Chemostratigraphic constraints on the time of deposition of carbonate rocks in the Karasjok Greenstone Belt, northern Norway

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The time of deposition of the rocks of the Karasjok Greenstone Belt, the most continuous and longest, linear Precambrian terrane in northern Norway, remains poorly dated. The belt is composed of several volcanic and sedimentary formations including a continuous unit of calcite and dolomite marbles. Most rocks have been strongly deformed and altered under amphibolite-facies metamorphism. Carbon isotope chemostratigraphy has been applied to provide an apparent depositional age of carbonate units and associated sedimentary and volcanic rocks. Thirty-eight samples representing marbles in four different locations in the northern part of the belt show high $\delta^{13}\text{C}_{\text{carb}}$ values, hence recording the Palaeoproterozoic Lomagundi–Jatuli isotopic event. The least-altered $\delta^{13}\text{C}_{\text{carb}}$ values suggest the presence of two isotopically different assemblages (+8 to +9‰ and +12 to +14‰) whose apparent time of deposition can be constrained between 2220 and 2140 Ma, and between 2220 and 2110 Ma, respectively.

Keywords: Palaeoproterozoic, Norway, marble, carbon isotopes

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Introduction

The Karasjok Greenstone Belt (KGB) represents a 160 km-long and 40 km-wide linear Precambrian structure situated in the northwestern part of the Precambrian Fennoscandian Shield in Finnmark, northern Norway (Fig. 1). In the north, the KGB is overlain by autochthonous Ediacaran (Vendian) to Cambrian sedimentary successions of the Dividal Group which are overthrust by the Caledonian nappes. The KGB extends for another 150 km to the south into northern Finland where it is named the Kittilä Greenstone Belt. The KGB has a potential for discovery of economic ore

deposits (e.g., Often, 1985). However, the lack of robust radiometric data constraining the time of deposition of its sedimentary and volcanic complexes (e.g., Krill et al., 1985; Marker et al., 2000) hampers chronostratigraphic correlations of the sedimentary and volcanic successions with those occurring in neighbouring Finland and elsewhere in the Fennoscandian Shield. In addition, this imposes problems on fully understanding the evolution of the greenstone belt and hampers the choice of prospecting and exploration strategy. Consequently, the goal of this study is to provide constraints on the age of some of the sedimentary formations by employing the C-isotope chemostratigraphic approach.

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Figure 1. Geological map of the eastern part of the Fennoscandian Shield emphasising the Early Palaeoproterozoic rocks and showing the location of the study area (in the northwest corner). The map is based on Koistinen et al. (2001).

Tectonic setting, depositional environments and main litho-/tectonostratigraphic units

The volcanosedimentary succession of the KGB overlies the Jergul Gneiss Complex in the west and dips away beneath the overthrust Tanaelv Migmatite Complex in the east (Figs. 2, 3). The KGB rocks underwent greenschist- to amphibolite-grade metamorphic alteration and were highly deformed during the Late Palaeoproterozoic Svecofennian orogeny. In the text that follows we purposely omit the prefix 'meta'. A robust stratigraphy for the volcanosedimentary succession is yet to be established. Two parallel, poorly correlated lithostratigraphic subdivisions have been proposed (Fig. 4) and their detailed descriptions summarised and documented by Siedlecka et al. (1985), Siedlecka (1985), Often (1985) and Davidsen (1994).

Litho-/tectonostratigraphy and depositional environment of the southern part of the KGB

Siedlecka et al. (1985), Siedlecka (1985) and Often (1985) subdivided the volcanosedimentary succession of the KGB into five tectonostratigraphic units (Fig. 4C). The three upper units comprise the Iddjávri Group, the type section for which is located in the southern part of the KGB where there is generally a low degree of outcrop and only sparse indicators of stratigraphic polarity. Hence, the identification of possible stratigraphic repetitions caused by isoclinal folding and/or thrusting is difficult to detect (cf., Siedlecka, 1985). A brief summary of the tectonostratigraphy of this southern area is given below.

The Vuomegielas Formation

The Vuomegielas Formation represents the lowermost structural unit whose lower boundary remains unknown. Lithologically homogeneous, it consists of

Figure 2. (A) Inset map showing location of the study area in Finnmark, northern Norway, and (B) schematic geological map of the northern and middle parts of the KGB with locations of the studied/sampled sections (denoted from 1 to 4). The geological map is based on Siedlecka & Roberts (1996). Detailed geological map of (C) area is shown in Fig. 3.

Figure 3. Detailed geological map of the northernmost part of the KGB with locations of the sampled profiles (denoted from 1 to 3). The map is based on Davidsen (1994), Siedlecka et al. (2011) and Roberts & Davidsen (2011). The geology of the northern part of this map has been described in detail in Davidsen (1994).

a foliated, finely crystalline amphibolite showing in places high Mg and low Ti contents, hence resembling komatiites (Siedlecka, 1985; Siedlecka et al., 1985). The amphibolites dip beneath the overlying terrigenous Skuvvanvárri Formation; however, their mutual contact has not been observed. Based on stratigraphic alternation

between finely crystalline amphibolite, conglomerate, hornblende schist and quartzite within a 250–300 m-thick transitional zone between these two formations, Siedlecka (1985) suggested a likely primary stratigraphic relationship.

The Skuvvanvárri Formation

The Skuvvanvárri Formation represents a dominantly sedimentary succession resting either on the Vuomegielas amphibolites or on the palaeoweathered, in places intensely dolomitised, gneisses of the Jergul Gneiss Complex (a detailed description is published in Siedlecka, 1985). The maximum observed formational thickness is close to 950 m. The upper contact represents a major thrust which, in places, has totally obliterated the formation (Siedlecka, 1985).

