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Middle Cambrian to Lower Ordovician faunas from the Chingiz Mountain Range, central Kazakhstan

T.JU. TOLMACHEVA, K.E. DEGTYAREV, J. SAMUELSSON AND L.E. HOLMER

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The middle Cambrian to Lower Ordovician back-arc sedimentary succession studied in the Kol'denen River and in the Zerbkyzyl Mountains of the central Chingiz Mountain Range is composed predominantly of siltstones, sandstones and volcanoclastic rocks with rare beds of micritic carbonates, black shales and cherts. Fossil assemblages including conodonts, lingulate brachiopods, arthropods, sponges and probable *Tasmanites* cysts were recorded both from the carbonate and chert beds showing that richly diverse marine environments existed directly adjacent to the volcanic arcs. The Kol'denen River localities contain a diverse upper Cambrian paraconodont assemblage of the open-sea affinity. The representatives of *Rossodus*, *Cordylodus*, *Drepanodus* and *Variabiloconus*, having an almost pandemic distribution and characteristic of basinal facies, dominate the Lower Ordovician conodont fauna. The Cambrian–Ordovician boundary transition is characterized by chert production that was more likely caused by a local productivity increase than by general changes in palaeoceanographic and palaeogeographical conditions.

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Key words: biostratigraphy, conodonts, siliceous deposits, upper Cambrian, Lower Ordovician, central Kazakhstan.

THE LOWER PALAEOZOIC stratigraphy and faunas of the Chingiz Mountain Range in central Kazakhstan are not well known, partly due to the original sedimentary successions having been strongly deformed leaving only rare eroded and remnant exposures. Detailed geological mapping of the area to the southeast of the Saryshokinskaya intrusive body (the Kol'denen River valley) carried out in the early 1970s revealed a tectonically disrupted thick succession of volcanoclastic and siliciclastic terrigenous rocks with sparse carbonates and cherts (Figs 1, 2). The first fossils

described from the region were middle Cambrian trilobites from the basal carbonate layer of the lowermost part of the succession considered to belong to the Chingiztau Formation (Ivshin *et al.* 1972). Regional mapping also placed the overlying >800 m thick succession in the Chingiztau Formation. Isolated siliceous and terrigenous strata with rare limestone beds attributable to the same unit are also exposed 65 km northwest from the Kol'denen River in the Zerbkyzyl Mountains.

The first biostratigraphic data from successions exposed in the Kol'denen River valley were obtained during a field trip in 1991 (Degtyarev *et al.* 1999). Findings of late Cambrian to Early Ordovician

Fig. 1. Geological maps of the Kol'denen River and Zerbkyzyl Mountain areas of the central part of the Chingiz Mountain Range showing the location of the studied sections.

conodonts and Early Ordovician graptolites in the middle part of the sedimentary succession indicated that the deposits represented a longer time span than previously thought (Fig. 2). Rare and poorly preserved Darriwilian trilobites were recovered from the tuffaceous siltstones (Degtyarev *et al.* 1999). A single brachiopod species, *Odontotreta mirabilis* Ushatinskaya,

1998, from the lowermost upper Cambrian part of succession has been illustrated and described (Ushatinskaya 1998). A new assemblage of fossils was collected from this succession during a field trip carried out by the authors in 2003. The field studies aimed to obtain more detailed sedimentological and palaeontological data from the siliceous parts of the succession.

Fig. 2. Generalized geological column of the Lower Palaeozoic sedimentary succession of the Kol'denen River and Zerbkyzyl Mountains areas.

In this paper, we document the upper Cambrian and Lower Ordovician fossil assemblages from the area of the Kol'denen River Basin and the Zerbkyzyl Mountains.

Geological setting and lithology
The lower Palaeozoic sedimentary rocks exposed in the central and northern part

of the Chingiz Mountain Range record part of the widespread and long-lived volcanic arc that bounded the margin of the Dzhyn-garo-Balkhash palaeobasin (Yakubchuk & Degtyarev 1993, Yakubchuk 1997). Sedimentary rocks within the Chingiz Mountain Range are represented mainly by marine volcanoclastic and siliciclastic terrigenous sediments with minor beds of carbonate and chert; the entire succession encompasses the interval from the mid-Cambrian to the Lower Silurian. Devonian sediments are very locally developed in this region and represented mainly by red beds and volcanic rocks of continental affinity.

Tectonically disrupted and intensively deformed volcanoclastic and terrigenous successions of mid-Cambrian to

mid-Ordovician age are located in the central part of the Chingiz Mountain Range (Degtyarev *et al.* 1999; Fig. 1). The succession is approximately 1200 m thick and is represented by four informal lithostratigraphical units that are briefly described in ascending order.

The lower unit, more than 100 m thick, overlies middle Cambrian volcanic rocks with a significant angular unconformity. The lower part of the unit is dominated by conglomerates and sandstones that grade upwards into calcareous siltstones and mudstones with carbonate nodules and thin beds of limestone (Figs 2, 3A). The succession is capped by coarse-grained, light pink, fossiliferous packstones, about 11 m thick yielding numerous middle Cambrian

Fig. 3. Outcrops of the Chingiztau Formation (KD04-2—the lower part) and (KD05) showing the character of limestone-siltstone intercalation (A) and the character of chert bedding (B).

trilobites and lingulate brachiopods. A layer of siltstones and sandstones in turn covers the carbonate beds.

