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# Glaciation and ~770 Ma Ediacara (?) Fossils from the Lesser Karatau Microcontinent, Kazakhstan

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ABSTRACT

The Cambrian explosion, c. 530–515 Ma heralded the arrival of a diverse assembly of multicellular life including the first hard-shelled organisms. Fossils found in Cambrian strata represent the ancestors of most modern animal phyla. In contrast to the apparent explosiveness seen in the Cambrian fossil record, studies of molecular biology hint that the diversification observed in Cambrian strata was rooted in ancestry extending back into the Ediacaran (635–542 Ma). Fossil evidence for this mostly cryptic phase of evolution is derived from the soft-bodied fossils of the Ediacaran biota found throughout the world and bilaterian embryos found in the Doushantuo lagerstätte in South China. The first appearance of Ediacara fauna is thought to have followed the last of the ~750–635 Ma Neoproterozoic glacial episodes by 20–30 million years. In this paper, we present evidence for the oldest discovery of the 'Ediacara' discoidal fossils *Nimbia occlusa* and *Aspidella terranovica* (?) that predate the early Cryogenian glaciations by more than fifty million years. There is considerable disagreement over the significance of discoidal Ediacaran fossils, but our findings may support earlier suggestions that metazoan life has roots extending deeper into the Proterozoic Eon. We also confirm the presence of a Late Cryogenian (e.g. "Marinoan") glaciation on the Lesser Karatau microcontinent including dropstones and striated clasts within the glacial strata.



## 1. Introduction

The Ediacaran Period is the most recent addition to the geologic time scale (Knoll et al., 2004). The Ediacaran type section is located in the Flinders Range of Australia and age estimates (635–542 Ma) are based on (a) the end of the "Marinoan" glaciation indirectly dated in Australia using correlations with well-dated sections found elsewhere throughout the world (Zhang et al., 2008) and (b) the base of the Cambrian as defined in Oman *(Amthor et al., 2003; Knoll et al., 2004; Jenkins, 2007; Shen and Schidlowski, 2010)*. Megascopic fossils of Ediacaran fauna are found in abundance beginning at around 570 Ma with a maximum diversity around 560–550 Ma *(Knoll and Carroll, 1999; Knoll et al., 2004; Martin et al., 2000)*. Although a few Ediacaran organisms persisted into the Cambrian, these enigmatic soft-bodied fossils are mostly absent in the Cambrian fossil record and were replaced by fauna that flourished in the Cambrian (Crimes et al., 1995; Crimes and McIlroy, 1999). Numerous explanations are proposed for the demise of the Ediacara and potential reasons for their extinction include an increase in predation and a change in the biodynamics of the seafloor (e.g. the 'substrate revolution' Bottjer et al., 2000). Less well known are the reasons for the rise of the metazoans although a common explanation is that environmental factors were more favorable for life following the severe Cryogenian glaciations including the "Gaskiers" glaciations at ~580 Ma (Bowring et al., 2003; McCall, 2006; Meert and Lieberman, 2008). Controversial accounts of much older (>635 Ma) Ediacaran type fossils or other metazoans are reported, but many of these discoveries are either poorly dated or the exact nature of the fossils is questioned (El Albani et al., 2010; McCall, 2006; Bengtson et al., 2007; Bengtson and Rasmussen, 2009; Malone et al., 2008; Meert and Lieberman, 2008). A recent report of a multicellular animals beneath the Marinoan glacial sequence in Australia suggests that there may be a more diverse (albeit poorly recognized) history of complex animal life in the Cryogenian (Maloof et al., 2010). El Albani et al. (2010) argue that the metazoan evolved in fits and starts as oxygen levels crossed critical thresholds at various times during the Proterozoic.

In this paper, we document the occurrence of early Cryogenian (>766 Ma) Ediacara fossils (*Nimbia occlusa and Aspidella terranovica* (?)) from the Lesser Karatau microcontinent in Kazakhstan. The fossils are found in the Kurgan Formation and lie well below a newly discovered glacial tillite (the Aktas tillite) of presumed Late Cryogenian age (~635 Ma; commonly referred to as Marinoan). U–Pb zircon dating

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of tuff layers above the fossils indicates an age for these discoidal impressions >766 Ma (Levashova et al., 2011). Our discovery may impact the evolutionary view of the Ediacara biota in several ways. First, if these are the remains of true metazoa, then the origins of some of the Ediacara must be extended back in time by at least an additional 100 million years. Secondly, our finding would lend support to the idea that the roots of the metazoa may extend deeper in geologic time and is perhaps tied to temporary increases in atmospheric oxygen (El Albani et al., 2010). At the very least, it would extend the record of *Nimbia* and *Aspidella terranovica* back to the early part of the Cryogenian. Lastly, we recognize the controversy surrounding these particular discoidal fossils. It is possible that our discovery of *Nimbia occlusa* and *Aspidella terranovica* (?) in sedimentary rocks during the early Cryogenian (>766 Ma) lends support the alternative hypotheses regarding these fossils and remove them from consideration as true metazoa (Grazhdankin and Gerdes, 2007; Retallack, 1994; Peterson et al., 2003; MacGabhann, 2007).

## 2. Geological setting

The location of our discovery lies within the micro-continental blocks that comprise the south-central part of the Eurasian continent in Kazakhstan (Fig. 1A). The Precambrian microcontinents within the Central Asian Orogenic Belt (CAOB) cluster mostly in the western part of Central Kazakhstan and to the south of the Siberian craton (Fig. 1A) (e.g., Rojas-Agramonte et al., 2011; Xiao and Kusky, 2009). Microcontinents with Precambrian basement are located in the western part of Central Kazakhstan and in the Central and North Tien Shan (Fig. 1A). These include the Kokchetav, Ulutau, Karatau-Talas, North Tien Shan, Aktau-Mointy sialic massifs and several smaller pieces of unknown affinity. The overall stratigraphic similarities between late Neoproterozoic–Cambrian sections of these microcontinents were noticed decades ago (e.g. Ankinovich, 1962; Zubtsov, 1971). The basement of these massifs consists primarily of Paleoproterozoic metamorphic complexes overlain by thick piles of younger Meso-Neoproterozoic sedimentary and metasedimentary sequences (Degtyarev and Ryazantsev, 2007). The Archean to Paleoproterozoic age of the crystalline basement is demonstrated for several blocks, including the Kokchetav, Ulutau, and North Tien Shan microcontinents, as compiled by Kröner et al. (2007). Detrital zircons with similarly old ages are found in sedimentary sequences from these microcontinents (Kröner et al., 2007). In some microcontinental blocks the available basement ages are considerably younger. For example an age of 880 ± 11 Ma was reported for the Aktau-Mointy microcontinent (multigrain zircon age; Kozakov et al., 1993), an age of 780 ± 20 Ma at Greater Karatau (multigrain zircon age; Kozakov et al., 1993). These younger ages do not preclude older rocks being present in the basement of these domains.

The Neoproterozoic (815–750 Ma) felsic and bi-modal volcanic series of several microcontinents are thought to be coeval and are typically correlated with each other (Chumakov, 2009). These include, for example, late Neoproterozoic volcanic rocks of the Koksu series from the Ulutau microcontinent, the Kainar Formation from Greater Karatau, the Bolshoy Naryn Formation from the Central Tien Shan, and the Altynsyngan Formation from the Aktau-Mointy and Junggar domains. Preliminary ages from Greater Karatau (~815 Ma) and Talas (~771 Ma) together with the data reported in Levashova et al. (2011) from Lesser Karatau support this correlation.

From ~750 Ma to ~550 Ma, the geologic record for most microcontinents of the Kazakhstan domain is fragmented. Sedimentary rocks overlie the basement and late Neoproterozoic volcanics on some microcontinents, for instance at the Ulutau microcontinent (Knipper, 1963), but ages are usually inferred on general grounds and often vary from publication to publication. In general, there appears to exist a regional ~200 Ma long hiatus, approximately until latest Neoproterozoic to earliest Cambrian time (Korolev and Maksumova, 1984).

The geological correlation between microcontinents improves in the terminal Neoproterozoic. The thick carbonate–clastic sequences of latest Neoproterozoic to early Paleozoic age are known on many CAOB microcontinents (Khain et al., 2003; Popov et al., 2009; Chumakov, 2009). The examples include the Karal and Basagin Fms. of the Aktau-Junggar massif and the Tamdy Series of the Lesser Karatau in Kazakhstan. These carbonate–clastic sequences show striking similarities (Ankinovich, 1962; Zubtsov, 1971). The most notable marker horizons are the late Neoproterozoic glacial diamictites, that are known at one or more stratigraphic levels on several microcontinents (Chumakov, 1978; Korolev and Maksumova, 1984; Chumakov, 2009) and phosphorite layers that occur at the Ediacaran–Cambrian boundary on some of these blocks (Korolev and Maksumova, 1984; Meert and Lieberman, 2008).

Overall stratigraphic, faunal and lithological similarities led many authors to hypothesize that these blocks originally constituted a continent-size “Kazakhstan” domain with a Paleoproterozoic basement and latest Neoproterozoic to early Paleozoic sedimentary cover (e.g., Degtyarev and Ryazantsev, 2007 and references therein).

The geology of the Maly Karatau section includes Paleoproterozoic basement rocks covered by Neoproterozoic to early Paleozoic sedimentary and volcaniclastic rocks that form a series of steep nappes (Fig. 1B and C). The names and spelling of these units are highly variable (see Eganov et al., 1986; Popov et al., 2009; Sergeev and Schopf, 2010) and we follow the hierarchy of Eganov et al. (1986) in this paper. The units are comprised of (in stratigraphic order from oldest to youngest): the Kokdzhot, Bolshekeroi, Zhanatass, Koksu, Malokaroi and Tamdy Series (See Eganov et al., 1986; Sergeev and Schopf, 2010; Sergeev, 1989). The older sequences (Kokdgot–Koksu Series) are not discussed in this paper although we note that zircons inherited in the tuffs of the Kurgan Formation yield ages of >2000 Ma for the basement of the region (see Levashova et al., 2011). Overlying the Koksu Group is the Malokaroi Series (Fig. 2) that consists of (from oldest to youngest) the sandstones and gravellites of the Aktugai Formation (~30–200 m thick); black mudstones, clastic rocks, cherts and limestones of the Chichkan Formation (up to ~120 m thick) followed by the volcano-sedimentary rocks of the Kurgan Formation (Eganov et al., 1986; Sergeev and Schopf, 2010). The Malokaroi Series is unconformably overlain by the carbonate-dominated Tamdy Series generally considered to be of Ediacaran-Lower Ordovician age. The uppermost Tamdy Series (phosphorites and carbonates) are well-exposed and although they are faulted and folded, stratigraphic relationships and unconformities are clearly recognized. In addition, several distinct fossil assemblages occur in the Chuluktau and Shabakty Suites that provide age constraints for the Middle and Upper parts of the Tamdy Series.

