87}Sr/^{86}Sr
y= -0.00003 * x + 0.7093
R^2= 0.89
δ^{18}O-^{87}Sr/^{86}Sr
y= -0.00004 * x + 0.7091
R^2= 0.95
Red lines:
seawater-meteoric water mixing model for the Yurtus Fm.
Crosses on the line:
Variable meteoric water−seawater ratios
Assumed seawater endmember
Meteoric water leached the Chigebrak Fm.

Figure 6

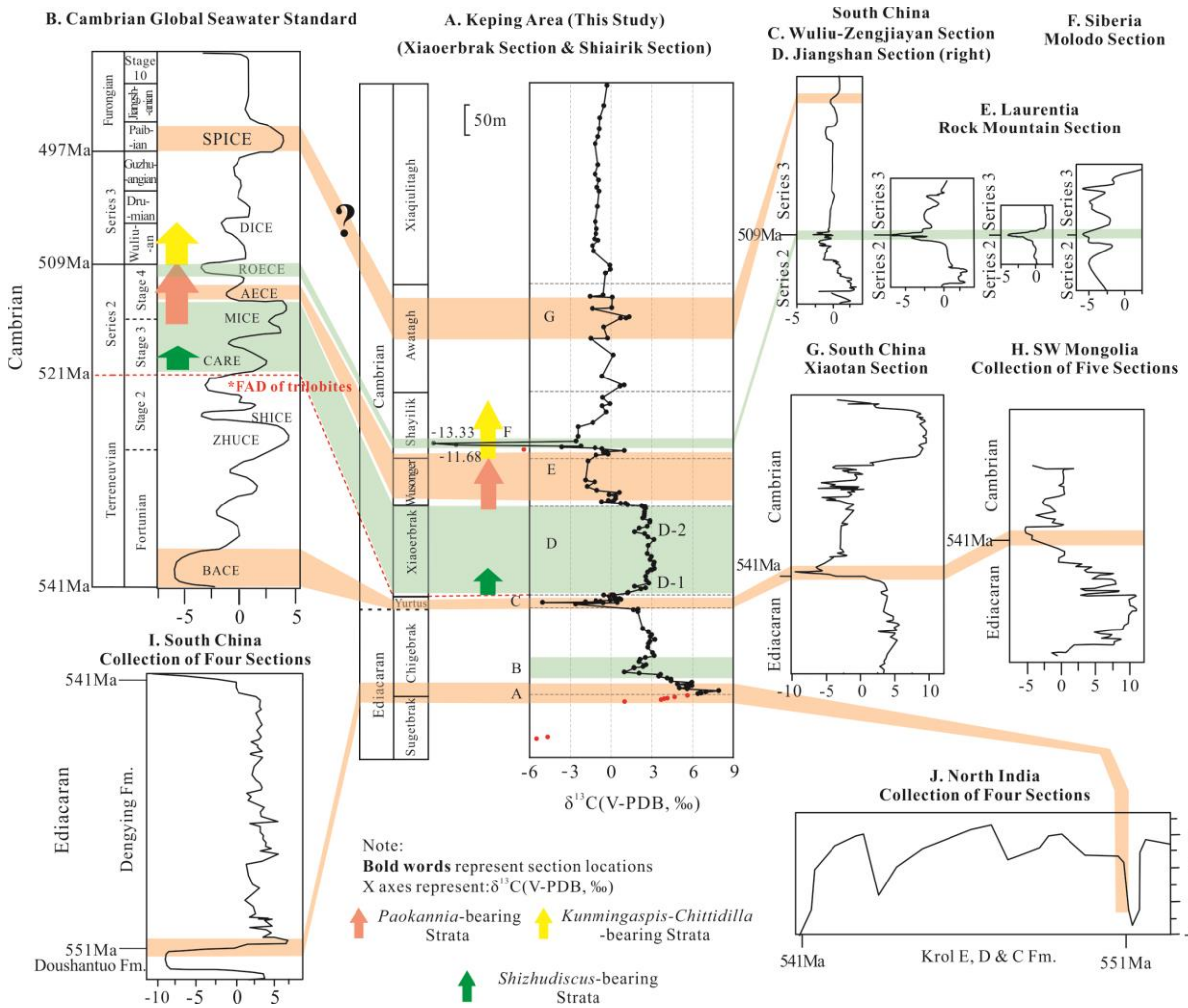

B. Cambrian Global Seawater Standard
A. Keping Area (This Study)
(Xiaoerbrak Section & Shiairik Section)
South China
C. Wuliu-Zengjiayan Section
D. Jiangshan Section (right)
F. Siberia
Molodo Section
E. Laurentia
Rock Mountain Section
G. South China
Xiaotan Section
H. SW Mongolia
Collection of Five Sections
I. South China
Collection of Four Sections
J. North India
Collection of Four Sections
Cambrian
Furongian
Series 3
Series 2
Terreneuvian
Stage 10
Jiangsh-anian
Paib-ian
Guzhu-angian
Dru-mian
Wuliu-an
Stage 4
Stage 3
Stage 2
Fortunian
497Ma
509Ma
521Ma
541Ma
SPICE
DICE
ROECE
AECE
MICE
CARE
*FAD of trilobites
SHICE
ZHUCE
BACE
-5
0
5
?
50m
Cambrian
Ediacaran
Xiaqiulitagh
Awatagh
Shayilik
Wusonger
Xiaoerbrak
Yurtus
Chigebrak
Sugetbrak
G
F
E
D
D-2
D-1
C
B
A
-13.33
-11.68
-6
-3
0
3
6
9
δ13C(V-PDB, ‰)
Series 3
Series 2
509Ma
Cambrian
Ediacaran
541Ma
-10
-5
0
5
10
551Ma
Ediacaran
Dengying Fm.
Doushantuo Fm.
541Ma
Krol E, D & C Fm.
551Ma
-10
-5
0
5
Note:
Bold words represent section locations
X axes represent:δ13C(V-PDB, ‰)
Paokannia-bearing Strata
Kunmingaspis-Chittidilla -bearing Strata
Shizhudiscus-bearing Strata