The second unit is about 50–70 m thick. It is composed of thin-bedded siltstones and mudstones with rare layers of grey or black cherts and grey micritic limestones. Carbonate breccias occur sporadically through the succession. In exposures with little tectonic deformation, the chert beds are dark grey to black and macroscopically homogeneous (Fig. 3B). In exposures with stronger tectonic deformation, they are generally light grey to whitish. The quantity of chert beds varies between exposures; however, in all localities they increase in abundance stratigraphically. The chert beds range from a few centimetres to about 10 m thick and grade into the underlying and overlying siliceous siltstone layers without sharp boundaries.

The third unit is up to 500 m thick and composed of sandstones, cherty red and green tuffaceous siltstones, and sparse pyroclastic tuffs. Some strata of this unit demonstrate graded bedding and are clearly turbiditic in origin. The base of the unit is marked by a 10 m thick layer of black shales, which yield rare lingulate brachiopods and graptolite and phyllocarid fragments. A few conodont elements of *Acodus* sp. and *Scandodus* sp. and poorly preserved *Tetragraptus* sp. cf. *T. quadribrachiatus* (Hall 1858) indicating an Early Ordovician age were recorded from this interval (Degtyarev *et al.* 1999). The volcanoclastic upper part of the succession is poorly fossiliferous but a few unidentified trilobites and brachiopods have been collected (Degtyarev *et al.* 1999).

The fourth unit is composed of fine- and coarse-grained sandstones with rare carbonate units in the uppermost part of the succession. The limestones contain poorly preserved echinoderm remains. The unit is potentially referable to the lower Darriwilian on the basis of its stratigraphic position.

Our field studies were focused mainly on fossils and lithology of the first and second units of the succession exposed along the Kol'denen River valley and Zerbkyzyl Mountain area. In the Kol'denen River valley, the strata of both units constitute the eastern and southern limbs of a syncline disrupted by small-amplitude faults. The best exposures of the succession are located along the right bank of the Kopa River, but they are rare and usually not more than 30 m in extent. The basal part of the studied section (outcrop KD04) is composed of red-brown calcareous siltstones and fine-grained sandstones intercalated with thin layers of micritic limestones. The entire thickness of this part of the section is about 20 m. Pinkish-grey bioclastic limestones 15 m thick overlie this section and, in turn, are overlain by 5 m of green calcareous siltstone and sandstone. The upper part of this sequence is well exposed 200 m northward of site KD05 (Fig. 1). Here, the pinkish grey limestones are overlain by a layer of carbonate breccias whose thickness varies significantly within the outcrop. Upwards, the sequence is composed of green calcareous siltstone and fine-grained sandstone with thin (up to 10–15 cm) lenses and layers of grey limestone. The thickness of the latter unit is 25–30 m (Fig. 4).

The higher stratigraphical interval is exposed in the upper part of the outcrop KD04 and at site KD010. The upper part of the section at KD04 consists of green siliceous and calcareous siltstones with thin layers and lenses of micritic limestones and a layer of carbonate breccia. The amount of carbonate decreases upward and completely disappears in the uppermost part of the section. In contrast, the content of siliceous material increases upward, and the uppermost part of the section is composed of dark grey siliceous siltstones with chert beds (Fig. 4). Similarly, the upper part of the section at KD010 is a succession of siliceous

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siltstones with gradual increase in silica content and is capped by a 20 m thick layer of black chert (Figs 3A, 4).

The same stratigraphic interval exposed on the left bank of the Kopa River is represented by more lightly coloured successions of siliceous siltstones and cherts. The uppermost part of the section is composed of light grey and green-banded cherts. They are overlain by cherry-coloured tuffaceous siltstones of the third unit of the succession. Similar strata of siliceous siltstones and cherts more than 80 m thick are exposed on the left bank of the Tomarash River (outcrop KD07) 5 km west of the Kopa River. However, this section differs from the successions of the Kopa River localities by the lack of carbonate lenses and beds (Fig. 4).

Fragments of the same succession are exposed in the Zerbkyzyl Mountains (Figs 1, 4) where outcrops in the southwestern limb and northwestern closure of a syncline are intensively disrupted by northwest-trending thrusts and strike-slip faults with extensive cleavage developed along the fault zones. Two well-exposed sections in the Zerbkyzyl Mountains area were described and sampled (Fig. 4). One of these (ZK02) incorporates a 25 m thick intercalation of grey-greenish siltstone, fine sandstone and thin beds of grey sandy limestones. Upward, above a small unexposed part the succession, black cherts and siliceous sandstones occur with rare carbonate lenses and beds. A 1 m thick bed of grey limestone occurs in the uppermost part of the section. The entire succession is more than 45 m thick. The second section in the Zerbkyzyl Mountains is located 1.5 km to the northwest (outcrop ZK01; Fig. 4). The exposed succession is 70 m thick and is composed of interbedded grey cherts and siltstone with rare lenses and thin layers of grey micritic limestones, which disappear in the upper part of the section.