The formation consists of pale-grey, matrix-supported conglomerate interbedded with green, fuchsite-bearing, and pink sandstones and mudstones (Siedlecka, 1985). The conglomerates are immature and polymict, lack any internal water-flow features and contain beds of mudstones with scattered pebbles. The sandstones are fine grained, well sorted and in contrast to the conglomerates show a high degree of textural maturity. The sandstones are quartzitic or feldspathic, the latter consisting mainly of albite of unknown origin (Siedlecka, 1985; Often, 1985). Both varieties exhibit large-scale, tabular, herringbone or tangential cross-bedding.

Lithofacies interpretation. Siedlecka (1985) reported a frequent variation in lithofacies distribution in the Skuvvanvárri Formation and described the conglomerates as debris-flow and sheet-flood deposits accumulated on alluvial fans. The sandstone lithofacies was interpreted as either fluvial channel deposits or water- or wind-transported coastal sediments.

The Gollebáiki Formation

The Gollebáiki Formation is the lowermost unit of the Idjávri Group, and has a thrust contact with the underlying Skuvvanvárri rocks. Due to abundant recumbent folding the control on total thickness is poor. The best profiles indicate a total thickness of 1000–1500 m (Often, 1985). The formation is composed of both volcanic and sedimentary units. The type-section described from the southern part of the belt starts with foliated amphibolites and chlorite-amphibole rocks containing several intervals of Mn-rich, banded iron formations, sulphide-bearing chert, massive sulphide, and subordinate clastic sedimentary and carbonate rocks. The middle part of the formation is dominated by mica schists and sulphide-bearing, poorly banded gneisses of dioritic to quartz-dioritic composition. The gneisses locally contain chalcopyrite and Cu–Au mineralisations (Bugge, 1978). The upper part of the formation consists of albite-rich (up to 70%; Elvebakk et al., 1985) feldspathic sandstones with minor haematite-bearing shales and mafic volcanic rocks.

Lithofacies interpretation. The formation exhibits a marked facies change between mafic lavas and greywackes. Primary textures are rarely preserved but a few exceptions have been recorded. In places, albite sandstones display well-preserved flaser bedding and herring-

bone cross-bedding. Shrinkage cracks are well preserved. All sedimentological features suggest a near-shore to tidal marine environment (Elvebakk et al., 1985). Amygdales in mafic volcanic rocks and volcanoclastic textures in ultramafic volcanites provide additional indications of shallow-water to subaerial conditions (Often, 1985). The nature of the dioritic to quartz-dioritic gneisses remains unknown.

Origin of the Na-rich sandstones. The origin of the high Na₂O content (6.4–7.9 wt.%) documented in the albite sandstones (Elvebakk et al., 1985) is puzzling. These sandstones are compositionally similar to rocks locally termed albite felsites (fine-grained, thinly laminated, albite-rich rocks) occurring in the Kautokeino Greenstone Belt, and to Na-rich Palaeoproterozoic rocks known from other places across northern Fennoscandia (Tuisku, 1985; Olerud, 1988; Ettner et al., 1994; Frietsch et al., 1997; Eilu, 1994; Mänttári, 1995; Vanhanen, 2001; Kyläkoski et al., 2012; Melezhik et al., 2012, 2013a). A volcanoclastic origin combined with diagenetic Na-metasomatism (e.g., Surdam & Boles, 1979) has been suggested as a possible explanation for the Gollebáiki Na-rich sandstones (Elvebakk et al., 1985; Often, 1985). Melezhik et al. (2015) have suggested that an abundant former presence of Palaeoproterozoic halite, subsequently dissolved and mobilised during the course of the post-depositional history, may represent the most probable source of sodium in metasomatically altered rocks in northern Fennoscandia.

The Báhkilvárri Formation

The lower boundary of the formation is transitional into the sedimentary rocks of the Gollebáiki Formation, but is commonly tectonically overprinted. The formation represents a monotonous succession of foliated, garnet-bearing amphibolites, originally tholeiitic basaltic lavas, tuffs and tuffites with an unknown number of dykes and sills (Henriksen, 1983; Often, 1985). Rare pillow structures have been observed in the amphibolites.

Chlorite-amphibole rocks, primary komatiites (Henriksen, 1983), occur in the middle and upper parts of the formation. Primary depositional features are common, showing a dominance of pyroclastic material with grain sizes ranging from fine tuff to breccias. Blocks of amygdaloidal lavas are common. Well-preserved pillows have been described in some localities (Often, 1985). Massive flows are uncommon. Henriksen (1983) described microscopic textures that might represent microspinifex.

Terrigenous sediments, mainly greywackes, are a minor constituent of the formation. Chemical precipitates, such as manganese-bearing banded iron formations, are associated with the komatiites. Sulphide-rich intervals represent either lateral extensions of the iron formations or separate layers within the mafic volcanites.

Lithofacies interpretation. Wennervirta (1969) recognised the effusive character of the komatiitic rocks and used

Figure 4. (A) Lithostratigraphic column of the KGB, and (B, C) its alternative correlation with litho-/tectonostratigraphic subdivisions previously advanced by Often (1985) (modified from Davidsen, 1994). The radiometric and chemostratigraphic dates are from: (*) – Krill et al. (1985), (**) – Marker et al. (2000), (***) – this study.

the term agglomerate in a broad sense, combining all coarse-grained volcanic breccias and conglomerates. The presence of pillows provides evidence of a subaqueous environment, whereas vesicles in the same pillows and the abundance of pyroclastic material are indicative of a relatively shallow-water setting (Often, 1985).

