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Integrated approaches to terminal Proterozoic stratigraphy: an example from the Olenek Uplift, northeastern Siberia

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Abstract

In the Olenek Uplift of northeastern Siberia, the Khorbusuonka Group and overlying Kessyusa and Erkeket formations preserve a significant record of terminal Proterozoic and basal Cambrian Earth history. A composite section more than 350 m thick is reconstructed from numerous exposures along the Khorbusuonka River. The Khorbusuonka Group comprises three principal sedimentary sequences: peritidal dolomites of the Mastakh Formation, which are bounded above and below by red beds; the Khatyspyt and most of the overlying Turkut formations, which shallow upward from relatively deep-water carbonaceous micrites to cross-bedded dolomitic grainstones and stromatolites; and a thin upper Turkut sequence bounded by karst surfaces. The overlying Kessyusa Formation is bounded above and below by erosional surfaces and contains additional parasequence boundaries internally. Ediacaran metazoans, simple trace fossils, and vendotaenids occur in the Khatyspyt Formation; small shelly fossils, more complex trace fossils, and acritarchs all appear near the base of the Kessyusa Formation and diversify upward. The carbon-isotopic composition of carbonates varies stratigraphically in a pattern comparable to that determined for other terminal Proterozoic and basal Cambrian successions. In concert, litho-, bio-, and chemostratigraphic data indicate the importance of the Khorbusuonka Group in the global correlation of terminal Proterozoic sedimentary rocks. Stratigraphic data and a recently determined radiometric date on basal Kessyusa volcanic breccias further underscore the significance of the Olenek region in investigations of the Proterozoic–Cambrian boundary.

1. Introduction

A chronostratigraphic framework is emerging for the terminal Proterozoic Eon. Biological events are important elements of this framework, but fossils are not as abundant or diverse in Neoproterozoic (1000–545 Ma) rocks as they are in younger strata, and turnover rates appear to have been lower. Therefore, paleontological data must be complemented by additional stratigraphic information. Secular variations in the C- and Sr-isotopic compositions of carbonates constitute an important and independent means of correlating rocks of this age (Kaufman and Knoll, 1995, and references therein), and preliminary results indicate that, in concert, fossils and isotopic chemostratigraphy provide reliable tools for intra- and interbasinal correlation of terminal Proterozoic successions (Knoll and Walter, 1992; Narbonne et al., 1994). Given the sensitivity of biostratigraphy to facies distributions and of chemostratigraphy to hiatuses, sequence stratigraphy constitutes an important lithostratigraphic framework for correlations.

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Terminal Proterozoic (Yudomian) and basal Cambrian (Nemakit–Daldyn and Tommotian) sedimentary rocks outcrop extensively along the Khorbusuonka River in the Olenek Uplift, northeastern Siberia (Fig. 1). Because of the wealth of paleontological data obtained from this succession, Sokolov (1985; Sokolov and Fedonkin, 1984) chose it to epitomize the development of Vendian equivalents on the Siberian Platform, respectively correlating the Mastakh, Khatyspyt, and Turkut formations (Figs. 2 and 3) with the Laplandian (Volhyn), Redkino, and Kotlin stages of the East European Platform. Khomentovsky and Karlova (1993) have similarly underscored the importance of these strata for Proterozoic–Cambrian boundary considerations.

The succession contains Ediacaran metazoans, vendotaenids, small shelly fossils, trace fossils, and acritarchs. It is rich in carbonates, facilitating chemostratigraphic correlation. Furthermore, Bowring et al. (1993) have recently provided a U–Pb age for zircons in the upper part of the succession. Thus, Khorbusuonka strata provide an unusual opportunity to explore the relationships among lithostratigraphic, chemostratigraphic, paleontological, and geochronometric data in a terminal Proterozoic/basal Cambrian succession.

2. Terminal Proterozoic and basal Cambrian stratigraphy of the Khorbusuonka River

Along the banks of the Khorbusuonka River and its tributaries (Fig. 1), outcrops have a structural dip of three to ten degrees to the north; because the river flows north-northwest, progressively higher stratigraphic levels are encountered downstream. A swarm of small-displacement, east–west-oriented high-angle normal faults cuts the Olenek Uplift (Krasilshchikov and Biterman, 1970). Although fault-displacement is low (less than a few tens of meters), regional dip is very low and equivalent stratigraphic levels may have lateral surface offsets of up to several kilometers. Consequently, while it is possible to obtain continuous measurements of individual formations at single locations (Figs. 1 and 2), the section presented in Fig. 3 is a composite pieced together from many sections over a distance of 80 km. Formational thicknesses and facies vary within the Olenek region (V.V. Khomentovsky in Sokolov and Fedonkin, 1985; S. Pelechaty, pers. commun., 1993).

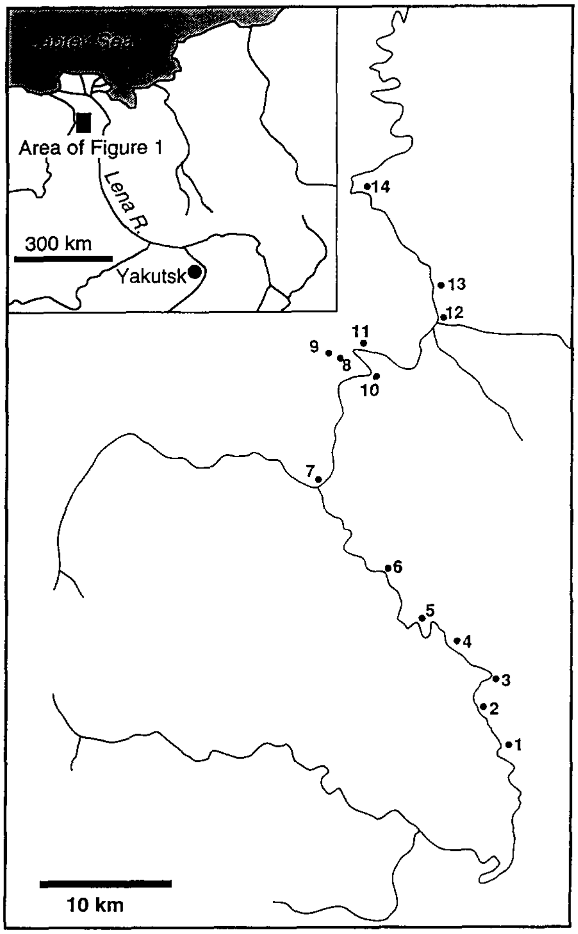


Fig. 1. Map of the Khorbusuonka River, Olenek Uplift, northern Siberia (see inset) showing the locations of the measured sections depicted in Fig. 2.

2.1. Riphean–Yudoma boundary and the Khaipakh Formation

The Riphean–Yudoma boundary is generally accepted as the contact between the Mastakh Formation and underlying units. At each of the four locations where we observed the contact, a different facies underlies the lithologically homogeneous Mastakh Formation (Fig. 2, sections 1–4). Shenfil (1992) interpreted this contact as an angular unconformity, with faulting and erosion exposing stromatolitic reefs, ooid shoals,

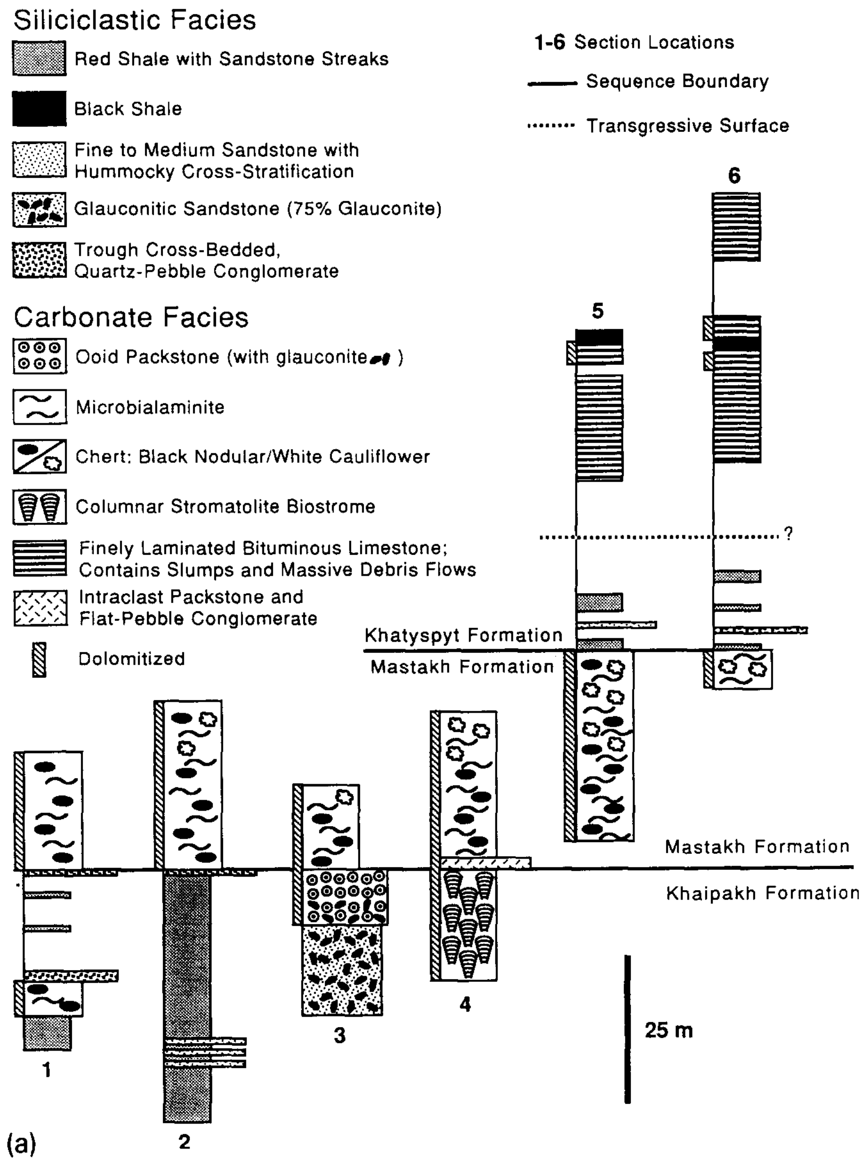


Fig. 2. Measured sections of terminal Proterozoic and basal Cambrian strata along the Khorbusuonka River. See Fig. 1 for section locations.

and outer-shoreface siliciclastic shelf facies of the Riphean Debengda and Khaipakh formations on the surface flooded by the Mastakh sea. In some sections (sections 1 and 2 in Fig. 2), these are abruptly overlain by quartz-pebble conglomerate that fills in along a scoured surface.

