Sedimentary-volcanic successions of the Alta–Kvænangen Tectonic Window in the northern Norwegian Caledonides: Multiple constraints on deposition and correlation with complexes on the Fennoscandian Shield

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Airborne geophysical data, sedimentological and geochemical characteristics of carbonate rocks, geochemical features of igneous rocks, carbon isotope chemostratigraphy and radiometric dating form a multiple approach applied for the reconstruction of depositional environments and the age of the sedimentary-volcanic succession exposed in the Alta–Kvænangen Tectonic Window (AKTW) in the northern Norwegian Caledonides. Aeromagnetic geophysical data confirm that the AKTW succession continues beneath the Caledonian nappe complexes and connects with the Kautokeino Greenstone Belt in the main part of the Fennoscandian Shield. The carbonate rocks, mainly dolostones, of the Kvenvik formation are markedly enriched in ¹³C ($\delta^{13}\text{C} = +7.4 \pm 0.7\%$, $n = 51$) and record a global positive excursion of carbonate carbon isotopes in sedimentary carbonates known as the Lomagundi–Jatuli isotopic event whose duration was constrained in the Fennoscandian Shield between c. 2220 and 2060 Ma. A radiometric date of 2146 ± 5 Ma (U–Pb, zircon) obtained from a gabbro comagmatic with mafic lavas provides a minimum age for the deposition of the ¹³C-rich, Lower and Upper dolostones and the accumulation age of the ¹³C-rich Uppermost dolostone. The carbonate rocks of the structurally overlying Storviknes formation show near-zero $\delta^{13}\text{C}$ ($+1.1 \pm 1.2\%$, $n = 41$). Carbon isotope chemostratigraphy suggests that their deposition post-dated 2060 Ma which, together with the above mentioned radiometric date, indicates a non-depositional break/hiatus of over 80 Myr separating accumulation of the two neighbouring formations. Sedimentological features of the carbonate rocks are consistent with deposition in a carbonate platform/shelf, whereas the depositional features of shales suggest accumulation in a shallow-water epeiric sea. The ¹³C-rich dolostones of the Kvenvik formation contain plentiful halite casts and sulphate pseudomorphs reflecting formation of abundant CaSO_4 and NaCl in time-equivalent successions across the Fennoscandian Shield, which may represent a source for Na and Cl metasomatism that affected Palaeoproterozoic rocks in the AKTW and across northern Fennoscandia.

Keywords: Caledonides, Norway, carbon isotopes, carbonates, evaporites, Palaeoproterozoic

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Introduction

Several tectonic windows in the northern Norwegian Caledonides represent an apparent northwestern continuation of the Precambrian Fennoscandian Shield beneath the Caledonian nappes. The Alta–Kvænangen Tectonic Window (AKTW) is one such window (Figs. 1 & 2). It contains several informally established formations with diverse lithologies of metamorphosed sedimentary

rocks associated with abundant mafic extrusive and intrusive rocks (Zwaan & Gautier, 1980; Pharaoh et al., 1983; Bergh & Torske, 1986, 1988).

There have been several attempts to correlate sedimentary and volcanic sequences of the AKTW with those occurring in other tectonic windows as well as in the main part of Fennoscandian Shield (e.g., Pharaoh et al., 1983; Siedlecka et al., 1985; Bergh & Torske, 1986, 1988). These correlations are largely based

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on lithological characteristics of sedimentary rocks and geochemical signatures of volcanic units. They demonstrated some similarities of the AKTW rock complex with those occurring across northern Norway, and consequently resulted in the general acceptance of their Palaeoproterozoic age. However, to date, there are no actual radiometric dates available which can provide robust constraints on the age of deposition of the AKTW rock successions.

Here, we use sedimentological constraints, carbon isotope chemostratigraphy, U–Pb radiometric dating of a gabbro, and airborne geophysical data to provide the first insight into the depositional history of the AKTW rock successions and their correlation with the sedimentary–volcanic formations across the Fennoscandian Shield. These correlations are used for placing the palaeogeographic and palaeotectonic positions of the AKTW rocks within the framework of the evolution of the shield in northern Fennoscandia.

Geological background

Sedimentary and volcanic rocks occurring in the AKTW constitute the Raipas Supergroup whose base remains unknown. The supergroup has been subdivided into four informal formations (Zwaan & Gautier, 1980). It starts with the Kvenvik formation which has a cumulative thickness of over 2200 m (Fig. 3). Based on lithological composition, the formation can be further subdivided into two units. The lower unit (>1000 m), defined by Vik (1985) as the Lower Kvenvik greenstone formation (Member A hereafter), is composed mainly of gabbro and subordinate dolostone, albite felsites, shale, albite-carbonate-magnetite rock, and mafic tuff and tuffite. The upper unit (c. 1200 m), defined by Vik (1985) as the Upper Kvenvik greenstone formation (Member B hereafter), consists mainly of mafic tuff, massive and pillowed tholeiitic basalt, subordinate dolostone, limestone and black shale (Bergh & Torske, 1988).

The Kvenvik formation is overlain by the Storviknes formation (300–600 m), composed of dolostones with stromatolites, dolostone breccias, and purple and grey siltstone (Bergh & Torske, 1988). The major dolostone unit occurring at Raipas and Borrás hosts copper deposits, and several copper showings also occur in the upper and thinner dolostone unit interbedded with siltstone (Fig. 3).

The Storviknes formation is succeeded by the Skoaddovárri formation, a c. 1700 m-thick unit composed mainly of sandstone containing interbeds of conglomerate, pebbly sandstone and shale (Zwaan & Gautier, 1980; Bergh & Torske, 1986).

The youngest unit in the Raipas Supergroup is the Luovosvárri formation (Zwaan & Gautier, 1980). It is composed of dolostone and sandstone. The sedimentary

and volcanic rocks of the Raipas Supergroup are tectonically deformed and metamorphosed at greenschist facies. The Raipas Supergroup is unconformably overlain by the Bossekop Group, comprising stromatolitic dolostone and siltstone, followed by the Borrás Group which consists of sandstone, siltstone, conglomerate and tillite of inferred Late Neoproterozoic age (Fig. 2; Zwaan & Gautier, 1980).

Several minor copper deposits, partly mined in the past, are known in the Raipas Supergroup. In the lower part of the Kvenvik formation, copper sulphide occurs as disseminations in albite felsites and as hydrothermal sulphide-quartz-carbonate veins in volcanic rocks throughout the formation. Sediment-hosted copper deposits are also widespread in the Storviknes formation (e.g., Vik, 1985). Here, copper sulphides occur as cement in breccias and disseminations in dolostone and shale lithologies (Fig. 3).

Analytical methods

Major and trace elements were analysed at the Geological Survey of Norway (NGU), Trondheim, by a PANalytical Axios at 4 kW X-ray spectrometer. The precision (1σ) is typically around 2% of the major oxide present. For marble samples, elemental concentrations were determined on acidified extracts of whole-rock (cold 10% HCl) by inductively coupled plasma-atomic emission spectrometry (ICP–AES) using a Thermo Jarell Ash ICP 61. Detection limits for Fe, Mg, Ca and Mn are 5, 100, 200 and 0.2 $\mu\text{g/g}$, respectively. The total analytical uncertainty including element extraction (1σ) is $\pm 10\%$ rel.

Stable carbon and oxygen isotope analyses were performed at the Scottish Universities Environmental Research Centre (SUERC) in Glasgow and at the Geological Institute of the Russian Academy of Sciences in Moscow. Approximately 1 mg powder was reacted overnight with phosphoric acid at 70 °C. Ratios were measured on PRISM III, AP2003 and DELTA V mass spectrometers. Repeat analyses of NBS–19 and internal calcite standards are generally better than $\pm 0.2\text{‰}$ for carbon and $\pm 0.3\text{‰}$ for oxygen, and interlaboratory differences are within these ranges. Carbon and oxygen isotopic values are reported and discussed in the conventional delta notation relative to V–PDB and V–SMOW, respectively.

Zircons for U–Pb geochronology were extracted from one crushed sample, using a water table followed by separation in heavy liquids and magnetic field. Zircon crystals selected for radiometric dating were mounted in epoxy together with the reference material, and polished to approximately half thickness. The grains were imaged individually with a panchromatic cathodoluminescence (CL) detector in a variable pressure Scanning Electron Microscope at NGU. U–Pb analyses were performed by Secondary Ion Mass Spectrometry (SIMS) with

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). Red crosses denote occurrences of scapolitised rocks after Frietsch et al. (1997).

Figure 2. (A) Inset map showing location of the study area in northern Norway, and (B) geological map of the AKTW with locations of sulphide deposits/occurrences and studied/sampled sections (denoted from 1 to 8).

Figure 3. Lithostratigraphic column of the Kvenvik and Storviknes formations occurring in the AKTW. Modified after Vik (1985).

the Cameca IMS 1280 instrument at the NORDSIM laboratory in Stockholm, with a primary oxygen beam of c. 15 μm in diameter. Analytical protocols and data reduction follow Whitehouse et al. (1999) and Whitehouse & Kamber (2005). Analyses were calibrated using the 91500 Geostandard reference zircon (1065 Ma, Wiedenbeck et al., 1995), measured at regular intervals. The analyses are corrected for common Pb using the ^{204}Pb signal, if this signal is above background. A Tera-Wasserburg (inverse) concordia diagram was prepared with the ISOPLOT macro for Microsoft Excel (, 2001).

The weighted average $^{207}\text{Pb}/^{206}\text{Pb}$ age of near concordant to concordant analyses is quoted with a 2 sigma error (decay constant uncertainties propagated; systematic uncertainties resulting from interlaboratory experiments not propagated).

Lithology and lithofacies interpretation of studied carbonate units

All studied carbonate rocks are within the Raipas Supergroup and form parts of the Kvenvik and Storviknes formations. In each formation, most of the carbonate units have been sampled in at least two separate sections located in geographically different areas.

The Kvenvik formation, Member A

There are two major carbonate units in Member A of the Kvenvik greenstone formation. The lowermost unit, the Lower dolostone hereafter, has been studied and sampled in the Bergmark area (Section 1; for location, see Fig. 2). Carbonate rocks representing the stratigraphically highest carbonate unit, the Upper dolostone hereafter, have also been studied and sampled in the Bergmark area (Sections 2–3) and along the western coast of Kåfjord (Section 4).

The Bergmark area, Section 1

The geology of the Bergmark area has been comprehensively studied by Vik (1985). In the following sections, we provide new sedimentological observations together with a review of previously published material.

Section 1 records the Lower dolostone unit. It is the lowermost unit in the Kvenvik formation (Fig. 3), and is only partially exposed in the core of a large anticline (Fig. 2). The base of the unit as well as its true thickness remain unknown. The exposed part of the Lower dolostone is less than 100 m thick and is composed of two different lithofacies.

The first lithofacies forms the lower part of the unit. It is typified by red, structureless, massive, dolerudites and dolostones (Fig. 4A, B) with rare, up to 1 m-thick intervals composed of calcitic rocks. All lithologies are recrystallised and intensively silicified. In places, dolostones contain sheets and rafts of green shales, and thin, intensely disrupted/dismembered mudstone layers (Fig. 4C). Several intervals contain clusters of centimetre-size quartz aggregates. Some have cube-shaped crystals (Fig. 4D), whereas others exhibit a spherical and wedge-shaped morphology (Fig. 4E). Red, massive and structureless dolostones are composed of cloudy dolospar (Fig. 4E).