The Rai'tegár'ži Formation

The Rai'tegár'ži Formation conformably overlies the Báhkilvárri Formation and consists of mica schists and amphibolites. The latter are indistinguishable from

those of the underlying Báhkilvárri Formation, and due to the limited exposure these two formations have not been mapped separately in most parts of the KGB. The Rai'tegár'ži mica schists and amphibolites have been strongly deformed and partly migmatized towards the Tanaelv Migmatite Complex (Fig. 2). No primary depositional features were observed. An apparent thickness of the formation is reported to be several kilometres. The upper boundary with the Tanaelv Migmatite Complex is tectonic.

Litho-/tectonostratigraphy and depositional environment of the northern part of the KGB

Braathen & Davidsen (2000) described the volcano-sedimentary succession of the KGB using a type section located in the northern part of the KGB. They stated that this section “*represents the best known and probably one of the most complete sections through the KGB*”. They subdivided the volcanosedimentary succession into four formations (Fig. 4A). A brief summary of their tectono-stratigraphy is given below.

The Lavttevárri Formation

The Lavttevárri Formation is 50–150 m thick and rests unconformably on Archaean granitic gneisses. Rocks occurring above the basal unconformity are carbonate-bearing mica schists (locally fuchsite-bearing) with conglomeratic intervals containing clasts of vein quartz and also granite and granitic gneiss from the basement. These are overlain by schistose tholeiitic amphibolites and sandstones, subordinate komatiites, possible rhyolitic volcanites, mica schists and marbles (Braathen & Davidsen, 2000).

The Čorgašávži Formation

The Čorgašávži Formation is 250(?)–700 m thick and interfingers with the underlying Lavttevárri rocks. Recently, the Čorgašávži Formation has been renamed as the Čorgaš Formation in the 1:50,000 mapsheets Skoganvarri and Lakselv, which were published after Braathen & Davidsen (2000). The formation is composed mainly of sandstones; however, the upper part of the formation includes tholeiitic, amygdaloidal lava flows intercalated with mica schists. Intervals of conglomerates and marbles are also present. Gabbros and minor pyroxenitic bodies are common (Braathen & Davidsen, 2000).

The Briittágielas Formation

The Briittágielas Formation is 300–700 m thick. The basal part of the formation is marked by the occurrence of a thick marble unit (calcite-dolomite-tremolite rocks). This is followed by kyanite-bearing, garnet-mica schists and possible calc-alkaline dacitic volcanites with intervals of graphite-bearing schist. The rest of the formation is composed of banded amphibolites (tholeiitic basalt) and volcanoclastic and pillowed komatiites occurring in four different units, each up to 70 m thick (Braathen & Davidsen, 2000). Several intervals of massive sulphide mineralisation occur throughout the stratigraphy.

The Fossestrand Formation

The Fossestrand Formation has a minimum thickness of 250–300 m. Two thick, pillowed lava flows occur in the lower part of the formation followed by agglomeratic/volcanoclastic deposits. The bulk of the formation is composed mainly of komatiitic and coarse-grained mafic volcanic rocks.

Lithofacies interpretation

Braathen & Davidsen (2000) inferred that, on the whole, the sedimentary-volcanic succession studied in the northern part of the KGB was deposited under subaqueous conditions, as based on the presence of the pillow lavas and graphite-bearing schists. The preserved depositional contacts against Archaean granitic basement indicate that at least the lowermost formation has been deposited in an intracratonic palaeotectonic setting. The abundance of thick sedimentary units dominated by material from granitoid sources indicates the presence of cratonic source areas in close proximity to the depositional site (Davidsen, 1994).

North-to-south lithostratigraphic correlation

Braathen & Davidsen (2000) reported that no significant tectonic breaks have been observed between any of their four formations. Based on this, Braathen & Davidsen (2000) have suggested that their subdivision is very likely to represent a lithostratigraphy (Fig. 4A). They also offered two alternative scenarios (Fig. 4B, C) for the correlation of their subdivisions with those previously advanced by Siedlecka (1985) and Often (1985). This correlation scheme suggests that the Gollebáiki and Báhkilvárri formations recognised by Often (1985) and Siedlecka (1985) in the southern part of the belt, correspond mainly to the Čorgaš and Briittágielas formations of Braathen & Davidsen (2000), defined in the northern part of the belt (Fig. 4).

Often (1985) stated that in the northern part of the belt the predominantly psammitic lithology of the Gollebáiki Formation terminates with a marble and mica schist unit that constitutes an excellent marker horizon (Crowder, 1959; Pharaoh, 1981, 1984). Indeed, the recently published 1:50,000 geological map demonstrates that there is a marble unit which is continuous over several kilometres in at least the northern part of the KGB (Fig. 3). In contrast, Braathen & Davidsen (2000) have reported two carbonate units (Fig. 4A). One occurs in the upper part of the Čorgašávži Formation whereas the second, much thicker unit, forms the base of the Briittágielas Formation. However, the 1:50,000 geological map which adopted Braathen and Davidsen's lithostratigraphy shows that the thickest carbonate unit occurs in the Čorgašávži Formation, not in the Briittágielas Formation. Moreover, the same map demonstrates that, in places, this thick carbonate unit crosses the boundary that divides these two formations (Fig. 3). This indicates that perhaps there is yet no clear understanding as to whether one or two carbonate units are present in the middle part of the KGB succession.