Material

The majority of sites expose rocks that are slightly cleaved in at least one direction along the bedding surfaces. The terrigenous rocks are mostly dark to light grey but become greenish in the upper tuffaceous part of the succession. Micritic and dense carbonates are medium to dark grey. Cherts are mostly dark grey to black and are clearly interstratified with siliceous siltstones. Petrographic and textural observation of rocks showed that they have not suffered severe deformation and are free of veins.

Microscopic examination of cherts reveals a generally homogenous fabric of microcrystalline quartz with diffuse pyrite crystals. The most abundant fossils found in thin-sections are sponge spicules, which occur in various states of preservation. In the lowermost siliceous beds they are usually pyritized, whereas in the samples from the Zerbkyzyl Mountain area they are preserved as moulds filled by quartz. Broken and deformed monaxons dominate the sponge spicule suite. Radiolarians remains are absent. Limestones are generally poor in bioclasts with variants grading into tuffaceous limestone with abundant volcanic quartz and others characterized by abundant terrigenous material.

Twenty-five samples of limestones and cherts were collected from the studied succession. Samples were cut parallel to bedding planes and then sectioned. The sections with conodonts were mounted on glass slides and ground on a rotary diamond lap until transmitted light penetrated through the chert. In translucent cherts, the completed thin-sections were approximately 0.3–0.5 mm thick. Many chert layers are strongly opaque due to tiny terrigenous inclusions; fossils were found only in several thin layers that are relatively translucent.

Carbonate samples, 1 kg each, were processed by the traditional method of

acetic acid digestion. Both carbonates and cherts are relatively poor in faunal remains. The samples were collected from almost all limestone layers of the studied interval and most proved to be barren of fossils. A few samples contained rare conodont elements. Only one sample yielded sufficient conodonts for precise identification.

Conodont biostratigraphy

Five samples from the limestone layers and seven samples from cherts produced conodonts (Fig. 4). Preservation of conodont elements in cherts is generally poor; their primary matter is damaged or replaced by iron oxides. Specimens showing stripes and cleavage are common. The conodonts from carbonates have recrystallized surfaces; their colour alteration index (CAI) varies between 4 and 4.5 indicating that the host rock reached at least 300°C.

Conodont identification is seriously hampered by the very sparse elements in the samples and by generally poor knowledge of the faunas and conodont biostratigraphy of the lower Palaeozoic in Kazakhstan. Although Cambrian and Ordovician conodonts were first recorded in Kazakhstan almost 30 years ago (Gridina & Mashkova 1977), few studies have illustrated and described the faunas (Abaimova 1978, Appolonov *et al.* 1984, Gridina 1991, Zhylkaidarov 1998, Dubinina 2000). Most papers only list identified conodonts; moreover some papers were published in Russian journals with a limited circulation (Dvoichenko & Abaimova 1987, Dubinina 1991, 1998).

The stratigraphically lowermost sample KD05 from the Kol'denen River area yields a poor conodont assemblage including the following taxa: *Phakelodus tenuis* (Müller 1959), *Ph. elongatus* (An *et al.* 1983), *Prooneotodus gallatini* (Müller 1959), *Furnishina primitiva* Müller, 1959, *Coelocerosodontus* sp. and several conodont elements

that can be identified only provisionally (Fig. 5A–M). The assemblage cannot be precisely dated, as it consists mainly of coniform paraconodont species with long stratigraphic ranges. Eoconodont species that can be confidently identified were not found in the sample. Two conodont elements that are provisionally assigned to *Proconodontus? savitzkyi* Abaimova, 1978 possibly represent a new paraconodont species.

Two chert samples collected higher in the section (KD04-2 and KD010) yielded a similarly diverse paraconodont fauna (Figs 4, 6A–N) that additionally includes '*Barnesodus*' *gibber* Dubinina, 2000 and *Furnishina* sp. A. Some paraconodonts are impossible to identify precisely, as identification is inhibited by the poor preservation of conodont elements in the cherts. Elements are optimally visible from their lateral sides only, whereas shapes of the cross-section of the basal cavity are impossible to see in most cases. Conodont assemblages from these two samples are virtually monospecific, containing abundant *Ph. tenuis*, with a subordinate number of other species. The most probable age for this stratigraphic level ranges from the *Westergaardodina amplicava* Zone to the *Hirsutodontus ani* Zone of the late Cambrian. This age is close to previously inferred ages from these beds (Degtyarev *et al.* 1999).

The occurrence of Early Ordovician conodonts has been reported previously from the Kol'denen River area (Degtyarev *et al.* 1999). The list of identified taxa indicates that the assemblage is not older than the *Paroistodus proteus* Zone. However, all the new samples that we collected from that interval appeared to be barren of conodonts.

The chert layers exposed near the Tomarash River valley (KD07) yielded a few conodont elements of *Paltodus* sp., *Drepanoistodus* sp. and *Prooneotodus* sp., suggesting a Tremadocian age (Fig. 6K, M, N).