Figure 4. Images of natural outcrops (A, D, G, H), polished slabs (B, C, F) and a thin-section (E) illustrating the main sedimentological features of carbonate rocks of Member A in the Kvenvik formation studied in Section 1 at Bergmark. Lower dolostone (red, massive and structureless dolostone lithofacies): (A) Red, massive, dolostone (sample JS1325; for chemical composition, see Table 4). (B) Red, massive, dolostone with dismembered layers of dark-brown mudstone (arrowed) (sample JS1324). (C) Dolostone containing sheets and rafts of dark-green shale originating from intensely disrupted/dismembered mudstone layers (sample JS1331). (D) Red, structureless dolarenite with abundant cube-shaped, quartz-pseudomorphed crystals of apparent halite (sample JS1325). (E) Photomicrograph in plane-polarised, transmitted light showing quartz-pseudomorphed crystals of probable Ca-sulphate embedded in a cloudy dolospar matrix (sample JS1325). Lower dolostone (thinly-laminated dolostone lithofacies): (F) Pale-yellow, thinly laminated dolostone (sample JS1326). (G) Rhythmically interbedded, decimetre-thick, silicate-rich (grey) and silicate-poor (pale-yellow) dolostone units (sample JS1327). (H) Millimetre-thick lamination in pale-yellow dolostone bed caused by alternation of laminae with a variable ratio of carbonate and non-carbonate components.

Figure 5. Images of natural outcrops (A, B, D, E, H, L–O, Q, R), polished slabs (F, I, P, S–U) and thin-sections (G, J, K) illustrating the main sedimentological features of carbonate rocks of Member A in the Kvenvik formation studied in Sections 2 and 3 at Bergmark. Section 2, Upper dolostone (impure, laminated, dolostone lithofacies): (A) Pale-grey, laminated, albitised, carbonate-silicate rock with abundant porphyroblastic dolomite (sample JS1312). (B) Enlarged image showing dolomite rhombs. (C) Trough cross-bedding in albitised dolarenite (sample JS1315). (D) A bed of massive dolarenite showing pale-brown colour on weathered surface (sample JS1313). (E) Carbonate-silicate rock with abundant nodules composed of albite-rich rims and dolomite-rich cores; modified after Vik (1985). (F) Impure, albitised dolostone with nodules showing dolomite-rich rims whereas the cores have a compositional similarity with the host rock (sample JS1314). (G) Albite felsites composed of an albite-biotite granular mass with porphyroblastic dolomite rhombs (sample JS1311). Section 2, Upper dolostone (bedded dolostone lithofacies): (H) Rhythmically interbedded dolostone-dolomarl-shale cycles typifying a turbiditic Bouma sequence (sample JS1321).

Figure 5 (continued). (I) Variegated dolarenite showing ripples and a small-scale erosional channel (arrowed) (sample JS1319). (J) Enlarged image of the erosional channel showing compositional differences between dark-grey, massive, intensively dolomitised dolarenite and pale-grey, bedded dolarenite filling the erosional channel (sample JS1319). (K) Photomicrograph of the erosional channel in polarised, transmitted light. Note that the channel is filled with fine-grained dolarenite which is devoid of albite, whereas dolarenite layers situated above and below the channel contain abundant albite porphyroblasts though differing in size (sample JS1319). Section 3, Upper dolostone (impure, laminated, dolostone lithofacies): (L) Thinly laminated quartz sandstone with asymmetrical ripples replaced by albite to form albite felsites. (M) Enlarged image detailing thin lamination preserved in the pervasively albitised quartz sandstone. (N) Ripple-marked quartz sandstone altered into albite felsites; hammer head for scale is 15 cm.

Figure 5 (continued). (O) Thin intervals of homogeneous dolostones (pale-brown) embedded in thin-bedded silicate-rich varieties (sample JS1336). (P) Dolarenite exhibiting small-scale erosional channels and small-scale trough cross-beds (sample JS1334). Section 3, Upper dolostone (bedded dolostone lithofacies): (Q) Rhythmically interbedded dolostone and sandy/shaly dolostone couplets (sample JS1341). (R) Fine, parallel lamination seen on weathered surface of a dolostone bed (sample JS1337). (S) Parallel-laminated dolostone appears massive on the polished surface (sample JS1337). (T) Small-scale erosional channel in pale-green dolostone (sample JS1339). (U) Dolostone showing flaser bedding (sample JS1342).

Figure 6. Lithostratigraphic columns and C-isotope profiles of sections studied in the Kåfjord area.

Lithofacies interpretation. Vik (1985) assigned this lithofacies to tidal channel deposits developed on a tidal flat. However, the lack of cross-bedding is in conflict with the above interpretation. Moreover, cube-shaped crystals of quartz resemble pseudomorphed halite crystals (Fig. 4D), whereas quartz aggregates with the spherical and wedge-shaped morphology are interpreted as crystals

and aggregates of Ca-sulphates (Fig. 4E). This, together with the general lack of bedding/layering, widespread red beds and intensely disrupted layers (desiccated?) are all indicative of deposition in an oxic, coastal sabkha environment.

The second lithofacies comprises the upper part of the

Lower dolostone. The contact with red, massive dolostones is not exposed. The lithofacies is typified by pale-yellow, thinly laminated dolostones (Fig. 4F). Some intervals exhibit a decimetre-thick, rhythmic bedding of silicate-rich and silicate-poor dolostone units (Fig. 4G). Similarly, the millimetre-thick lamination is also expressed by a variable ratio of carbonate and non-carbonate components (Fig. 4H). The laminated dolostones contain numerous, 0.1–1.0 m-thick, fine-grained, thinly laminated, albite-rich beds locally termed albite felsites. They are concordant with the host sedimentary rocks (Vik, 1985). The laminated dolostones are impure and contain mica, rounded and angular quartz grains and abundant albite appearing as a late growth rather than as clastic grains. Rare beds composed of sparitic calcite with minor clastic quartz, albite porphyroblasts and small fragments of quartz sandstone are present.

Lithofacies interpretation. Vik (1985) interpreted the laminated dolostone lithofacies as intertidal deposits accreted on a carbonate platform. Sedimentological features briefly described above corroborate the intertidal depositional setting; however, a high degree of dolostone impurity is more consistent with a carbonate shelf significantly affected by siliciclastic input. Vik (1985) noted that the origin of the albite felsites remains enigmatic.

The Bergmark area, Section 2

Section 2 records the Upper dolostone unit. It is separated from the Lower dolostone by a c. 50 m-thick gabbro sill fringed at its top by c. 10 m-thick layer of albite felsite (Fig. 3). The Upper dolostone was sampled in the eastern limb of a syncline close to its hinge (Fig. 2) where the exposed part of the Upper dolostone is over 70 m thick. Vik (1985) recognised two main lithofacies in the Upper dolostone: (i) an impure, laminated dolostone (c. 35 m thick), forming the base of the unit; and (ii) a bedded dolostone (c. 40 m thick) representing the upper part of the succession.

The impure, laminated dolostone lithofacies consists of pale-grey rocks with millimetre-scale lamination (Fig. 5A) and abundant porphyroblastic dolomite rhombs (Fig. 5B). Vik (1985) reported that the parallel lamination is caused by a gradual transition between two mineral parageneses: (i) dolomite+albite+biotite±(quartz, calcite, chlorite); and (ii) quartz+dolomite+biotite±(calcite, chlorite). Planar and trough cross-bedding (Fig. 5C), and convoluted lamination (Vik, 1985) have also been observed. Some sandy beds show nicely preserved ripple marks (Vik, 1985). The microfabric of the rock is defined by a microcrystalline, granular texture. The impure, laminated dolostones are commonly albitised and may contain numerous thin beds of albite felsites. Some intervals incorporate decimetre-thick beds of a pale-brown, massive dolarenite (Fig. 5D). A c. 3 m-thick interval, occurring in the middle part of the section, contains numerous carbonate-silicate concretions. One type of nodule has an albite-rich rim and dolomite-rich

core (Fig. 5E), whereas another has a dolomite-rich rim with a core showing no mineralogical differences with respect to the host rock (Fig. 5F). The albite felsites are composed of albite and biotite intergrown into a microcrystalline, granular mass with porphyroblastic dolomite rhombs (Fig. 5G). The rocks have a millimetre-scale parallel and wavy lamination.

The bedded dolostone lithofacies is typified by centimetre-to decimetre-thick cycles composed of dolostone-dolomarl-shale triplets resembling a turbiditic Bouma A–B sequence (Fig. 5H). Thicker and more homogeneous dolostone beds are also present. The uppermost part of the section contains dolostone beds alternating with siltstones overlain by limestones and graphitic schists (Vik, 1985). The dolostones contain numerous beds of albite felsite. The bedded dolostone lithofacies is characterised by weak, parallel bedding and graded bedding. Current ripples, small-scale erosional troughs and a flat-laminated, probable stromatolitic structure are other features of this lithofacies (Vik, 1985).

The bedded dolostones are composed of dolomite and biotite intergrown into a microcrystalline, granular mass containing abundant porphyroblasts of albite. In some beds, the density of porphyroblasts and their size are controlled by primary sedimentological features such as lamination and channelling (Fig. 5I–K). This suggests that the distribution and size of the porphyroblasts were apparently inherited from those of original clastic grains during albitisation. Vik (1985) also reported beds with magnetite impregnations and scapolite porphyroblasts. The beds with scapolite porphyroblasts are 0.5–1 m thick, and the amount of visually determined scapolite is up to 50% by volume.

Facies interpretation. Vik (1985) suggested that the two lithofacies of the Upper dolostone represent a continuous sequence of carbonate accumulation. The slope of a carbonate platform was inferred for deposition of the dolostone-shale lithofacies with Bouma cycles, whereas a deeper water setting, distant from the ‘carbonate factory’, was suggested for accumulation of the impure, laminated dolostone lithofacies. Dolostone beds with probable stromatolitic structures terminating the carbonate-shale sequence were assigned to intertidal algal mat deposits. A source for the non-carbonate components and the high degree of impurity of all carbonate rocks were interpreted by Vik (1985) to relate to a probable synchronous volcanism or redeposition and long transport of volcanic ash material.

The Bergmark area, Section 3

Section 3 is a river-bed section located in the eastern limb of the syncline, 1.8 km to the south of Section 2 (Fig. 2). Here, rocks dip steeply to the east implying tectonic inversion of the limb. The two lithofacies documented and sampled in Section 3 are somewhat similar to those described above in Section 2.

Figure 7. Images of natural outcrops illustrating the main sedimentological features of carbonate rocks and associated lavas and tuffs in Member B of the Kvenvik formation studied in Section 4 at Kåfford. (A) Sandstone-siltstone-shale with lenticular bedding (black bar) and bedded tuff (red bar) overlain by amygdaloidal, tholeiitic basalt (white bar); Hammer head for scale is 15 cm. (B) Purple dolarenite (white bar) overlain by pale-brown, sandy dolarenite with slumped mudstone-siltstone beds; lens cap for scale is 5 cm in diameter. (C) Massive dolarenite overlain by thickly bedded dolarenite passing upward into thinly bedded varieties with siltstone interlayers. (D, E) Impure, sandy dolostones showing graded bedding, small-scale trough cross-lamination, and abundant, small-scale channelling; lens cap for scale is 5 cm in diameter. (F) Dolarenites (red bars) interbedded with purple siltstone/shale; note that the dolarenites occur as continuous beds, ripples and channel infills, whereas the interbedded siltstones are characterised by low-angle cross-lamination. (G) Variegated siltstone with parallel bedding. (H) Graded bedding in pink sandstone-siltstone.