The Tamdy Series consists of a lower terriginous and dolomitic Kyrshabakty Suite, middle phosphatic and carbonaceous Chuluktau Suite and the very thick carbonate-dominated Shabakty Suite (Fig. 2). Direct geochronologic constraints are lacking on most of the sequence with the exception of the upper part of the Tamdy Series (Chuluktau and Shabakty Suites) where fossil correlations provide a Cambrian–Ordovician ages (Eganov et al., 1986). The Shabakty Suite, unconformably overlies the Chuluktau Suite and contains trilobite fossils including *Hebediscus orientalis*, *Ushbaspis limbata, Redlichia-chinensis and Kootenia gimmerlfarbi* (Popov et al., 2009) and is assigned to the Botomian–Amgan Stages (Geyer and Shergold, 2000). The middle part of the Tamdy Series, the Chuluktau Suite, contains small-shelly fossils of the *Protohertzina anabarica* zone in the so-called “Lower dolomite” (Eganov et al., 1986; Popov et al., 2009), *Pseudorthotheca costata* in the phosphorite zone (Tommotian stage), *Rhombocorniculum cancellatum* and *Bercutia cristata* in the uppermost dolomite of the Chuluktau (Atdabanian stage). The *Protehertzina anabarica* zone is constrained in the Meishucun section of South China to ~535 Ma (Zhu et al., 2009). In the Meishucun section, the major phosphorite beds also occur around the same time and we therefore tentatively correlate this part of the

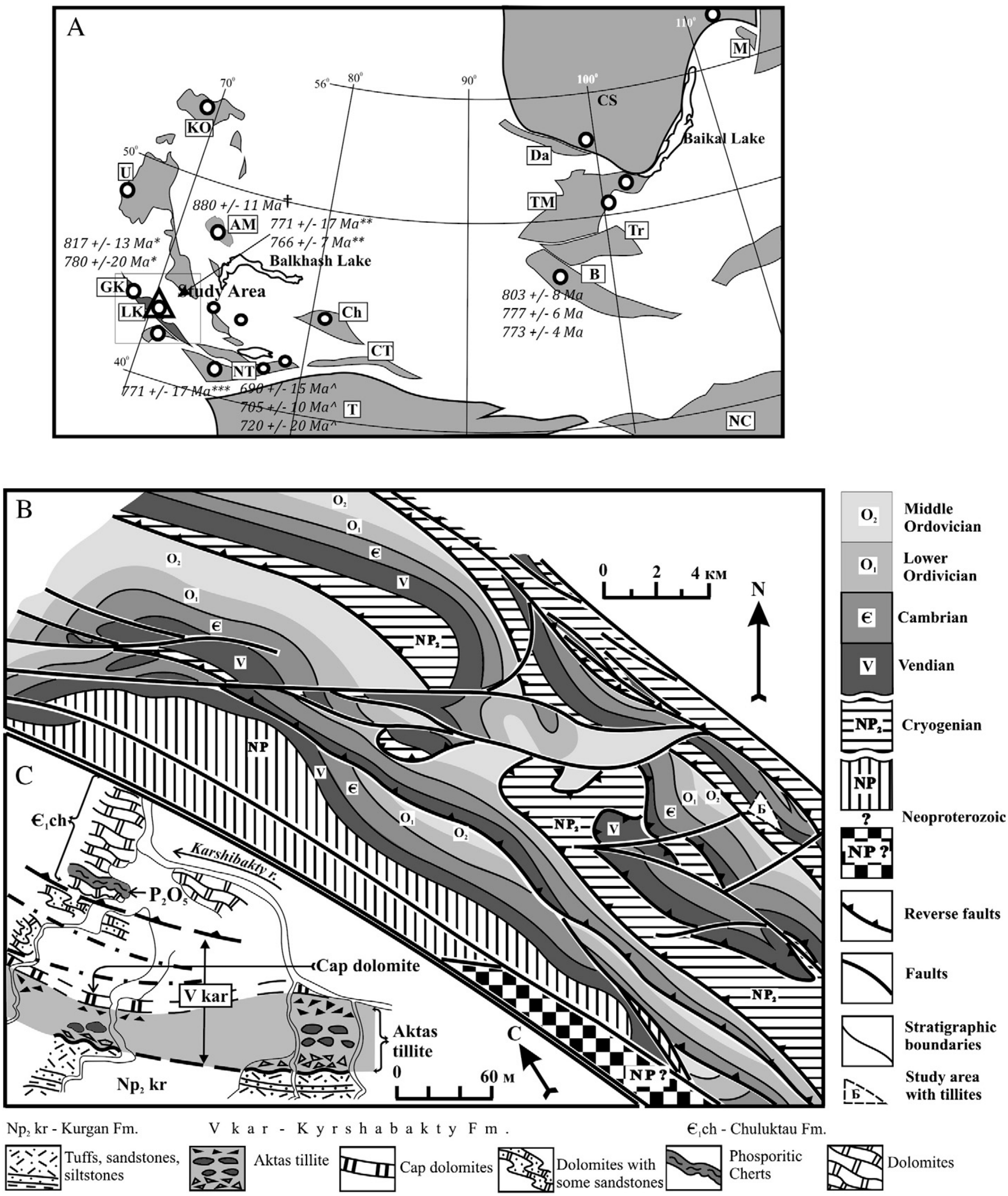


**Fig. 1.** (A) Location of microcontinental fragments and tillite deposits in the Ural-Mongol fold belt, KO-Kokchetav; AM = Aktau-Mointy, U = Ulutau; LK = Lesser Karatau; Ch = Chy-Ili, CT = Central Tien Shan; NT = Northern Tien Shan; TM = Tuva-Mongolia; Tr = Tarbogatay; B = Baydaric; NC = North China; T = Tarim; Da = Derbi Arzube; M = Mui; S = Siberia. Lesser Karatau tillites are denoted by triangles with circles and other tillites by circles. Rectangle shows the study area in Kazakhstan. (B) Geologic sketch map of the the Lesser Karatau region (small triangle represents area shown in detail in Fig. 1 (C) closeup geological map of the tillite discovery at Lesser Karatau. U–Pb ages for the volcanic sequences at Greater Karatau*, Lesser Karatau**, Talas***, Dzhetym^, Aktau–Mointy† and the Baydaric microcontinents are taken from Levashova et al. (2010, 2011), Pradhan (personal communication), Sudorgin, 1992; Korolev and Maksumova, 1984; Kiselev, 2001; Kozakov et al., 1993; Sovetov, 2008; Zhao et al., 2006.