Figure 7

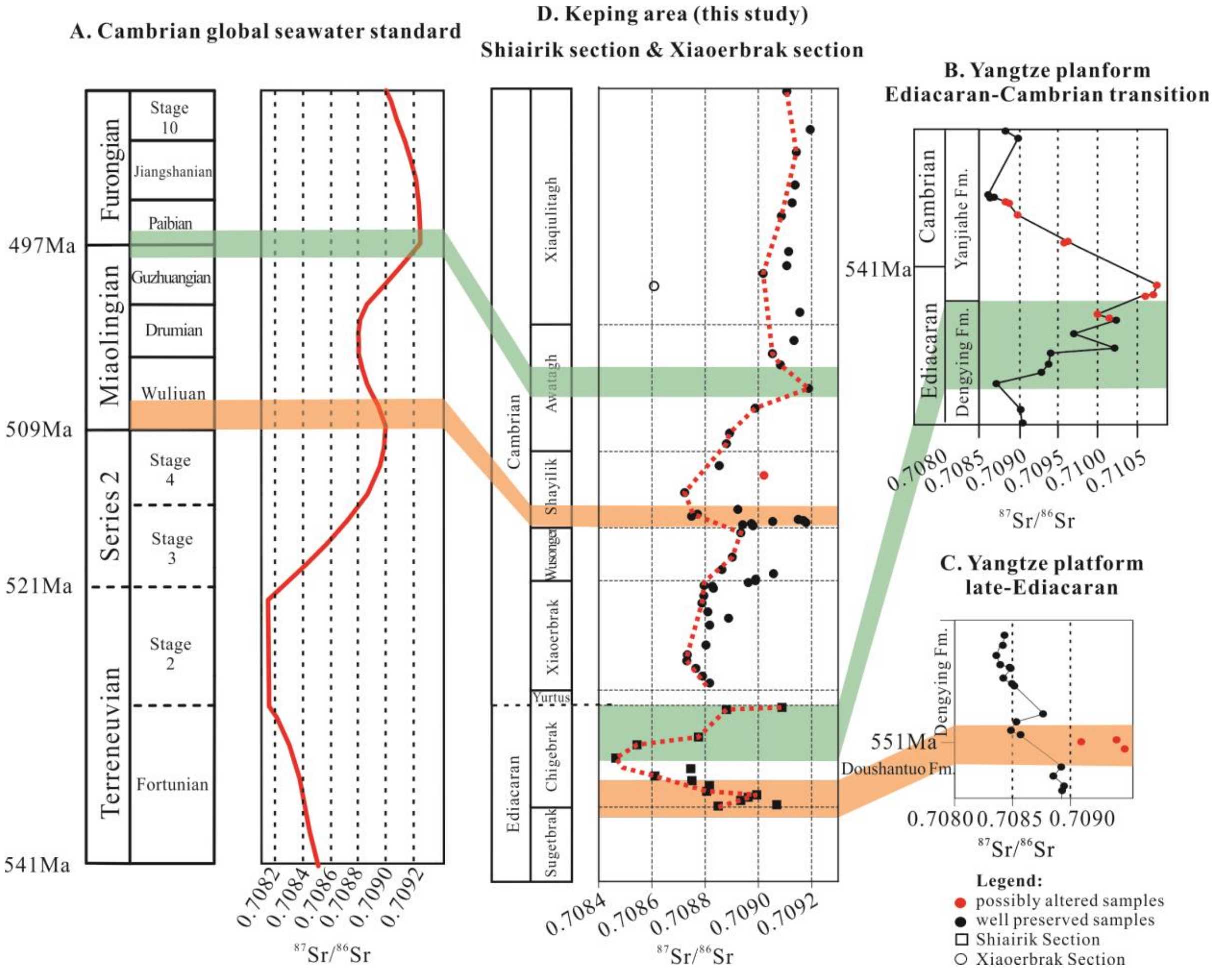