The impure, laminated, dolostone lithofacies rests with a sharp base on albite felsites. The latter show a thin, parallel bedding with some intervals preserving asymmetrical ripples (Fig. 5L–N). The impure, laminated dolostones are pale-grey, beige and pale-green rocks with millimetre-scale lamination. This lithofacies contains numerous beds of homogeneous dolostones (5 to 15 cm thick) embedded into thinly bedded silicate-rich varieties (Fig. 5O). Their primary sedimentological features are defined by numerous small-scale erosional channels and stacks of small-scale, trough cross-beds (Fig. 5P).

The bedded dolostone lithofacies is characterised by rhythmically bedded units composed of 2–10 cm-thick, dolostone-sandy/shaly dolostone couplets (Fig. 5Q). They contain irregularly spaced, thicker beds of dolostones showing a fine, parallel lamination on weathered surfaces (Fig. 5R) with homogeneous, massive fabrics in polished

slabs (Fig. 5S). Such dolostone beds contain intervals with small-scale, erosional channels, cross-bedding (Fig. 5T) and flaser bedding (Fig. 5U).

Lithofacies interpretation. The sedimentological features observed in Section 3 suggest depositional settings similar to those inferred for rocks documented in Section 2.

Kåfjord western coast, Section 4

Section 4 intersects a dolostone unit located in the uppermost part of Member A of the Kvenvik formation. This has been studied on the western coast of Kåfjord (Fig. 6, Profile A). According to Vik (1985), the studied succession represents a time equivalent of the Upper dolostone unit studied in Sections 2 and 3. The sedimentary unit represents a mixed carbonate-siliciclastic succession composed of several carbonate-siltstone cycles of

variable thicknesses, which pass upward into calcareous greywacke-chlorite schist cycles. The carbonate-dominated cycles are separated from the siliciclastic ones by a mafic lava flow (Fig. 7A; Fig. 6, Profile C). The total thickness of the carbonate-siliciclastic succession is close to 100 m; however, it is only partly accessible and its suitability for sampling is limited to Profile D (Fig. 6).

Section 4 is a composite section consisting of two marginally overlapping profiles on the western side of Kåfjord (Fig. 2). The lowermost part of the sampled interval, Profile B hereafter, located on a partially accessible cliff, is underlain by a siltstone breccia which is intruded by a gabbro body (Fig. 6, Profile A). The upper part of Section 4, Profile C hereafter (Fig. 6), was sampled on the Kåfjord coast. The top of Profile B marginally overlaps with the base of Profile C; however, a strike distance of over several hundred metres between the two prevents their precise correlation.

Profile B represents a succession typified by several centimetre-thick, grey dolarenite/pink mudstone/siltstone couplets showing lenticular, wavy and flaser bedding. Sampled dolarenite units are either massive or bedded and may show indistinct vertical grading in clast size, whereas the mudstone-siltstone beds are characterised by small-scale, low-angle cross-lamination. One bed displays soft-sediment deformation structures caused by slumping (Fig. 7B).

Profile C starts with the grey dolarenite/pink mudstone/siltstone succession described above. This is followed by a 1 m-thick bed of grey, thick-bedded greywacke sharply overlain by a c. 12 m-thick unit composed mainly of both thick- and thin-bedded dolarenites with two intervals of thinly interbedded dolostone-siltstone, each 1 to 2 m thick. The primary textural pattern of the dolarenites varies from massive through thick-bedded to thin-bedded with siltstone interlayers (Fig. 7C). Many intervals show graded bedding, small-scale, trough cross-lamination and abundant, small-scale, channelling (Fig. 7D, E). All dolarenite beds are sharp-based. Thick-bedded dolostones display flat top surfaces, whereas the thin-bedded varieties with siltstone interbeds show rippled surfaces. Some dolarenite beds retain their thickness over several metres, whereas others occur as ripples or channel infills. Interbedded siltstones are characterised by low-angle cross-lamination, parallel bedding and graded bedding (Fig. 7F–H).

Facies interpretation. The observed sedimentological features of the sampled dolarenite-siliciclastic succession suggest that sedimentation occurred in a carbonate shelf environment transitional from the lower intertidal (massive to thickly bedded dolarenite) to the upper intertidal (interbedded dolarenite-siltstone with lenticular, wavy and flaser bedding) zone.

The Kvenvik formation, Member B

The Badderer area, Section 5

The carbonate rocks sampled at Badderer (Fig. 2, Section 5) represent a c. 100 m-thick succession of carbonate rocks with minor shale beds exposed along a road-cut. This sedimentary succession, the Uppermost dolostone hereafter, is sandwiched between two thick piles of mafic lava flows. The carbonate rocks occurring at the base of the unit are composed of microcrystalline dolomite with abundant quartz particles, tremolite needles and biotite porphyroblasts. The rocks display parallel bedding which is visible only on weathered surfaces (Fig. 8A), or show only a faint parallel lamination on polished surfaces (Fig. 8B).

The carbonate rocks of the middle and upper parts of the Uppermost dolostone are composed of microsparitic dolomite and calcite, and contain abundant biotite porphyroblasts, tremolite needles and quartz particles. The rocks preserve parallel bedding and a faint, millimetre-scale, low-angle cross lamination (Fig. 8C, D). Associated beds of siliciclastic rock show well-developed parallel bedding expressed by 1–2 cm-thick siltstone layers draped by thinner units of dark-grey mudstone (Fig. 8E).

Thick piles of mafic volcanic rocks underlying and overlying the Uppermost dolostone imply deposition in an active volcanic environment. The overall sedimentological features of both carbonate and siliciclastic rocks suggest deposition in a deep-marine setting, apparently from turbidity currents.

The Storviknes formation

Three major carbonate units of the Storviknes formation have been sampled in three different locations: (i) at Storvik, Section 6; (ii) at Raipas, Section 7; and (iii) at Borras, Section 8 (Fig. 2). Although these three sections are located in different areas, they represent correlative, supposedly time-equivalent successions (Vik, 1985).

Storvik, Section 6

At Storvik, the carbonate-dominated succession, the Dolostone member hereafter, rests on mafic tuffs of the Kvenvik formation and is overlain by grey siltstones. The reported total thickness of the member may reach 200 m (Fig. 6), but it is not possible to make accurate measurements. Section 6 represents only the lower part of the Dolostone member (Fig. 6, Profile E) where the dominant rocks are bedded dolostone, syndepositional breccia and purple siltstone. The bedded dolostone and purple siltstone are present throughout the section whereas the dolostone breccia has only a localised development. The sedimentary succession is gently folded and all lithologies are variably recrystallised.

The bedded dolostone (Fig. 9A) preserves a large array of primary sedimentological features including parallel

Figure 8. Images of natural outcrops (A, D, E) and polished slabs (B, C) illustrating the main sedimentological features of carbonate rocks of Member B in the Kvenvik formation studied in Section 5 at Badderen. (A) Parallel bedding in microcrystalline dolostone visible on a weathered surface (sample JS1310). (B) A polished slab of microcrystalline dolostone showing faint bedding (sample JS1308). (C) Green, microcrystalline dolostone characterised by faint, millimetre-scale, low-angle cross-lamination and a 'roll structure' caused by bedding-parallel infiltration of a post-depositional fluid (sample JS1306). (D) Pale-grey, microcrystalline dolostone with thick, parallel bedding (sample JS1303). (E) Brownish-grey, siltstone beds draped by dark-grey mudstone.

and wavy lamination, flaser bedding, silicified flat-laminated stromatolites, ripple marks and chert nodules. Several dolostone beds reveal a clastic texture and are composed of sand-size particles of dolostone, quartz and

red jasper, hence representing a dolarenite. Interbeds of purple siltstone are 0.5 to 5 cm thick, partially silicified, and commonly exhibit wavy and low-angle cross lamination.

Figure 9. Images of natural outcrops (A–F) and a thin-section (G) illustrating the main sedimentological features of carbonate rocks of the Storviknes formation studied in Section 6 at Storvik (Profile E in Fig. 6). (A) Bedded dolostone forming the bulk of the Storviknes formation in the Storvik area. (B) Clast-supported breccias composed of angular, unsorted fragments of dolostone (pink, brown and white) and chert (red, black and grey). (C) Matrix-supported breccias composed of blocks of crystalline dolostone. (D) Carbonate breccias-greywacke lithofacies represented by two sets of trough cross-beds. Note upward grading from a breccia through a grit-sand into a silt-mud. (E) Breccia beds each separated by thin, black, hematite-rich, mud drapes. (F) A slump structure in a sand bed sandwiched between two beds of breccias. (G) Photomicrograph in polarised, transmitted light showing sandstone composed of rounded grains of quartz and chert embedded in a quartz-sericite matrix.

A large, c. 35 m-thick, lensoidal body of syndepositional breccias (Fig. 9B–G) occurs in the lower part of the sampled succession. A quarry with near-vertical smooth walls cut through the breccias provides an excellent exposure even though the basal contact of the lens is not exposed. Two main lithofacies are recognised in the breccia body: (i) carbonate debris lithofacies; and (ii) carbonate breccia lithofacies. The former is c. 2 m thick and forms the base of the breccia body. This lithofacies is composed of angular, unsorted blocks and fragments of two rock types. The majority of the clasts are pink, brown and white, crystalline, massive dolostones. Some dolostone clasts show pink rims (Fig. 9B). Clasts of red, black and grey cherts are less abundant. The clasts range in size from less than 1 cm to 25 cm (Fig. 9B, C). Although the majority of clasts appear to be sub-rounded, some of the fragments are angular. In general, the carbonate debris is clast-supported and poorly sorted (Fig. 9B, C).

The carbonate breccia lithofacies is represented by numerous sets of decimetre-thick, trough cross-beds. Many individual beds show breccias grading upwards into a grit-sand and then into a silty mud (Fig. 9D). Where beds do not show grading they are separated by thin, black, hematite-rich, mud drapes (Fig. 9E). Several beds display soft-sediment deformation structures associated with slumping (Fig. 9F). The clast composition in the breccias and gritstones remains unchanged vertically and laterally. These are mainly dolostones with minor cherts. However, sandstone and siltstone members in all cross-beds are largely composed of rounded grains of quartz and chert supported by a quartz-sericite matrix (Fig. 9G).

Facies interpretation. Primary sedimentary structures and textures preserved in the dolostones are consistent with their accumulation on a carbonate shelf/platform. The carbonate breccias can be confidently assigned to a channel deposit. Two distinctly different sources of clastic material (dolostone intraclasts and quartz sand), a low degree of roundness and the unsorted nature of dolostone intraclasts can be reconciled with a depositional model involving the formation of a high-relief, shore-to-basin, fault scarp followed by the development of a channel, with subsequent, long-distance transport of quartz sand from the continent-basin margin, which was redeposited together with locally derived dolostone intraclasts on the carbonate shelf/platform. The red colour of the dolostones and shales suggests oxic environments.