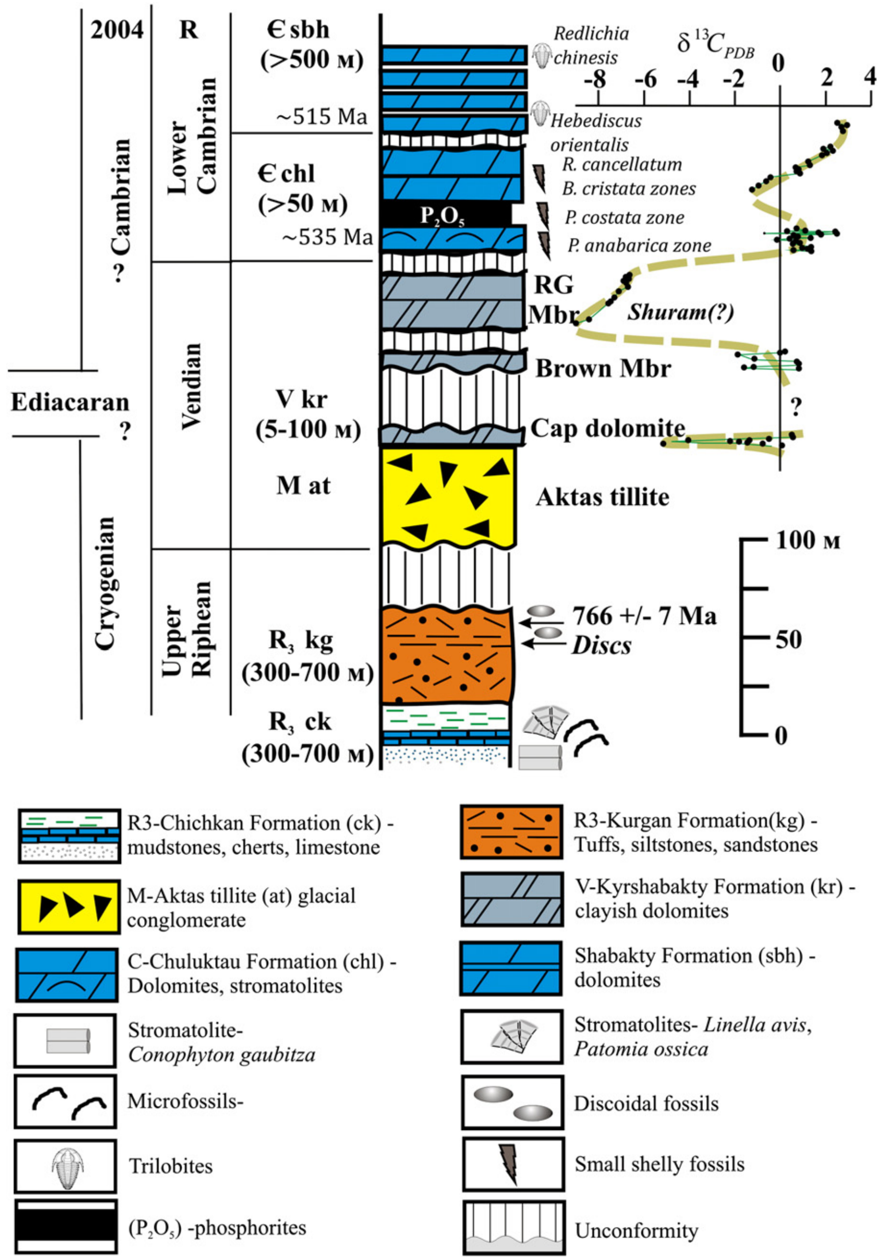


**Fig. 2.** Stratigraphic column for the Lesser Karatau sequence showing the relative locations of the fossil discoveries *Nimbia occlusa* and *Aspidella terranovica* along with previously described stromatolites, small shelly fossils such as *Protohertzina anabarica Miss. and Unguliformis Miss.*) and trilobitic (*Kootenia* and *Redlichia*) fossils in the upper part of the section. The basal section of the Shabakty formation also hosts abundant phosphatic sclerite fossils of *Mongolitubulus* and *Microdictyon.* We also show the $\delta^{13}C$ stable isotopic curve along with our interpretation of that curve. The glacial sequence is interpreted as Marinoan in age (~635–650 million years old) and the large negative excursion below the phosphate layer is assigned to the Shuram/Wonoka anomaly. GTS = ICS2004 time scale, R2000 = Russian 2000 time scale nomenclature. Thickness ranges are given in meters.

Meishucun section to the Chuluktau Suite of the Tamdy Series in Maly Karatau.

The basal part of the Tamdy Series is represented by the Kyrshabakty Suite (Fig. 2). The exact nature of the boundary between the Kyrhsabakty Suite and the overlying Chuluktau Suite are problematic. Part of the confusion arises from the fact that a fossiliferous dolomite called either the “Lower Dolomite” or the “Berkuty Member” has been assigned by various authors to the uppermost Kyrshabakty Suite or the lowermost Chuluktau Suite (Popov et al., 2009; Eganov et al., 1986). Eganov et al. (1986) assigned the Kyrshabakty to the “Late Vendian” whereas Mambetov (1993) placed the uppermost Kyrshabakty Suite (Berkuty Member) in the Nemakit-Daldyn based on the presence of *Protohertzina anabarica*. We will argue for an older (~635 Ma) age for the basal part of the Kyrshabakty Suite. Eganov et al. (1986) note that the basal Kyrshabakty Suite rests unconformably over the underlying Kurgan Formation. The thickness and quality of exposure of the Kyrshabakty Suite is variable with measured sections ranging from just a few meters to over 150 m. Eganov et al. (1986) describe the stratigraphy of the Kyrshabakty Suite as follows. The Suite is floored by a basal conglomerate containing poorly rounded clasts of tuffaceous and other materials from the underlying Proterozoic rocks in a mudstone-sandstone sized matrix. This basal unit was never positively identified as having a glacial origin; however, Eganov and Sovetov (1979) mentioned these as ‘tilloids’. The ‘tilloid’ is overlain by a widespread *pink dolomite* with a thickness ranging from 10 to 12 m. The pink dolomite is overlain by a terriginous and carbonate mixed sequence. According to Eganov et al. (1986), the upper part of the Kyrshabakty Suite is conformable with the "Lower Dolomite" of the Chuluktau Suite; however, in other cases the Kyrshabakty is absent or the conformable nature of the contacts is difficult to ascertain.

As discussed later, the lowermost Kyrshabakty Suite exhibits characteristics of a glacial origin and is capped by relatively thin pink dolomites. One of the hallmark traits of many Cryogenian glacial sequences around the globe is glacial deposits overlain by pink 'cap' dolomites (Hoffman and Schrag, 2002).

### 2.1. Lower Kyrshabakty glacials

Tillites are found in the Aksay member of the Kyrshabakty Suite at an outcrop along the Kyrshabakty River (Fig. 1C; 40°32′02.4″N, E69°57′15.3″E) where the following section is exposed. Five to seven meters of indurated angular debris of green tuffs of the Kurgan formation are followed by an ~30-meter thick well-indurated diamictite member. This member consists of chaotically distributed clasts that are found among small angular debris and sand in a green clayish matrix (Fig. 3A). Some of the clasts are typical dropstones with a keel or undulating surface. The rear parts of dropstones are very uneven and show no rounding and sets of scratches and minor troughs are well developed on other surfaces (Fig. 3B). An upper clastic bed, about 10–15 m thick, consists of chaotically distributed angular clasts (2–5 cm in size) that account for ~20% of rock volume. This upper clastic bed is followed by a 30–40 cm thick transitional zone from tillites to pink cap dolomites. The zone includes a thin layer of green slate with lenses of sandy dolomite and finely laminated red slates with a bed of massive fine-grained pink dolomite at the top. The main part of cap-dolomite is represented by dense porcelain-like layered rock, sometimes with a wavy pattern that may be layered stromatolites. In this location, the cap dolomite is 1 to 3 m thick and is followed, with a hiatus, by a brownish fine-grained dolomite.

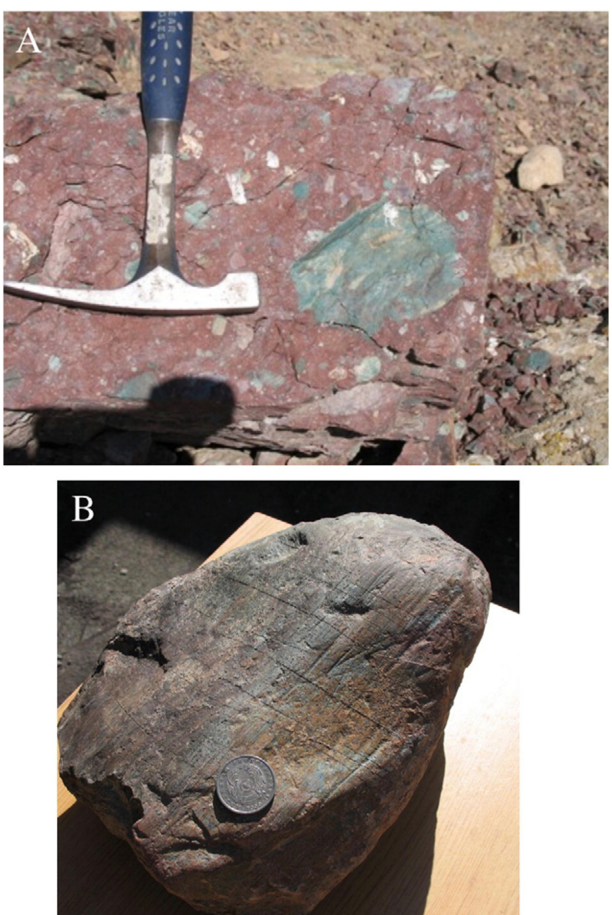


**Fig. 3.** (A) Diamictite deposit from the Aktas tillite showing angular clasts of the underlying Kurgan Series, (B) glacially striated dropstone from the Aktas tillite.

## 3. $\delta^{13}C$ studies

Samples for carbon isotopic work were micro-drilled from pristine carbonate samples with a 2 mm drill bit. Samples were collected along a transect that included the pink dolomites of the Kyrshabakty Suite into the overlying Chuluktau and Shabakty Suites. Because of the nature of the outcrop in the region samples were taken at intervals that ranged from less than 0.5 m to more than 2 m (a total of 80 samples, Table 1). Samples were not collected from the phosphorite member of the Chuluktau Suite or from regions where outcrops were covered or missing.

Carbonate samples were dissolved in ortho-phosphoric acid for two hours at 70 ° C and then linked directly to a semi-automatic Gas Bench II device connected to a Finnigan MAT-253 mass-spectrometer. $\delta^{13}C$ and $^{18}O$ were measured with a precision of not less than 0.1‰ and 0.3‰, respectively. $^{18}O$ measurements are compared to SMOW (Standard Mean Ocean Water) and PDB (Pee Dee Belemnite). The precision was monitored with international (NBS-19), Russian (VNIIYGG, DVGI) and laboratory (Ca-770) standards (Table 1).

$\delta^{13}C$ values within the stratigraphic section varied between −8.95‰ to 2.93‰. A cross-plot of $\delta^{13}C$ and $\delta^{18}O$ show no correlation ($R^2 = 0.5686$) suggesting that little diagenetic alteration of the primary $\delta^{13}C$ values has occurred (See Fairchild et al., 1990; Fig. 4). Derry (2010) pointed out that most studies of glacial sequences lump all analyses into a single comparison of $\delta^{13}C$ and $\delta^{18}O$ and therefore miss some important correlations (particularly in regions where the $\delta^{13}C$ values are extremely negative). We have performed this comparison (not illustrated) for all stratigraphic intervals and found no significant correlation.

Fig. 5A–D shows the trends in the $\delta^{13}C$ values in the Kyrshabakty, Chuluktau and lowermost Shabakty Suites. The signal begins with a slightly positive value at the base of the Kyrshabakty cap dolomite followed by mostly negative values in the first 3 m (total thickness of the cap in this locality). In the unconformably overlying brown dolomite, the $\delta^{13}C$ values range between +0.85‰ and −2.25‰. The brown dolomite is overlain by a red and grey layer with mixed carbonate/terrigenous sedimentary rock. Values of $\delta^{13}C$ in this section are the most negative in the entire sequence and range from a low of −8.95‰ at the base of the layer to −6.6‰ at the top.

Although there is some discussion regarding the global synchroneity and veracity of carbon isotopic values in the Neoproterozoic (Meert, 2007; Halverson et al., 2005; Knauth and Kennedy, 2009; Macdonald et al., 2010; Derry, 2010; Hebert et al., 2010; Frimmel, 2010), we compare our results to the most recent 'composite' $\delta^{13}C$ curve of Macdonald et al. (2010). The glacial Aktas tillite lies unconformably above the Kurgan Formation that is dated (near the top) to 766 Ma (Levashova et al., 2011). It lies below the Berkuty dolomite of the Chuluktau Suite containing the *Protohertzina anabarica* small-shelly fossil (SSF) assemblage. This SSF assemblage is constrained in the South China Meishucun section to be 535 Ma and younger. Based on its stratigraphic position, the Aktas tillite could be of Sturtian (~716 Ma), Marinoan (~650 Ma), Gaskiers (~580 Ma), or Baykonurian (terminal Ediacaran; Chumakov, 2009, 2010) age. While we cannot *a priori* reject a Sturtian age for the glaciation, we find this a less-likely scenario based on the deep erosional and angular unconformable relationship with the underlying 766 Ma Kurgan Formation.