Fig. 5. Scanning electron micrographs of Late Cambrian conodonts and sponge spicules from carbonate beds of the Kol'denen River sections. A, B, *Coelocerodontus* sp.: A, Morphotype 'latus', CM 5/13174, sample KD05, $\times 96$; B, Morphotype 'bicosatus', CM 6/13174, sample KD05, $\times 75$. C, K, L, *Furnishina primitiva* Müller, 1959: C, CM 15/13174, sample KD05, $\times 70$; K, Deformed specimen, CM 16/13174, sample KD05, $\times 75$; L, CM 17/13174, sample KD05, $\times 77$. D, *Prooneotodus gallatini* (Müller, 1959), CM 31/13174, sample KD05, $\times 105$. E, *Prooneotodus* sp. cf. *P. gallatini* (Müller, 1959), CM 32/13174, sample KD05, $\times 125$. F, *Coelocerodontus*? sp., CM 11/13174, sample KD05, $\times 130$. G, *Phakelodus* sp., cluster, CM 27/13174, sample KD05, $\times 93$. H, *Phakelodus tenuis* (Müller, 1959), CM 25/13174, sample KD05, $\times 116$. I, *Proconodontus? savitzkyi* Abaimova, 1978, CM 29/13174, sample KD05, $\times 110$. J, *Prosagittodontus* sp., CM 36a/13174, sample KD05, $\times 101$. M, Sponge spicule, CM 45/13174, sample KD05, $\times 56$.

In the Zerbkyzyl Mountain area, conodonts were obtained from a few samples of carbonates interbedded within the cherty succession and from three chert layers (Fig. 4). Stratigraphically, the lowermost cherty sample ZK01-4 contains *Phakelodus tenuis*, *Ph. elongatus*, *Coelocerodontus bicosatus* van Wamel, 1974 and *Variabiloconus* sp. (Fig. 4). The most abundant (40

conodont elements) assemblage is late Tremadocian in age yielding *Cordylodus* sp. cf. *C. angulatus* Pander, 1856, *C. bicosatus*, *Ph. tenuis*, *Ph. elongatus*, *Rossodus manitouensis* Repetski & Ethington, 1983, *Variabiloconus* sp. cf. *V. bassleri* (Furnish, 1938) and *Polycostatus* sp. This sample was recovered from the limestone bed in the middle part of the section (sample ZK01-2; Figs 4, 7A–O).

The presence of *R. manitouensis* indicates that this stratigraphic level is not older than the *R. manitouensis* Zone. One element of *Cordylodus* is very small and too poorly

preserved for confident specific identification, but the concave anterior margin of the basal cavity in lateral view indicates that it may be assigned to *C. angulatus*.

The uppermost chert layer in the section ZK01 (ZK01-11) yielded 25 conodont elements including *Prioniodus* sp. cf. *P. oepiki* (McTavish, 1973; Fig. 8A, B, F), *Drepanodus arcuatus* Pander, 1856 (Fig. 8E, J, K), *Drepanoistodus* sp. (Fig. 8H), *Variabiloconus* sp. (Fig. 8I), *Cornuodus* sp. (Fig. 8L) and several paraconodont elements: *Ph. tenuis*, *C. bicostatus* (Fig. 8C, G) and *Hertzina?* sp. (Fig. 8D). The occurrence of paraconodont species in the Tremadocian sediments may indicate that the assemblage contains reworked late Cambrian conodonts. However, it is possible that paraconodont species had a longer stratigraphic distribution in pelagic open-oceanic facies.

Section ZK02 is less rich in conodonts; only a few specimens of *Ph. tenuis*, *C. bicostatus* and *Variabiloconus* sp. were obtained from three carbonate layers. The cherts of section ZK02 are more opaque relative to those of the section ZK01 due to the presence of fine terrigenous material and numerous sponge spicules (Fig. 8M).

Palaeogeographic implications of the Chingiz Range conodont faunas

Given that there are only a few records of the Cambrian and Ordovician conodonts in Kazakhstan, the pattern of their palaeobiogeographic distribution in the region is poorly known. We can compare three Cambrian occurrences: one from

the Batyrbai section of Malyi Karatay in south Kazakhstan (Dubinina 2000), and two isolated occurrences in the Selety River valley (Gridina 1991) and in the Agyrek Mountains (Tchelinograd region) in the northern part of central Kazakhstan (Abaimova 1978). The common constituents of those conodont assemblages are *Ph. tenuis* and representatives of *Furnishina* and *Westergaardodina* that have an almost cosmopolitan distribution. The rest of the assemblages are represented by endemic species that were only recorded in one or two localities in Kazakhstan.

Lowermost Ordovician conodonts were well documented only from the Malyi Karatay Batyrbai section (Dubinina 2000). This fauna is characterized by a mixture of endemic species, species typical of warm and shallow water environments, and species with a wide geographic distribution that commonly occur in slope and basin environments. Although a single known occurrence of Early Ordovician conodonts in Malyi Karatay obviously cannot form a basis for a thorough understanding of the biogeographic setting of Kazakhstan, the assemblage is best referable to the palaeobiogeographic province that includes south-eastern China and eastern Australia (Dubinina 2000).