Preliminary inspection of seismic reflection profiles for the Olenek Uplift shows continuous reflections for strata immediately beneath the Mastakh Formation. The seismically continuous strata are thin and onlap faulted subjacent strata. This suggests that sedimentary rocks immediately beneath the Mastakh Formation might alternatively be interpreted as a complex mosaic of mixed carbonate–siliciclastic facies, featuring reef-ooid shoals within a broad siliciclastic shelf. In the

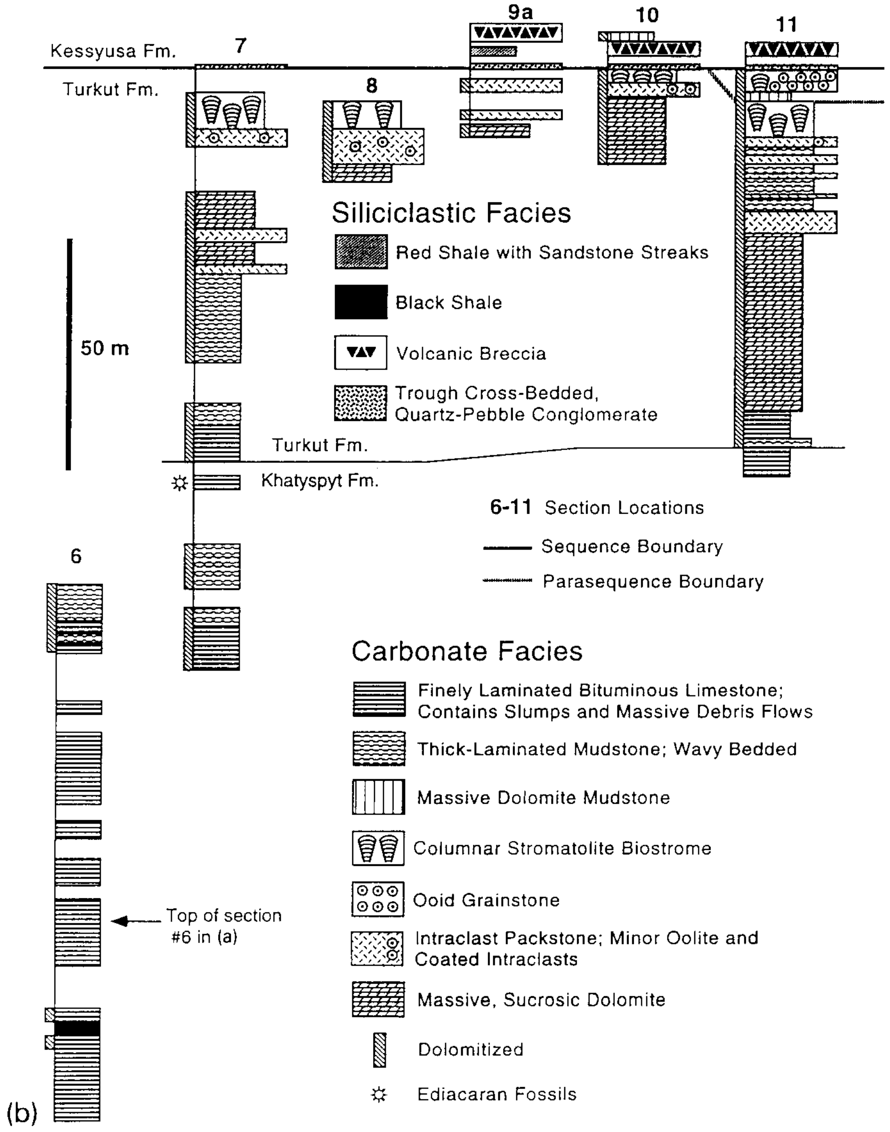
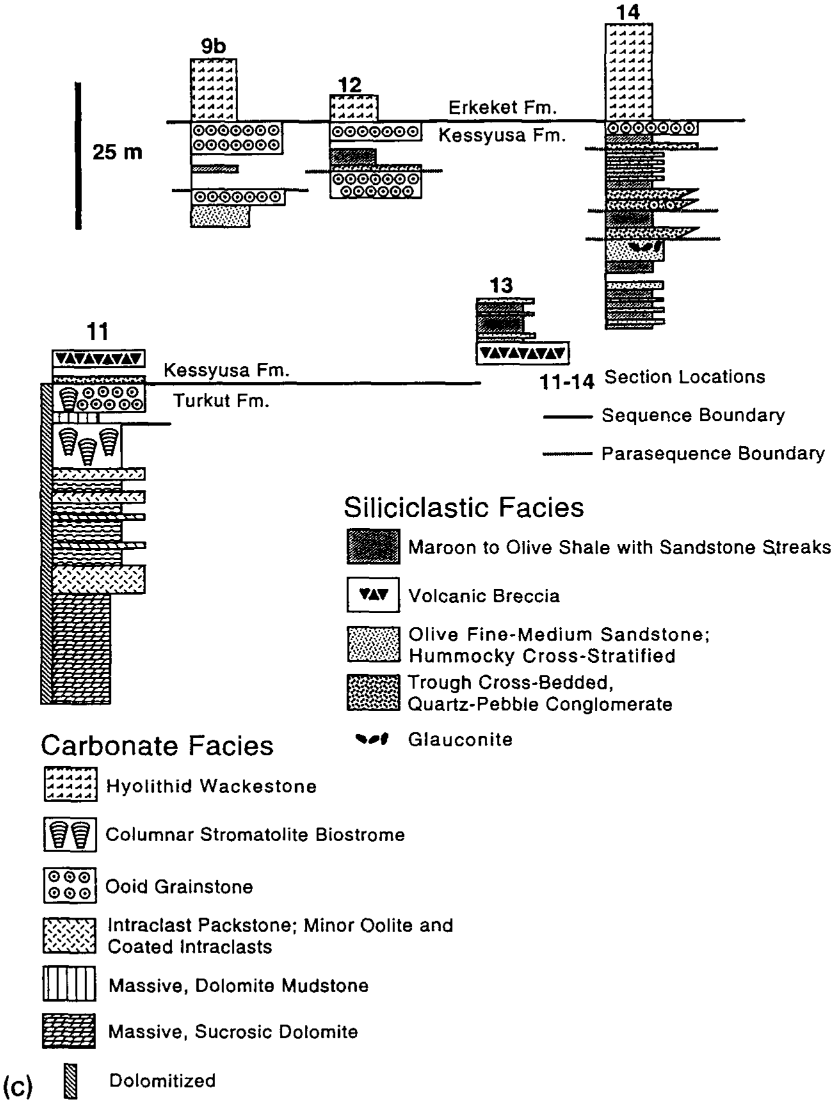


Fig. 2. (continued).

absence of more extensive exposure, sedimentological interpretation is difficult.

2.2. Mastakh Formation

The Mastakh Formation is ~40 m thick. Of the several sections observed (Fig. 2), section 5 is the most completely exposed, although the basal few meters were not observed at the point of measurement. The base is exposed a few hundred meters along strike, allowing our estimate of its placement to within about five meters. The Mastakh Formation consists of desiccation-cracked, microbially laminated, fine-grained dolostones with abundant black chert nodules and white "cauliflower" chert nodules. The black cherts predate compaction; cauliflower chert nodules also predate compaction but locally replace black cherts. Black chert nodules are distributed through the entire formation, whereas cauliflower chert nodules are restricted to its upper half (Fig. 2). In most sections, cherty



microbial laminites (stratiform and rare domal to low columnar stromatolites) rest abruptly on underlying Riphean rocks; however, at section 4 (Fig. 2) an intervening 2 m unit of intraclast packstone and flat-pebble conglomerate is developed. The upper contact of the Mastakh Formation is also sharp; red, sandy siltstones of the basal Khatyspyt Formation sit abruptly on Mastakh dolostones.

Sedimentologically, the Mastakh Formation is interpreted to represent deposition on regionally extensive, tidally exposed mud flats. The fine but uneven and often irregular lamination of most Mastakh dolomites and poorly preserved filamentous fossils in black chert nodules demonstrate that these mud flats were colonized by microbial mats. Extensive precipitation of chert is consistent with concentration of dissolved silica in an evaporative tidal-flat setting (Maliva et al., 1990). Although neither evaporite minerals nor mineral pseudomorphs were observed, cauliflower cherts may have replaced anhydrite (Milliken, 1979). Mg-rich dolomitizing fluids may have been generated by evaporative concentration following precipitation of anhydrite

(Adams and Rhodes, 1960) or, depending on the calcium to bicarbonate ratio of seawater at that time, precipitation of calcium carbonate (Grotzinger, 1989; Fairchild et al., 1991; Grotzinger and Kasting, 1993).