Raipas, Section 7

At Raipas, the Storviknes formation has a thickness of over 500 m. The formation is composed of purple and grey shales with a c. 200 m-thick unit of dolostones (the Dolostone member) in the middle (Fig. 10). The studied and sampled part of the formation (Section 7) represents a 100 m-thick upper part of the Dolostone member. It includes purple siltstone lithofacies and three carbonate lithofacies: (i) a layered dolostone; (ii) a massive dolostone;

and (iii) a stromatolitic dolostone. All rocks are variably recrystallised.

The layered dolostone is a dominant carbonate lithofacies occurring as a series of planar beds 2–15 cm thick. Individual beds are separated by thin layers of purple and grey shales (Fig. 11A). The layered dolostones display a planar, gently buckled and wrinkled lamination resembling that observed in flat-laminated algal mats (e.g., Demicco & Hardie, 1994). The shale interlayers are intensely silicified (Fig. 11B) and preserve abundant ripples. The overall structural pattern of the bedded dolostone lithofacies is suggestive of stromatolitic biostromes.

The massive dolostone lithofacies occurs as several cupola-like bodies within the layered dolostones. The cupola-like bodies are sharp-based and separated from the layered varieties by thin beds of purple and grey shales with ripple marks. The upper contact is equally sharp but may lack the shale intercalations (Fig. 11B). The massive appearance, lack of lamination and clotted fabrics of this lithofacies superficially resemble those commonly observed in thrombolitic bioherms (cf., Aitken, 1967).

The stromatolitic dolostone lithofacies occurs as columnar and domal stromatolites forming biostromes. Large columns of tightly packed stromatolitic columns with poorly preserved fabrics are separated from each other by intensely silicified shales (Fig. 11C). The biostromes are 1 to 2 m thick and show a gradual vertical replacement by the layered dolostone lithofacies.

The purple and grey shale lithofacies associated with the dolostone lithofacies occurs as lenses (ripples) and thin discontinuous layers and beds with a maximum thickness of 50 cm (Fig. 11D). The shale units have sharp contacts with the host dolostones and are characterised by a combination of planar parallel lamination, wavy and low-angle cross-lamination.

Facies interpretation. Vik (1985) suggested that deposition of the carbonate lithofacies in the Raipas area occurred in intertidal settings. Our observations are in accord with this previous interpretation.

Borras, Section 8

The Upper dolostone at Borras is only 25 m thick. As reported by Vik (1985), the unit is generally composed of layered and massive, crystalline dolostones with siltstone breccias at the base. The area is poorly exposed; consequently, samples obtained in Section 8 represent only two separate outcrops and actually comprise crystalline limestones.

Figure 10. (A) Lithostratigraphic column of the Storviknes formation. (B) Lithological column detailing the sampled portion of the Storviknes formation combined with stratigraphic variations of Mn/Sr, Mg/Ca, C- and O-isotopic ratios.

Radiometrically dated igneous rock from the Kvenvik formation at Kåfjord

In the AKTW, all igneous rocks occur exclusively in the Kvenvik formation (Fig. 3). These are mainly mafic lavas and mafic tuffs as well as intrusive gabbro-dolerite sills and ultramafic bodies. Most of the gabbro bodies are intruded into the lower part of the formation, whereas mafic lavas are abundant in its upper part. The upper part of the formation studied in detail by Bergh & Torske (1988) contains massive, amygdaloidal and pillowed lavas, pillow breccias and hyaloclastic rocks (Fig. 12A). Volcaniclastic facies are represented by ash, accretionary lapilli tuff and tuff beds, and volcaniclastic rocks (Fig. 12B). The volcanic rocks are intruded by sills

of gabbro (Fig. 12C). Bergh & Torske (1988) classified the Kvenvik volcanic rock association as a MORB-type, tholeiitic basalt. They demonstrated that the lavas and volcaniclastic rocks were deposited in cyclically repeated couplets whose deposition was assigned to a continental rift setting in shallow-water to subaerial environments (Bergh & Torske, 1988).

Mineralogically, all newly obtained and analysed samples are actinolitized and exhibit a variable degree of saussuritisation and albitisation of primary plagioclase, resulting in a high Na_2O content observed in some of the samples (BBF -031, JS1411; Table 1). The albitised samples also show enrichment in Fe. Chemically, all analysed magmatic rock samples are tholeiitic basalts (Fig. 13A) with moderate MgO , TiO_2 , Ni and Cr contents

(Tables 1 & 2). On a Ti–Zr–Y discrimination diagram, all analysed samples plot as ‘MORB’, transitional to ‘within-plate’ and ‘island-arc’ tholeiites, thus providing no unique solution (Fig. 13B). On a chondrite-normalised rare-earth element (REE) diagram, two gabbro samples show a positively sloped pattern suggesting a depleted mantle source (Fig. 13C). Other samples exhibit a negatively sloped pattern, hence enrichment in light REEs. This group also includes a dated gabbro collected in a road-cut along the western shore of Kåfjord. The geochemical similarity between lavas and the dated gabbro suggests that they originated from the same parental magma. The positive Eu anomaly visible in the REE pattern of the dated gabbro is consistent with its plagioclase-rich nature and plagioclase accumulation. The overall enrichment in light REEs in both lavas and gabbro could arise either from fractional crystallisation, a small degree of partial melting or crustal contamination.

The radiometric U–Pb age determination on zircon was conducted on a thick gabbro body hosting the historical Kåfjord copper mine and exposed on the western side of Kåfjord (Figs. 2 & 3). Here, a large road-cut exposure exhibits a leucocratic, coarse-grained gabbro showing an unfoliated ophitic texture (Fig. 12C). Plagioclase laths, commonly 1 cm long, are embedded in a coarse-grained interstitial amphibole matrix. The latter shows a significant alteration. Mafic minerals are partially replaced by calcite, epidote and chlorite. The rock contains accessory chalcopyrite and magnetite.

The sample collected for zircon separation and age determination represents a homogeneous gabbro, though containing a few calcite-chalcopyrite veinlets. Interestingly, Bergh & Torske (1988, their fig. 3 on p. 230) mapped this gabbro as a “*coarse-grained, massive (mafic) lava*”.

U–Pb radiometric age of the gabbro and coeval mafic lavas of the Kvenvik formation

With 100 ppm zirconium, the gabbro sample collected on the western side of Kåfjord (BBF–031) contains prismatic to stubby zircon crystals up to 200 µm long. All of the twenty-four non-metamict zircon crystals extracted show a consistent morphology and are characterised by a weak, planar oscillatory zoning and absence of a core-rim texture. Some crystals have cracks filled with luminescent zircon. Thirteen U–Th–Pb analyses of ten oscillatory-zoned zircon crystals are concordant to near-concordant (concordance > 96%) and yield a well-clustered $^{207}\text{Pb}/^{206}\text{Pb}$ weighted average age of 2146 ± 5 Ma (Fig. 14, Table 3). Two additional analyses were probably affected by some radiogenic Pb loss and therefore are not included in the average calculation. The oscillatory zoning and Th/U ratio in the range between 0.3 and 1.5 are characteristic of magmatic zircon, and therefore the age of 2146 ± 5 Ma is interpreted as the magmatic crystallisation age of the gabbro body and the age of the co-magmatic lavas higher in the stratigraphy of the Kvenvik formation (Member B).

Table 1. Content of major elements and loss on ignition in igneous rocks of the Kvenvik formation based on XRF analyses.

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	LOI	Sum
wt%												
Gabbro												
JS1411	49.90	10.20	22.80	2.33	2.95	4.50	5.73	0.28	0.10	0.18	0.08	99.00
JS1412	48.20	14.10	14.70	1.22	6.79	7.29	3.75	0.81	0.22	0.09	2.07	99.20
JS1413	47.60	13.70	13.30	1.30	6.41	7.64	4.35	0.94	0.22	0.09	3.44	99.00
JS1417	50.50	13.70	13.40	1.50	6.44	9.34	2.49	0.53	0.20	0.12	1.64	99.90
JS1419	48.70	13.40	12.40	0.65	8.22	9.41	3.22	0.20	0.32	0.04	2.41	99.00
BBF-031	49.00	11.80	18.00	2.00	4.12	6.23	5.54	0.61	0.15	0.15	1.31	98.90
Tholeiitic basalt and mafic tuff*												
JS1415	47.10	14.00	15.10	1.25	6.46	8.80	3.13	0.23	0.25	0.09	2.11	98.50
JS1416*	48.50	15.50	15.10	1.01	6.70	2.99	1.89	2.03	0.16	0.06	5.62	99.60
JS1418	49.30	13.30	13.60	1.56	6.37	11.30	1.68	0.32	0.19	0.11	1.85	99.60
JS1420	50.10	13.70	13.80	1.60	6.38	7.00	4.10	0.30	0.21	0.12	1.98	99.30
JS1430	48.60	13.00	14.00	1.60	6.01	10.80	2.17	0.26	0.20	0.12	1.96	98.70
JS1431	49.00	13.90	13.30	1.43	6.92	9.73	2.39	0.51	0.20	0.12	2.00	99.50

[illegible]

The Kvenvik formation

Geochemical characteristics of the studied carbonate formations are based on 52 whole-rock analyses (Table 4). Fifty out of the fifty-two analysed samples from all dolostone units in the Kvenvik formation contain a significant amount of silica ($\text{SiO}_2 = 29 \pm 13$ wt.%; Fig. 15A). The albitised varieties in the Bergmark area show high Na₂O contents (Fig. 15C, Table 4).

The Lower dolostone unit is composed of slightly calcitised dolostones ($\text{Mg}/\text{Ca} = 0.55\text{--}0.58$, $n = 7$) and minor limestones ($\text{Mg}/\text{Ca} = 0.009\text{--}0.01$, $n = 3$). Mn/Sr ratios are elevated in the dolostones (15–41) and low in the limestones (0.8–2.8). One of the limestone samples exhibits over 3 wt.% Ba in the form of barite.

The Upper dolostone is composed mainly of calcitised dolostones ($\text{Mg}/\text{Ca} = 0.34\text{--}0.68$, averaging at 0.48 ± 0.08 , $n = 20$) with elevated Mn/Sr ratios (19–41). Two samples show a Mg/Ca ratio of 0.73 and 0.83 (Table 4, Fig. 15C), hence exceeding the 0.62 ratio of stoichiometric dolomite, and thus may contain a magnesite phase. These samples have high Mn/Sr ratios (35–72).

The Uppermost dolostone unit consists of intensely calcitised dolostones ($\text{Mg}/\text{Ca} = 0.03\text{--}0.46$, $n = 10$) having low Mn/Sr ratios (4–13, Table 4). It remains unresolved whether or not two samples with Mg/Ca ratios of 0.03 and 0.05 represent original limestones or intensely calcitised dolostones.

All carbonate rocks in the Lower, Upper and Uppermost dolostone units are rich in ^{13}C . The $\delta^{13}\text{C}$ values cluster tightly between +5.4 and +8.5‰ (+7.4 $\pm$ 0.7‰, n = 51) with one outlier at +1.7‰ (Fig. 15D, Table 4). In contrast, $\delta^{18}\text{O}$ values are low (15.9 $\pm$ 2‰, n = 51) and show a significant fluctuation between 13 and 23.3‰. The lowest $\delta^{13}\text{C}$ of +1.7‰ is coupled with a low $\delta^{18}\text{O}$ value of 14.1‰. If considered either all together or at the level of an individual formation/section, neither $\delta^{13}\text{C}$ nor $\delta^{18}\text{O}$ values show any significant correlation with SiO_2 and Na_2O abundances, nor with Mg/Ca or Mn/Sr (Fig. 15). There is no significant correlation between $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$.