Therefore we can tentatively assign this glaciation to either the Late Cryogenian (~635 Ma);) the Gaskiers glaciation (~580 Ma) or the Baykonurian glaciations (~550–540 Ma). The $\delta^{13}C$ values reach

**Table 1**
Carbon and oxygen isotopic compositions.

| Stratigraphic level | (M)[a] | $\delta^{13}C$ | $\delta^{18}O$ (PDB) | $\delta^{18}O$ (SMOW) |
|---|---|---|---|---|
| *Kyrshabkty suite* | | | | |
| Cap dolomite | 0 | 0.59 | −3.07 | 27.69 |
| Cap dolomite | 0 | −5.61 | −10.73 | 19.8 |
| Cap dolomite | 0.1 | −4.05 | −11.38 | 19.12 |
| Cap dolomite | 0.4 | −0.85 | 1.2 | 32.1 |
| Cap dolomite | 0.4 | −1 | −0.25 | 30.6 |
| Cap dolomite | 0.4 | −1.3 | −1.51 | 29.3 |
| Cap dolomite | 1 | 0.55 | −3.4 | 27.35 |
| Cap dolomite | 1.5 | 0.06 | −4.13 | 26.6 |
| Cap dolomite | 2 | −0.8 | −4.04 | 26.7 |
| Cap dolomite | 2.1 | −1.44 | −4 | 26.74 |
| Cap dolomite | 2.7 | −1.8 | −3.84 | 26.9 |
| Cap dolomite | 3 | −2.25 | −4.66 | 26.06 |
| Brown dolomite | 3.5 | 0.85 | −2.49 | 28.29 |
| Brown dolomite | 4 | −1.61 | −4.35 | 26.38 |
| Brown dolomite | 4.7 | −0.11 | −1.87 | 28.93 |
| Brown dolomite | 6 | 0.8 | −3.45 | 27.3 |
| Brown dolomite | 7 | 0.76 | −7.94 | 22.67 |
| Brown dolomite | 8.5 | −0.11 | −7.65 | 22.98 |
| Brown dolomite | 10.6 | −1.85 | −5.04 | 25.67 |
| Brown dolomite | 10.8 | 0.02 | −3.99 | 26.75 |
| Brown dolomite | 11 | 0.2 | −4.23 | 26.5 |
| Red member | 15.5 | −8.95 | −8.91 | 21.67 |
| Red member | 17.5 | −8.36 | −10.32 | 20.22 |
| Red member | 25.5 | −7.57 | −9.41 | 21.15 |
| Red member | 26.5 | −7.46 | −9.15 | 21.43 |
| Red member | 27.5 | −7.37 | −9.13 | 21.44 |
| Red member | 28.5 | −7.27 | −8.86 | 21.73 |
| Red member | 31.5 | −7.09 | −8.97 | 21.61 |
| Red member | 33.5 | −6.67 | −8.59 | 22 |
| Grey member | 35.5 | −6.8 | −8.89 | 21.7 |
| Grey member | 36.5 | −6.92 | −8.47 | 22.13 |
| Grey member | 37.5 | −6.66 | −8.79 | 21.8 |
| Grey member | 38.5 | −6.8 | −8.69 | 21.9 |
| Grey member | 39.5 | −6.6 | −8.69 | 21.9 |
| *Chuluktau suite* | | | | |
| Berkuty dolomite | 3.3 | 1.32 | −2.83 | 27.94 |
| Berkuty dolomite | 3.4 | 1.19 | −2.4 | 28.38 |
| Berkuty dolomite | 3.5 | 0.6 | −3.87 | 26.87 |
| Berkuty dolomite | 3.8 | 1.27 | −1.78 | 29.02 |
| Berkuty dolomite | 4.5 | 0.98 | −2.88 | 27.89 |
| Berkuty dolomite | 4.7 | 1.19 | −2.59 | 28.2 |
| Berkuty dolomite | 5 | 1.32 | −2.92 | 27.85 |
| Berkuty dolomite | 5.5 | 0.61 | −3.34 | 27.42 |
| Berkuty dolomite | 6.5 | 0.86 | −2.87 | 27.9 |
| Berkuty dolomite | 7 | 0.72 | −3.9 | 26.84 |
| Berkuty dolomite | 8 | −0.1 | −3.22 | 27.54 |
| Berkuty dolomite | 8.5 | 0.43 | −3.39 | 27.37 |
| Berkuty dolomite | 9 | 1.3 | −4.04 | 26.7 |
| Berkuty dolomite | 9.4 | 0.68 | −3.82 | 26.92 |
| Berkuty dolomite | 9.8 | 0.86 | −4.68 | 26.04 |
| Berkuty dolomite | 10 | 0.56 | −4.35 | 26.37 |
| Berkuty dolomite | 10.8 | 0.63 | −4.24 | 26.49 |
| Berkuty dolomite | 11 | 1.81 | −4.21 | 26.52 |
| Berkuty dolomite | 11.2 | 2.45 | −3.84 | 26.9 |
| Berkuty dolomite | 11.7 | −0.74 | −5.04 | 25.66 |
| Berkuty dolomite | 12 | 1.77 | −4.01 | 26.73 |
| Berkuty dolomite | 12.2 | 2.4 | −3.65 | 27.1 |
| Berkuty dolomite | 12.8 | 0.1 | −5.01 | 25.7 |
| Berkuty dolomite | 13 | 1.1 | −4.13 | 26.6 |
| Berkuty dolomite | 13.2 | 0.68 | −4.62 | 26.1 |
| Ferro-Mn carbonate | 34 | −1.2 | −6.17 | 24.5 |
| Ferro-Mn carbonate | 36 | −0.96 | −7.79 | 22.83 |
| Ferro-Mn carbonate | 38 | −0.6 | −6.95 | 23.7 |
| Ferro-Mn carbonate | 40 | −0.4 | −5.39 | 25.3 |
| Ferro-Mn carbonate | 42 | 0.9 | −4.71 | 26 |
| Brown dolomite | 43 | 0.74 | −4.51 | 26.21 |
| Brown dolomite | 44 | 0.79 | −4.06 | 26.67 |
| Brown dolomite | 45 | 0.73 | −4.5 | 26.22 |
| Brown dolomite | 47 | 1.28 | −4.85 | 25.86 |
| Brown dolomite | 49 | 1.25 | −5.18 | 25.52 |
| Brown dolomite | 52 | 1.85 | −5.41 | 25.28 |
| Brown dolomite | 53 | 2.05 | −4.32 | 26.4 |
| Brown dolomite | 54 | 2.3 | −4.42 | 26.3 |
| *Chuluktau suite* | | | | |
| Brown dolomite | 55 | 1.9 | −3.98 | 26.75 |
| Brown dolomite | 56 | 2.2 | −4.13 | 26.6 |
| *Shabakty series* | | | | |
| Member A | 57 | 2.8 | −4.81 | 25.9 |
| Member A | 59 | 2.76 | −4.24 | 26.49 |
| Member A | 60 | 2.93 | −4.12 | 26.62 |
| Member A | 61 | 2.48 | -5.47 | 25.22 |

[a] As measured from the base of cap dolomite (for the Kyrshabakty Series) or from the base of the Berkuty Dolomite (Chuluktau & Shabakty Series).

maximum negative values of −5.6‰ in the Aktas cap carbonates to more positive values in the so-called overlying brown dolomite member of the Kyrshabakty (Figs. 2 and 5A). The magnitude of the shift is similar to the Maiberg cap anomaly (Namibia, Macdonald et al., 2010) although the thickness of the pink cap dolomite overlying the Aktas tillite is much smaller than other cap dolomites around the globe.

A very large negative shift in $\delta^{13}C$ values is observed in the red and grey dolomites of the Kyrshabakty Suite (up to −9‰; Figs. 2, 5B). The trend shows a steady, but small increase in values from the lower part of the section to the uppermost where $\delta^{13}C$ values are −6.5‰. If our assessment of the age for the Aktas tillite is correct, then this large negative excursion observed in the upper part of the Kyrshabakty Suite may correspond to the Shuram/Wonoka anomaly (Le Guerroué et al., 2006; Le Guerroué and Cozzi, 2010; Le Guerroué, 2010) or alternatively it may correlate with the negative $\delta^{13}C$ anomaly in South China (e.g. the DOUNCE anomaly; Zhu et al., 2007).

The Shuram anomaly is argued to be both the most negative carbon excursion in the past 600 million years and also the longest in duration. Le Guerroué (2010) estimates that from the onset until the return to neutral or positive $\delta^{13}C$ values may span over 50 million years and includes the time interval of the Gaskiers glaciation. Because of the long duration of the Shuram excursion and the overall lack of robust age constraints on the event, it is difficult to estimate the age of the red and grey dolomites at Maly Karatau. An extremely tentative comparison of the Maly Karatau values to the curve of Le Guerroué (2010) indicates an age of ~580–570 Ma for the anomaly in the red and grey dolomites. The existence of this anomaly in the sedimentary sequence above the Aktas glaciation supports (though it does not confirm) a Late Cryogenian age (~635 Ma Ghaub/Nantuo/Marinoan glaciation) for the tillites.