Available radiometric ages

The komatiites in the Báhkilvárri Formation have been dated to 2085 ± 85 Ma, using the whole-rock Sm–Nd method (Krill et al., 1985). Marker et al. (2000) reported a Sm–Nd model age (T_{DM}) of 3445 Ma for the terrigenous metasedimentary rocks of the Skuvvanvárri Formation, thus suggesting an Archaean source for the detritus. They also reported that detrital zircons from metasedimentary rocks of the Gollebáiki Formation dated by Secondary Ion Mass Spectrometry (SIMS) at the NORDSIM show a spread of concordant ages between 1971 and 2438 Ma with one grain at 3188 Ma. The four youngest grains define an intercept at 1947 ± 88 Ma. In addition, a granodiorite body intruded into the Báhkilvárri amphibolites was dated conventionally by a U–Pb technique. In this study, five fractions defined a concordia intercept at 1974 ± 8 Ma, which provides a minimum depositional age for the Báhkilvárri Formation (Marker et al., 2000).

Carbonate rocks, sampled localities

The carbonate unit (CU hereafter) occurring in the Čorgašávži/Briittáielas stratigraphic interval (or in the Gollebáiki Formation of Often (1985)) has been sampled in four different localities designated as profiles 1 to 4 in Figs. 2 and 3.

Profile 1 (samples 1–17) is located in the northeastern corner of the studied area where the CU is exposed in several isolated outcrops. The sampled interval extends over c. 500 m along strike, and has a visible structural thickness of more than 25 m. The CU is tightly folded and composed of banded, calcite-dolomite-tremolite rocks with numerous 0.5–1 m-thick beds of pale-grey dolomite marbles and rare calcite marbles (Fig. 5A, B).

Profile 2 (samples 18–27) is located in the central part of the studied area where two tightly spaced CUs separated by a plagioclase-mica schist are exposed in several isolated localities. It remains uncertain as to whether these two CUs represent two stratigraphically different intervals or only one duplicated by folding. The CUs are composed mainly of pale-grey, thinly banded dolomite and calcite marbles which, in places, are mylonitised (Fig. 5C). In one case the marble is red coloured.

Profile 3 (samples 28–36) is located in the western part of the studied area and represents the most exposed section of the CU. Samples were collected from a c. 40 m-high cliff (Fig. 5 D). Here, the CU is composed of pink, red and pale-grey dolomite marbles. The marbles show centimetre to decimetre banding with intercalations of thin layers of calc-silicate rocks (Fig. 5E, F).

Profile 4 (samples 37–39) represents the southernmost known locality of the CU (Fig. 2B). This is a road-cut exposing white, banded, calcite marbles in a unit a few metres in thickness.

Analytical methods

Major and trace elements were analysed by X-ray fluorescence spectrometry (XRF) at the Geological Survey of Norway (NGU), Trondheim, using a PANalytical Axios at 4 kW X-ray spectrometer. The precision (1s) is typically around 2% of the major oxide present. Acid-soluble Fe, Ca, Mg and Mn extracted with 10% cold HCl acid were analysed by ICP–AES at NGU using a Thermo Jarrell Ash ICP 61 instrument. Detection limits for Fe, Mg, Ca and Mn are 5, 100, 200 and 0.2 ppm, respectively. The total analytical uncertainty including element extraction (1 σ) is $\pm 10\%$ rel.

Oxygen and carbon isotope analyses of whole-rock marble samples were carried out at the Scottish Universities Environmental Research Centre, Glasgow, using the phosphoric acid method of McCrea (1950) as modified by Rosenbaum & Sheppard (1986) for operation at 100°C. Carbon and oxygen isotope ratios in carbonate constituents of the whole-rock samples were measured on a VG SIRA 10 mass spectrometer. Analyses were calibrated against NBS 19, and precision (1s) for both isotope ratios is better than $\pm 0.2\%$. Oxygen isotope data for dolomites were corrected using the fractionation factor 1.00913 recommended by Rosenbaum & Sheppard (1986). The $\delta^{13}\text{C}$ data are reported in per mil (‰) relative to V–PDB and the $\delta^{18}\text{O}$ data in ‰ relative to V–SMOW.

Major and trace element geochemistry, and C and O isotope ratios of marbles

Major and trace element data and C and O isotopic values obtained from the marbles of the KGB are presented in Table 1. Geochemical characteristics of the studied carbonate and carbonate-bearing rocks are based on 38 whole-rock analyses (Table 1). Geochemically, all analysed samples obtained from the CU can be divided into four groups: Group 1 – dolomite marbles ($\text{Mg}/\text{Ca}_{\text{carb}}$ (carbonate phase) = 0.44–0.65, $\text{MgO}/\text{CaO}_{\text{wr}}$ (whole rock) = 0.48–0.70, $n = 22$); Group 2 – marbles with a mixed calcite-dolomite mineralogy ($\text{Mg}/\text{Ca}_{\text{carb}}$ = 0.003–0.34, $\text{MgO}/\text{CaO}_{\text{wr}}$ = 0.19–0.37, $n = 4$); Group 3 – dolomite-calcite-tremolite rocks ($\text{Mg}/\text{Ca}_{\text{carb}}$ = 0.18–0.35, $\text{MgO}/\text{CaO}_{\text{wr}}$ = 0.40–0.94, $n = 4$); and Group 4 – calcite marbles ($\text{Mg}/\text{Ca}_{\text{carb}}$ = 0.003–0.13, $\text{MgO}/\text{CaO}_{\text{wr}}$ = 0.02–0.16, $n = 8$). A large discrepancy between $\text{Mg}/\text{Ca}_{\text{carb}}$ and $\text{MgO}/\text{CaO}_{\text{wr}}$ ratios is observed in Group 2 rocks in Profile 1 (samples 6–8) and Group 3 rocks in Profile 2 (samples 19 and 24;