Figure 11. Images of natural outcrops illustrating the main sedimentological features of carbonate rocks of the Storviknes formation studied in Section 7 at Raipas. (A) Layered and structureless dolostones separated by dark-grey shale and calcitised interval (white). The layered dolostone consists of planar beds separated by thin layers of purple and grey shales (red arrowed). (B) A sharp-based, cupola-like body of structureless dolostone sandwiched between layered dolostones. The cupola-like body shows a massive appearance and resembles thrombolitic bioherms. The layered dolostone contains several interlayers of intensely silicified shale (red arrowed). (C) A biostrome composed of tightly packed, columnar and domal stromatolites with poorly preserved primary fabrics. Note that individual stromatolitic columns are separated from each other by intensely silicified shales (white). (D) Ripple-marked shale (overgrown with lichen) in sharp contact with a cross-bedded dolarenite containing shale layers with rippled surfaces.

Table 3. SIMS zircon U–Pb geochronological data for gabbro sample BBF-031, corrected for common Pb.

Spot #	Lab #	Sel	U	Th	Pb	$\frac{^{206}\text{Pb}}{^{204}\text{Pb}}$	$f_{^{206}}$	$\frac{^{207}\text{Pb}}{^{235}\text{U}}$	σ	$\frac{^{206}\text{Pb}}{^{238}\text{U}}$	σ	rho	Disc.	$\frac{^{238}\text{U}}{^{206}\text{Pb}}$	σ	$\frac{^{207}\text{Pb}}{^{206}\text{Pb}}$	σ	$\frac{^{206}\text{Pb}}{^{238}\text{U}}$	σ		
(a)	(b)	(c)	ppm	ppm	ppm	measured	%	(d)	%	%	%	(e)	(f)	%	%	%	Ma	Ma			
K3-01a	n4855	NS	29	12	14	4872	0.38	6.9656	1.69	0.3880	0.97	0.58	0.7	2.577	0.97	0.1302	1.38	2101	24	2114	18
K3-02a	n4855	S	337	502	210	67996	0.03	7.3862	1.05	0.3995	0.99	0.94	0.8	2.503	0.99	0.1341	0.36	2152	6	2167	18
K3-03a	n4855	S	143	90	75	33544	0.06	7.3453	1.13	0.4006	0.98	0.87	1.9	2.496	0.98	0.1330	0.56	2138	10	2172	18
K3-04a	n4855	S	244	273	141	135657	0.01	7.2167	1.65	0.3941	1.58	0.96	0.3	2.538	1.58	0.1328	0.47	2136	8	2142	29
K3-06a	n4855	S	171	164	95	>1e6	{0.00}	7.2636	1.14	0.3947	1.02	0.89	0.0	2.534	1.02	0.1335	0.52	2144	9	2144	19
K3-06b	n4855	S	174	247	104	160331	{0.01}	7.1987	1.09	0.3895	0.97	0.89	-1.7	2.567	0.97	0.1340	0.50	2152	9	2121	18
K3-08a	n4855	S	24	8	11	87543	{0.02}	7.2102	1.71	0.3863	1.06	0.62	-3.4	2.589	1.06	0.1354	1.35	2169	23	2106	19
K3-08b	n4855	S	126	105	70	68971	{0.03}	7.5127	1.19	0.4058	1.01	0.86	2.2	2.464	1.01	0.1343	0.61	2155	11	2196	19
K3-14a	n4855	S	167	146	93	129982	{0.01}	7.4380	1.10	0.4039	0.99	0.90	2.3	2.476	0.99	0.1336	0.48	2145	8	2187	18
K3-15a	n4855	S	388	293	213	82760	0.02	7.5117	1.03	0.4084	0.97	0.94	3.5	2.449	0.97	0.1334	0.36	2143	6	2207	18
K4-01a	n4864	S	100	141	60	157421	{0.01}	7.2686	1.01	0.3920	0.79	0.79	-1.4	2.551	0.79	0.1345	0.62	2157	11	2132	14
K4-03a	n4864	S	39	24	21	31973	{0.06}	7.4600	1.15	0.4044	0.81	0.71	2.3	2.473	0.81	0.1338	0.81	2148	14	2189	15
K4-03b	n4864	NS	51	38	27	58000	{0.03}	7.3382	1.07	0.4049	0.80	0.75	4.1	2.470	0.80	0.1314	0.70	2117	12	2192	15
K4-05a	n4864	S	159	192	95	>1e6	{0.00}	7.4416	0.89	0.4037	0.79	0.89	2.2	2.477	0.79	0.1337	0.40	2147	7	2186	15
K4-09a	n4864	S	415	478	243	147271	0.01	7.3545	0.83	0.4000	0.79	0.95	1.5	2.500	0.79	0.1334	0.26	2143	4	2169	15

(a) Analysis identifier.

(b) Nordsim laboratory sample identifier.

(c) (S) analysis selected for weighted average $^{206}\text{Pb}/^{207}\text{Pb}$ age calculation; (NS) analysis not selected.(d) Proportion in % of ^{206}Pb of common Pb derivation, value between bracket indicates that ^{204}Pb is at background level and that no common Pb correction was performed.

(e) Error correlation in conventional concordia space.

(f) Errors are 2-sigma, propagated using the algorithms of Schmitz and Schoene (2007).

GPS coordinates, UTM WGS84, zone 34: 578330–7761120; Geodetic: long: 23.04707°, lat: 69.94559°

Figure 12. Selected images of igneous rocks comprising the Kvenvik formation in the Kåffjord area. (A) Tholeiitic pillow lava. Hammer head for scale is 14 cm. (B) Cross-bedded, volcaniclastic sandstone sandwiched between two lava flows. (C) A gabbro sill sampled for the radiometric age determination.

The Storviknes formation

Geochemical characteristics of the Storviknes formation carbonate rocks are based on 41 whole-rock analyses (Table 4). In contrast to the Kvenvik formation, all analysed samples contain much less silica ($\text{SiO}_2 = 7.3 \pm 6.5$; Fig. 15A) and are devoid of Na_2O , excluding two samples having 0.19 and 0.43 wt.% Na_2O . In the Kåffjord and Raipas areas, all analysed samples are dolostones with an average Mg/Ca ratio of 0.62 ± 0.01 ($n = 36$) which corresponds to that of stoichiometric dolomite. In contrast, all analysed samples from the Borrás area are calcitic with Mg/Ca ratios clustering tightly between 0.006 and 0.013 ($n = 5$). Mn/Sr ratios in all lithologies from all studied areas are low (10.6 ± 5.4 , $n = 40$) with one dolostone outlier at 122.

The carbonate rocks in the Storviknes formation are characterised by $\delta^{13}\text{C}$ fluctuating around zero (-3.1 to $+2.8\text{‰}$) with an average value of $+1.1 \pm 1.2\text{‰}$ ($n = 41$, Table 4). Three out of four negative $\delta^{13}\text{C}$ values are associated with the Borrás limestones (-3.1 , -2.3 and -1.9‰), and one with the Raipas dolostone (-0.9‰). The latter is marked by the highest Mn/Sr ratio of 122

(Fig. 15E). The $\delta^{18}\text{O}$ values in all dolostones are high ($22.3 \pm 1.1\text{‰}$, $n = 36$) which also represents a significant difference from the dolostones of the Kvenvik formation which are generally ^{18}O -depleted and ^{13}C -rich (Fig. 15D). The Borrás limestones are marked by low $\delta^{18}\text{O}$ values ranging between 14.7 and 19.3‰ ($16.5 \pm 2.3\text{‰}$, $n = 5$). The $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values do not exhibit any significant correlation except for the Borrás limestones. In all studied cases, neither of these isotopic compositions correlates significantly either with SiO_2 abundances or with Mg/Ca or Mn/Sr ratios (Fig. 15).

Geochemical screening of C isotopes for post-depositional alteration

Diagenetic, hydrothermal and metamorphic fluids have low water/rock ratios for carbon, but high water/rock ratios for oxygen (e.g., Hudson, 1977; Banner & Hanson, 1990; Nabelek, 1991; Land, 1992; Jacobsen & Kaufman, 1999). In rare cases, depletion in both oxygen and carbon isotope values has been reported from carbonate rocks that experienced a significant deformation (e.g., Guerrero

Figure 13. Geochemical characteristics of igneous rocks of the Kvenvik formation. (A) An AFM diagram from Irvine & Baragar (1971). (B) A Ti–Zr–Y diagram from Pearce & Cann (1973) (C) A REE diagram with chondrite-normalised values from Sun & McDonough (1989).

Figure 14. (A) Tera-Wasserburg (inverse) concordia diagram with SIMS analyses of zircon from the gabbro on the western side of Kåfjord, intruded into the Kvenvik formation. The analyses in blue are selected for the average $^{207}\text{Pb}/^{206}\text{Pb}$ age calculation, while the analyses in yellow are probably affected by minor Pb loss. (B) Cathodoluminescence (CL) image of a selected zircon crystal with the site of the U–Pb analyses and corresponding $^{207}\text{Pb}/^{206}\text{Pb}$ age.

et al., 1997). Nevertheless, as a rule, oxygen isotopes are commonly much more easily affected by exchangeable oxygen derived from either meteoric water or interstitial fluids at elevated temperatures (e.g., Fairchild et al., 1990), whereas the carbon system is far more resilient during post-depositional, open-system recrystallisation (excluding organic diagenesis) because it is effectively buffered by the dissolving carbonate precursor. Consequently, in the following discussion, no attempt was made to reconstruct depositional $\delta^{18}\text{O}$ values; rather, the oxygen isotope values and $\delta^{13}\text{C}$ – $\delta^{18}\text{O}$ cross-plots have been utilised for tracking a 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 ($r = 0.96$, $>99\%$, $n = 5$) only for the Borras limestones (Fig. 15D), hence suggesting that only in this location could both isotope systems have been affected by alteration. Consequently, only the highest $\delta^{13}\text{C}$ value at c. 1‰ is accepted as the least altered and as a proxy for the depositional C-isotope value.

There is no significant correlation between $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ in all other studied carbonate rocks; however, the

Figure 15. Various cross-plots demonstrating C-isotope and geochemical differences between carbonate rocks of the Kvenvik and Stortviknes formations.

Figure 16. $(\text{MgO}/\text{CaO}_{\text{wr}}) - (\text{Mg}/\text{Ca}_{\text{carb}})$ versus $\delta^{13}\text{C}$ cross-plots demonstrating a lack of correlation between the parameters for the Stortviknes formation carbonates (A), and a strong negative correlation for the Kvenvik formation (B).

wide spread in oxygen isotope values recorded in each individual formation (Fig. 15D) reflects resetting of oxygen isotopes during post-depositional recrystallisation. This is particularly well pronounced in isotopically heavy carbonates from the Bergmark and Badden areas. The observed limited spread in $\delta^{13}\text{C}$ of 2–3‰ in each studied unit/section suggests that the alteration affect on the C-isotope system was rather limited.