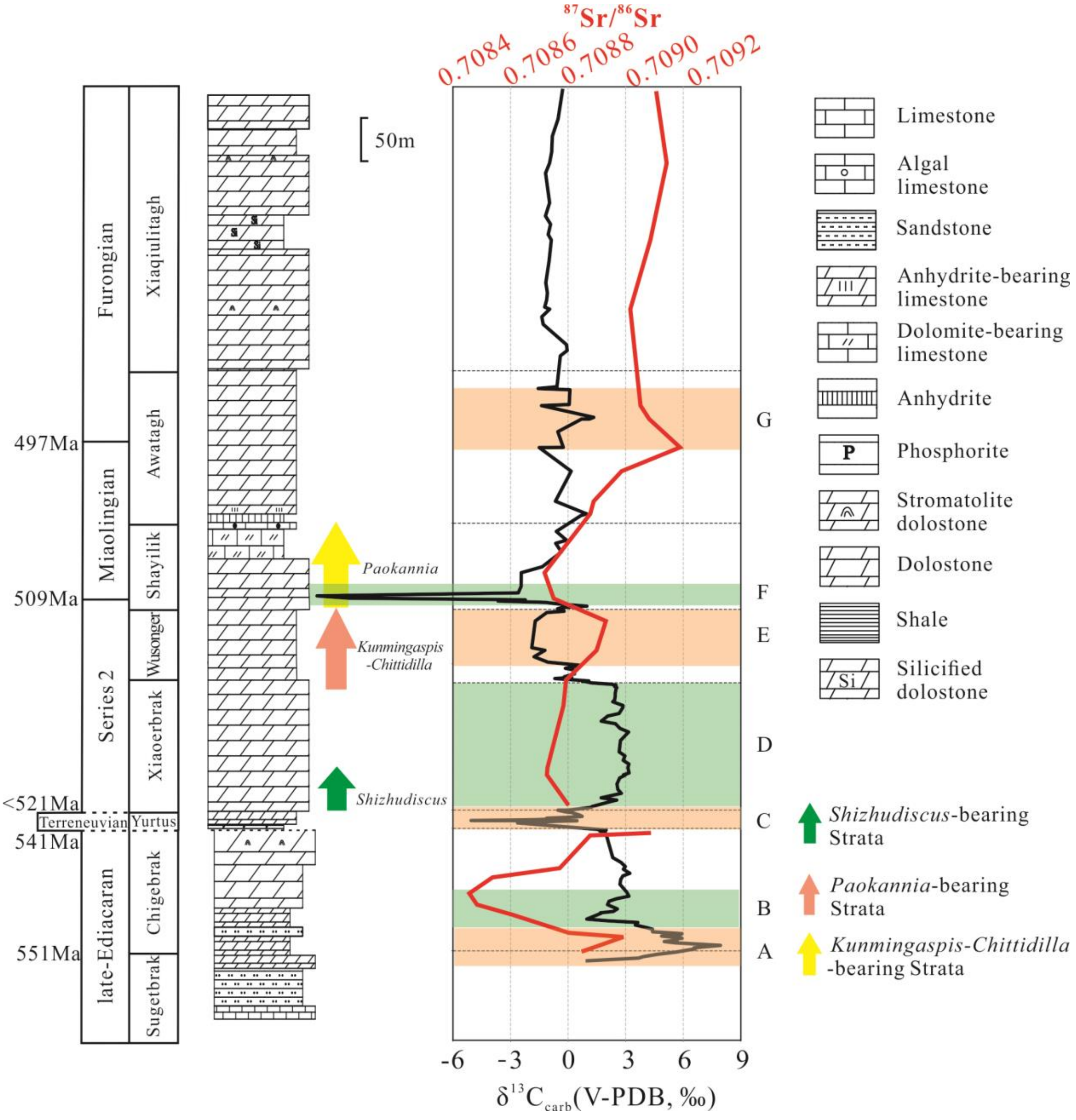