In contrast, Chumakov (2009, 2010) correlates numerous glacial sequences in Kazakhstan and Krygyzstan to end Ediacaran glaciation

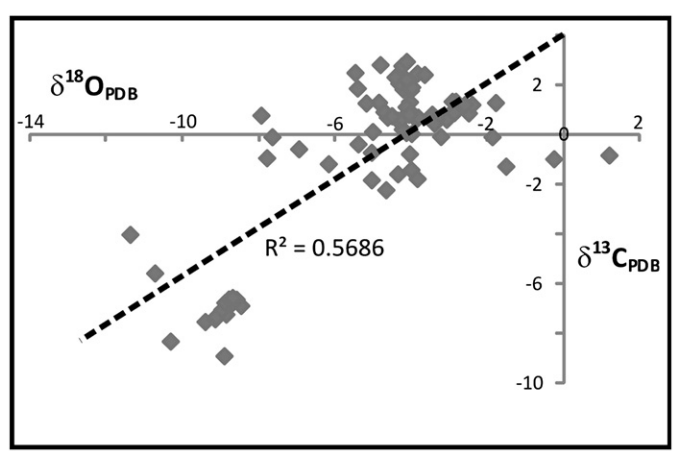


**Fig. 4.** Crossplot of $\delta^{18}O_{PDB}$ versus $\delta^{13}C_{PDB}$ along with the best fit line (R2 = 0.5686) through the data showing no significant covariance of the two signals.

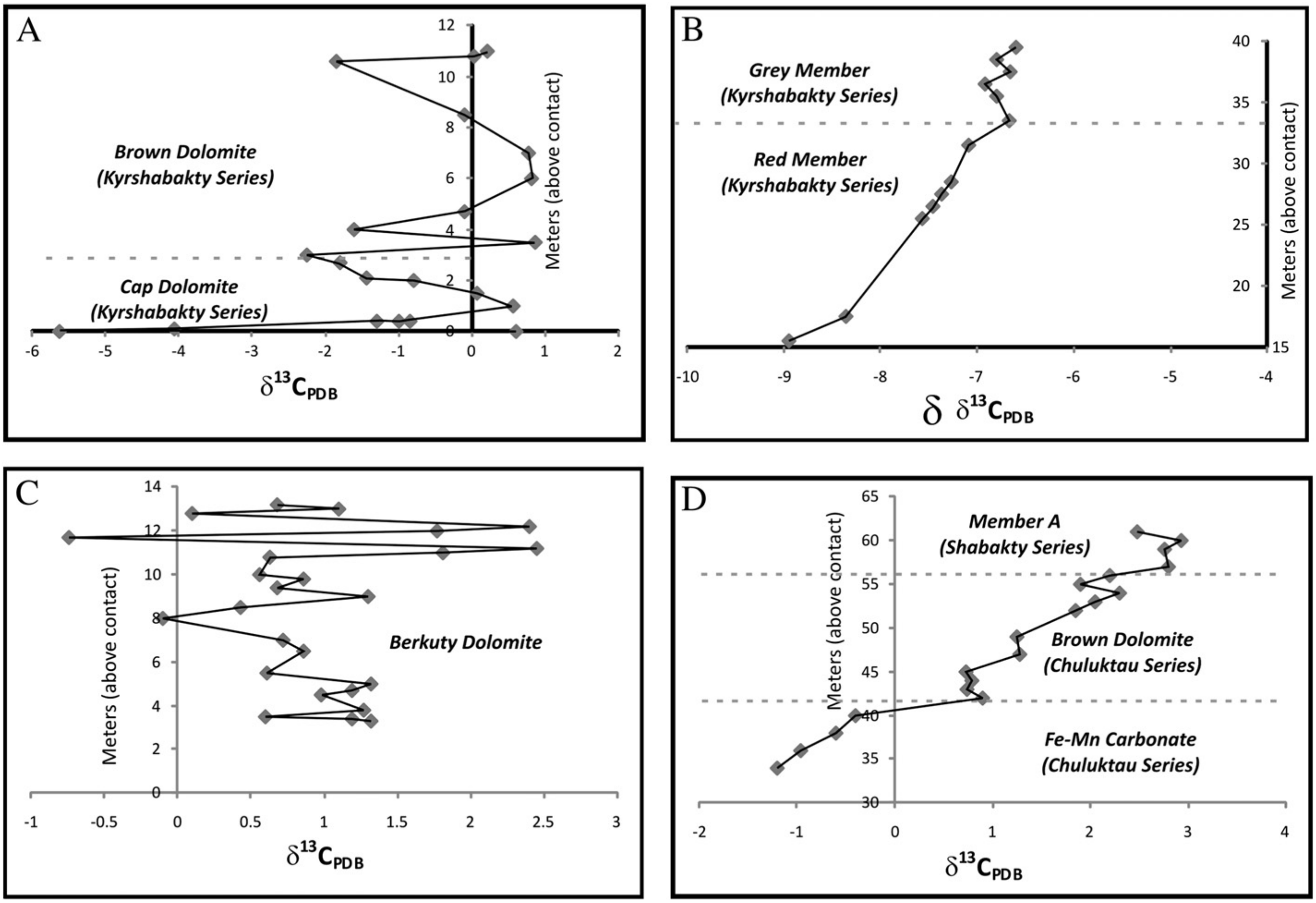


**Fig. 5.** Carbonate profiles versus stratigraphic level (A) $\delta^{13}C_{PDB}$ profile above the Aktas tillite (B) $\delta^{13}C_{PDB}$ profile in the Kyrshabakty Series red and grey members above the Aktas tillite showing very large negative values corresponding to the Shuram/Wonoka anomaly. (C) $\delta^{13}C_{PDB}$ in the Berkuty dolomite and (D) $\delta^{13}C_{PDB}$ in the Chuluktau Series and the lower part of the Shabakty Series.

(<Gaskiers age) that he called Baykonurian. The Baykonurian glaciations in Kazakhstan and Kyrgyzstan are further correlated with tillites in the Luoquan Formation (North China), the Hankalchough Formation (Tarim), Zabit Formation (East Sayan) and the Pourpree de l'Ahnet Group (NW Africa). Chumakov (2010) assigns a Late Vendian-Nemakit-Daldyn age to this glaciation (550–540 Ma). Age assignments and correlations between these glacial units are tentative. Most of the glacial deposits lie above volcanic/igneous 'basement' dated between 815 and 750 Ma as do the tillites in our study. Direct geochronologic dating of the glacial rocks in Kazakhstan and Kyrgyzstan is lacking and so Chumakov (2009, 2010) uses the presence of small-shelly fossils above the tillite as indicating deposition very late in the Ediacaran to syn-Nemakit-Daldyn (550–540 Ma). Other than our work in Maly Karatau, no stable isotopic studies have been completed on any of the glacial sequences so we view the Baykonurian glacials as constrained to only between ~750 Ma and 542 Ma.

We tentatively argue for a Late Cryogenian age for the Aktas tillite corresponding to the Nantuo (S. China), Ghaub (Namibia) and Marinoan (Australia) glaciations. We base this on the similarity of our negative excursion in the cap carbonate above the Aktas tillite and the very negative $\delta^{13}C$ anomaly in the Red-Grey member of the Kyrshybakty Suite that shows a monotonic increase towards the top of the section. We feel that this negative trend is similar in magnitude and character to the Shuram–Wonoka anomaly as described in Le Guerroué and Cozzi (2010). This would also suggest that the other glacial deposits in Kazakhstan and Kyrgyzstan are of Late Cryogenian in age. If we are correct, then the Kazakhstan and Kyrgyzstan glacials (Chumakov, 2009) correlate to the Nantuo–Marinoan–Ghaub glaciations. Furthermore, the Aktas tillite may be synchronous with glaciations recently described on other microcontinental blocks in Central Asia such as the Khongoryn diamictite (Mongolia, MacDonald et al., 2010).

$\delta^{13}C$ values in the Chuluktau Suite (Berkuty Dolomite layer) are mostly positive with values ranging from 2.45‰ to −0.74‰ (Figs. 2 and 5C). Samples were not collected from the phosphorite layers; however a few samples were collected from the top of a ferro-manganese carbonate layer and the overlying brown dolomite. These showed a progression from slightly negative values −1.2‰ in the Fe-Mn carbonate to +2.3‰ in the brown dolomite (Figs. 2 and 5D). Four samples from the lowermost Shabakty Suite exhibited a very narrow range of $\delta^{13}C$ from +2.93‰ to +2.48‰ (Figs. 2 and 5D).

## 4. Review of age constraints on the Kurgan Formation

Samples K2006-2 and K2006-4 were collected for U–Pb geochronology from a rhyolite tuff sequence within the Kurgan Formation and the ages and methods are given in Levashova et al. (2011). K2006-2 is a reworked tuffaceous sandstone that yielded two distinct zircon populations. The first population of grains yielded U–Pb and Pb–Pb ages between ~1950 and 2828 Ma with a distinct clustering at 2032 ± 14.0 Ma (Fig. 6A). A second population of larger, euhedral to subhedral-shaped, zircons from K-2006-2 yield a concordia age of 831 ± 15 Ma (2σ, Fig. 6B). Sample K2006-4, collected from near the top portion of the tuffaceous sequence, yielded a single population of small (~100–200 μm) euhedral zircons with a concordant age of 766 ± 7 Ma (2σ, Fig. 6C). The