Stratigraphically, the Mastakh Formation is interpreted as an incomplete sequence, containing evidence for only highstand deposition. The intraclast breccia at the base of section 4 (Fig. 2) probably represents the reworking of sediments initially deposited during transgression of the sequence boundary atop the Riphean formations. However, with the exception of the basal transgressive deposit, sedimentation is virtually restricted to the aggradation of tidal mud-flats, consistent with transgressive to early highstand deposition (Sarg, 1988).

2.3. *Khatyspyt Formation*

The Khatyspyt Formation can be divided informally into lower and upper members. The lower member consists predominantly of poorly exposed, thinly interbedded red siltstones, fine sandstones, and intercalated medium-to-coarse sandstones and quartz-pebble conglomerates (see sections 5 and 6, Fig. 2). We did not observe the contact between the lower and upper members, but estimate that it occurs about 20 to 25 m above the base of the formation. Sandstones are fine- to medium-grained and characterized by hummocky cross-stratification (HCS). Coarser sandstones and pebble conglomerates are massive or trough cross-stratified.

The upper member of the Khatyspyt Formation consists of gray to dark gray or black, finely laminated, bituminous limestones and minor dolostones, intercalated with massive limestones (Fig. 2) and uncommon black shales. The fine laminae of the limestones characteristically define low-relief hummocks and swales with amplitudes of less than 5–10 cm and wavelengths of 50–100 cm. Locally, these sediments show extensive, penecontemporaneous deformation structures, including recumbent folds and small thrusts. Massive limestone beds are 50 to 100 cm thick and often contain intraclasts of finely laminated limestone dispersed in a fine matrix; intraclasts may show normal grading. Organic content (TOC) of the bituminous limestones locally exceeds 11 mg C/g, or 1.1%.

The upper member is on the order of 150 to 160 m thick. Its thickness is difficult to estimate because the lower contact was not exposed and the upper contact is gradational with the overlying Turkut Formation and defined principally on the basis of dolomite/limestone content. This latter feature appears to be highly variable, but we place the contact at the top of the last bed of limestone or dolomitic limestone greater than 1 m thick. This definition allows for minor occurrences of limestone within the lower Turkut Formation. We note that the finely laminated, bituminous facies of the Khatyspyt Formation extend up into the lower Turkut Formation where they are overprinted by dolomite (see sections 7 and 11, Fig. 2). The transition from mostly limestone to mostly dolomite, which marks the Khatyspyt–Turkut contact, also coincides with a subtle change in stratification style from fine to thick lamination.

The Khatyspyt Formation is interpreted to record a significant increase in accommodation space over the Olenek platform. An overall transgression and deepening-upward sequence is represented by flooding of the exposed Mastakh surface, deposition of lower Khatyspyt outer-shoreface siliciclastic sediments, and continued submergence to basinal depths where the finely laminated, bituminous limestones of the upper Khatyspyt accumulated. The presence of HCS and wave ripples in sandstones of the lower Khatyspyt indicate deposition above storm wave-base (Nottvedt and Kreisa, 1987). However, the hummocky and swaley surfaces in the finely laminated upper Khatyspyt limestones probably formed by differential compaction and pressure solution rather than tractional current reworking of clastic sediment particles. Grain size in these micrites is outside of the stability field for tractionally produced HCS (J. Southard, pers. commun., 1993), and wave or current ripples that commonly accompany tractionally produced HCS are absent. The finely laminated limestones probably represent fall-out of suspended carbonate muds in a distal, basinal setting. Intercalated structureless limestone beds are interpreted as debris flows deposited on the slope or floor of the basin. The presence of intraclasts of finely laminated sediment and the absence of fragments of shallower-water lithologies suggest that debris flows were initiated on the slope where only deeper-water sediments were present (Cook, 1979; Krause and Oldershaw, 1979). The absence of shallow-water breccia fragments in the debris flows also implies that the morphology of the carbonate platform was a ramp rather than rimmed shelf (Read, 1985; Grotzinger, 1989). Stratigraphically, the Khatyspyt and Turkut formations are considered to constitute a single sequence, lacking only lowstand systems tract deposits. The lower Khatyspyt Formation is interpreted as part of a transgressive systems tract, dominated by siliciclastic sedimentation. It is possible that siliciclastic sediments were introduced into the basin as lowstand deposits associated with subaerial exposure of the Mastakh Formation, and subsequently reworked during transgression (Nummendal and Swift, 1987). The upper Khatyspyt Formation is likely part of a continued transgressive to initial highstand systems tract. The transition from lower to upper Khatyspyt deposition probably coincides with a transgressive surface as shown in sections 5 and 6 of Fig. 2. The transition into the Turkut Formation is gradational; we suggest that the Turkut Formation represents highstand deposition (see below). We further suggest that the transition from deposition of transgressive red siltstones and sandstones of the lower Khatyspyt Formation to black, organic-rich limestones of the upper Khatyspyt Formation may indicate deposition beneath a stratified water column that promoted basinal anoxia.

2.4. Turkut Formation

The Turkut Formation consists of thickly laminated dolostones and minor finely laminated limestones, overlain by dolomitic intraclast–ooid grainstone and stromatolitic biostromes. The formation is ~60 to 70 m thick (based on our placement of the formation's base, as discussed above). Finely laminated limestones are similar to those of the Khatyspyt Formation, although much less abundant. Thick-laminated dolostones with wavy layering and fine grain size are common in the lower Turkut. The thick-laminated facies passes upward into dolomitic intraclast–ooid grainstone, although this transition is typically obscured by fabric-obliterative sucrosic dolomite. The sucrosic overprinting makes estimations of bedding thicknesses difficult, but the grainstones are clearly interstratified with the thick-laminated dolostones as shown in section 11 of Fig. 2. Grainstones typically consist of coarse sand- to pebble-size intraclasts, commonly with surficial oolitic coatings. Pores are filled with coarse spar, indicating high initial porosity; some grains are leached, providing evidence for secondary porosity as well. The upper Turkut Formation is marked by stromatolitic mounds several meters wide and with up to one meter of synoptic relief. Internally, the mounds consist of small, irregular, columnar stromatolites a few centimeters wide, with common linkage of laminae and poor inheritance. Dolomitization has resulted in considerable recrystallization, and microfabrics are vague.

The top of the stromatolitic unit is marked by an exposure surface, characterized by karstic brecciation of stromatolites and underlying grainstones. This surface is overlain by a unit of fine-grained, massive dolostone that passes upward into medium-grained ooid grainstone and infrequent stromatolite bioherms. The dolostones above the exposure surface contrast with those below it and are distinguished on the basis of their fabric-retentive, fine-grained and slightly darker dolomite, and the finer-grained and well-sorted aspect of the ooid grainstone. At several localities, primary and secondary pores are filled with bitumen. These grainstones are cut by a major erosional surface which is overlain by quartz-pebble conglomerates and shales of the overlying Kessyusa Formation.

The Turkut Formation is interpreted to represent shallowing from a basinal to shallow ramp environment (Grotzinger, 1989). Thick-laminated dolostones probably represent settling of suspended micrite and calcisiltite transported to a distal, down-dip position on the ramp. Finely laminated sediments accumulated in the most distal, basinal position. The upward transition to shallow-water grainstones and stromatolitic biostromes records progradation of more proximal, up-dip ramp environments, including shoals. Shallowing culminated in the development of a widespread, karstic subaerial exposure surface. A brief return to shallow-water carbonate sedimentation is shown by local development (section 11, Fig. 2) of a thin sequence of ooid grainstone at the top of the formation. The top of this sequence, and the formation boundary, is marked by another karstic, subaerial exposure surface. This surface apparently merges with the stratigraphically lower karstic surface between sections 11 and 10 (Fig. 2). It is likely that during exposure and karst development, undersaturated meteoric fluids produced secondary porosity in incompletely lithified ooid and intraclast grainstones (Meyers, 1988; Pelechaty et al., 1991). It is significant that these porous, reservoir-quality grainstones occur stratigraphically between underlying hydrocarbon source rocks (Khatyspyt bituminous limestones) and overlying reservoir-sealing shales of the Kessyusa Formation. That hydrocarbons did migrate and were trapped is shown by the occurrences of bitumen in the grainstones. The Turkut Formation represents predominantly highstand systems tract progradation. Collectively, the Khatyspyt and Turkut formations form a single stratigraphic sequence. The karstic surfaces near and at the top of the Turkut Formation indicate the position of the upper sequence boundary. The thin interval, bounded by karstic surfaces near the top of section 11 (Fig. 2), may be a regionally discontinuous parasequence formed near the end of highstand deposition under conditions of rapidly diminishing accommodation space. The sequence boundary at the top of the Turkut Formation is notable in that it closely approximates the Precambrian–Cambrian boundary.