In post-depositional processes, the Mn and Sr contents in carbonates would be partially shifted towards equilibrium with the ambient diagenetic/metamorphic fluids. Commonly, an increasing degree of post-depositional alteration by meteoric waters leads to Sr depletion and Mn enrichment (Veizer, 1983; Kaufman & Knoll, 1995). Hence, Brand & Veizer (1980) and Derry et

al. (1992) reported that Mn and Sr abundances can serve as another tool for calibration of the relative diagenetic rank of lithostratigraphic successions. The common practice adopted in the exercise of extracting the least-altered carbon isotope values is to selected samples with the lowest Mn/Sr ratios (e.g., Kaufman & Knoll, 1995). However, in some cases precaution needs to be taken when assessing the effect of meteoric diagenesis in Palaeoproterozoic carbonates, which may have formed in an anoxic water enriched in Fe^{2+} and Mn^{2+} (e.g., Veizer et al., 1992; Bekker et al., 2001).

The absence of negative correlation between $\delta^{13}\text{C}$ and Mn/Sr ratios in the studied carbonates (Fig. 15E) does not indicate any apparent alteration, with just a few exceptions. The latter includes the two outlying samples from Bergmark and Raipas areas marked by high Mn/Sr ratios (72 and 122, respectively; Fig. 15E) and showing a marked ^{13}C depletion ($\delta^{13}\text{C} = +1.7\text{‰}$ and -0.3‰ , respectively). Thus, both samples are considered to be significantly altered.

In metamorphosed carbonate rocks another possible cause of alteration of C- and O-isotope systems can be associated with the greenschist-facies metamorphism, marked by the reaction:

The dolomite + calcite $\pm$ talc $\pm$ actinolite paragenesis observed in some carbonate beds indicates that the aforementioned volatilisation reaction could have been

Figure 17. $\delta^{13}\text{C}$ histogram showing distinctive differences in $\delta^{13}\text{C}$ between carbonate rocks of the Kvenvik and Storviknes formations.

Figure 18. $\delta^{13}\text{C}$ variability in Palaeoproterozoic carbonate formations based on U–Pb and Re–Os data. Modified after Karhu & Holland (1996) and Martin & Condon (2013). The horizontal pink and blue bars represent the range of $\delta^{13}\text{C}$ for the Kvenvik and Storviknes carbonates, respectively. Red vertical line denotes the radiometric age for the gabbro emplacement, and consequently the upper age limit for the deposition of ^{13}C -rich carbonates of the Kvenvik formation. The blue arrow indicates the start of deposition of ^{13}C -rich carbonates constrained in the Fennoscandian Shield, hence setting the lower age limit for accumulation of the ^{13}C -rich carbonates of the Kvenvik formation. The black arrow indicates the termination of deposition during the Lomagundi–Jatuli isotopic event in the Fennoscandian Shield and worldwide, thus providing the lower age limit for the deposition of the Storviknes formation carbonates. The horizontal orange bar projects the apparent depositional time-range of ^{13}C -rich carbonate rocks of the Kvenvik formation.

the pathway 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). A case study research of carbonates metamorphosed under high-temperature greenschist-facies conditions demonstrated that $\delta^{13}\text{C}$ may be reset in the order of -1‰ in the precursor dolomite, up to -3‰ in the metamorphic calcite, whereas $\delta^{18}\text{O}$ can be depleted up to 6‰ in both newly formed carbonate components and their precursor (e.g., Melezhik et al., 2003). The significance of such a reaction and consequently its effect on the C-isotope system is commonly identified by the presence of: (i) talc or tremolite; (ii) discrepancy between whole-rock $\text{MgO}/\text{CaO}_{\text{wr}}$ and $\text{Mg}/\text{Ca}_{\text{carb}}$ bound to carbonate phases; (iii) a negative correlation between $\delta^{13}\text{C}$ and SiO_2 ; and (iv) a positive correlation between $\delta^{13}\text{C}$ and Mg/Ca ratio.

The amount of tremolite or talc is insignificant in the dolostones of the Storviknes formation. The correlations stated in (iii) and (iv) are not seen in Fig. 15A, B, and the discrepancy described in (ii) is insignificant (Fig. 16A); hence, all collectively suggest that the effect of alteration specified in (1) is also insignificant.

A moderate amount of tremolite is a characteristic feature of the dolostones from the Kvenvik formation occurring in the Badderen area. A strong negative correlation between $\delta^{13}\text{C}$ and $(\text{MgO}/\text{CaO}_{\text{wr}}) - (\text{Mg}/\text{Ca}_{\text{carb}})$ observed in the dolostones of the Kvenvik formation (Fig. 16B; $r = -0.45$, $>99.9\%$, $n = 51$ but one extreme value of $+1.7\text{‰}$ is excluded) is in accord with the petrographic observations. Consequently, depositional $\delta^{13}\text{C}$ values of some dolostones with elevated differences in $(\text{MgO}/\text{CaO}_{\text{wr}}) - (\text{Mg}/\text{Ca}_{\text{carb}})$ might have been slightly lowered (by $\leq 1\text{‰}$; e.g., Melezhik et al., 2003).

In a $\delta^{13}\text{C}$ histogram, the isotopically normal and heavy values exhibit two distinct, separate, unimodal distributions (Fig. 17). The former is skewed with a thin negative tail suggesting that all samples with $\delta^{13}\text{C} < 0\text{‰}$ are very likely altered. The distribution of the isotopically heavy values shows only an insignificant skew; hence, the depositional isotopic ratios show an overall better preservation.

Discussion

Depositional age constraint of the studied carbonate formations

The radiometric date of 2146 ± 5 Ma obtained from the gabbro intruding dolostones of the Kvenvik formation provides a minimum age limit for the deposition of the ^{13}C -rich carbonates of the Lower and Upper dolostone units in Member A. Moreover, being co-magmatic with tholeiitic basalts of Member B of the same formation, this date also constrains the deposition of the ^{13}C -rich carbonates in the Uppermost dolostone unit which is sandwiched between such lavas (Fig. 3). In addition, being co-magmatic with tholeiitic basalts of Member B of the Kvenvik formation, the date of 2146 ± 5 Ma constrains a maximum age limit for the deposition of the isotopically normal carbonates of the Storviknes formation (Fig. 3).

A further constraint on the deposition of both groups of carbonate rocks can be provided by means of the carbon isotope chemostratigraphy. The Early Palaeoproterozoic is known for a well-documented global positive excursion of carbonate carbon isotopes in sedimentary carbonates, namely the Lomagundi–Jatuli isotopic event (reviewed in Melezhik et al., 2013a; Fig. 18).

Worldwide compilations of available radiometric dates allow us to constrain the isotopic event between 2300–

2220 and 2060 Ma (Martin & Condon, 2013; Martin et al., 2013). Recently, Martin et al. (2014) have suggested that a maximum permitted range could be placed between 2306 ± 9 and 2057 ± 1 Ma, and a minimum permitted range between 2221 ± 5 and 2106 ± 8 Ma. On the Fennoscandian Shield alone, 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. 18). Consequently, the latter constraint combined with the obtained isotopic composition of carbonate rocks and available radiometric date of 2146 ± 5 Ma permit us to place the depositional age of the ^{13}C -rich carbonates of the Kvenvik formation between c. 2220 and 2146 ± 5 Ma (Fig. 18).

Based on similar reasoning, the deposition of isotopically normal carbonate rocks of the Storviknes formation should either pre-date or post-date the Lomagundi-Jatuli isotopic event (Fig. 18). However, red beds (Fig. 9) and stromatolitic carbonate rocks (Fig. 11) have not been documented in the Fennoscandian Shield prior to 2220 Ma. In contrast, these lithologies are known in sedimentary successions younger than 2060 Ma (Melezhik & Hanski, 2012a; McLoughlin et al., 2013), and the 2146 ± 5 Ma age on the underlying Kvenvik formation gabbro and co-magmatic lava flows also suggests that the Storviknes carbonates should post-date the LJE. Consequently, the deposition of the Storviknes carbonate formations is very likely younger than 2060 Ma, and currently cannot be constrained more precisely.

Litho- and chronostratigraphic correlation with the Palaeoproterozoic sedimentary and volcanic successions of the main part of the Fennoscandian Shield

Several attempts have been made to provide detailed correlations between the sedimentary-volcanic successions of the AKTW and those occurring in the main part of the Fennoscandian Shield in northern Norway (e.g., Pharaoh et al., 1983; Bergh & Torske, 1986, 1988). Due to the frequent lithofacies variations and the lack of reliable age determinations, all correlation schemes have remained only provisional (e.g., Bergh & Torske, 1988). However, aeromagnetic maps revealed that the supracrustal sequences of the AKTW are connected underneath the Caledonian nappes with those in the Kautokeino Greenstone Belt occurring to the south (Åm, 1975; Olesen & Solli, 1985; Olesen et al. 1990; Fig. 19).

A recent project (ICDP Fennoscandian Arctic Russia – Drilling Early Earth Project, Melezhik (2012a)) provided comprehensive information on litho- and chronostratigraphy, depositional ages and palaeotectonic evolution of the Fennoscandian Shield in the Early Palaeoproterozoic (Hanski & Melezhik, 2012; Melezhik & Hanski, 2012a). This represents a valuable database which

Figure 19. Aeromagnetic anomaly map over the AKTW and the northern part of the Kautokeino Greenstone Belt (Nasuti et al., 2015). Note that the sedimentary-volcanic formations occurring in the western and central parts of the AKTW are traceable beneath the Caledonian Kalak Nappe Complex into the Kautokeino Greenstone Belt.

allows the correlation of Palaeoproterozoic sedimentary-volcanic successions across the Fennoscandian Shield.

First of all, sedimentary successions bearing ^{13}C -rich carbonate formations whose depositional ages have been constrained in northern Fennoscandia to 2220–2060 Ma provide an invaluable tool for chemo- and chronostratigraphic correlation. Such correlation suggests that the ^{13}C -rich carbonate units of the Kvenvik formation have chemo- and chronostratigraphic correlatives in all major greenstone belts and depositional sites across the Finnish, Swedish and Russian parts of the Fennoscandian Shield. Fig. 20 exemplifies such a correlation.

The ^{13}C -rich carbonate units of the Fennoscandian Shield show diverse lithofacies including lacustrine red-coloured dolarenites, shallow-marine stromatolitic dolostones, beds of dissolution breccias, and deep-water dolarenites, many of which have been documented in the

Figure 20. Correlation of the AKTW sedimentary-volcanic succession with those occurring in other belts and palaeobasins in the main part of the Fennoscandian Shield. The correlation is based on chemostratigraphic data (¹³C-rich sedimentary carbonates denoted by pale-pink field) and published radiometric dates. The radiometric dates are from: 1 – Amelin et al. (1995), 2 – Vrevsky et al. (2010), 3 – Gärtner et al. (2014), 4 – Brasier et al. (2013), 5 – Martin et al. (2013), 6 – Melezhik et al. (2007), 7 – Hanski et al. (2014), 8 – Hannah et al. (2006), 9 – Skuf'in & Bayanova (2006), 10 – this study, 11 – Ovchinnikova et al. (2007), 12 – Hannah et al. (2008), 13 – Puchtel et al. (1992, 1998).