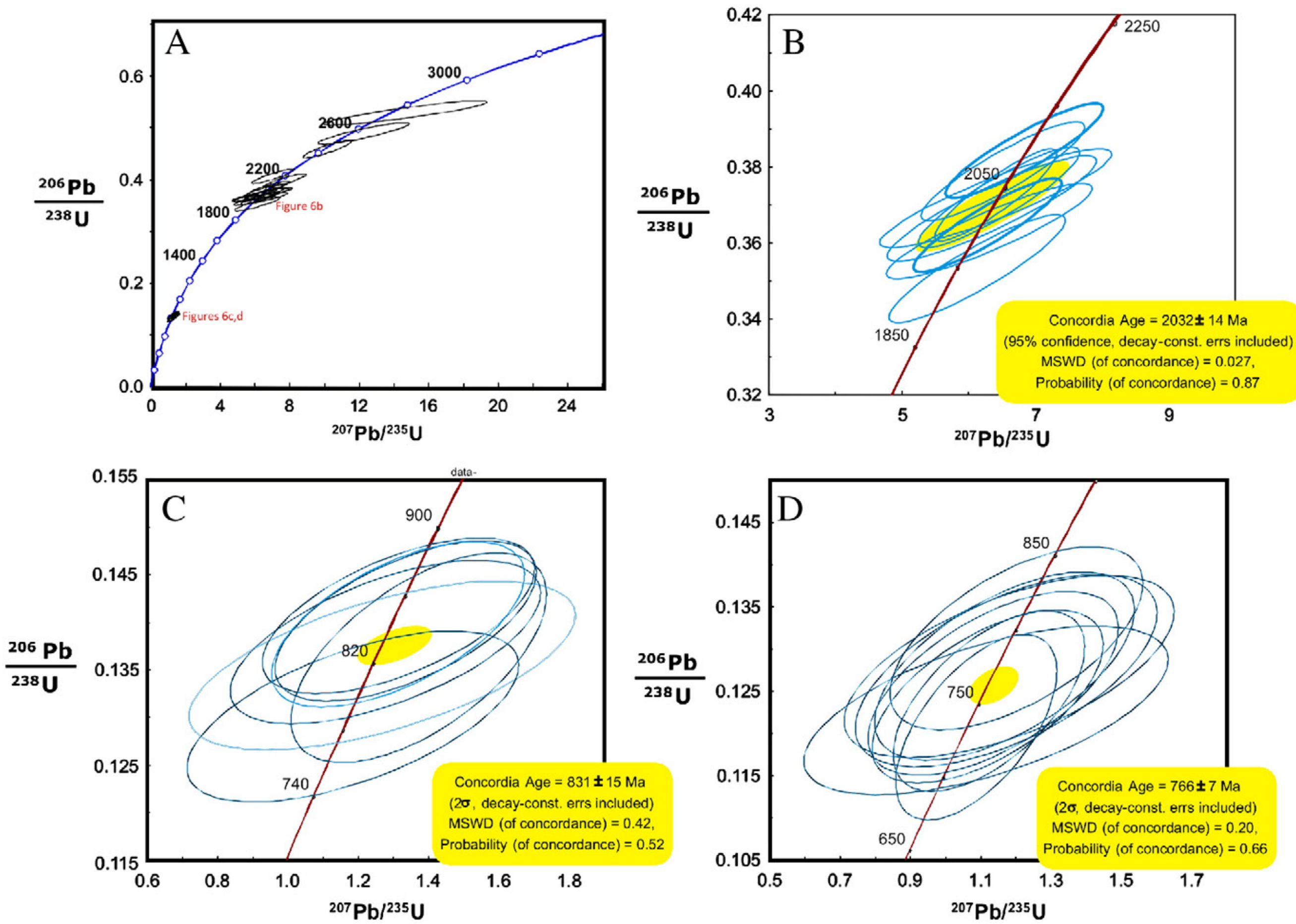


**Fig. 6.** (A) Concordia diagram for all measurements from samples K2006-2 and K2006-4. (B)Concordia diagram for inherited zircons in sample K2006-2 with an age of 2032 ± 14 Ma (2σ) (C) Concordia diagram for the lowermost rhyolitic tuff yielded an age of 831 ± 15 Ma (2σ) (D) Concordia diagram for the uppermost tuff (K2006-4) yielded an age of 766 ± 7.0 Ma (2σ). Note error ellipses for individual zircons are shown at the 1σ level.

Neoproterozoic U–Pb age of euhedral zircons from K2006-4 is interpreted to be a crystallization age for rhyolitic tuffs in the Kurgan Formation. Sovetov (2008) also report an age from the same formation of 771 ± 17 Ma. The U–Pb ages thus constrain the Kurgan fossils to be older than 750 Ma. The Paleoproterozoic–Archean zircons from K2006-2 are inherited from Lesser Karatau basement rocks. Our geochronologic results for the Kurgan Formation provide important constraints on the age of the fossils discovered in the Malokoroi Series at Lesser Karatau. The stromatolite fossil *Linella avis* (Krylov, 1967; Grey and Blake, 1999) was previously described from the Chichkan Formation (stratigraphically below the Kurgan formation; Fig. 2). *Linella avis* is part of the larger *Acaciella australica* assemblage that is considered, based on a detailed analysis of Australian strata, to have developed between ~850 and 800 Ma (Grey and Blake, 1999). Sergeev and Schopf (2010) report a rich microfossil assemblage from cherts within the Chichkan Formation including diverse microscopic eukaryotes, vase-shaped testate amoeba, spiny phytoplanktonic unicells, megasphaeromorphic acritarchs and sausage shaped blue-green algae. In addition, Sergeev and Schopf (2010) describe *Conophyton gaubitza, Linella avis and Potomic ossica* stromatolites from the Chichkan Formation. Based on the older 831 Ma ages from the Kurgan Formation and its conformable relationship with the underlying Chichkan, we assign an age of ~850 Ma to these fossils.

## 5. Ediacara (?) fossils

We report the discovery of the discoidal fossils strongly resembling *Nimbia occlusa* and invaginate type morphs of *Aspidella terranovica* (Billings, 1872; Fedonkin, 1980; Gehling et al., 2000) in siliclastic units within the Kurgan Formation below the Aktas tillite (Figs. 2 and 7A–C). These siliclastic beds contain 'elephant-skin' wrinkle structures commonly seen in other Ediacaran (and older) fossil sites (Gehling, 1999; Gerdes et al., 2000). The wrinkle structures are thought to represent trace remains of microbial mats into which Ediacara organisms could attach their holdfasts. This oxygen-poor environment also enhanced preservation potential for Ediacara fossils (Gehling, 1999; Hagadorn and Bottjer, 1997; Fig. 7B). Previous *Nimbia* discoveries are reported from Ediacaran localities such as the Krol-Tal Belt-India, West Africa, the Flinders Ranges-Australia, the Great Basin-W. U.S., the Digermul Peninsula (Norway), the Avalonian blocks of Newfoundland and the White Sea Region-Russia (Fig. 8A–E; Crimes et al., 1995; Crimes and McIlroy, 1999; Hofmann et al., 1990; Gehling et al., 2000; Shanker et al., 2004; Fedonkin, 1980; Bertrand-Sarfati et al., 1995). The oldest previously reported occurrence of *Nimbia occlusa* was observed in the Twitya formation beneath the Marinoan tillites in NW Canada and the youngest documented occurrence is from Upper Cambrian strata in Ireland (Crimes et al., 1995; Crimes and McIlroy, 1999; Hofmann et al., 1990).

Our specimens were discovered in a brown-red shale within the largely siliclastic Kurgan Formation (Fig. 7C) and are preserved in both hypo and epi-relief. They are circular to oval-shaped impressions that surround either a smooth interior (in positive relief, Fig. 9A, B and F) or circular to oval indentations with a smooth interior (in negative relief; Fig. 9C and D). The long axis of the oval ranges from 3 mm up to 25 mm, the ratio of long/short axis from 1.2 to 1.6 and relief of the raised oval ranges from 0.05 to 1 mm. These are similar in size and shape to Nimbia fossils described in younger strata (Hofmann et al.,

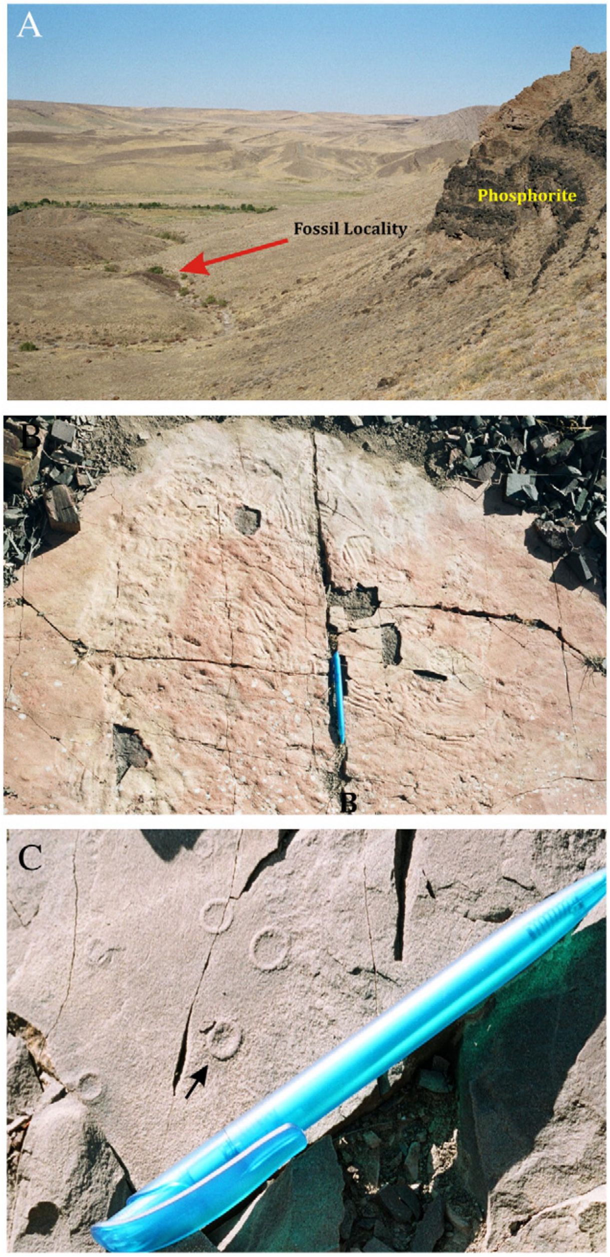


**Fig. 7.** (A) Panoramic view of the fossil locality at Lesser Karatau. The main phosphorite layer in the Chuluktau Series can be seen in the right part of the photo. (B) Bedding surface of the fossil site in the Kurgan Formation showing 'elephant-wrinkle' structures that are similar to moder-day microbial mat structures. (C) Close-up of the bedding surface with oval impressions of *Nimbia occlusa* and *Aspidella terranovica* (invaginate morph-small arrow) fossils all in positive relief.

1990; Hagadorn and Waggoner, 2000; Fedonkin, 1980). In many previously reported discoveries, *Nimbia* impressions also contain a slightly raised central tubercle, but this was not clearly observed in any of the specimens we recovered. We used imaging software to 'invert' the negative relief image in Fig. 9D and, although faint, there is a hint of a raised central tubercle (see white arrow Fig. 9D). Because our fossils are derived from a rock sequence that includes numerous volcaniclastic facies we also note that none of the specimens in our collection show any vertical deformation in cross-section that might indicate gas or fluid escape structures (Cloud, 1960; see Fig. 9E and F).