2.5. Kessyusa Formation

The Kessyusa Formation, ~45 to 50 m thick along the Khorbusuonka River, comprises a diverse assemblage of mixed siliciclastic, carbonate and volcanic facies (Fig. 2). Its boundaries are well-defined by a basal quartz-pebble conglomerate and an abrupt, karstic contact with overlying maroon, hyolithid wackestones of the Erkeket Formation. The basal quartz-pebble conglomerate varies in thickness from 0.2 to 3.5 m and fills in residual topography atop the Turkut Formation. Above this conglomerate, the lower part of the formation consists predominantly of maroon siltstones with intercalated, thin, hummocky cross-stratified, fine- to medium-grained sandstone beds. Siltstones contain several 5 to 10 cm layers of flat-pebble conglomerate formed of fine calcisiltite and calcite-cemented siltstone. These siltstones contain shelly invertebrate fossils representing the *Anabarites trisulcatus* zone of Missarzhevsky (1989a, b). A volcanic breccia occurs in the lower part of the siltstone interval, ~2 to 3 m above the base of the formation. The breccia varies in thickness from 0.5 to 15 m and consists of pumice, dolomite and siltstone rock fragments, and pumice-coated rock fragments in a fine, dolomitic matrix. Magmatic zircons isolated from the breccia have an age of 543.9 ± 0.3 Ma (Bowring et al., 1993).

The siltstone interval is overlain by a ~5 m unit of amalgamated, glauconitic HCS located near the middle of the formation at section 14 (Fig. 2). This unit has an upper, scoured contact and is abruptly overlain by 2 to 3 m of trough cross-bedded pebble conglomerate. This conglomerate is overlain by a thin sequence of sandy siltstones which, in turn, are scoured by a second quartz-pebble conglomerate containing one prominent and possibly two amalgamation surfaces. In contrast to the first conglomerate, the second conglomerate contains abundant ooids; it is possible that the erosional surface at its base has resulted in the removal of oolitic facies which are still preserved in section 12 (Fig. 2). A significant flooding surface separates the second conglomerate unit from overlying siltstones. This ~10 m interval contains intercalated fine- to medium-grained sandstone beds with HCS; it is abruptly overlain by a third quartz-pebble conglomerate near the top of the formation (see section 14, Fig. 2). The top of the Kessyusa Formation is marked by one or more oolites separated by thin, maroon shale beds. Oolites may be amalgamated; they contain abundant fragments of Tommotian shelly invertebrates. The upper contact of the Kessyusa is marked by a karst surface developed on the upper oolite unit, as observed at sections 9, 12 and 14 (Fig. 2). Karst-related dissolution features, including potholes and grikes, extend up to 1.5 m below the boundary with the overlying Erkeket Formation. In most locations, particularly at section 14 (Fig. 2), ferruginous stromatolites and accretionary hardgrounds encrust the karstic surface; it is overlain by a thin lag deposit of glauconite and rounded intraclasts.

The siltstones and hummocky cross-stratified sandstones of the Kessyusa Formation indicate deposition on the outer shoreface of a siliciclastic shelf (Dott and Bourgeois, 1982; Nottvedt and Kreisa, 1987). Deposition was close to storm wave-base, as shown by the occurrence of HCS in sandstones and occasional flat-pebble conglomerates. These facies are interpreted to indicate episodic impingement of storm-induced, high-velocity oscillatory shear currents on the seafloor where early-cemented siltstone and fine calcisiltite layers (hardgrounds?) were reworked into lag deposits. The breccia near the base of the formation provides evidence of explosive volcanism and is probably related to numerous kimberlite and diatreme pipes that cut the Turkut Formation but not the Kessyusa (Shpunt and Shamshina, 1989). The composition and coarse clast size of the breccia suggest that it was the result of phreatomagmatic eruption.

The quartz-pebble conglomerates were deposited under conditions of sustained, possibly unidirectional, traction currents, as shown by the dominance of trough cross-bedding. In comparison to enclosing shales and sandstones, the conglomerates are discordant with respect to both grain size and stratification style. Consequently, we suggest that they were emplaced during relative lowstands in sea level. A reduction in accommodation space is an efficient way for coarse, upper shoreface or fluvial sediments to be shunted down-dip into more distal positions (Jervey, 1988; Posamentier and Vail, 1988). As such, the conglomerates would be underlain by disconformable surfaces and would represent lowstand deposits (Vail et al., 1977).

Oolites of the upper Kessyusa probably formed as isolated shoals within a siliciclastic-dominated shelf (Beukes, 1977, 1983). The top of the formation is marked by a regionally persistent oolite. Minor changes in relative sea level would have promoted downward shifts in siliciclastic sedimentation and, possibly, incision of previously deposited oolites. At least, local incision of the oolites is supported by the occurrence of conglomerate that scours oolite at section 12 (Fig. 2) and the occurrence of detrital ooids within the middle conglomerate of section 14 (Fig. 2).

Stratigraphically, the Kessyusa Formation is interpreted as a single depositional sequence, subdivisible into four parasequences (section 14, Fig. 2). The bases of all four parasequences are marked by quartz-pebble conglomerates deposited during accommodation minima. Vertical transitions do not reveal net long-term trends in parasequence thickness, grain size, or facies. Consequently, the Kessyusa sequence probably records deposition during maximum flooding of the shelf, or possibly initial highstand progradation (Van Wagoner et al., 1988). Neither transgressive nor strongly progradational (highstand) sediments were preserved. Thus, the Kessyusa sequence is truncated beneath the Erkeket Formation, indicating that the formation boundary is also a significant sequence boundary.

2.6. Erkeket Formation

Only the basal 10 m of the Erkeket Formation were examined. At several locations (sections 9, 12 and 14, Fig. 2) this interval consists of monotonous, maroon- to mauve-colored, argillaceous, fossiliferous, lime mudstone and wackestone. Large hyolithids are locally abundant enough to form packstone lags. Bioturbation is characteristic and large, simple, bedding-parallel traces are particularly common. The lower Erkeket Formation represents deposition on an open, shallow-marine carbonate platform. The considerable bioturbation prevents detailed analysis of fairweather stratification, but it is likely that storms resulted in significant reworking of the seafloor, such that fossils were concentrated as lag deposits by winnowing. The Erkeket Formation represents regional onlap of the Olenek platform, following subaerial exposure of upper Kessyusa rocks. Erkeket carbonates may have accumulated during a highstand in relative sea level.

3. Biostratigraphy

3.1. Prokaryotes, protists, and stromatolites

As noted in the previous section, silicified carbonates of the Mastakh Formation contain poorly preserved sheaths of mat-forming cyanobacteria. This abundant but taxonomically depauperate assemblage supports sedimentological inferences about Mastakh depositional environments, but does little to constrain its age.

More informative microfossils occur stratigraphically higher, in the Kessyusa Formation. Lower Kessyusa siltstones contain rare leiospherid acritarchs. Dark green siltstones in the middle to upper part of the formation contain abundant leiospherids (*Leiospheridia* spp.), as well as *Granomarginata prima*, *G. squamacea*, *Leiomarginata simplex*, *Tasmanites tenellus*, *Ceratophyton vernicosum*, *Michrystridium tornatum*, and *Cephalonyx* sp. (Ogurtsova, 1975; Rudavskaya and Vasileva, 1985; Kolosov, 1989; Pyatiletov, 1989; Table 1). This assemblage implies correlation with the lower Cambrian Lontova Horizon in the East European Platform (Ogurtsova, 1975; Kiryanov, 1987). Rare, relatively large (up to 40 μm) acanthomorphic acritarchs in the uppermost Kessyusa Formation were referred to the "Chuskun" assemblage by Kiryanov (1987) and interpreted as intermediate in age between the Lontova and overlying Talsy horizons of the East European Platform. *Skiagia ciliosa* and *Globosphaeridium cerinum* occur in the Erkeket Formation, suggesting Talsy or Vergale equivalence (Pyatiletov, 1989).

Table 1
Ediacaran metazoans identified from the Khatyspyt Formation along the Khorbusuonka River, northern Siberia

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Nemiana simplex Palij
Ediacaria flindersi Sprigg
Kullingia concentrica Foyn and Glaessner [a]
Hiemalora stellaris Fedonkin
Hiemalora pleiomorphus Vodanjuk
Glaessnerina longa (Glaessner and Wade)[a]
Aspidella costata Vodanjuk
Aspidella hatyspytia Vodanjuk
Anabylia improvisa Vodanjuk
Charnia masoni Ford
Ovatoscutum concentricum Glaessner and Wade
Paliella patelliformis Fedonkin
Khatyspytia grandis Fedonkin
Beltanelloides sp.
Cyclomedusa sp.

Data from Fedonkin (1987) and Vodanjuk (1989).
[a] Species identified provisionally.

Calcified microbes also occur in Kessyusa carbonates. Riding and Voronova (1984) reported *Renalcis*, *Botomaella*, *Botominella*, *Girvanella*, *Rothpletzia*, and *Korilophyton* in the upper part of the formation. According to V.A. Luchinina (in Meshkova et al., 1973), *Proaulopora glabra* and *Renalcis polymorphus* occur about 25 m above the base of the formation in outcrops along the Olenek River. This distribution is consistent with an extensive body of data which indicates that the abundance of calcified cyanobacteria increased greatly near the beginning of the Cambrian Period (e.g., Riding, 1991); in this context, it must be emphasized that *Girvanella*-like filaments and other calcified microorganisms have been reported from Neoproterozoic carbonates (Knoll et al., 1991; Turner et al., 1993).