Figure 5. Images illustrating some main sedimentological features of carbonate rocks of the CUs in the KGB. (A) A natural outcrop of tightly folded, thinly banded, calcite marble from Profile 1. (B) A photograph of banded dolomite marble composed of alternating tremolite-poor (white line) and tremolite-rich (red line) bands from Profile 1; a coin for scale is 1 cm in diameter. (C) A small vertical cliff exposing a mylonitised, calcite marble from Profile 2. (D) A c. 40 m-high cliff exposing dolomite marbles from Profile 3. (E) Pink and white dolomite marble with thin interlayers of red calc-silicate bands (white arrowed). (F) Thin alternation of dolomite (pale pink) and calc-silicate (pale grey) bands. Scale-bars in (A), (C), (E) and (F) with cm divisions.

Table 1. Chemical, and carbon- and oxygen-isotopic composition of marbles from the CUs in the KGB.

Sample #	SiO ₂	Al ₂ O ₃	Fe ₂ O ₃	TiO ₂	MgO	CaO	Na ₂ O	K ₂ O	MnO	P ₂ O ₅	Fe*	Mn*	Mg*	Ca*	Sr*	Mg/Ca	MgO/ CaO	Mn/Sr	δ ¹³ C	δ ¹⁸ C
wt. %											μg g ⁻¹									
Profile 1																				
Dolomite marble																				
3	6.25	0.35	1.90	0.01	19.6	29.3	<0.1	<0.01	0.11	0.03	10200	690	112000	193000	133	0.58	0.67	5.2	8.0	19.1
4	0.76	0.05	0.36	<0.01	20.9	32.5	<0.1	<0.01	0.05	0.03	2210	329	128000	215000	175	0.60	0.64	1.9	8.5	18.9
5	6.99	0.33	0.42	0.01	19.7	30.7	<0.1	0.02	0.06	0.09	1970	340	108000	202000	155	0.53	0.64	2.2	8.3	19.1
12	9.37	0.94	1.08	0.04	18.4	29.6	0.47	<0.01	0.07	0.15	4770	433	97700	194000	118	0.50	0.62	3.7	7.5	17.4
13	2.63	0.31	0.68	0.01	20.0	31.5	<0.1	<0.01	0.06	0.04	3510	388	120000	211000	168	0.57	0.63	2.3	3.8	13.4
14	0.76	0.13	0.97	<0.01	21.2	30.6	<0.1	<0.01	0.06	<0.01	5790	362	132000	205000	206	0.64	0.69	1.8	3.3	14.3
15	3.30	0.18	1.05	<0.01	18.8	32.2	<0.1	<0.01	0.08	0.05	5560	517	109000	211000	162	0.52	0.58	3.2	8.0	15.5
16	1.02	0.12	1.10	<0.01	20.8	30.8	<0.1	<0.01	0.07	0.04	6710	443	129000	204000	176	0.63	0.68	2.5	4.5	13.6
Calcite-dolomite marble																				
1	6.01	0.45	2.04	0.02	11.3	39.7	<0.1	<0.01	0.31	0.07	4980	1880	52200	265000	121	0.20	0.28	16	6.6	14.9
2	1.48	0.07	0.506	<0.01	14.7	39.4	<0.1	<0.01	0.14	0.02	2840	811	85300	254000	146	0.34	0.37	5.6	8.0	19.1
Dolomite-calcite-tremolite rock																				
6	18.4	0.71	0.283	0.03	16.4	29.3	0.20	0.03	0.06	0.25	989	343	68000	192000	117	0.35	0.56	2.9	7.6	18.7
7	27.9	2.63	0.651	0.09	16.6	25.8	0.32	0.05	0.08	0.17	815	454	27900	157000	69	0.18	0.64	6.6	5.6	16.1
8	13.8	0.95	0.47	0.04	13.9	34.6	0.13	0.03	0.09	0.13	1420	558	50300	230000	98	0.22	0.40	5.7	6.2	16.3
Calcite marble																				
9	6.07	0.15	0.66	<0.01	4.76	47.7	<0.1	<0.01	0.04	0.07	482	262	17300	321000	73	0.05	0.10	3.6	6.1	16.2
11	1.73	0.12	0.17	<0.01	7.82	47.6	<0.1	<0.01	0.14	0.02	462	862	41900	317000	118	0.13	0.16	7.3	7.5	18.5
Graphite-bearing albite sandstone																				
10	57.6	15.4	4.63	0.81	2.74	1.84	6.33	0.04	0.06	0.05										
Profile 2																				
Dolomite marble																				
21	2.61	0.28	0.61	0.01	20.0	31.4	<0.1	0.11	0.11	0.04	3270	710	117000	211000	60	0.55	0.64	12	11.2	18.7
22	0.93	0.21	0.27	0.01	19.4	33.3	<0.1	0.13	0.11	0.04	1420	717	120000	225000	53	0.53	0.58	13	13.9	20.7
23	2.46	0.45	0.23	0.02	19.0	33.1	<0.1	0.19	0.14	0.03	790	913	113000	223000	71	0.51	0.57	13	11.7	20.0
25	3.22	0.47	0.33	0.02	19.2	32.5	<0.1	0.01	0.07	0.07	753	459	109000	217000	59	0.50	0.59	7.8	11.6	21.7
27	<0.5	0.03	1.03	<0.01	21.0	30.6	<0.1	<0.01	0.11	0.03	6660	706	135000	209000	76	0.65	0.69	9.3	13.2	17.5

Table 1. Continued.

