

Temporal constraints on the Paleoproterozoic Lomagundi-Jatuli carbon isotopic event

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ABSTRACT

The Paleoproterozoic Lomagundi-Jatuli positive $\delta^{13}\text{C}$ excursion in sedimentary carbonates represents an event whose magnitude and duration is unique in Earth history, although precise absolute chronology of this event remains poorly constrained. In northeastern Fennoscandia, an ~1300-m-thick sedimentary-volcanic succession of the Pechenga Greenstone Belt records decline of this isotopic excursion. Zircons from sedimentary rocks that occur within the decline have yielded $^{207}\text{Pb}/^{206}\text{Pb}$ dates at 2058 ± 2 Ma (± 6 Ma including U decay constant uncertainties) and provide the first maximum age constraint on the termination of the Lomagundi-Jatuli event. Combined with existing constraints, these data indicate an ~140 m.y. interval characterized by ^{13}C -rich carbonate accumulation.

Keywords: Fennoscandia, Paleoproterozoic, dolostone, carbon, isotopes, age.

INTRODUCTION

One of the most remarkable events known from the Archean-Paleoproterozoic transition (ca. 2500–2000 Ma) is the largest-magnitude (e.g., Karhu and Holland, 1996) positive excursion of $\delta^{13}\text{C}$ in sedimentary carbonates known in Earth history. The excursion has been termed the Lomagundi-Jatuli event after the first-studied ^{13}C -rich formations (Galimov et al., 1968; Schildowski et al., 1976). The Lomagundi-Jatuli event appears to be an integral part of a series of other prominent environmental events that occurred during the Archean-Paleoproterozoic transition (Fig. 1); however, the interrelationship of such events remains intriguing and unresolved. Given existing constraints it is evident that the Lomagundi-Jatuli event developed on all continents after the “Huronian-age” glaciation (Kopp et al., 2005; Melezhik et al., 2005b) and post-dates the initial accumulation of atmospheric O_2 (Bekker et al., 2004; Hannah et al., 2004). It is also evident that the global appearance of terrestrial “red beds” and abundant Ca sulfates coincides with the Lomagundi-Jatuli event (Melezhik et al., 2005c) and that the Shunga event, an unprecedented accumulation of C_{org} -rich sediments, occurred in its aftermath (Fig. 1).

Originally, the causative mechanism responsible for the Lomagundi-Jatuli event was linked to increased carbon burial (Baker and Fallick, 1989a, 1989b). Melezhik and Fallick (1996) recognized the Lomagundi-Jatuli event as a paradox because of the absence of geological

evidence for enhanced carbon burial rate. At present, there is no consensus on the causative mechanism(s) responsible for the Lomagundi-Jatuli event. Several sets of authors have continued with the traditional model (Karhu and Holland, 1996; Bekker et al., 2001, 2003a, 2003b) inferring that the necessary geological record has not been preserved. Other models include fermentative diagenesis (Yudovich et al., 1991), and decoupling of the P and C cycles in a redox-stratified ocean (Aharon, 2005). The diagenetic model recently has been revisited and developed further by Hayes and Waldbauer (2006), who linked the Lomagundi-Jatuli event to the development of methanogenic diagenesis in deeper levels of the sediment as the response to O_2 increasing. The termination of the event was assigned to either (1) decreasing supply of organic material to methanogens, or (2) destabilization of the tight association between stromatolitic producers and methanogenic consumers.

Whether this large-magnitude positive $\delta^{13}\text{C}$ shift was contemporaneous in various basins, and the possible role of local factors in amplification of a global signal (+28‰; Bekker et al., 2003a), remain unresolved (Melezhik et al., 1999; Bekker et al., 2003a, 2003b; Shields and Veizer, 2002). Neither the onset nor the termination of the Lomagundi-Jatuli event has been reliably constrained. Data from the Fennoscandian Shield indicate that the accumulation of ^{13}C -rich carbonates was under way by ca. 2200 Ma and continued until ca. 2060 Ma (Karhu, 1993). In addition, Buick et al. (1998) reported a positive isotopic shift at ca. 2330 Ma (Hannah et al., 2004) from the Transvaal Supergroup. However, it remains unclear whether this indicates an ear-