Fig. 9A and C show discoidal impressions with a central invagination similar in appearance to 'junior morphs' of the fossil *Aspidella terranovica* (Fig. 8F; Gehling et al., 2000; Billings, 1872). Varieties of *Aspidella* junior morphs are also known from deposits in the Podolia (Ukraine), the Ural Mountains, NW Canada, the Avalon Peninsula, Newfoundland and Namibia (compilation in Gehling et al., 2000). Based on a detailed analysis of Ediacara fossil occurrences in Newfoundland, Gehling et al. (2000) concluded that many discoidal Ediacara fossils may represent a variety of stages in the life of *Aspidella terranovica.* The 'invagination' type morph can be observed in Fig. 9A and c. Strong arguments are forwarded for a biological origin of *Aspidella terranovica* and traditionally, *Aspidella terranovica* and its multiple morphs have been placed in the Phylum Cnidaria (Fedonkin, 1980; Hofmann et al., 2008). A detailed examination of *Aspidella terranovica* morphotypes suggest that they are basal impressions (holdfasts) of collapsible bulb or frondose-shaped organisms (Gehling et al., 2000). Fig. 10A and B shows a possible stem extension emanating from the discoid fossil (in negative relief-10a and schematically-10b) from one of our samples.

## 6. Paleogeographic setting of the Kurgan Formation

Levashova et al. (2010, 2011) discuss the Neoproterozoic paleogeographic setting of microcontinents that now comprise large tracts of the Eurasian continent. Paleomagnetic studies were conducted on the Kurgan Formation at Lesser Karatau and the Cambrian–Ordovician carbonates of the Tamdy Series. Levashova et al. (2011) note a strong overprint of probable Late Paleozoic age in the Kurgan rocks, but also a pre-folding magnetization that yielded an average inclination of 54° corresponding to a paleolatitude of 34° for the Kurgan formation.

The younger section of the Lesser Karatau sequence (e.g. Cambrian and Ordovician Tamdy Series) consists of a thick sequence of limestones and dolostones. Very limited paleomagnetic data from these units indicates tropical latitudes (Pradhan, personal communication). A low-latitude for the central Asian microcontinents during the Late Ediacaran to Cambrian was also suggested by Kravchinsky et al. (2001, 2010) Eganov et al. (1986) suggest a shallow water, passive-type margin, depositional setting for the Tamdy Series.

The Lesser Karatau microcontinent may have been part of a larger domain and Levashova et al. (2011) note the similarity in ages and lithologies between the Baydaric microcontinent (Mongolia, Fig. 1) and Lesser Karatau. They suggest that these microcontinental domains may have been in proximity to the Tarim and South China cratons during the Neoproterozoic (~750–800 Ma) and closer to the Siberian craton during Ediacaran–Cambrian time as part of an island archipelago known as "Paleo Polynesia". Popov et al. (2009) place Karatau (along with most other Kazakh terranes) at low latitudes in the vicinity of South China and Tarim during the mid-Ordovician. Chumakov (2010) notes the stratigraphic and faunal similarities between many of the Kazakhstan and Kyrgyzstan microcontinental domains and suggest that they may have been in close proximity during the Late Neoproterozoic and Cambrian time.

## 7. Conclusions

We confirm previous suggestions that the basal Kyrshabakty Suite from the Lesser (Maly) Karatau microcontinent in Kazakhstan contains a glacially derived tillite. The Aktas tillite lies well above discoidal, Ediacara-type (?), fossils in the ~770 Ma Kurgan Formation. At other Ediacara fossil sites throughout the world, there is a common association of *Nimbia occlusa* with well described Ediacara fauna such as *Cyclomedusa sp*., *Ediacaria*, *Tribachidium*, *Eoporita*, *Belanella sp*. and *Spriggina. Aspidella terranovica* is known to occur along with *Hiemalora*, *Triforilloni costellae Charnia, Charnodiscus* and *Blackbrookia* (McCall, 2006; Gehling et al., 2000; Hofmann et al., 1990; Bertrand-Sarfati et al., 1995; Narbonne and Aitken, 1995).

Whatever the exact classification of these discoidal fossils, our finding suggests that the range of both *Nimbia occlusa* and *Aspidella*

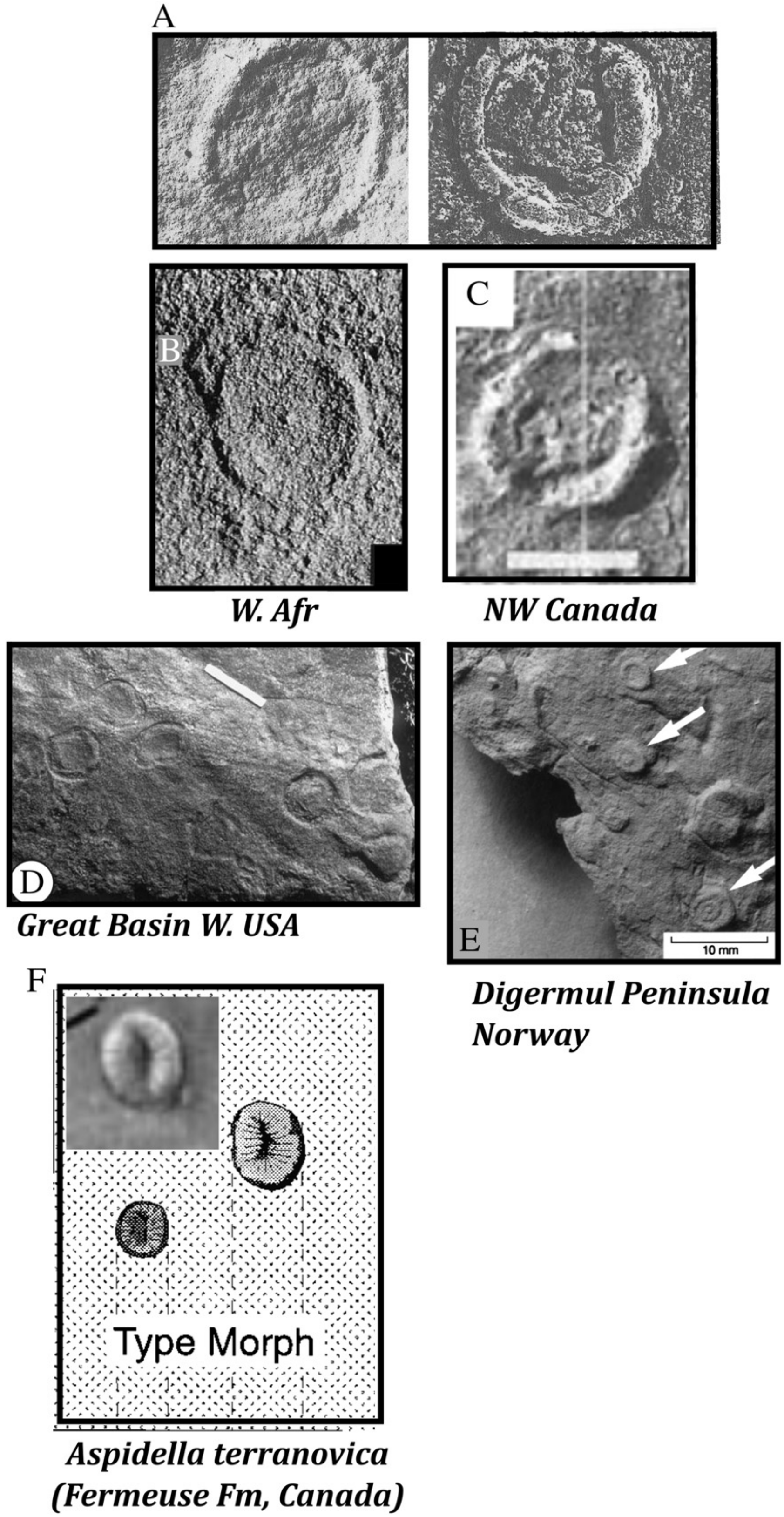


**Fig. 8.** Examples of *Nimbia* fossils and *Aspidella terranovica* from around the globe. (A) Type fossils from the White Sea (Fedonkin, 1981); (B) *Nimbia occlusa* from the Taoudenni Basin, West Africa. *Nimbia* fossils are found in occurence with *Medusinites sp.* (Ediacaran ~580 Ma; Bertrand-Sarfati et al., 1995); (C) *Nimbia occlusa* impression from the Twitya Formation, northwestern Canada.). In this locale, *Nimbia* occurs in Neoproterozoic-age intertillite beds along with *Irridinitus* and *Vendella* (Ediacaran, ~650 Ma Hofmann et al., 1990); (D) *Nimbia* impressions from the Stirling Quartzite in the Great Basin, western United States. *Nimbia* occurs with *Swartpuntia, Archaeichnium, Cloudina, Corumbellaand Onuphionella* (Ediacaran ~545 Ma; Hagadorn and Waggoner, 2000); (E) *Nimbia* fossils from the Digermul Peninsula in Norway. Here Nimbia is found in both Neoproterozoic strata with *Cyclomedusa, Ediacaria, Beltanella, Vendotaenia* and *Hiemalora* and in Cambrian strata with *Tirasiana sp.* (Late Ediacaran–Early Cambrian — 550–540 Ma; Crimes and McIlroy, 1999); (F) *Aspidella terranovica* fossil and sketch from the Fermeuse Formation, Canada (Ediacaran, <565 Ma; Gehling et al., 2000).