Kvenvik formation (Figs. 4–8). Importantly, the ^{13}C -rich carbonate units in many studied sites in Fennoscandia and worldwide share one common feature: they contain barite, pseudomorphed calcium sulphates and, in places, halite casts (Fig. 21A–F; Melezhik et al., 2013a; Strauss et al., 2013). Such features are also documented in the Bergmark area (Fig. 4D, E). All collectively allow a strong assumption of widespread former sulphates and halite coeval with the Lomagundi–Jatuli isotopic event. Recently, this assumption has been justified by the discovery of an over 190 m-thick bed of halite and an over 200 m-thick formation of massive anhydrite preserved at a depth of 3000 m below the surface and recovered in cores of a parametric well drilled in the Onega basin (Fig. 1; Morozov et al., 2010; Krupenik et al., 2011a). These beds of halite and Ca-sulphates, preserved at the base of a ^{13}C -rich dolostone formation (Figs. 20 & 21G, H), have been taken as robust evidence for the establishment of a sizeable seawater sulphate reservoir formed in response to a significant oxidation of terrestrial atmosphere and hydrosphere (e.g., Melezhik et al., 2005; Reuschel et al., 2012; Kump et al., 2013). For the first time in Earth's history, Ca-sulphates were widespread and forming in sedimentary basins worldwide (e.g., Strauss et al., 2013).

Apparent nature and fluid source for albitisation and scapolitisation

The bulk of the sedimentary and volcanic rocks spatially associated with Cu-sulphide occurrences and deposits in the Bergmark area (Sections 1 and 3) show a significant degree of albitisation (Table 4, Fig. 3). In places, the albitisation resulted in the formation of Na_2O -rich rocks, locally termed 'albite felsites'. Some intervals have been intensively scapolitised (e.g., Vik, 1985). This alteration phenomenon, however, does not represent a small-scale regional exception. Widespread chlorine and sodium metasomatism in Palaeoproterozoic volcano-sedimentary and igneous rocks has been documented throughout northern Fennoscandia (Fig. 1). In northern Sweden and Finland and adjacent parts of Norway and Russia, the areal extent of scapolitised and albitised Palaeoproterozoic rocks covers over several hundred square kilometres, and represents the largest known scapolite-bearing Precambrian terrane in the world (e.g., Frietsch et al., 1997). In places, the formation of scapolite and albite is accompanied by calcitisation and tourmalinisation. Interestingly, a number of epigenetic Cu-(Au) sulphide and Fe oxide deposits in northern Fennoscandia show a spatial and genetic relationship to Cl–Na metasomatism (e.g., Frietsch et al., 1997; Eilu et al., 2007).

Frietsch et al. (1997) reported that scapolite is mainly represented by a Cl and CO_3 species with small amounts of SO_4 and F, thus indicating high Na and Cl activity at the time of crystallisation. Fluid inclusion data obtained from several Cu–Au deposits indicate ore deposition

from highly saline aqueous solutions under temperatures of $\approx 300^\circ\text{C}$ (e.g., Ettner et al., 1994). Frietsch et al. (1997) speculated that the source of Cl, Na, SO_4 and F might derive from 2500–2000 Ma hypothetical evaporitic sequences or high-salinity brines formed in rift basins with active mafic volcanism. However, they acknowledged that the ultimate source of saline fluids remains enigmatic. They also admitted that evaporites and brines might have been mobilised through multiphase igneous activities and regional metamorphism and transported to their present positions during multiple metasomatic phases.

The recent discovery of over 200 m-thick beds of Palaeoproterozoic halite and massive anhydrite in the Onega basin (Figs. 20 & 21G, H) may shed light upon the most probable source of Cl, Na and SO_4 in metasomatically altered rocks in northern Fennoscandia. These thick deposits of halite and anhydrite in the Onega basin represent a uniquely preserved salt accumulation which apparently expresses only a small fraction of what had originally been accumulated on a large scale. Northern Fennoscandia holds the best preserved rock record characterising this period (e.g., Melezhik, 2012a).

It appears that these Early Palaeoproterozoic salts, having once been accumulated in great abundance, had been largely dissolved and remobilised during orogenic, magmatic and metamorphic alteration processes throughout a 2100 Myr post-depositional history, and as a rule largely lost from the rock record. However, they left behind abundant pseudomorphs, casts and dissolution-collapse breccias as well as diagenetic, metasomatic and hydrothermal albitisation, calcitisation and scapolitisation. All such features have been abundantly recorded across northern Fennoscandia in the Bergmark (Vik, 1985; this study), Kautokeino (Olerud, 1988; Ettner et al., 1994), Kiruna (Frietsch, et al., 1997), Central Lapland (Tuisku, 1985; Eilu, 1994; Mänttari, 1995), Kuusamo (Vanhanen, 2001), Peräpohja (Kyläkoski et al., 2012) and Onega (Melezhik et al., 2012, 2013b) areas. The isotopic compositions of the components, e.g., $^{34}\text{S}/^{32}\text{S}$ known from the Onega Parametric Hole ($\delta^{34}\text{S} = +4$ to $+8\text{‰}$, e.g., Krupenik et al., 2011b; Strauss et al., 2013), could provide a test of this hypothesis.

Depositional history of the Alta-Kvænangen succession in the Fennoscandian palaeogeographic framework

Based on the carbon isotope stratigraphy and reported radiometric date, the depositional history of the Alta-Kvænangen succession can be placed somewhere between 2220–2146 and <2060 Ma (Fig. 18). This succession, however, does not represent a continuous rock record. The date of 2146 ± 5 Ma constrains the emplacement of the gabbro. Importantly, this date also constrains the timing of coeval extrusive volcanism terminating the deposition of the Kvenvik formation. The deposition of the

Figure 21. Selected images of massive anhydrite and halite, and pseudomorphed sulphates documented in Early Palaeoproterozoic sedimentary rocks of the Fennoscandian Shield. (A) Dolomarl-hosted, quartz-pseudomorphed gypsum rosettes. (B) Back-scattered electron image of anhydrite relics preserved in quartz-pseudomorphed sulphate nodule. (C) Mudstone-hosted, quartz-pseudomorphed sulphate nodules with hollow core. (D) Former sulphate nodule (white) hosted by interlaminated red mudstone and white sulphates exhibiting enterolithic folds. (E) Sandstone-hosted former sulphate nodules replaced by quartz. (F) Former halite crystals replaced by dolomite. (G) Sawn core of massive anhydrite bed from the Onega parametric hole in the Onega palaeobasin (for the hole location, see Fig. 1). (H) Sawn core of massive halite with numerous inclusions of anhydrite and magnesite (white) and shale (grey) from the Onega parametric hole in the Onega palaeobasin. (A–D, F–H) Tulomozero Formation in the Onega palaeobasin; (E) Kuetsjärvi Sedimentary Formation in the Pechenga Greenstone Belt. (A, D) modified from Melezhik et al. (2005), (C) modified from Melezhik et al. (2013c), (E) modified from Melezhik & Hanski (2012b), (F) modified from Melezhik et al. (2013d), (G, H) courtesy of D. Rychanchik.

Figure 22. Major stages of palaeotectonic and palaeogeographic evolution of the Fennoscandian Shield in the Mid Palaeoproterozoic (2220–1950 Ma). Due to the lack of robust palaeomagnetic data, the time-slice reconstructions were made based on the present-day locations. Modified after Melezhik & Hanski (2012a).

structurally overlying carbonate units of the Storviknes formation is younger than 2146 ± 5 Ma and most likely post-dated 2060 Ma (Fig. 18). Consequently, a hiatus of over 80 Myr or more separates the depositional history of the two adjacent formations (Fig. 3).

The ^{13}C -rich dolostones forming the base of the Kvenvik formation (Lower and Upper dolostone in Member A; Fig. 3), and whose deposition is older than 2140 ± 5 Ma, occur in a large array of lithofacies representing depositional environments including an oxic coastal sabkha, a carbonate shelf, and the slope of a carbonate platform. The lithology, the depositional settings, and the ^{13}C -rich nature of the carbonate rocks fit well with the major environmental components of c. 2220–2060 Ma period inferred for the Fennoscandian Shield (Fig. 22A). This period was marked by (i) the development of an epeiric sea with several carbonate platforms; (ii) deposition of diverse carbonate lithologies and accumulation of abundant Ca-sulphates (oxic seawater); (iii) widespread ‘red beds’ (O_2 -rich atmosphere); and (iv) formation of ^{13}C -rich sedimentary carbonates representing the Lomagundi–Jatuli perturbation of the global carbon cycle (Melezhik & Hanski, 2012a). In the Fennoscandian palaeogeographic context, the ^{13}C -rich carbonates of the Kvenvik formation may represent a part of the Vesterålen carbonate platform (Fig. 22A).

The ^{13}C -rich dolostones, forming the middle part of the Kvenvik formation (the Uppermost dolostone in Member B; Fig. 3), whose deposition was coeval with the 2140 ± 5 Ma intensive mafic volcanism, accumulated in a deep-water shelf setting. This is again consistent with the general palaeotectonic evolution of the Fennoscandian Shield whose geotectonic regime was marked by the break-up of the continental crust at c. 2100 Ma. This resulted in the opening of the Kola Ocean between the Kola and Karelian cratons (Fig. 22A).

The 2140–2060 Ma period is missing in the Alta–Kvænangen succession. Sedimentological features of the Storviknes sedimentary rocks suggest that the shale accumulated in an oxic environment, whereas the carbonate deposition was in an intertidal setting within a channelled carbonate platform and its slope. Such an environment closely resembles the depositional setting reconstructed for time-equivalent successions documented elsewhere in the Fennoscandian Shield. The Il'mozero Sedimentary Formation from the Imandra/Varzuga Greenstone Belt represents such an example (Melezhik, 2012b; Fig. 22B).

Conclusions

The aeromagnetic map demonstrates that the sedimentary-volcanic succession exposed in the Alta–Kvænangen Tectonic Window (AKTW) in the northern Norwegian Caledonides is connected beneath the Kalak Nappe Complex with the main part of the Fennoscandian Shield, forming a northern extension of the Kautokeino Greenstone Belt.

Carbon isotope chemostratigraphy combined with the zircon U–Pb date for a gabbro constrains the deposition of the Kvenvik formation containing ^{13}C -rich carbonates to between <2220 and 2146 ± 5 Ma, whereas the accumulation of the Storviknes formation sedimentary succession with isotopically ‘normal’ carbonates is younger than 2146 ± 5 Ma and most likely occurred after 2060 Ma. The deposition of these two formations is separated by an over 80 Myr non-depositional break/h hiatus.

The deposition of the Kvenvik formation, which contains ^{13}C -rich carbonates, records a global positive excursion of carbonate carbon isotopes in sedimentary carbonates known as the Lomagundi–Jatuli isotopic event.

The deposition of ^{13}C -rich carbonates in the AKTW is marked by accumulation of Ca-sulphates and halite which accords with an abundant formation of CaSO_4 and NaCl in time-equivalent successions across the Fennoscandian Shield. These sedimentary sulphates and halite are inferred to represent the most plausible source for the Na and Cl metasomatism that affected Palaeoproterozoic sedimentary and volcanic rocks in the AKTW and across northern Fennoscandia.

Geochemical and petrological characteristics of volcanic rocks of the older part of the AKTW succession suggest an intraplate rift environment. Sedimentological features of carbonate rocks in all formations are consistent with deposition within a carbonate platform/shelf, whereas siliciclastic sedimentary rocks were apparently accumulated in a shallow-water epeiric sea.

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Table 4. Chemical composition of carbonate rocks of the Kvenvik and Storviknes formations.