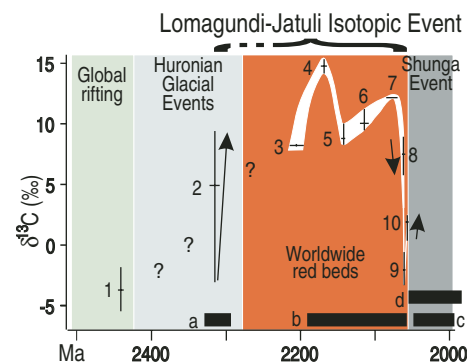


Figure 1. $\delta^{13}\text{C}$ variations in Early Paleoproterozoic sedimentary carbonates based on high-precision U-Pb and Re-Os age data (Table DR1). Vertical bars represent $\delta^{13}\text{C}$ variation, and horizontal bars indicate uncertainty in age. 1—Seidorechka Sedimentary Formation; 2—Duitschland Formation; 3—Sericite Schist Formation; 4—Dunphy and Portage Formations; 5—Pistolet Subgroup; 6—Lower Vistola Formation; 7—Siltstone Formation; 8—Kuetsjärvi Sedimentary Formation; 9—Houtenbek Formation; 10—Kolasjoki Sedimentary Formation. a—end of mass-independent fractionation of sulfur isotopes; b—abundant marine sulfates; c—first sedimentary phosphorites; d—breakup of Archean supercontinent. Arrows show stratigraphic $\delta^{13}\text{C}$ trend within the formation.

lier initiation of the Lomagundi-Jatuli event or represents a separate isotopic shift (Bekker et al., 2001). Furthermore, in the Pechenga Greenstone Belt, Melezhik et al. (1999) defined two apparently separate pulses of ^{13}C enrichment divided by return to 0‰. The overall absolute chronology of the Lomagundi-Jatuli event is still to emerge (Lindsay and Brasier, 2002; Hannah et al., 2004). A general lack of robust age constraints, which could highlight internal structure of this long-lasting interval characterized by ^{13}C -rich sediment accumulation, hampers our understanding of the role of the Lomagundi-Jatuli event in the Archean-Paleoproterozoic transition. Here we report U-Pb zircon ages for a volcanic formation that separates isotopically heavy and isotopically normal sedimentary carbonates from the Pechenga Greenstone Belt in northwestern Russia and thus constrain the termination of the Lomagundi-Jatuli event.

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STRATIGRAPHY AND AGE OF THE PECHENGA GREENSTONE BELT

The Pechenga Greenstone Belt is part of a larger belt distributed over a distance of 1000 km in northeastern Fennoscandia (Fig. 2A). The Pechenga Greenstone Belt rocks comprise the North Pechenga and South Pechenga Groups. The North Pechenga Group is composed of four paired sedimentary-volcanic cycles, each separated either by a nondepositional unconformity or by a fault (Figs. 2B and 2C). Each cycle begins with sedimentary rocks and ends with a thicker unit of volcanic rocks (for formational names see Figs. 2B and 2C). The cumulative stratigraphic thicknesses of the sedimentary and volcanic rocks are 1500 m and 6500 m, respectively. The rocks have undergone metamorphism from prehnite-pumpellyite to amphibolite facies. The maximum age of the North Pechenga Group is constrained by the presence of gabbro-norite pebbles in the basal Neverskruck conglomerates, which unconformably overlie the Mount Generalskaya gabbro-norite intrusion (Fig. 2C) dated to 2505 ± 1.6 Ma (Amelin et al., 1995). The minimum age of the group is 1970 ± 5 Ma (U-Pb zircon; Hanski et al., 1990), obtained from rhyolitic tuffs of the Pilgajärvi Volcanic Formation (Fig. 2C). The available ages of the Ahmalahti Formation and Kuetsjärvi Volcanic Formation are based on whole-rock Rb-Sr isotopic data (2324 ± 28 Ma and 2215 ± 54 Ma, respectively; Balashov, 1995).

CARBON ISOTOPE STRATIGRAPHY

The Kuetsjärvi Sedimentary Formation (Figs. 2B, 2C, and 3) records the Lomagundi-Jatuli event in the Pechenga Greenstone Belt (Karhu, 1993). Previous studies have shown that the Kuetsjärvi Sedimentary Formation carbonates display invariably high $\delta^{13}\text{C}$ values ($+5.8\text{‰}$ to $+8.9\text{‰}$, $n = 95$). Stratigraphic variations of the least-altered $\delta^{13}\text{C}$ values are expressed by significant $\delta^{13}\text{C}$ lowering in the uppermost part of the succession (Fig. 3), associated with invasion of seawater (Melezhik et al., 2005a).