The late Cambrian conodont fauna from the Chingiz Mountain Range recorded herein comprises abundant *Phakelodus* and diverse coniform paraconodonts. Elements

◀
 Fig. 6. Late Cambrian and Early Ordovician conodonts from cherts of the Kol'denen River sections photographed in transmitted light except where indicated. **A, G, I, ?** *Coelocerosodontus bicostatus* van Wamel, 1974: **A**, Morphotype 'latus', CM 2/13174, sample KD010, $\times 95$; **G**, Morphotype 'bicostatus', CM 3/13174, sample KD010, $\times 108$; **I**, Morphotype 'latus', CM 4/13174, sample KD010, $\times 80$. **B**, '*Barnesodus*' *gibber* Dubinina, 2000, CM 1/13174, sample KD010, $\times 120$. **C**, *Prooneotodus?* sp., CM 36/13174, sample KD010, $\times 95$. **D**, *Phakelodus tenuis* (Müller, 1959), CM 24/13174, sample KD010, $\times 75$. **E, F**, *Furnishina* sp. **A**: **E**, Cluster composed of 7 elements, CM 18/13174, sample KD010, $\times 98$; **F**, Posterior view, CM 19/13174, sample KD010, $\times 109$. **H, J, L**, *Prooneotodus* sp.: **H**, CM 33/13174, sample KD010, $\times 100$; **J**, CM 34/13174, sample KD010, $\times 120$; **L**, CM 35/13174, sample KD010, $\times 115$. **K**, Gen. et sp. indet., photographed in reflected light, CM 1/13174, sample KD07, $\times 65$. **M, N**, *Paltodus* sp., CM 21/13174: **M**, Photographed in transmitted light, $\times 70$; **N**, Photographed in reflected light, $\times 82$.

of *Westergaardodina* that are usually common in most upper Cambrian conodont occurrences were not recorded in the

localities studied. *Westergaardodina* representatives are also missing in contemporaneous deep-water ribbon-banded cherts of

the Burubaital Formation in central Kazakhstan that further favours an open-sea affinity for the late Cambrian conodont assemblage of the Chingiz Mountain Range.

The conodont fauna from the Lower Ordovician part of the succession is dominated by representatives of *Cordylodus*, *Variabiloconus* and *Drepanodus*—genera of almost cosmopolitan distribution that are commonly found in shallow-sea deposits, and slope and basin facies bounding continental margins (Seo *et al.* 1994, Pyle & Barnes 2001, Miller *et al.* 2006). Too few elements of other euconodont taxa were recovered in this study to establish the palaeogeographic affinity of the Chingiz Range fauna. However, an occurrence of *Rossodus* elements supports the palaeogeographical reconstruction of Kazakhstan in generally low latitudes during the late Cambrian and Ordovician (Collins *et al.* 2003).

Other palaeontological data

Besides conodonts, both the carbonate and chert samples examined produced lingulate brachiopods, sponge spicules, various arthropods and algal cysts. These fossil assemblages are typical of pelagic and deeper-water benthic environments, being widely recorded in the offshore slope and basin black shale and siliceous facies of all continents (Ethington 1981, Repetski 1997).

The fossil assemblages recovered from cherts and carbonate beds of the same stratigraphic level are almost identical, varying only in the abundance of rock-forming sponge spicules that are common in the cherts but rare in the carbonates. However, the abundance of spicules varies, even in the chert samples; cherts range from true spiculites to transparent glass-like microcrystalline quartzites.

Brachiopods

The lowermost sample KD04 yields relatively abundant mid-Cambrian lingulate brachiopods including *Odontotreta galinae* (Popov, Holmer & Gorjansky, 1996) [= *O. mirabilis* Ushatinskaya, 1998], *Kotylotreta undata* Koneva, 1990, *Stilpnoretta* sp., *Erbotreta* sp., '*Homotreta*' sp. and unidentified acrothelids and lingulids.

Two samples from carbonate layers in the Zerbkyzyl Mountain area (ZK01, ZK02-5) have produced a few brachiopods assignable to *Eurytreta* sp. (Fig. 9G, H) and *Eoscapheasma* sp. (Fig. 9K, L). Unidentifiable juvenile obolids numerically dominate the residues (Fig. 9I, J).

Algal cysts

The most abundant component of the fossil fraction of the residues besides sponge spicules are small spheroidal bodies. All of the spheroids are translucent and dark brown, of regular shape with a smooth