Stromatolites are not abundant in this succession. *Boxonia grumulosa* and *Stratifera irregularis* occur in the Mastakh Formation; an interval in the upper Turkut Formation contains *Colleniella singularis*, *Paniscollenia emergens*, *Boxonia* (?)*grumulosa*, and *Stratifera irregularis* (Komar, 1966). According to Semikhatov (1976), this assemblage is characteristic of terminal Proterozoic (Vendian) carbonates. Consistent with the abundance of calcified cyanobacteria, thrombolites occur in upper Kessyusa carbonates. Vendotaenids are locally abundant in the carbonaceous limestones of the Khatyspyt Formation, but have not been described in detail.

3.2. Ediacaran fossils

Sokolov and Fedonkin (1984), Fedonkin (1985a, 1987), and Vodanjuk (1989; see also Yakschin and Vodanjuk, 1986) have identified fifteen taxa of Ediacara-type soft-bodied fossils in the Khatyspyt Formation (Table 2). Most come from a single meter-thick horizon near the top of the formation, although rare specimens occur throughout the formation (Fig. 3). The fossils stand out in positive relief on the tops and bottoms of mm-scale laminae of carbonaceous limestone. Preservation appears to be related, at least in part, to the early cementation of fossiliferous beds.

Identified taxa are widespread components of Ediacaran faunas. Most are simple "medusiform" remains; *Charnia masoni* is exceptional in terms of its complex, frond-like morphology.

Table 2
Skeletonized invertebrates identified from the Kessyusa Formation along the Olenek and Khorbusuonka rivers

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Upper Turkut Formation (terminal Proterozoic):
Cambrotubulus sp.

Lower Kessyusa Formation (lower Nemakit–Daldyn):
Anabarites trisulcatus, *Cambrotubulus decurvatus*, *Protohertzina anabarica*, sabelliditids

Upper Kessyusa Formation (upper Nemakit–Daldyn and lower Tommotian):
Cambrotubulus sibiricus, *Cambrotubulus conicus*, *Anabarites signatus*, *Anabarites tripartitus*, *Anabarites kelleri*, *Anabarites latus*, *Kugdatheca voluta*, *Aldanella attleborensis*, *Aldanella plana*, *Aldanella costata*, *Aldanella* sp. cf. *A. crassa*, *Hyolithellus tenuis*, *Hyolithellus vladimirovae*, *Spinulitheca rotunda*, *Spinulitheca billingsi*, *Anabarithellus hexasulcatus*, *Ladatheca blanda*, *Ladatheca dorsocava*, *Tiksitheca licis*, *Tiksitheca* sp., *Lobiochrea natella*, *Securiconus* sp., *Crosbitheca arcuaria*, *Barskovia hemisimmetrica*, *Latouchella sibirica*, *Bemella jacutica*, *Turcutheca crassecochlea*, *Turcutheca rugata*, *Khetatheca cotuiensis*, *Selindeochrea ternaria*, *Selindeochrea* sp., *Jakutiochrea tristicha*, *Fomitchella infundibuliformis*, *Halkiera costata*, *Halkiera sacciformis*, *Halkiera* sp., *Chancelloria* sp., *Siphogonuchites* sp., *Tommotia koslowzkii*, *Girospina araniformis*, *Syssospine irregularis*, *Heraultipegma* sp.

Data from Val'kov (1987), Missarzhevsky (1989a) and Bokova and Vasilyeva (1990).

3.3. Trace fossils

Ichnofossils are unknown from the Mastakh Formation and are rare in the Khatyspyt and Turkut foro-

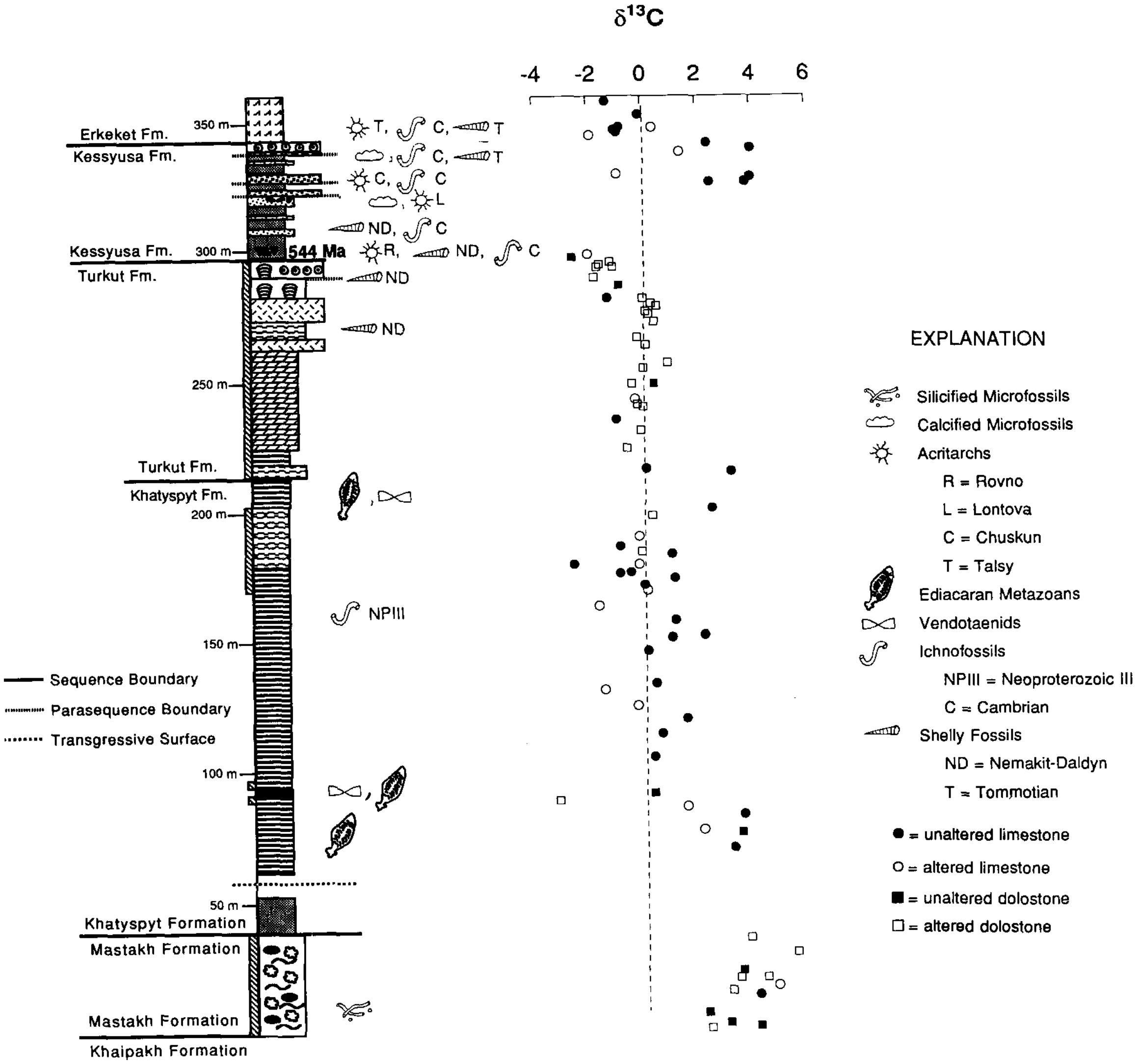

Fig. 3. Carbon-isotopic and biostratigraphic data for the terminal Proterozoic and basal Cambrian succession along the Khorbusuonka River, Olenek Uplift, Siberia.

mations. Fedonkin (1985b) recorded *Palaeopascichnus delicatus* and *Nenoxites* sp. in Khatyspyt carbonates, and we have additionally noted the presence of *Yelovichnus gracilis*; all occur in horizons containing Ediacaran metazoans on the East European Platform (Fedonkin, 1985b, 1992). Fedonkin (1985b) described *Gordia* sp., *Phycodes* sp., *Olenichnus irregularis*, *Planispiralichnus grandis*, and *Protospiralichnus circularis* from the lower Kessyusa Formation. Runnegar (1992) considers the last three to be pseudofossils, although all are accepted as true ichnofossils by Crimes and Droser (1992). Trace fossil density increases markedly through the Kessyusa Formation, with *Planolites* sp., and *Didymaulichnus* spp. occurring prominently in its upper part (Fedonkin, 1985b). A still higher density of mostly horizontal traces marks the Erkeket Formation. This is consistent with the observations of Droser and Bottjer (1988), who described increasing bioturbation from terminal Proterozoic through early Cambrian rocks.