The next stratigraphic unit containing carbonate rocks is the Kolasjoki Sedimentary Formation (Fig. 2C), which is separated from the ^{13}C -rich dolostones by the 900 m thick Kuetsjärvi Volcanic Formation (Fig. 3). The sedimentary formation is 80 m thick and fills paleovalleys incised into the underlying alkaline felsic volcanites, which are in places affected by incipient paleoweathering. The basal part of the formation consists of fluvial volcanoclastic siltstone-sandstone cycles and deltaic volcanoclastic siltstones followed by deltaic gritstones and pebble- and matrix-supported conglomerates. Deltaic sandstones, containing fragments of Kuetsjärvi Volcanic Formation volcanites, constitute the middle part. The upper part of the formation is a 27-m-thick micritic and stromatolitic dolostone unit with a thin jasper bed at the base. A black shale unit caps the succession.

Analyses of drill core (drill hole IX) material shows that the Kolasjoki Sedimentary Formation carbonates are impure, rich in clastic quartz

($<0.1\text{--}36$ wt% SiO_2), variably de-dolomitized dolostones ($\text{Mg}/\text{Ca} = 0.36\text{--}0.58$, $n = 30$). The rocks have moderate concentrations of Mn and Sr, and show rather homogeneous $\delta^{13}\text{C}$ ($+1.9 \pm 0.3\text{‰}$) and $\delta^{18}\text{O}$ ($18.4 \pm 0.7\text{‰}$) values (Appendix DR1 and Table DR2 in the GSA Data Repository¹). $\delta^{13}\text{C}$ irregularly increases upwards from $+1.2\text{‰}$ to $+2.4\text{‰}$ (Fig. 3) and exhibits a significant correlation with stratigraphic height ($r = +0.74$, $n = 28$, $>99.9\%$). $\delta^{18}\text{O}$ values show random stratigraphic distribution, implying a certain degree of postdepositional resetting. There is no correlation between $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$, nor between Mn/Sr ratios and $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$. The overall $\delta^{18}\text{O}$ data fall within the range of "normal", low-temperature greenschist-facies-grade, Paleoproterozoic dolostones in general, and within the range of ^{13}C -rich Kuetsjärvi dolostones in particular (Melezhik et al., 2005a). Thus, the Kolasjoki Sedimentary Formation dolostones very likely retain $\delta^{13}\text{C}$ values close to syndepositional and contrast with those from the Kuetsjärvi Sedimentary Formation by being much less enriched in ^{13}C (Table DR2). Combined, these units record the decline and termination of the Lomagundi-Jatuli event (Fig. 3).

SAMPLING FOR GEOCHRONOLOGICAL DATING

The main target for dating was the Kuetsjärvi Volcanic Formation separating ^{13}C -rich and isotopically normal carbonates (Fig. 3). Samples were collected from volcanoclastic, fluvial, and deltaic sedimentary rocks of the Kolasjoki Sedimentary Formation (overlying the Kuetsjärvi Volcanic Formation) since erosion and reworking of volcanic products would have concentrated any comagmatic zircon in the overlying sedimentary succession, and a "young" zircon population (in the sense of the existing geochronological constraints) would provide maximum age constraints on the stratigraphic interval sampled as well as an inferred age constraint upon the preceding/underlying Kuetsjärvi Volcanic Formation. In order to constrain a maximum age for the Kuetsjärvi Sedimentary Formation, material was collected from the sedimentary formation overlying the Ahmalahti basaltic andesites (Fig. 3).