Fig. 7. Scanning electron micrographs of Early Ordovician conodonts from carbonate beds of the Zerbkyzyl Mountains. **A**, *Phakelodus elongatus* (An, 1982), CM 23/13174, sample ZK01-2, $\times 45$. **B**, Cluster of *Phakelodus tenuis* (Müller, 1959) elements, CM 26/13174, sample ZK02-1, $\times 85$. **C**, **D**, *Coelocerodontus bicostatus* van Wamel, 1974: **C**, Morphotype 'bicostatus', CM 9/13174, sample ZK01-2, $\times 55$; **D**, Morphotype 'bicostatus', CM 10/13174, sample ZK01-2, $\times 45$. **E**, *Proconodontus? savitzkyi* Abaimova, 1978, CM 30/13174, sample ZK01-2, $\times 95$. **F**, *Phakelodus* sp., CM 27a/13174, sample ZK01-2, $\times 85$. **G**, *Cordylodus* sp. cf. *C. angulatus* Pander, 1856, CM 12/13174, sample ZK01-2, $\times 84$. **H**, *Rossodus?* sp., M element, CM 37/13174, sample ZK01-2, $\times 80$. **I**, **J**, *Rossodus manitouensis* Repetski & Ethington, 1983: **I**, Sb element, posterior view, CM 38/13174, sample ZK01-2, $\times 75$; **J**, Sd element, CM 38a/13174, sample ZK01-2, $\times 68$. **L**, *Drepanoistodus* sp., CM 14/13174, sample ZK01-2, $\times 65$. **K**, **M**, **N**, *Variabiloconus* sp. cf. *V. bassleri* (Furnish, 1938): **K**, CM 39/13174, sample ZK01-2, $\times 61$; **M**, CM 40/13174, sample ZK01-2, $\times 76$; **N**, CM 41/13174, sample ZK01-2, $\times 72$. **O**, *Polycostatus* sp., CM 28/13174, sample ZK01-2, $\times 69$.

Fig. 8. Early Ordovician conodonts and sponge spicules from cherts of the Zerbkyzyl Mountains photographed in transmitted light. **A, B, F**, *Prioniodus* sp. cf. *P. oepiki* (McTavish 1973): **A**, S element, CM 28a/13174, sample ZK01-11, $\times 75$; **B**, P element, CM 28b/13174, sample ZK01-11, $\times 80$; **F**, M element, CM 28c/13174, sample ZK01-11, $\times 78$. **C, G**, *Coelocerodontus bicostatus* van Wamel, 1974: **C**, Morphotype 'latus', CM 7/13174, sample ZK01-11, $\times 64$; **G**, Morphotype 'flat', CM 8/13174, sample ZK01-11, $\times 75$. **D**, *Hertzina*? sp., CM 20/13174, sample ZK01-11, $\times 70$. **E, J, K**, *Drepanodus arcuatus* Pander, 1856: **E**, CM 13a/13174, sample ZK01-11, $\times 70$; **J**, CM 13b/13174, sample ZK01-11, $\times 63$; **K**, CM 13c/13174, sample ZK01-11, $\times 80$. **H**, *Drepanoistodus* sp. CM 13d/13174, sample ZK01-11, $\times 70$. **I**, *Variabiloconus* sp., CM 13e/13174, sample ZK01-11, $\times 75$. **L**, *Cornuodus* sp., CM 13/13174, sample ZK01-11, $\times 65$. **M**, Sponge spicules in grey cherts, CM 44/13174, sample ZK01-11, $\times 20$.

Fig. 9. Fossils from the carbonate beds of the Zerbkyzyl Mountains. **A, B**, *Liangshanella* sp. Huo, 1956: **A**, Carapace in dorsal view, CM 49/13174, sample ZK01-2, $\times 45$; **B**, Carapace in left lateral view, CM 50/13174, sample ZK01-2, $\times 40$. **C, F**, *Tasmanites*? sp.: **C**, CM 43/13174, sample ZK01-2, $\times 110$; **F**, CM 43a/13174, sample ZK01-2, $\times 115$. **D, E**, Arthropod fragments: **D**, Cell-like structure on the carapace surface, CM 52/13174, sample ZK01-2, $\times 550$; **E**, CM 53/13174, sample ZK01-2, $\times 41$. **G, H**, *Eurytreta* sp.: **G**, Dorsal-lateral view, CM 46/13174, sample ZK01-2 $\times 48$; **H**, Ventral view, CM 47/13174, $\times 50$. **I, J**, Juvenile obolid brachiopod, CM 51/13174, sample ZK01-2: **I**, $\times 40$; **J**, $\times 38$. **K, L**, *Eoscapheasma* sp., CM 48/13174, sample ZK01-2: **K**, $\times 35$; **L**, $\times 35$.

outer surface lacking natural openings. Their diameters vary between 100 and 250 μm . Broken specimens show that the spheroids are hollow with walls about 1 μm thick (Fig. 9C). Some spheroids are creased (Fig. 9F) or almost flattened, but their surfaces do not bear any cracks, showing that the walls of the spheroidal bodies were flexible. The internal structure of the walls was not studied, so we cannot be sure if canals perforate the walls of the spheres. Cysts of the prasinophyte *Tasmanites* are relatively common in similar faunal assemblages, e.g. in black shale associations, and have been reported from many localities in the world from the Proterozoic to Holocene (Schieber & Baird 2001). In the Lower Ordovician *Caryocaris* Shale in Nevada, they have been found in an almost identical fossil assemblage with caryocarid arthropods, small obolid and acrotretid brachiopods, conodonts and unidentified fragments of phosphatic plates (Ethington 1981). Therefore, we conclude that sphaeromorphs from the Zerbkyzyl Mountain area are probably *Tasmanites* cysts.