Sample #	SiO ₂	Al ₂ O ₃	Fe ₂ O ₃	TiO ₂	wt. %										μg g ⁻¹				MgO/Ca	Mn/Sr	δ ¹³ C	δ ¹⁸ C
					MgO	CaO	Na ₂ O	K ₂ O	MnO	P ₂ O ₅	Fe*	Mn*	Mg*	Ca*	Sr*	‰						
Calcite-dolomite marble																						
19	44.0	0.05	0.09	<0.01	7.36	25.9	<0.1	<0.01	0.03	0.04	49.7	148	580	185000	25	0.003	0.28	6.0	5.9	20.0		
Carbonate-bearing, albite sandstone																						
24	53.7	13.70	10.6	1.48	5.40	5.76	6.61	0.20	0.04	0.14	1470	63.4	2270	9350	4.5	0.24	0.94	14	11.4	20.7		
Calcite marble																						
17	3.86	0.19	0.11	0.01	1.89	52.2	<0.1	<0.01	0.19	0.05	154	1220	1080	352000	195	0.003	0.04	6.3	5.9	18.3		
18	3.95	0.05	0.06	<0.01	2.15	52.4	<0.1	<0.01	0.15	0.03	82.8	962	1030	356000	188	0.003	0.04	5.1	5.6	18.9		
20	2.76	0.25	1.07	<0.01	6.94	46.9	<0.1	<0.01	0.16	0.02	1280	943	33300	312000	63	0.11	0.15	15	7.7	17.3		
26	4.72	0.95	2.00	0.03	7.04	44.9	<0.1	0.18	0.11	0.03	1890	632	29000	304000	59	0.10	0.16	11	9.7	16.7		
Profile 3																						
Dolomite marble																						
28	10.2	0.94	0.48	0.03	17.8	29.0	<0.1	0.20	0.07	0.13	2230	402	106000	201000	99	0.53	0.61	4.1	8.3	19.5		
29	13.3	0.55	0.85	0.03	17.5	27.0	<0.1	0.14	0.07	0.09	4510	429	108000	184000	92	0.59	0.65	4.7	7.7	18.3		
30	6.99	0.93	1.09	0.04	19.1	28.7	0.15	0.04	0.10	0.13	4580	429	109000	183000	91	0.60	0.67	4.7	5.9	16.6		
31	4.73	0.07	0.18	<0.01	16.3	34.2	<0.1	<0.01	0.04	0.04	1000	240	104000	234000	143	0.44	0.48	1.7	8.5	18.4		
32	15.4	0.44	0.25	0.02	15.0	29.7	0.11	0.08	0.04	0.04	1400	297	93500	204000	133	0.46	0.51	2.2	8.5	18.7		
33	12.0	0.19	0.30	<0.01	18.3	27.6	<0.1	0.03	0.04	0.07	1880	285	115000	186000	139	0.62	0.66	2.1	7.4	17.5		
34	7.86	0.38	0.37	0.02	19.4	28.8	<0.1	0.06	0.07	0.12	2080	396	122000	199000	118	0.61	0.67	3.4	8.5	18.2		
35	12.6	0.05	0.14	<0.01	14.9	31.2	<0.1	0.01	0.03	0.03	818	191	95800	214000	137	0.45	0.48	1.4	8.7	19.7		
36	50.6	0.11	0.17	<0.01	10.4	14.8	<0.1	0.05	0.03	0.01	1040	196	67100	105000	70.3	0.64	0.70	2.8	9.0	19.5		
Profile 4																						
Calcite marble																						
37	38.2	0.12	0.34	<0.01	0.499	33.2	<0.1	0.01	0.10	0.03	282	681	2510	238000	85	0.01	0.02	8.0	7.5	17.0		
38	0.57	0.14	0.56	0.01	2.08	53.3	<0.1	0.04	0.11	0.05	468	724	12700	366000	92	0.03	0.04	7.9	10.1	15.9		
Calcite-dolomite marble																						
39	<0.5	0.09	0.53	<0.01	8.78	45.1	<0.1	0.03	0.60	0.01	2710	3630	55700	301000	76	0.19	0.19	48	12.0	16.6		

Table 1). These five samples plot well above a best fit-line for other samples in a $\text{Mg}/\text{Ca}_{\text{carb}}$ - $\text{MgO}/\text{CaO}_{\text{wr}}$ cross plot (Fig. 6A).

The carbonate and carbonate-bearing rocks have a variable SiO_2 content ranging from <0.5 to 54 wt.%. All Group 1, 2 and 3 marbles, having SiO_2 contents over 1 wt.%, contain a variable amount of tremolite. Abundances of Al_2O_3 are below 1 wt.% with two exceptions (2.6 and 13.7 wt.%; samples 7 and 24). The rock with the highest SiO_2 and Al_2O_3 content (sample 24) is also rich in Na_2O (6.6 wt.%). It represents a carbonate-bearing, albite sandstone, which is somewhat similar in composition to another albite sandstone, though graphite rich (sample 10, Na_2O = 6.3 wt.%), found in the CU in Profile 1 (Table 1). The rest of the rocks have Na_2O abundances commonly below 0.1 wt.% with a few exceptions (0.11–0.47 wt.%, $n = 6$). All rocks are low in K_2O (<0.01–0.2 wt.%) and have low to moderate abundances of Sr (25–206 ppm; 110 ppm on average). The Mn/Sr ratio is generally low (6.1 ± 4 , $n = 37$, with one outlier at 48 in sample 39).