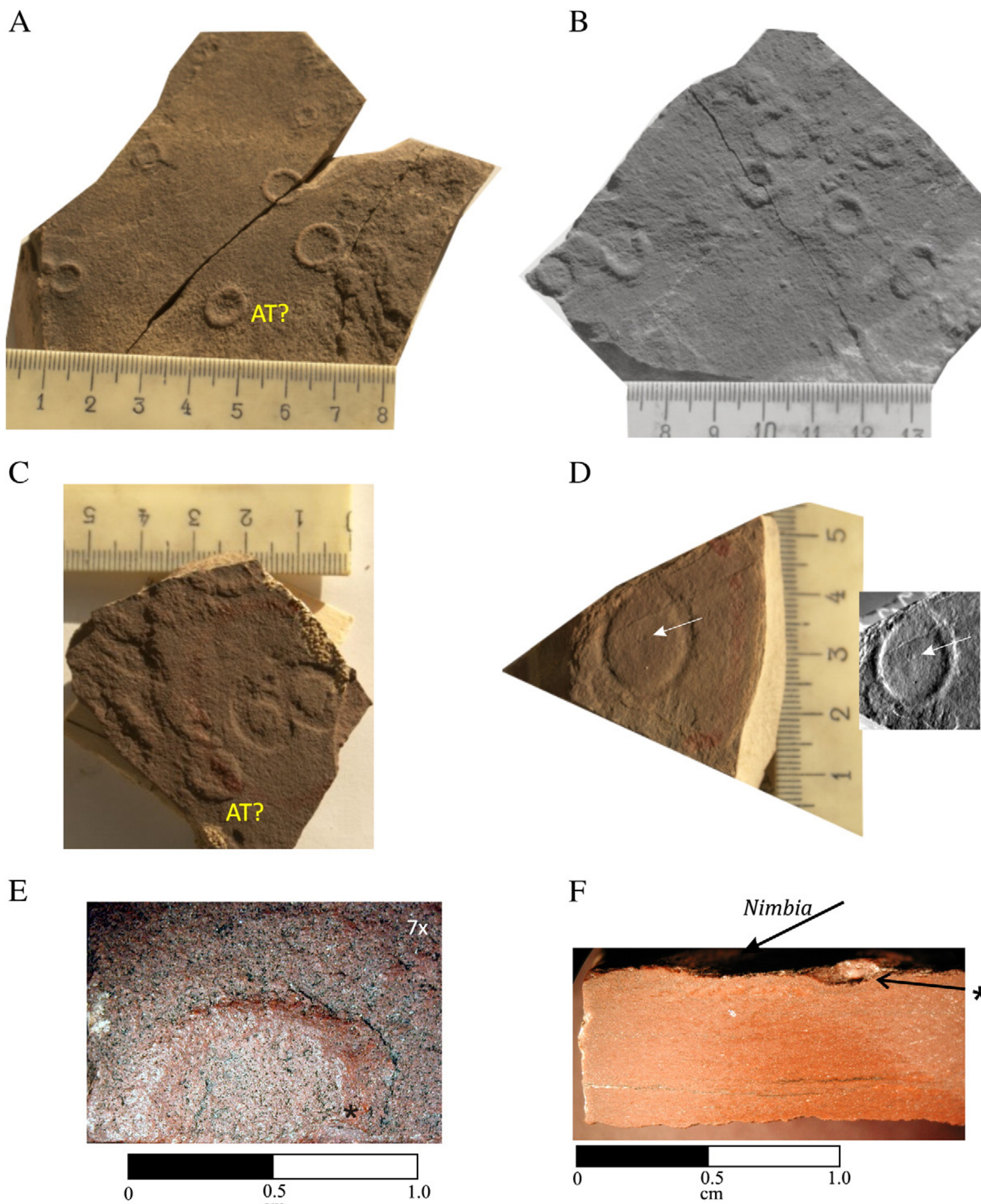


**Fig. 9.** (A) closeup of the fossils in photo (7c) shown with a scale bar (cm). Here *Nimbia* fossils are in positive epi-relief and AT marks the occurrence of an invaginate morph of *Aspidella terranovica*; (B) B&W photo of *Nimbia* impressions shown mostly in positive epi-relief; (C) *Nimbia* fossils in negative relief with an invaginate morph of *Aspidella terranovica* (?) in positive relief; (D) large ~2 cm (long axis) × ~1.5 cm (short axis) negative hypo-relief impression of *Nimbia occlusa* and an 'inversion' of the photo in 9d as the fossil might appear in positive epi-relief. The white arrow points to a possible raised central nodule observed in Nimbia occlusa fossils elsewhere (E) 7× magnification of a small *Nimbia* disk cross-sectioned for Fig. 9(F). The * marks the location of the rim visible in Fig. 9F.; (F) cross-section through *Nimbia occlusa* showing no deformation (planar laminae) beneath the fossil. Arrow points to the center depression of the fossil and * marks the location of one of the raised rims of the fossil.

*terranovica* should extend deeper into the Cryogenian and prior to the "Sturtian" (~716 Ma) glaciations.

In the grand history of life on Earth, the fossil record contains only a muted account of the transitions between prokaryotic to eukaryotic to metazoan life (Fedonkin, 2003; Knoll et al., 2006; Budd, 2008). The age of the oldest eukaryotic organisms is debated, but recent estimates place the eukaryotic stem group in the Paleo-Mesoproterozoic and a subsequent Meso-Neoproterozoic divergence of the eukaryotes (Knoll et al., 2006). Evidence for eukaryotic fossils at Lesser Karatau are described in detail in Sergeev and Schopf (2010). They note that the underlying Chichkan Formation biota may be linked to meiosis-based sexuality in eukaryotic organisms thought to have started around

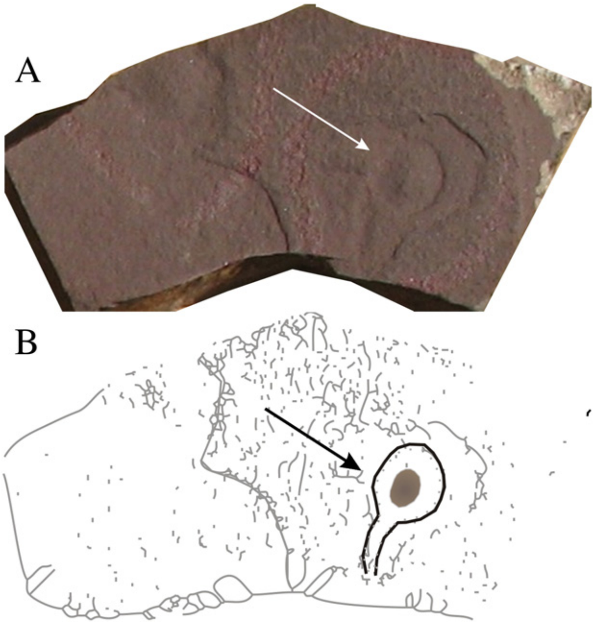


**Fig. 10.** (A) Discoidal impression (negative relief) with possible stem-like extension. (B) Sketch showing the location of discoidal fossil and possible stem.

1000 Ma. In their view, the Chichkan fossils provide a useful link between Early Cryogenian evolutionary patterns and the advent of multi-cellular organisms in the Ediacaran.

Numerous explanations for the post-Snowball arrival of metazoans have been put forth in the literature (Meert and Lieberman, 2008). Nearly all of the suggested triggers for metazoan evolution are presumed to precede their appearance in the fossil record by only a few million years to at most 20–30 million years. Molecular clock analyses have yielded various estimates for the divergence of the metazoans, but the more recent analyses have also placed this divergence near the end of the Snowball Earth episodes (Peterson et al., 2005).

The fact that our fossils are identical to those observed some 100–150 Ma later also suggests that the organisms that they represent passed though the myriad tectonic, climatic, oceanic and atmospheric events of the Neoproterozoic and Early Cambrian without significant change in bauplan (Crimes and McIlroy, 1999; Narbonne and Aitken, 1995). If these fossils represent metazoan remains, then the impact of the multiple glaciations may have been muted somewhat for these particular taxa.

Although a few Ediacara and other purported metazoan fossils have been found below the Marinoan glaciation (Maloof et al., 2010; Hofmann et al., 1990), we report the observation of discoidal fossils in strata that precede the oldest of the global "Snowball" Neoproterozoic glacial sequences (e.g. Sturtian at 716 Ma, MacDonald et al., 2010) including *Nimbia occlusa* and morphotypes of the body fossil *Aspidella terranovica*. Other discoveries of metazoan fossils (or large trace fossils) predating the Ediacaran Period are controversial because the relationships between the strata, the fossils and the methods use to date the fossils are not straightforward (Malone et al., 2008; Seilacher et al., 1998; Bengtson et al., 2007; De, 2006; Bengtson et al., 2007; Bengtson and Rasmussen, 2009).

Earliest Neoproterozoic to Late Mesoproterozoic occurrences of Ediacara or "Ediacara-like" organisms are reported from the Purana basins of India (Malone et al., 2008; Seilacher et al., 1998; Bengtson et al., 2009; De, 2006), but because of the >1000 Ma ages for the strata in these basins, the finds are viewed with skepticism (Malone et al., 2008). Even older occurrences of metazoan life (2.0–1.8 Ga) were reported from the Stirling Range in Australia (Bengtson et al., 2007) although these are reinterpreted as traces made by large eukaryotic organisms (Bengtson and Rasmussen, 2009). We understand that the exact placement of the Ediacara-type organisms in phylogenetic space is both problematic and controversial (Gehling, 1991; Grazhdankin and Gerdes, 2007; Retallack, 1994; Peterson et al., 2003; MacGabhann, 2007; Hofmann et al., 2008). It is possible (even probable?) that the discoidal fossils described in this study represent something other than metazoa and our >766 Ma age will lend support to those who wish to argue for a bacterial, lichen, Vendobiont or non-biogenic origin for the impressions (Grazhdankin and Gerdes, 2007; Retallack, 1994; Peterson et al., 2003; MacGabhann, 2007).

It is axiomatic to say that additional work into the taxonomic position of *Nimbia occlusa* and *Aspidella terranovica* are needed; however, there is no doubt as to the Early Cryogenian age of our discovery. Furthermore, our findings, coupled with recent research indicating a richer Meso-Neoproterozoic metazoan biota, may indicate that the roots of metazoan life extend back to an interval of time before the "Sturtian" glaciations or more controversially to a time shortly after the development of eukaryotes. This early evolution of metazoan life was possibly modulated by temporary increases in oxygen (El Albani et al., 2010; Canfield et al., 2007). These recent discoveries should result in an increasingly careful search for older metazoans.

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