3.4. Shelly invertebrates

The lowermost small shelly fossils in the Khorbusuonka River sections are *Cambrotubulus* sp. specimens collected a reported 30 m below the Turkut–Kessyusa boundary (Karlova, 1987; Khomentovsky and Karlova, 1993). A more diverse fauna that includes *Anabarites trisulcatus, Cambrotubulus decurvatus, Protoherzina anabarica*, and sabelliditids occurs just above the base of the Kessyusa Formation (Val'kov, 1987; Missarzhevsky, 1989a). In the biostratigraphic scheme of Missarzhevsky (1989a, b), this assemblage defines the *Anabarites trisulcatus* zone, the lowermost of his four zones in the basal Cambrian Nemakit–Daldyn (Manykai) Stage. Khomentovsky and Karlova (1993) also recognize an *A. trisulcatus* zone that constitutes the lower of their two Nemakit–Daldyn zones. (Khomentovsky and Karlova, 1993, place the base of the Cambrian at the beginning of the succeeding Tommotian Stage.) Rozanov and Zhuravlev (1992) recognize a broad *A. trisulcatus* zone and consider it to be coextensive with the Nemakit–Daldyn Stage (again, considered by these authors to lie beneath the Proterozoic–Cambrian boundary). Although the three concepts of the *Anabarites sulcatus* zone differ with respect to its duration, all three place its *beginning* at the base of the Nemakit–Daldyn Stage. Thus, small shelly fossils place the beginning of the Nemakit–Daldyn Stage at or just beneath the base of the Kessyusa Formation. All three sets of authors seem to regard the placement of the Proterozoic–Cambrian boundary as a matter of opinion, but it is not. Now that the Newfoundland GSSP for the boundary has been ratified, boundary placement in other sections is a question of *correlation.* In the following section, we summarize reasons why we concur with other workers who place the Proterozoic–Cambrian boundary at or near the base of the Nemakit–Daldyn Stage in the Olenek region (e.g., Narbonne et al., 1987; Landing et al., 1989; Landing, 1992). The upper, carbonate-bearing segment of the Kessyusa Formation contains a rich fauna (Val'kov, 1987; Missarzhevsky, 1989a; Bokova and Vasilyeva, 1990; Table 3). The thickness of rock eroded at the Kessyusa/Erkeket disconformity varies within the Olenek region (Missarzhevsky, 1989a). Where we collected along the Khorbusuonka River, the lower half of these upper Kessyusa beds includes fossils that mark the upper two zones (*Aldanella crassa* and *Anabarella plana*) of Missarzhevsky's Nemakit–Daldyn (Manykai) Stage (equivalent to the *Purella antiqua* zone of Khomentovsky and Karlova, 1993). Small shelly fossils in the uppermost 10–15 m of the formation mark the *Heraultipegma–Lapwothella tortuosa* zone (sensu Missarzhevsky, 1989a, b) of the lower Tommotian Stage (*Aldanocyathus sunnaginicus* zone of Khomentovsky and Karlova (1993). Missarzhevsky (1989a, b), Rozanov and Zhuravlev (1992) and Khomentovsky and Karlova (1993) concur in assigning the basal beds of the overlying Erkeket Formation to the upper Tommotian Stage (*Lapworthella bella* zone of Missarzhevsky, 1989a, b). Fallotaspid trilobites occur 14 m above the base of the Erkeket limestones.

4. Carbon isotope stratigraphy

4.1. Methods and sample evaluation

Methods used in this study are identical to those described in detail by Kaufman et al. (1991), Derry et al. (1992), and Kaufman and Knoll (1995) and are not reiterated here. Plane-light and cathodoluminescence (CL) petrographic summaries, as well as elemental and isotopic compositions of Khorbusuonka Group, Kessyusa, and Erkeket carbonates and total organic carbon (TOC) are reported in Table 3. Most samples consist of fine-grained microspar or dolomicrospar which are non- to moderately-luminescent under CL. A few of the very fine-grained, organic-rich mudstones of the Mastakh and Khatyspyt formations were not examined in thin section; for these samples only whole-rock (WR) values are reported.

Elemental and isotopic abundances of the Olenek carbonates vary widely. In this report, least-altered samples are considered to have $\mathrm{Mn/Sr} < 8$ and $\delta^{18}\mathrm{O} > -8$. In addition, for samples containing more than trace amounts of organic matter (> 0.5 mg C/g), a measured difference between carbonate and TOC $\delta^{13}\mathrm{C}$ values ($\Delta\delta$) of < 30 is taken to indicate the effects of organic diagenesis. This value of $\Delta\delta$ exceeds the value of 28.5 ± 2 empirically determined for other Neoproterozoic sections in which C-isotopic values for carbonate carbon and TOC covary smoothly (e.g., Knoll et al., 1986). Our choice reflects the observation that $\Delta\delta$ values in Olenek samples containing abundant and well preserved organic matter are 32–36. The signifi-

Table 3
C- and O-isotopic and elemental compositions of Khorbusuonka carbonates

number	sample[1]	height[2] (m)	formation	$\delta^{13}C$	$\delta^{18}O$	$\delta^{13}C_{org}$	$\Delta\delta$[3]	TOC (mg C/g)	Mn/Sr	Mg/Ca
				(‰, PDB)						
KG-30	MLDM	0.5	Mastakh	2.3	-5.5	-25.7	28.0	0.24	9.00	0.405
KG-38	MLDM	1.5	"	4.1	-5.2	-	-	-	3.97	0.526
KG-31	NLDM	2.5	"	3.0	-5.7	-	-	-	4.08	0.600
KG-32	NLDM	6.5	"	2.2	-6.2	-25.1	27.3	1.56	2.54	0.600
KG-40	NLM	13.5	"	4.1	-4.3	-25.6	29.7	0.66	3.80	0.159
KG-35	MLDM	15.0	"	3.1	-5.7	-	-	-	12.40	0.600
KG-41	MLM	17.0	"	4.8	-7.4	-	-	-	12.90	0.138
KG-36	MLDM	20.0	"	3.4	-4.8	-25.3	28.7	0.22	15.63	0.600
KG-42	MLDM	20.0	"	4.4	-2.0	-26.7	31.1	0.44	10.46	0.600
KG-43	WR/D	23.0	"	3.5	-4.9	-	-	-	7.75	0.600
KG-45	WR/D	30.0	"	5.5	-7.0	-	-	-	20.72	0.472
KG-47	WR/D	35.5	"	3.8	-7.3	-	-	-	23.28	0.600
KG-61	WR/L	70.0	Khatyspyt	3.2	-2.6	-32.6	35.8	1.38	0.04	0.010
KG-63	NLDM	76.0	"	3.5	-5.4	-31.0	34.5	1.01	1.81	0.600
KG-56	NLM	77.0	"	2.1	-9.2	-	-	-	1.11	0.022
KG-65	WR/L	83.0	"	3.6	-5.0	-31.4	35.0	2.04	0.05	0.005
KG-66	WR/L	86.0	"	1.5	-8.0	-29.4	30.9	1.01	0.66	0.125
KG-67	WR/D	88.0	"	-3.2	-4.0	-30.3	27.1	0.30	27.29	0.396
KG-68C	NLM	91.0	"	0.3	-5.0	-	-	-	0.80	0.414
KG-72	WR/L	105.0	"	0.3	-6.0	-31.9	32.2	2.79	0.18	0.080
KG-75	WR/L	114.0	"	0.6	-7.1	-32.0	32.5	1.06	0.08	0.050
KG-76	WR/L	120.0	"	1.5	-7.1	-32.7	34.2	11.07	0.02	0.006
KG-78	MLM	125.0	"	-0.3	-7.4	-27.3	27.0	1.23	0.24	0.007
KG-79	NLM	131.0	"	-1.5	-6.3	-29.0	27.5	0.97	0.38	0.003
KG-80	WR/L	133.5	"	0.4	-6.1	-30.7	31.1	2.70	0.09	0.097
KG-85	WR/L	146.0	"	0.1	-7.1	-32.5	32.6	1.57	0.06	0.043
KG-70	WR/L	151.0	"	1.0	-7.2	-32.8	33.8	10.05	0.06	0.008
KG-87	WR/L	152.0	"	2.2	-5.6	-33.9	36.1	3.30	0.02	0.006
KG-88	WR/L	157.5	"	1.1	-3.9	-32.4	33.5	1.38	0.10	0.012
KG-96	NLM	163.0	"	-1.7	-0.5	-29.7	28.0	0.48	0.13	0.044
KG-89	NLM	169.0	"	0.1	-1.7	-29.0	29.1	0.64	0.26	0.185
KG-98A	NLM	171.0	"	-0.0	-7.6	-34.6	34.6	0.98	0.17	0.004
KG-90	NLM	173.5	"	1.1	-3.2	-31.9	33.0	0.81	0.10	0.012
KG-91	NLM	175.5	"	-0.9	-3.2	-33.2	32.3	1.75	0.04	0.014
KG-99	MLM	176.0	"	-0.5	-5.1	-	-	-	0.07	0.031
KG-92	NLM	179.0	"	-2.6	-1.6	-	-	-	0.23	0.101
KG-100	NLM	179.0	"	-0.2	-9.8	-32.3	32.1	0.51	0.10	0.021
KG-93	NLM	183.0	"	1.0	-2.3	-	-	-	0.05	0.021
KG-101	MLDM	184.0	"	-0.1	-10.7	-30.1	30.0	0.92	2.80	0.584
KG-94	WR/L	186.0	"	-0.9	-4.0	-	-	-	0.03	0.008
KG-103	MLM	190.0	"	-0.2	-10.0	-28.5	28.3	0.40	0.21	0.017
KG-105	MLDM	198.0	"	0.3	-8.7	-30.0	30.3	0.39	3.01	0.600
KG-106	MLM	201.0	"	2.5	-3.0	-	-	-	0.65	0.160
KG-108	NLM	215.0	Turkut	3.2	-7.9	-30.0	33.2	3.44	0.06	0.019
KG-158	NLM	216.0	"	0.1	-3.4	-34.3	34.4	n.d.	0.06	0.017
KG-160	MLDM	224.0	"	-0.6	-10.3	-35.2	34.6	1.91	1.21	0.573
KG-162	MLDM	231.0	"	-0.1	-9.8	-	-	-	2.21	0.238
KG-233	NLM	235.0	"	-1.0	-6.4	-26.2	25.2	0.20	2.93	0.007
KG-165	MLDM	240.0	"	0.0	-11.2	-	-	-	3.29	0.600
KG-113	MLDM	241.0	"	-0.2	-8.5	-32.7	32.5	3.49	0.89	0.200
KG-166	NLM	243.0	"	-0.3	-8.0	-	-	-	2.00	0.143