All carbonate and carbonate-bearing rocks of the CU are rich in ^{13}C . Both the $\delta^{13}\text{C}$ and the $\delta^{18}\text{O}$ values show significant fluctuations. $\delta^{13}\text{C}$ ranges between +3.3 and +12.9‰ ($+8.1 \pm 2.4$ ‰, $n = 38$), whereas the $\delta^{18}\text{O}$ values fluctuate between 13.4 and 21.7‰ ($+17.8 \pm 2$ ‰, $n = 38$). The lowest three $\delta^{18}\text{O}$ values of 13.4–14.3‰ are coupled with the lowest $\delta^{13}\text{C}$ values of +3.3 to +4.5‰ (samples 13, 14 and 16).

If considered either altogether or at the level of individual profiles, neither $\delta^{13}\text{C}$ nor $\delta^{18}\text{O}$ values show any significant correlation with SiO_2 abundances and Mn/Sr ratios. If the analyses are considered altogether, there is a significant positive correlation between $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values

($+0.58$, >99%, $n = 38$; Fig. 6B). This correlation becomes stronger if only Groups 1 and 3 rocks are considered ($+0.89$, >99%, $n = 25$). If the three samples with the lowest $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values are excluded, there is also a positive correlation between $\delta^{13}\text{C}$ values and $\text{Mg}/\text{Ca}_{\text{carb}}$ ratios ($+0.37$, >95%, $n = 35$, Fig. 6C). This correlation appears to be stronger if Group 1 and Group 2 rocks are considered separately ($+0.81$, 99%, $n = 12$; and $+0.86$, >99%, $n = 11$, respectively). Similar correlations are observed between $\delta^{13}\text{C}$ values and $\text{Mg}/\text{Ca}_{\text{wr}}$ ratios.

Geochemical screening of C isotopes for post-depositional alteration

In diagenetic and metamorphic processes, the Mn and Sr contents in carbonates are commonly partially shifted towards equilibrium with the post-depositional fluids, and may lead to Sr depletion and Mn enrichment (Veizer, 1983; Kaufman & Knoll, 1995). Consequently, Mn and Sr abundances can serve as possible tools for calibration of the relative diagenetic rank of sequences (Brand & Veizer, 1980; Derry et al., 1992).

In general, diagenetic, hydrothermal and metamorphic fluids have high water/rock ratios for oxygen, but low water/rock ratios for carbon (e.g., Hudson, 1977; Banner & Hanson, 1990; Nabelek, 1991; Land, 1992; Jacobsen & Kaufman, 1999). Consequently, as a rule, the carbon system is effectively buffered by the dissolving carbonate precursor; hence, it is far more resilient during post-depositional, open-system recrystallisation (excluding organic diagenesis) with respect to the oxygen system. The latter is commonly much more easily affected by

Figure 6. Various cross-plots demonstrating C-isotope and geochemical variability in carbonate rocks from different profiles. (A) $\text{MgO}/\text{CaO}_{\text{wr}}$ versus $\text{Mg}/\text{Ca}_{\text{carb}}$ cross-plot; three samples from Profile 1 (encircled black diamonds) and two samples from Profile 2 (red diamonds) above the best-fit line (solid black) experienced quartz-dolomite metamorphic reaction, formation of tremolite and degassing of CO_2 enriched in ^{13}C and ^{18}O . (B) $\delta^{13}\text{C}$ versus $\delta^{18}\text{O}$ cross-plot illustrating a positive correlation between the two values in marbles from Profiles 1 and 3; the most altered samples from Profile 1 are encircled. (C) $\text{Mg}/\text{Ca}_{\text{carb}}$ versus $\delta^{13}\text{C}$ cross-plot; dashed lines show apparent alteration trends.

exchangeable oxygen derived from either meteoric water or interstitial fluids at elevated temperatures (e.g., Fairchild et al., 1990). In some cases, carbonate rocks that experienced a significant deformation may exhibit depletion in both oxygen and carbon isotope values (e.g., Guerrero et al., 1997). In the following discussion, oxygen isotope values and $\delta^{13}\text{C}$ – $\delta^{18}\text{O}$ cross-plots have been utilised for tracking the degree of post-depositional recrystallisation.

For the studied carbonate rocks the $\delta^{13}\text{C}$ – $\delta^{18}\text{O}$ cross-plot shows significant positive correlation, particularly for Groups 1 and 3 carbonates (Fig. 6B), hence suggesting that both isotope systems have been affected by alteration. This is also supported by variable Mn/Sr ratios of these rocks (Table 1). The $\delta^{13}\text{C}$ – $\delta^{18}\text{O}$ cross-plot suggests that the three samples with lowest $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ should be excluded from further consideration as they are definitely severely altered. A large discrepancy between $\text{Mg}/\text{Ca}_{\text{carb}}$ and $\text{MgO}/\text{CaO}_{\text{wr}}$ ratios observed in these three samples (Fig. 6A) and the presence of tremolite is indicative of an alteration process associated with a decarbonation reaction and removal of CO_2 enriched in ^{13}C and ^{18}O :

The aforementioned volatilisation reaction could have been the mechanism influencing the $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ compositions of residual dolomite and newly formed calcite (Shieh & Taylor, 1969; Bucher & Frey, 2002), and could explain the observed depletion in ^{18}O and ^{13}C with respect to the least altered samples.