Arthropods

Abundant trilobites collected from the limestone bed of the Chingiztau Formation (section KD04) in the early 1970s showed that this unit belongs to the Russian–Kazakhian mid-Cambrian Mayan Stage. The trilobite collection was never illustrated, but the taxa were identified by N.K. Ivshin and cited in several publications (Ivshin *et al.* 1972).

One carbonate sample from the Zerbkyzyl Mountains (ZK01-2) yielded a few tiny bradoriid arthropods (Fig. 9A, B) that possibly belong to *Liangshanella* Huo, 1956 (identified by L.M. Mel'nikova). Small fragments of chitinous plates ornamented with small spines are undoubtedly the remains of arthropod carapaces (Fig. 9D, E).

However, the high degree of fragmentation makes closer taxonomic discrimination impossible.

Depositional history and stratigraphic framework

The available data provide some constraints on the timing of depositional and tectonic events during the Cambrian and Ordovician in the Chingiz Mountain Range. Carbonate sedimentation gradually decreased from the middle to the late Cambrian; the end of the Cambrian was characterized by extensive siliceous deposition. This trend from slope margin to basin facies is most probably related to regional tectonism. The upper Cambrian cherts and siltstones contain only a minor admixture of volcanic ash. At the end of the Early Ordovician, the sequences of pelagic and hemipelagic sediments including cherts grade upward to thick proximal turbidite assemblages of volcanoclastic rocks indicating that an active volcanic arc persisted nearby.

Siliceous sedimentary rocks are widespread in the Ordovician of central Kazakhstan. They are usually prominent in ophiolite assemblages (Tolmacheva *et al.* 2001) and in successions lying basinward of carbonate shelves (Nikitin 1972). Petrographic studies and macroscopic observation of different types of cherts in Kazakhstan (Zhylkaidarov 1998, Tolmacheva *et al.* 2004) indicate that they are composed essentially of a mixture of siliceous bioclasts (radiolarians and sponge spicules). Cherts and jaspers of a hydrothermal origin are less common and are usually restricted to volcanic sequences. The cherts in the studied area of the Chingiz Mountain Range are composed primarily of sponge spicules concentrated in beds within predominantly clastic slope successions that were deposited near the shallow Ordovician level of carbonate compensation (CCD).

The latter assumption is confirmed by numerous carbonate beds in the succession. During the latest Cambrian and Early Ordovician, the input of siliciclastic material was very limited, leading to an accumulation of biogenic sediments dominated by sponge spicules. However, deposition of silica was episodic evident from the patchy stratigraphic distribution of chert beds and their variable thickness from a few centimetres to several tens of metres.

During the Cambrian and Early Ordovician, sedimentation in the Chingiz Mountain Range area took place in the back-arc basin that flanked the active volcanic island arc to the southwest. Much of the middle Cambrian to Early Ordovician history in the region involves deposition of non-fossiliferous clastic and volcanoclastic sediments with only rare siliceous and carbonate layers. Conodont-based biostratigraphy indicates the presence of the paraconodont-dominated zone of the upper Cambrian overlain by *R. manitouensis* and *Prioniodus oepiki* zones of the Tremadocian, although the Cambrian–Ordovician boundary is impossible to recognize definitively. We can only conclude that the boundary lies within an interval approximately 50 m thick of presumably clastic sedimentation with minor chert layers that precedes the interval characterized by the main pulses of siliceous sedimentation (Fig. 4).

Taxonomic notes

The following remarks apply to the most abundant conodont elements. The majority of conodont elements in our collection are identified in open nomenclature, as the paucity of conodonts in all the samples coupled with their poor preservation in chert makes specific identification uncertain. Isolated elements and clusters of *Phakelodus* species are numerous in the chert and limestone samples. Clusters of these conodonts occur more commonly in the

carbonates and are usually as abundant as isolated elements.

Minute paraconodont elements are diverse in both the cherts and limestones, but most are inadequately preserved for confident specific or generic determination. Most paraconodont elements in cherts lack visible basal cavities and outlines of the basal margins. Each kind of element is found in very small numbers, making their identification difficult. Euconodont elements are scarce in both chert and limestone samples, but they do provide some data for at least an approximate determination of age and generalized biostratigraphy.

Figured specimens (Figs. 5–9) are housed in the collection of the F.N. Chernyshev Central Research Geological Exploration Museum (CNIGR Museum) of the Russian Research Geological Institute, St. Petersburg, Russia. Conodonts and sponge spicules in thin-sections were photographed in transmitted and reflected light with a Nikon Coolpix E5000 digital camera.

Phakelodus taxa (Figs 5G; 6D; 7A, B, F)

Elongated and slender elements of *Ph. elongatus* are easily distinguished from similar elements of *Ph. tenuis* by the posterior keel and lacrimiform cross-section. However, when fossils are viewed in two dimensions in cherts, these taxa are nearly identical in lateral view and cannot be reliably differentiated.