A. Cambrian global seawater standard
D. Keping area (this study)
Shiairik section & Xiaoerbrak section
B. Yangtze planform
Ediacaran-Cambrian transition
C. Yangtze platform
late-Ediacaran
497Ma
509Ma
521Ma
541Ma
Furongian
Miaolingian
Series 2
Terreneuvian
Stage 10
Jiangshanian
Paibian
Guzhuangian
Drumian
Wuliuan
Stage 4
Stage 3
Stage 2
Fortunian
0.7082
0.7084
0.7086
0.7088
0.7090
0.7092
^{87}Sr/^{86}Sr
Cambrian
Ediacaran
Xiaqiulitagh
Awatagh
Shayilik
Wusonger
Xiaoerbrak
Yurtus
Chigebrak
Sugetbrak
0.7084
0.7086
0.7088
0.7090
0.7092
^{87}Sr/^{86}Sr
Cambrian
Ediacaran
Yanjiahe Fm.
Dengying Fm.
541Ma
0.7080
0.7085
0.7090
0.7095
0.7100
0.7105
^{87}Sr/^{86}Sr
Dengying Fm.
551Ma
Doushantuo Fm.
0.7080
0.7085
0.7090
^{87}Sr/^{86}Sr
Legend:
● possibly altered samples
● well preserved samples
☐ Shiairik Section
○ Xiaoerbrak Section

Figure 8



$^{87}Sr/^{86}Sr$
0.7084
0.7086
0.7088
0.7090
0.7092
Furongian
Xiaqiulitagh
Awatagh
497Ma
Miaolingian
Shayilik
509Ma
Series 2
Wusonger
Xiaoerbrak
<521Ma
Terreneuvian
Yurtus
541Ma
late-Ediacaran
Chigebrak
551Ma
Sugetbrak
50m
Paokannia
Kunmingaspis-Chittidilla
Shizhudiscus
G
F
E
D
C
B
A
-6
-3
0
3
6
9
$\delta^{13}C_{carb}$(V-PDB, ‰)
Limestone
Algal limestone
Sandstone
Anhydrite-bearing limestone
Dolomite-bearing limestone
Anhydrite
Phosphorite
Stromatolite dolostone
Dolostone
Shale
Silicified dolostone
Shizhudiscus-bearing Strata
Paokannia-bearing Strata
Kunmingaspis-Chittidilla -bearing Strata

Figure 9