Table 3 (continued)

number	sample[1]	height[2] (m)	formation	$\delta^{13}C$	$\delta^{18}O$	$\delta^{13}C_{org}$	$\Delta\delta$[3]	TOC (mg C/g)	Mn/Sr	Mg/Ca
				(‰, PDB)						
KG-115	MLDM	249.0	Turkut	0.4	-7.1	-33.1	33.5	0.47	0.93	0.361
KG-168	MLDM	249.0	"	-0.4	-8.2	-	-	-	2.27	0.240
KG-170B	NLDM	255.0	"	-0.0	-11.6	-28.6	28.6	n.d.	1.72	0.591
KG-117	NLDM	257.0	"	0.9	-9.5	-32.8	33.7	0.20	3.00	0.269
KG-171	MLDM	258.0	"	0.6	-8.8	-26.8	27.4	0.07	2.73	0.050
KG-173	MLDM	264.0	"	0.1	-10.2	-	-	-	2.60	0.600
KG-174	NLDM	267.0	"	-0.2	-8.9	-	-	-	2.40	0.208
KG-176	NLDM	273.0	"	0.4	-10.1	-	-	-	2.87	0.548
KG-177	NLDM/P	276.0	"	0.2	-12.4	-	-	-	3.16	0.574
KG-126	NLDM	277.0	"	0.1	-9.0	-	-	-	2.56	0.383
KG-178	NLDM	279.0	"	0.5	-9.5	-33.1	33.6	0.52	3.40	0.600
KG-122	NLDM/P	280.0	"	0.3	-8.3	-	-	-	1.81	0.205
KG-128	NLM	282.0	"	-1.3	-3.2	-29.2	27.9	0.27	6.16	0.181
KG-179	MLDM	282.0	"	0.0	-9.4	-32.4	32.4	0.98	2.93	0.600
KG-123	NLDM	287.0	"	-0.9	-2.8	-	-	-	4.18	0.337
KG-188	MLDM	290.0	"	-1.8	-3.0	-31.4	29.6	0.32	22.39	0.288
KG-183	NLDM	294.0	"	-1.1	-4.5	-	-	-	8.91	0.600
KG-189	MLDM	294.0	"	-1.7	-3.2	-	-	-	16.91	0.277
KG-133	MLDM	295.0	"	-1.6	-6.3	-34.8	33.2	n.d.	22.11	0.595
KG-190	MLDM	296.0	"	-1.2	-5.3	-	-	-	21.51	0.230
KG-185	NLDM	298.0	"	-2.6	-4.9	-	-	-	5.67	0.600
KG-194	MLDM	299.0	"	-2.0	-7.8	-	-	-	20.20	0.005
KG-187	NLM	327.0	Kessyuse	3.8	-6.1	-26.4	30.2	0.62	5.77	0.010
KG-227	NLM	327.0	"	2.5	-7.4	-	-	-	6.81	0.021
KG-140	MLM	329.0	"	4.0	-7.5	-	-	-	5.01	0.012
KG-228	MLM	330.0	"	-0.9	-9.9	-27.1	26.2	0.20	22.45	0.009
KG-229	MLM	338.5	"	1.4	-6.2	-	-	-	13.71	0.003
KG-145	NLM	340.0	"	4.0	-5.7	-23.7	27.7	0.27	0.82	0.014
KG-231	MLM	342.0	"	2.4	-7.9	-25.5	27.9	0.44	6.66	0.020
KG-147	MLM	345.0	Erkeket	-1.9	-8.4	-25.3	23.4	n.d.	4.01	0.014
KG-149	NLM	346.2	"	-0.9	-7.0	-	-	-	3.86	0.009
KG-234	NLM	347.0	"	-1.0	-6.7	-	-	-	2.47	0.004
KG-151	NLM	348.0	"	-0.8	-6.6	-	-	-	3.41	0.017
KG-224	MLM	348.0	"	0.4	-9.7	-24.2	24.6	0.19	2.71	0.022
KG-225	MLM	353.0	"	-0.1	-7.8	-23.2	23.1	0.30	1.79	0.040
KG-226	NLM	358.0	"	-1.3	-6.6	-25.7	24.4	0.26	2.13	0.016

[1] Sample abbreviations: WR = whole rock; NLM = non-luminescent microspar; NLDM = non-luminescent dolomicrospar; MLM = moderately luminescent microspar; MLDM = moderately luminescent dolomicrospar; /L = limestone; /D = dolostone; /P = pisoids.

[2] Height measured above base of Mastakh Formation.

[3] $\Delta\delta = \delta^{13}C - \delta^{13}C_{org}$.

cance of these high $\Delta\delta$ values is at present unresolved.

From the Mn/Sr determinations it is apparent that a number of carbonates in the Mastakh, Turkut, and Kessyusa formations have been altered during dolomitization. However, few of these samples have excessively depleted $\delta^{18}O$, as might have been expected. Dolomites in Proterozoic successions commonly record more positive $\delta^{18}O$ values than do interbedded limestones, presumably reflecting early dolomite formation in fluids isotopically similar to seawater followed by calcium carbonate neomorphism at depth (Tucker, 1983). Some sucrosic dolomites in the

Turkut Formation exhibit the converse pattern: markedly depleted $\delta^{18}O$ and low Mn/Sr. This is consistent with outcrop and petrographic evidence for stratigraphically restricted hydrothermal fluid flow through deeply buried sediments.

Perhaps the least-altered samples come from the Khatyspyt Formation. Samples commonly exhibit minimally altered $\delta^{18}O$, high $\Delta\delta$ and extremely low Mn/Sr. Samples whose C-isotopic composition falls significantly below the curve for all Khatyspyt carbonates are clearly altered, as indicated by reduced $\Delta\delta$ and, in the most strongly anomalous sample, KG91-67, by an Mn/Sr value of 27.3. We note that during preliminary investigations, we examined both least-altered and more obviously altered petrographic phases of individual Khatyspyt samples. In nearly every case, the $\delta^{13}C$ values of the more altered samples were depleted by several permil, presumably due to the incorporation of ^{13}C-depleted carbon from remineralized organic matter into diagenetic carbonates.

Despite the presence of some altered carbonates, the consistency of $\delta^{13}C$ values between most such samples and stratigraphically contiguous samples that appear to be little altered suggests that measured C-isotopic compositions can be used to define secular trends in $\delta^{13}C$ throughout the succession.

4.2. Stratigraphic results

From the base of the Mastakh Formation $\delta^{13}C$ values rise from values around +2 to a high of +5.5 near the top of the unit. Above a short interval in which there are no carbonate rocks, basal Khatyspyt limestones record values of $\delta^{13}C$ around +3.5, but higher in the unit values range widely between −4.2 and +2.5. As noted in the previous section, $\Delta\delta$ values strongly suggest that negative $\delta^{13}C$ values in these bituminous limestones and dolostones reflect the early diagenetic oxidation of organic matter and the subsequent precipitation of neoformed carbonate (cf. Kaufman et al., 1992). In light of these observations, we suggest that primary $\delta^{13}C$ values for most Khatyspyt limestones lay between 0 and +2. The exception is several values in the 176–186 m interval of the composite section (samples KG91-91 to KG91-94; Fig. 3; Table 3). These samples pass all screening tests for alteration, but $\delta^{13}C_{TOC}$ data for the most negative sample, KG91-92, are not available. Thus, while these samples may define a stratigraphically useful isotopic excursion, confidence in this interpretation is limited by evidence for hydrothermal fluid flow in ambient strata. Most Turkut dolostones record $\delta^{13}C$ abundances near 0 until near the top of the formation where carbonates trend to more isotopically depleted values near −3, a shift reflected in $^{13}C_{TOC}$ data. A significant excursion to positive $\delta^{13}C$ values near +4 is noted in upper Kessyusa samples, following which lower Erkeket values return to slightly negative $\delta^{13}C$ values.

5. Integration of litho-, bio- and chemostratigraphy

Khorbusuonka fossils and C-isotopic compositions each display stratigraphic distributions similar to those recorded elsewhere. More to the point, the stratigraphic covariation *between* the two data sets closely resembles those established in other terminal Proterozoic and basal Cambrian sections.

In sections containing Varanger glaciogenic rocks, immediately post-glacial carbonates exhibit significant depletion in ^{13}C (see summary in Kaufman and Knoll, 1995). The absence of negative $\delta^{13}C$ values in Mastakh carbonates, coupled with the stratigraphic increase of $\delta^{13}C$ values through the unit from +2.2 to +5.5, indicates that the Mastakh transgression substantially postdated the melting of Varanger ice sheets. On the Siberian Platform, Yudomian strata commonly rest unconformably on older Riphean rocks. Although there has been a tendency to correlate transgressive packages across the platform, isotopic chemostratigraphy demonstrates that basal Yudomian strata can vary significantly in age. For example, along the Kotuikan River on the western flank of the Anabar Uplift, Yudomian carbonates of the Staraya Rechka Formation that rest unconformably on older Riphean rocks correlate *only* with uppermost Turkut rocks in the Olenek region (Pokrovsky and Missarzhevsky, 1993).