The negative correlation of $\delta^{13}\text{C}$ values with both $\text{Mg}/\text{Ca}_{\text{wr}}$ and $\text{Mg}/\text{Ca}_{\text{carb}}$ ratios (Fig. 6C) suggests that a dedolomitisation process has been involved in the depletion of carbonate rocks in ^{13}C . This is particularly the case for Groups 1, 2 and 4 carbonates; completely dedolomitised rocks (e.g., calcite marbles) are depleted with respect to the dolomite marbles by 3–8‰ (Fig. 6C). If the $\delta^{13}\text{C}$ values of the least dedolomitised samples are considered to represent the best available proxy for the isotopic composition of seawater, then two different rock assemblages can be recognised. The first is characterised by the least-altered $\delta^{13}\text{C}$ values of +8 to +9‰ and includes marbles from Profiles 1 and 3 (Fig. 6C). The second assemblage has least-altered $\delta^{13}\text{C}$ values of c. +12 to +14‰ and comprises marbles from Profiles 2 and 4.

Implication of $\delta^{13}\text{C}$ values of the CU of the KGB for depositional age constraint

The Sm–Nd, whole-rock, imprecise date of 2085 ± 85 Ma obtained from the komatiites in the Báhkilvárri Formation (Krill et al., 1985), and the U–Pb zircon age of 1974 ± 8 Ma obtained from the granodiorite intruded into the Báhkilvárri amphibolites (Marker et al., 2000) provide the minimum depositional age for the supracrustal rocks of the KGB (Fig. 4A). The U–Pb date of 1947 ± 88 Ma on detrital zircons from the Gollebaiki Formation (Marker et al., 2000) sets the upper age limit.

A further constraint on the deposition of CU marbles and associated rocks can be provided by means of the carbon isotope chemostratigraphy. The Early Palaeoproterozoic history of the Earth's evolution is marked by the earliest known perturbation of the global carbon cycle (Baker & Fallick, 1989a, b) known now as the Lomagundi–Jatuli isotopic event (reviewed in Melezhik et al., 2013b). The event is expressed as a global-scale positive excursion of carbonate carbon isotopes in sedimentary carbonates (Fig. 7). Recent compilations of all available radiometric dates suggest that the excursion very likely occurred between 2300–2220 and 2060 Ma (Martin & Condon, 2013; Martin et al., 2013, 2014). However, in the Fennoscandian Shield, the Lomagundi–Jatuli isotopic event has been confidently constrained in several places between c. 2220 and 2060 Ma (Karhu, 2005; Melezhik et al., 2007; Fig. 7). Projection of the least-altered carbon-isotopic values on to the $\delta^{13}\text{C}$ reference curve (Fig. 7) suggests that the CUs of the KGB were apparently deposited between 2220 and 2110 (Profiles 2 and 4) and 2220 and 2140 Ma (Profiles 1 and 3). The obtained chemostratigraphic ages partially overlap within error with the Sm–Nd date of 2085 ± 85 Ma; however, they definitely cannot be younger than 2060 Ma, and very likely are older than 2110–2140 Ma.

We consider that the obtained apparent depositional ages of the protolith of the studied marbles are applicable to the entire belt stratigraphy despite uncertainties involved in correlation between its northern and southern parts. The obtained C-isotopic data also allow correlation of the studied carbonate rocks with ^{13}C -rich dolostones ($\delta^{13}\text{C} = +7.4 \pm 0.7\text{‰}$, $n = 51$; Melezhik et al., 2015) constituting the Kvenvik formation, which occurs in the Alta–Kvænangen Tectonic Window near Alta, c. 80 km west of the northern KGB. In both locations, the carbonate rocks are markedly enriched in ^{13}C and record the Lomagundi–Jatuli positive excursion of carbonate carbon isotopes in sedimentary carbonates. However, high- $\delta^{13}\text{C}$ carbonate rocks have not yet been found in either the Kautokeino Greenstone Belt (Melezhik, unpublished data) or the Repparfjord Tectonic Window northeast of Alta (Melezhik & Fallick, 1996).

Figure 7. A $\delta^{13}\text{C}$ variability in Palaeoproterozoic carbonate formations based on U-Pb and Re-Os data. Modified after Karhu & Holland (1996) and Martin & Condon (2013). The long horizontal violet and blue bars represent the range of the least altered $\delta^{13}\text{C}$ values in marbles from Profiles 2 and 4, and Profiles 1 and 3, respectively. Pink vertical bar denotes the Sm-Nd radiometric age for the Båhkvárri Formation komatiites. The red arrow indicates the start of deposition of ^{13}C -rich carbonates reported from Precambrian rocks of the Fennoscandian Shield, hence setting the lower age limit for accumulation of the ^{13}C -rich carbonates of the KGB. The black arrow indicates the termination of deposition of the Lomagundi-Jatuli isotopic event in the Fennoscandian Shield and worldwide, thus providing the lower age limit for the deposition of the CUs. The violet and blue horizontal bars placed along the age axis denote the projection of the apparent depositional time-range of carbonate rocks from Profiles 2 and 4, and from Profiles 1 and 3, respectively.

Conclusions

1. The marble unit occurring in the middle part of the tectonostratigraphic succession of the Karasjok Greenstone Belt shows anomalously high $\delta^{13}\text{C}_{\text{carb}}$ values and records a global positive excursion of carbonate carbon isotope ratios in sedimentary carbonates known as the Lomagundi-Jatuli isotopic event.
2. The marble unit sampled in the northern part of the belt contains two isotopically different assemblages occurring in geographically different locations.
3. Carbon isotope chemostratigraphy constrains the deposition of these two ^{13}C -rich carbonate assemblages to between <2220 and 2110 Ma ($\delta^{13}\text{C}_{\text{carb}} = +12$ to $+14\text{‰}$) and 2220 and 2140 Ma ($\delta^{13}\text{C}_{\text{carb}} = +8$ to $+9\text{‰}$).

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