The sample of the Kuetsjärvi Sedimentary Formation was obtained from an arkosic sandstone. In the Kuetsjärvi Volcanic Formation, the samples were collected from intraformational, matrix-supported and clast-supported volcani-

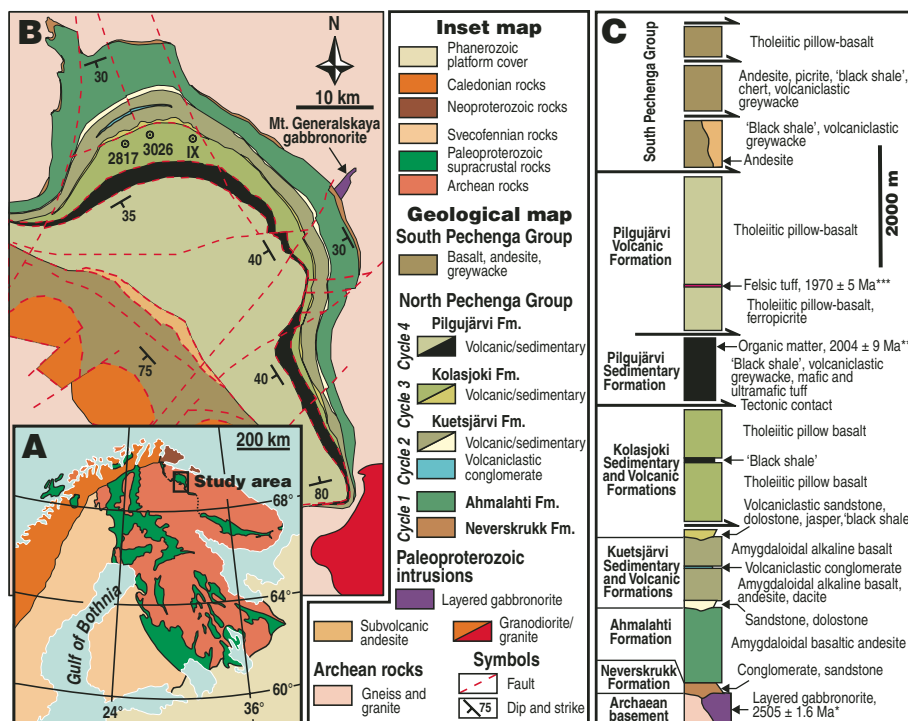


Figure 2. A: Geographic position of the study area. B: Simplified geological map of the Pechenga Greenstone Belt. C: Lithostratigraphic column of the Petsamo Supergroup. *—age data from Amelin et al. (1995); **—age data from Hannah et al. (2006); ***—age data from Hanski et al. (1990).

¹GSA Data Repository item 2007162, Appendix DR1 (analytical procedures), Table DR1 (compilation of available $\delta^{13}\text{C}$ and radiometric ages), Table DR2 (C and O isotope data), Tables DR3–DR4 (U-Th-Pb SIMS and U-Pb ID-TIMS data), and Figure DR1 (histogram plots of $^{207}\text{Pb}/^{206}\text{Pb}$ concordant dates), is available online at www.geosociety.org/pubs/ft2007.htm, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301, USA.