The positive $\delta^{13}C$ values in the Mastakh Formation are similar to values reported in the Zaris Formation of the Nama Group in Namibia (Kaufman et al., 1991), the Doushantou Formation in South China (Lambert et al., 1987), the uppermost Sheepbed Formation in the Mackenzie Mountains of northwestern Canada (Narbonne et al., 1994), the middle Pertatataka Formation of central Australia (Walter et al., 1995), and the lower

Rodda Beds and upper Wonoka Formation in South Australia (Jenkins et al., 1992; Walter et al., 1995). The Namibian and northwestern Canadian strata contain the oldest known diverse Ediacaran metazoans. Metazoan trace fossils and possible body impressions also occur in the South Australian horizons thought to correlate chemostratigraphically with the Mastakh Formation (Jenkins, 1995). Thus, while Mastakh facies are unlikely candidates for Ediacaran fossil preservation, it is not surprising that occasional animal traces and body fossils occur only slightly higher, in Khatyspyt horizons that lie well below the principal fossiliferous beds near the top of the formation.

A sequence boundary marked by coarse fluviatile sandstones and finer-grained siliciclastic lithologies separates the Mastakh and Khatyspyt formations, but the lowermost 15 m of Khatyspyt limestones continue the trend of positive $\delta^{13}C$ values. Above the 85 m mark (composite section), there is a succession in which $\delta^{13}C$ values range between 0 and +1, followed by an interval of more variable $\delta^{13}C$. While we cannot eliminate the possibility that this variation includes an element of true secular change in seawater composition, most and possibly all of the variation may be explained by the diagenetic and hydrothermal processes noted above. Equivalent horizons in the Kharulakh Mountains do not exhibit the variation recorded in the Khorbusuonka sections (S. Pelechaty, pers. commun., 1993). Thus, we cautiously accept that upper Khatyspyt carbonates with $\delta^{13}C$ values of +1 to +2 most closely approximate depositional compositions. This interval includes the principal Ediacaran fossil-bearing horizon and extends to the top of the formation (as understood in this paper). The coincidence of diverse Ediacaran metazoans and moderately positive C-isotopic values is also seen in Australia (Jenkins et al., 1992), northwestern Canada (Narbonne et al., 1994), South China (Lambert et al., 1987), and Namibia (Kaufman et al., 1991). The interval of limited C-isotopic variation continues through most of the main Turkut succession, with $\delta^{13}C$ values lying near 0. A comparable stratigraphic trend is seen in the Mackenzie Mountains, where carbonates of the Risky Formation carbonates continue the trend of slightly positive $\delta^{13}C$ values seen in the Ediacara-fossil-bearing Blueflower Formation (Narbonne et al., 1994). The Olenek and Mackenzie Mountain successions both record the terminal Proterozoic excursions from these moderate values to markedly negative values in the upper Turkut and Ingta formations, respectively.

The markedly positive $\delta^{13}C$ values recorded in upper Kessyusa carbonates imply a strong isotopic excursion through the interval of the Kessyusa Formation that contains Nemakit–Daldyn fossils. The return to negative $\delta^{13}C$ values approximates the Nemakit–Daldyn/Tommotian boundary recognized by Missarzhevsky (1989a, b) on the basis of fossil invertebrates. Similar stratigraphic variation characterizes the type Nemakit–Daldyn succession and overlying strata in the western Anabar region, northern Siberia (A.J. Kaufman et al., unpubl. data).

Proterozoic–Cambrian boundary secessions in southern Siberia also contain similar C-isotope excursions (Magaritz et al., 1986). Here, small shelly fossils place the base of the Nemakit–Daldyn Stage just above the latest Proterozoic negative excursion, at a point midway in the section where C-isotopic compositions rise to values as high as +3 (Magaritz et al., 1986; Kirschvink et al., 1991; Khomentovsky and Karlova, 1993; Brasier et al., 1993). The initial Tommotian boundary is marked by a rapid descent of the C-isotope curve from its peak to moderately negative values that characterize much of the Tommotian interval. In Khorbusuonka sections, the Tommotian interval is thin, contains abundant siliciclastic lithologies in its lower part, and includes a major sequence boundary. Nonetheless, the stratigraphic pattern of C-isotopic compositions for this interval permits correlation to southern Siberia.

In the Mackenzie Mountains, *Phycodes pedum* and the lowermost reported, small shelly fossils occur just above a marked negative C-isotopic excursion in the lower Ingta Formation (Narbonne et al., 1994; Narbonne and Aitken, 1995). A Proterozoic/Cambrian boundary excursion from negative to positive (up to +6) values also occurs in the Anti-Atlas of Morocco (Magaritz et al., 1991), although paleontological control is less secure in this section.

The coincidence of chemostratigraphic and paleontological data in the Mackenzie Mountains is particularly helpful in the attempt to locate the Proterozoic–Cambrian boundary in the Olenek region. Direct correlation with the GSSP in Newfoundland is difficult because the type section has yielded no acritarchs or small shelly fossils near the boundary point (Narbonne et al., 1987; Landing, 1992), neither does the type section contain bedded carbonates in the boundary interval. Strauss et al. (1992) reported C-isotopic data for organic carbon preserved in the stratotype section, but most reported samples contain very low TOC abundances, and the data are consequently difficult to interpret in terms of seawater isotopic composition. The principal criterion used in placing the GSSP was the first appearance of a distinctive ichnofossil assemblage characterized by *Phycodes pedum*.

Phycodes sp. has been reported from the lower Kessyusa Formation, suggesting that the GSSP should correlate with a horizon at or near the Turkut–Kessyusa boundary. Small shelly fossils of Cambrian aspect also imply that the erathem boundary lies with upper Turkut or basal Kessyusa strata, but the absence of comparable fossils in GSSP boundary beds leaves uncertain the stratigraphic relationship between the first appearance of small shelly fossils and the Proterozoic–Cambrian boundary.

Correlation between Newfoundland and northwestern Canada is possible using *Phycodes pedum* and other trace fossils. From the Mackenzie Mountains, one can continue the boundary correlation to Siberia using C-isotopic data. As noted above, isotopic data support correlation of the initial boundaries of the Nemakit–Daldyn and Tommotion stages in their stratotypes with the base of the Kessyusa and a point in the uppermost Kessyusa or basal Erkeket Formation, respectively. In the Mackenzie Mountains, the first appearances of both *Phycodes pedum* and small shelly fossils occur within a C-isotope excursion inferred to be the one that marks the beginning of the Nemakit–Daldyn Stage. As noted above, Narbonne et al. (1987), Landing et al. (1989) and Landing (1992) and have correlated the Nemakit–Daldyn Stage with the lower part of the basal Cambrian Placentian Stage in North America.

From this exercise flow two conclusions: (1) the base of the Cambrian System in northeastern Siberia lies at or near the transgressive surface that marks the beginning of the Kessyusa Formation—an interpretation consistent with the placement of the initial Nemakit–Daldyn boundary in its type section some 1200 km to the west; and (2) even though the Proterozoic–Cambrian boundary has been fixed at a point stratigraphically below the beginning of the Tommotian Stage, there are *still* pre-Cambrian small shelly fossils of Cambrian aspect. *Cambrotubulus* specimens 30 m below the top of the Turkut Formation predate the initial Nemakit–Daldyn boundary in its type area and correlate chemostratigraphically to strata beneath the interpreted base of the Cambrian in northwestern North America.

6. Conclusions

Relative to sections in Newfoundland, northwestern Canada, and South Australia, the terminal Proterozoic succession in the Olenek region is thin. Despite this, the northeastern Siberian section displays nearly all of the paleontological and C-isotopic events recorded in the much thicker sections. For the most part, bio- and chemostratigraphic events recorded in the Olenek sections do not coincide with sequence boundaries, suggesting that sequence development does not exert principal control on the stratigraphic distribution of fossils and isotopic excursions.

Olenek data contribute to a rapidly growing set of data which demonstrate that paleontology and isotopic chemostratigraphy together provide a useful framework for the correlation of terminal Proterozoic strata. The Olenek section is not a good candidate for the initial boundary GSSP of the terminal Proterozoic (Neoproterozoic III of Plumb, 1991) period. The section contains neither a Varanger tillite nor the distinctive negative C-isotopic excursion that characterizes immediately post-Varanger carbonates; neither has it yielded the distinctive post-Varanger acritarch biota found elsewhere (e.g., Zang and Walter, 1992).

On the other hand, its combination of a well-defined sedimentological framework, abundant small shelly fossils and trace fossils, acritarchs, carbonates that yield reliable chemostratigraphic data, and radiometric dates makes the Olenek Uplift a significant reference region for studies of the *terminal* boundary of this period (the Proterozoic–Cambrian boundary; see also Khomentovsky and Karlova, 1993).

Olenek data additionally support subdivision of the post-Varanger/pre-Cambrian interval into three units that can be correlated among basins: a basal interval marked by a distinctive acritarch flora and a C-isotopic excursion from negative to markedly positive values; an interval characterized by simple acritarchs, diverse Ediacaran faunas, and $\delta^{13}C$ values of $\sim +1$ to $+2$; and a unit of sparse medusoids and other soft-bodied metazoans, distinctive trace fossils, simple acritarchs, and a $\delta^{13}C$ plateau near 0 that ends in a negative excursion. These units coincide broadly (but not precisely) with the Volhyn, Redkino, and Kotlin intervals of the East European Platform (Sokolov and Fedonkin, 1984; Sokolov, 1985; Knoll and Walter, 1992).

As the stratigraphic correlation of terminal Proterozoic successions improves, so will our understanding of the profound biological and geological events that mark the transition to the Phanerozoic world. Terminal Proterozoic sections in Siberia, and particularly in the Olenek region, have much to offer in this endeavor.

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