One group of elements (Figs 5G; 7F) should probably be recognized as a new species of *Phakelodus* mainly due to their occurrence in clusters. The *Phakelodus* sp. elements are more laterally compressed and broader than those of *Ph. elongatus* but are similar to them in having a smooth anterior side and well-developed posterior keel. The elements of *Phakelodus* sp. also differ slightly from *Ph. tenuis* and *Ph. elongatus* elements by having their maximum

curvature closer to the element tip. In the latter species, the anterior margin forms more or less a continuous curve from the aboral margin to the element tip. All three *Phakelodus* taxa are found in uni-membrate clusters that contain elements of one or another taxon.

'Barnesodus' gibber Dubinina, 2000 (Fig. 6B)

A few very small paraconodont elements are referable to '*B.*' *gibber* described from the upper Cambrian of the Malyi Karatay (Dubinina 2000). The diagnosis of *Barnesodus* encompasses laterally compressed simple-cone paraconodont species. Three of four described *Barnesodus* species are non-geniculate, whereas the elements of '*B.*' *gibber* have a geniculate morphology. The geniculate forms are extremely rare among Cambrian euconodonts and paraconodonts. In fact, '*B.*' *gibber* is the first known geniculate paraconodont whose assignment to paraconodonts is proven by its mode of growth. The base of juvenile elements is relatively low; whereas adult and gerontic elements have significantly higher bases, so the ratio of the base and cusp sizes changes during element ontogeny.

Furnishina sp. A (Fig. 6E, F)

A few small paraconodonts with a long thin cusp and low flared basal section that is distinctly set off from the cusp are referred to this informal species of *Furnishina*. The elements of *Furnishina* sp. A were found only in chert where their morphology cannot be established with confidence due to poor preservation. The number of keels on the base is an important character for discrimination of *Furnishina* species, but keels and the outline of basal margins are not visible on minute elements in the cherts. One cluster of *Furnishina* sp. A comprising six elements was discovered.

Coelocerodontus bicostatus van Wamel, 1974 (Figs 6A, G, I; 7C, D; 9C, G)

More than 10 elements in the collection are referred to *C. bicostatus*. Simple cone acostate elements of this species are easily recognizable by their hook-like apical portion. Costate elements have distinct keels on one or both lateral sides. In Kazakhstan, elements of *C. bicostatus* are very common in all conodont assemblages from upper Cambrian and lowermost Ordovician siliceous sedimentary rocks.

Rossodus manitouensis Repetski & Ethington, 1983 (Fig. 7I, J)

Five elements in our collection are referred to *R. manitouensis*, as three of the more complete examples are alate with distinctive carinas on their anterior faces. The elements are dark grey and almost opaque, obscuring the pattern of white-matter distribution. A single M element in the collection (Fig. 7H) is only tentatively assigned to *Rossodus* apparatus, as it has a less inclined cusp and significantly shorter anterior extension than is characteristic of *R. manitouensis* M elements. The elements of *R. manitouensis* are more commonly found in North America (Repetski & Ethington 1983) and western Newfoundland (Ji & Barnes 1994), although they have also been reported in Gondwanan terrains (Seo *et al.* 1994). *Rossodus manitouensis* has not been recorded previously in Kazakhstan.

Prioniodus sp. cf. **P. oepiki** (McTavish 1973) (Fig. 8A, B, F)

Three Sc, two P and two M prioniodontid elements agree in gross morphology with the published descriptions of *P. oepiki* (McTavish 1973). This species has been reported from Western Australia (McTavish 1973) and western Newfoundland (Stouge & Bagnoli 1988), and it is also common in Kazakhstan (Tolmacheva *et al.* 2004).

However, as Sd and Sa elements have not been found in our samples, we cannot be precisely sure that the available elements belong to *P. oepiki*. *Prioniodus oepiki* from Kazakhstan differs from the Newfoundland material (Stouge & Bagnoli 1988) by the morphology of the M elements that in the Kazakhstani population have a shorter and non-denticulate anterior extension of the base (Fig. 8F). However, in our opinion, the Kazakhstani and Newfoundland morphotypes do not represent separate taxa, but rather represent intraspecific variation within a wide-ranging population.

Conclusions

The sedimentary succession in the central Chingiz Range is composed of siltstones, sandstones and volcanoclastic rocks with rare beds of micritic carbonates, black shales and cherts, and, as defined by conodont biostratigraphy, encompasses the middle Cambrian to Lower Ordovician (*P. oepiki* Zone) interval.

Sponge spicules are the most common fossils in the sediments and occur with less abundant conodonts, lingulate brachiopods, arthropods and *Tasmanites*? cysts. The conodont assemblages recorded from both carbonate and chert layers within the volcanoclastic and siliciclastic successions have an open oceanic affinity.

Siliceous sediments of the latest Cambrian and Early Ordovician were deposited in an open marine environment in the back-arc setting adjacent to an active island arc. Chert production was most likely caused by local productivity increase connected with the volcanic activity. Numerous carbonate beds and layers of carbonate breccias occur within the predominantly siliciclastic intervals, indicating that sedimentation during the late Cambrian and Early Ordovician occurred in slope and deeper basin environments near the shallow Ordovician carbonate compensation depth (CCD).

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