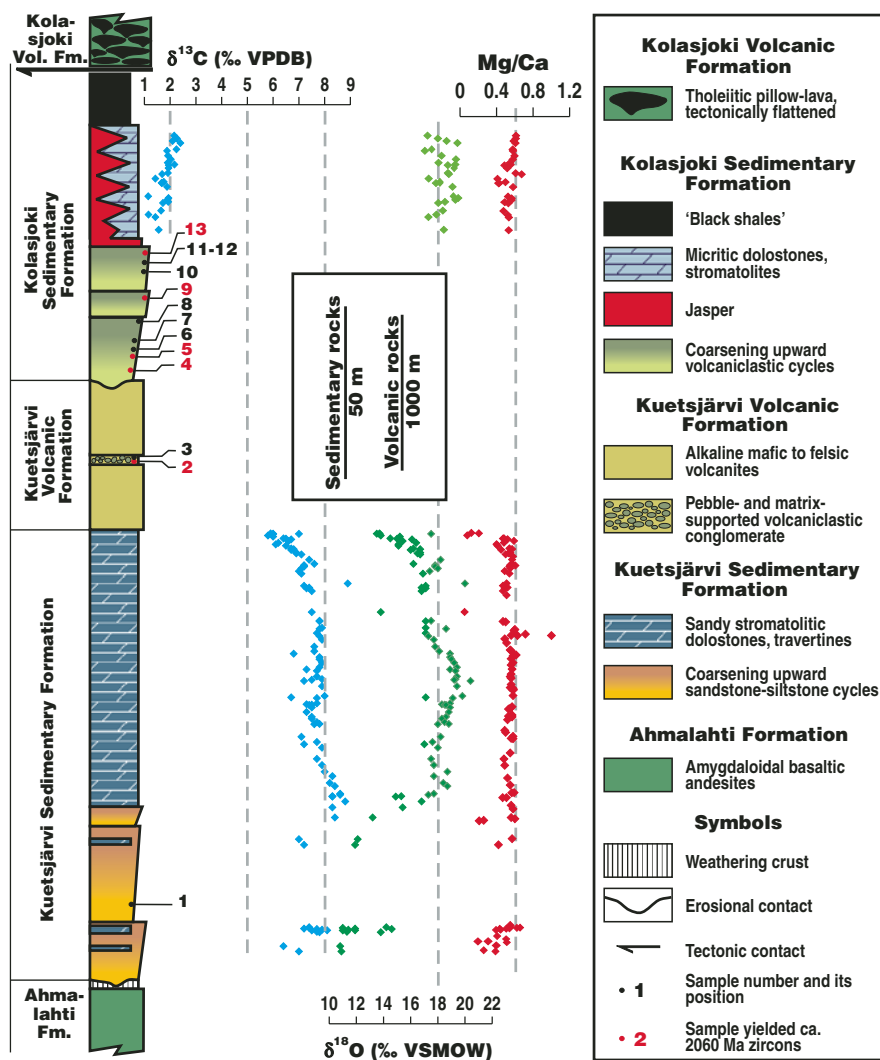


Figure 3. Lithostratigraphic column, carbon isotopic record of the end of the Lomagundi-Jatuli event, and position of samples for geochronological dating.

clastic conglomerates that rest with erosional contact upon the irregular paleosurface of alkaline felsic lava-breccia in the middle of the formation (Fig. 3). In the Kolasjoki Sedimentary Formation (Fig. 3), the samples were obtained from both surface outcrops and drill cores (drill holes 2817 and 3026). The entire stratigraphic section was sampled starting from alluvial and deltaic volcanoclastic siltstones through deltaic, fine-pebble, volcanoclastic conglomerates and volcanoclastic graywackes with large scattered fragments of Kuetsjärvi tuffs and lavas (Tables DR3 and DR4).

GEOCHRONOLOGICAL RESULTS AND INTERPRETATION OF AGE

The results for all samples dated are reported in Tables DR3 and DR4. Only $^{207}\text{Pb}/^{206}\text{Pb}$ dates from concordant analyses have been considered. The youngest population (ca. 2060 Ma) is displayed on concordia plots (Fig. 4), whereas all concordant analyses are shown in cumulative histograms (Fig. DR1).

The youngest age measured from zircons in the Kuetsjärvi Sedimentary Formation indicates an Archean age of ca. 2700 Ma. Thus, the attempt to obtain the age of the underlying Ahmalahti volcanites has proved unsuccessful. The youngest zircon from the Kuetsjärvi Volcanic Formation has been dated to 2049 ± 28 Ma (Table DR3). The overlying Kolasjoki Sedimentary Formation yielded several zircons with similar $^{207}\text{Pb}/^{206}\text{Pb}$ dates at ca. 2060 Ma ($n = 10$) (Fig. 4). In all processed samples, the younger population of zircons is distinguished by small (<100 mm), euhedral, prismatic crystals showing no sign of roundness. The older zircons are commonly >100 mm in size and rounded. The absence of any igneous activity of this age in the area other than Kuetsjärvi volcanism argues for the underlying volcanic rocks being the only probable source of these zircons. Given the consistency of the secondary ion mass spectrometry (SIMS) data with the isotope dilution-thermal ionization mass spectrometry (ID-TIMS) data (Figs. 4A and 4B), we suggest that the $^{207}\text{Pb}/^{206}\text{Pb}$ date of $2058 \pm$

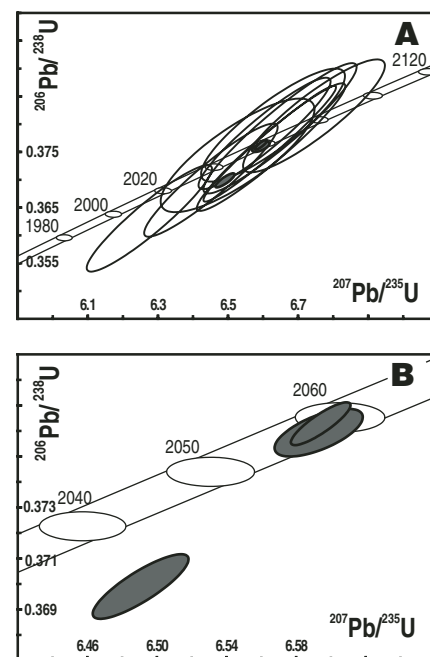


Figure 4. U-Pb concordia plots of SIMS and ID-TIMS analyses at ca. 2060 Ma. A: Plot of all analyses in the 1980–2120 Ma range from the Kuetsjärvi Volcanic Formation and Kolasjoki Sedimentary Formation. SIMS analyses are plotted as open ellipses; ID-TIMS analyses are filled. B: Detail of concordia showing the three 2060 Ma ID-TIMS analyses from the upper Kolasjoki Sedimentary Formation. Concordia are plotted showing uncertainties in the U decay constants. Ellipses represent 2σ (95% confidence) uncertainties in both plots.

2 Ma (± 6 Ma when U decay constant uncertainties are included), the youngest concordant ID-TIMS analyses, provides a robust maximum age constraint on the Kolasjoki Sedimentary Formation and inferentially constrains the age of part of the Kuetsjärvi Volcanic Formation.

IMPLICATION FOR THE LOMAGUNDI-JATULI EVENT

Previous estimates of the absolute timing and duration of the Lomagundi-Jatuli event have been ca. 2350–2000 Ma (Shields and Veizer, 2002) or ca. 2220–2060 Ma (Karhu and Holland, 1996). Existing constraints from the Fennoscandian Shield for the age of ^{13}C -rich carbonates indicate that accumulation was under way by ca. 2200 Ma based upon the U-Pb age of 2206 ± 9 Ma on sills that intrude ^{13}C -rich carbonates in the Kuusamo Schist Belt (Karhu, 1993). Previously, the termination of the isotopic event was inferred at ca. 2060 Ma by lithostratigraphic correlation of dated volcanic rocks (Koivusaari Formation) with Vaivani volcanites overlying the ^{13}C -rich Petonen Formation in southern Finland (Karhu, 1993). A felsic porphyry of the Väystäjä Formation dated to 2050 ± 8 Ma provides similar inference due to its unclear stratigraphic position within the Paakkola Group, which itself shows allochthonous relationship with ^{13}C -rich dolo-

stones of the Peräpohja Schist Belt (Perttunen and Vaasjoki, 2001). The newly obtained age of 2058 ± 2 (± 6) Ma provides the first precise and reliable time constraint acquired within a stratigraphic section that records the shift from high to normal $\delta^{13}\text{C}$ values, and thus the end of the Lomagundi-Jatuli event (Fig. 3). Consequently, the Fennoscandian Shield data suggest an ~140 m.y. interval characterized by ^{13}C -rich carbonate accumulation, although the onset of the excursion remains to be constrained (Fig. 1). The newly obtained age separates the Kolasjoki carbonates from the Lomagundi-Jatuli event; thus, the previously suggested dissection of the long-lasting continuous excursion into two separate excursions in the Pechenga Greenstone Belt (Melezhik et al., 1999) is not verified and should be discarded. The Pechenga data combined with South African data (Hountebek Formation) suggest an apparent sharp negative to positive fluctuation of $\delta^{13}\text{C}$ at the end of the Lomagundi-Jatuli event (Fig. 1), though further data are required in addition to these constraints to determine the global pattern.

The newly obtained age for the termination of the Lomagundi-Jatuli event indicates its temporal association with final dispersal of Paleoproterozoic supercontinents (Fig. 1). In Fennoscandia ca. 2050 Ma, the tectonic regime was changing from rifting to passive margin thermal subsidence resulting in paleogeographic restructuring of major landmasses and associated decrease in stromatolite abundance and diversity. Large shallow-water carbonate platforms and shelves colonized by stromatolite-forming cyanobacteria drowned and disappeared on several continents in North and South America (Bekker et al., 2003b; Bekker and Eriksson, 2003) and Fennoscandia (Wanke and Melezhik, 2005). This linkage suggests a potential causative relationship between the two events; thus, the model suggested by Hayes and Waldbauer (2006), invoking the destabilization of the association between stromatolitic producers and methanogenic consumers, looks plausible. However, whether the fermentative diagenetic model of Hayes and Waldbauer (2006) as a whole is valid remains to be seen.

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