Sample ##	SiO ₂	Al ₂ O ₃	Fe ₂ O ₃	TiO ₂	MgO	CaO	Na ₂ O	K ₂ O	BaO	Fe	Mg	Ca	Mn	Sr	MgO/CaO	Mg/Ca	Mn/Sr	δ ¹³ C	δ ¹⁸ O
wt. %																			
µg g ⁻¹																			
‰																			
Storviknes formation																			
Borras, Section 8																			
At - 1	2.83	0.68	2.82	0.08	0.59	51.6	<0.1	0.05	<0.025	5700	2000	340000	1020	189	0.01	0.006	5.4	-3.1	15.1
At - 2	4.75	1.17	3.98	0.10	0.99	48.9	<0.1	0.08	<0.025	7560	3040	337000	1310	267	0.02	0.009	4.9	-1.9	14.8
At - 3	10.9	1.52	1.72	0.19	1.10	46.7	<0.1	0.42	<0.025	1400	2210	325000	590	64.2	0.02	0.007	9.2	1.2	19.3
At - 4	9.35	0.43	0.41	0.03	0.97	49.0	<0.1	0.07	0.027	1820	4550	342000	976	75.2	0.02	0.01	13	1.1	18.8
At - 6	7.87	3.21	7.84	0.39	1.49	42.9	<0.1	0.02	<0.025	12100	3040	296000	1480	237	0.03	0.01	6.2	-2.3	14.7
Storviknes formation																			
Raipas, Section 7																			
RP_14	11.3	0.98	2.52	0.05	17.1	26.5	0.43	0.07	<0.025	11300	113000	194000	5970	49.0	0.65	0.58	122	-0.3	20.9
RP_1	2.08	0.88	0.57	0.05	20.7	30.2	<0.1	0.36	<0.025	3360	134000	217000	820	52.2	0.69	0.62	16	1.9	22.3
RP_2	18.7	4.49	2.80	0.26	15.7	21.9	<0.1	2.00	<0.025	11500	95900	161000	858	45.2	0.72	0.60	19	2.1	22.0
RP_3	20.1	4.87	3.19	0.29	15.1	21.0	<0.1	2.09	<0.025	12700	92100	155000	821	42.3	0.72	0.59	19	2.0	22.2
RP_4	21.0	4.15	2.83	0.23	15.1	21.3	0.19	1.69	<0.025	9800	93700	155000	1390	47.2	0.71	0.60	29	1.8	22.0
RP_5	12.1	1.73	1.08	0.09	18.0	26.1	<0.1	0.72	<0.025	4810	117000	186000	1060	43.6	0.69	0.63	24	1.7	22.3
RP_6	5.65	0.86	0.54	0.05	20.1	29.0	<0.1	0.37	0.042	2380	130000	206000	454	47.6	0.69	0.63	9.5	2.6	22.2
RP_7	2.12	0.76	0.74	0.04	20.6	29.9	<0.1	0.28	0.033	4000	135000	215000	822	64.8	0.69	0.63	13	2.3	21.7
RP_8	10.0	1.43	1.13	0.08	18.7	27.0	<0.1	0.57	<0.025	4160	121000	193000	704	46.3	0.69	0.63	15	2.6	22.4
RP_9	7.01	0.98	0.65	0.06	19.4	28.2	<0.1	0.40	0.081	2800	127000	202000	589	55.9	0.69	0.63	11	2.5	22.1
RP_10	1.46	0.66	0.54	0.03	20.9	30.5	<0.1	0.25	<0.025	3020	135000	214000	441	66.1	0.69	0.63	6.7	0.8	22.0
RP_11	4.05	0.73	0.70	0.05	20.2	29.3	<0.1	0.26	<0.025	3640	134000	213000	795	57.0	0.69	0.63	14	0.9	21.9
RP_12	3.24	0.72	0.86	0.04	20.3	29.6	<0.1	0.25	<0.025	5430	131000	210000	1210	68.5	0.69	0.62	18	0.5	21.7
RP_13	1.90	0.83	0.71	0.05	20.7	29.9	<0.1	0.31	<0.025	2640	136000	213000	578	60.9	0.69	0.64	9.5	0.8	22.4
Storviknes formation																			
Viknes, Section 6																			
RP_25	2.13	0.23	0.73	0.01	20.7	30.2	<0.1	0.07	<0.025	4280	137000	219000	645	110	0.69	0.63	5.9	0.6	19.2
RP_26	3.29	0.58	0.51	0.03	20.4	29.5	<0.1	0.22	<0.025	3010	137000	218000	483	64.7	0.69	0.63	7.5	1.1	21.3
RP_27	6.93	0.46	0.46	0.03	19.7	28.8	<0.1	0.10	<0.025	3080	130000	209000	399	52.5	0.68	0.62	7.6	1.5	21.6

Table 4 (continued).

Sample ##	SiO ₂	Al ₂ O ₃	Fe ₂ O ₃	TiO ₂	MgO	CaO	Na ₂ O	K ₂ O	BaO	Fe	Mg	Ca	Mn	Sr	MgO/CaO	Mg/Ca	Mn/Sr	δ ¹³ C	δ ¹⁸ O
wt. %																			
μg g ⁻¹																			
‰																			
Storviknes formation																			
Viknes, Section 6																			
RP_28	13.2	2.46	2.03	0.14	17.5	24.7	<0.1	1.05	<0.025	3910	117000	185000	551	38.8	0.71	0.63	14	2.7	22.1
RP_29	9.31	0.12	0.44	<0.01	19.4	28.0	<0.1	0.03	<0.025	3010	132000	210000	427	45.8	0.69	0.63	9.3	1.9	21.8
RP_30	3.34	0.09	0.26	<0.01	20.7	29.7	<0.1	0.03	<0.025	3940	122000	194000	298	52.6	0.70	0.63	5.7	1.2	22.1
RP_31	12.3	0.79	0.68	0.09	18.6	26.5	<0.1	0.32	0.089	3990	125000	198000	301	53.1	0.70	0.63	5.7	1.8	22.8
RP_32	6.43	1.10	0.88	0.09	19.4	28.0	<0.1	0.47	<0.025	3000	132000	211000	442	44.2	0.69	0.63	10	0.6	22.0
RP_33	33.7	0.89	0.68	0.07	13.7	19.5	<0.1	0.40	0.354	2530	92200	148000	364	48.5	0.70	0.62	7.5	0.7	21.6
RP_34	2.24	0.64	0.59	0.05	20.6	29.8	<0.1	0.27	<0.025	2690	139000	222000	384	48.0	0.69	0.63	8.0	0.6	22.0
RP_35a	2.83	0.12	0.41	0.01	20.8	29.9	<0.1	0.05	<0.025	1430	141000	221000	270	32.6	0.70	0.64	8.3	1.7	22.7
RP_35b	4.12	0.13	0.30	0.01	20.4	29.4	<0.1	0.05	0.232	1510	137000	216000	303	43.9	0.69	0.63	6.9	1.6	22.4
RP_35c	3.09	0.33	0.50	0.03	20.7	29.8	<0.1	0.13	0.123	2310	139000	219000	314	39.5	0.69	0.63	7.9	1.7	22.4
RP_35d	2.56	0.20	0.34	0.02	20.7	29.8	<0.1	0.08	0.032	1890	139000	218000	346	33.1	0.69	0.64	10	1.8	22.2
At - 10	7.60	1.31	1.15	0.11	19.1	27.4	<0.1	0.57	<0.025	2500	114000	190000	270	42.2	0.70	0.60	6.4	0.3	22.8
At - 11	1.91	0.25	0.36	0.03	20.8	29.8	<0.1	0.05	0.421	1880	123000	198000	212	40.5	0.70	0.62	5.2	0.6	23.9
At - 12	4.49	0.42	0.66	0.05	20.2	29.1	<0.1	0.13	0.051	1040	119000	197000	371	39.6	0.69	0.60	9.4	0.6	24.9
At - 13	3.08	0.24	0.52	0.04	20.5	29.7	<0.1	0.05	<0.025	2800	123000	205000	235	29.4	0.69	0.60	8.0	1.4	23.6
At - 14	2.34	0.24	0.42	0.03	20.7	29.9	<0.1	0.07	0.148	2230	122000	201000	272	37.1	0.69	0.61	7.3	1.5	26.4
At - 15	8.53	1.63	1.81	0.18	17.2	25.0	<0.1	0.67	<0.025	2110	109000	179000	323	32.3	0.69	0.61	10	1.2	22.1
At - 16	3.93	1.52	1.39	0.23	20.1	26.6	<0.1	0.41	<0.025	2170	112000	186000	360	32.7	0.76	0.60	11	1.4	22.7
At - 17	5.69	1.78	1.87	0.29	17.9	25.7	<0.1	0.72	<0.025	2050	112000	187000	332	32.8	0.70	0.60	10	1.3	22.4
Kvenvik formation. Member B																			
Badderer, Uppermost dolostone, Section 5																			
JS1301	49.7	12.9	4.04	0.47	9.33	9.07	<0.1	2.22	0.066	2020	3410	69200	238	48.2	1.03	0.05	4.9	6.5	13.6
JS1302	44.6	9.25	3.04	0.41	7.46	16.4	<0.1	1.36	0.038	2170	3770	119000	359	84.7	0.45	0.03	4.2	6.6	14.1
JS1303	42.1	9.53	3.28	0.36	9.37	15.1	<0.1	1.39	0.038	4430	17400	108000	377	65.5	0.62	0.16	5.8	7.2	15.1
JS1304	44.7	10.4	3.40	0.40	8.55	12.8	<0.1	2.04	0.061	5490	20300	93500	419	48.3	0.67	0.22	8.7	7.4	15.5
JS1305	39.2	8.70	3.43	0.32	9.59	15.9	<0.1	1.71	0.054	7880	32000	114000	525	50.8	0.60	0.28	10	7.4	15.6
JS1306	46.0	10.9	3.00	0.43	8.50	11.1	<0.1	2.51	0.081	6610	28300	79700	355	37.4	0.77	0.36	9.5	7.4	15.8

Table 4 (continued).

Sample ##	SiO ₂	Al ₂ O ₃	Fe ₂ O ₃	TiO ₂	MgO	CaO	Na ₂ O	K ₂ O	BaO	Fe	Mg	Ca	Mn	Sr	MgO/CaO	Mg/Ca	Mn/Sr	δ ¹³ C	δ ¹⁸ O
wt. %																			
μg g ⁻¹																			
‰																			
Kvenvik formation. Member B																			
Badderen, Uppermost dolostone, Section 5																			
JS1307	41.3	8.38	3.25	0.33	9.83	13.5	<0.1	2.11	0.074	10100	43000	97200	493	38.6	0.73	0.44	13	7.6	16.3
JS1308	44.1	10.8	2.92	0.40	8.38	11.3	0.11	3.07	0.117	9380	38500	82900	393	35.8	0.74	0.46	11	7.8	16.4
JS1309	42.4	9.52	3.04	0.38	8.83	12.8	<0.1	2.64	0.095	9930	40100	91400	424	38.5	0.69	0.44	11	7.1	15.7
JS1310	44.3	11.7	3.71	0.44	8.23	10.3	0.12	3.32	0.159	8610	33500	76200	376	30.7	0.80	0.44	12	7.8	16.7
Kvenvik formation. Member A																			
Kåfjord, Upper dolostone, Section 4																			
Profile B																			
RP_24	20.5	5.25	2.52	0.20	12.5	23.1	2.55	0.54	<0.025	14600	74100	169000	1810	37.2	0.54	0.44	49	7.4	13.9
RP_23	12.1	2.41	2.40	0.10	14.9	27.3	1.27	0.16	<0.025	15600	98900	198000	1850	34.9	0.55	0.50	53	7.4	14.5
RP_22	16.2	3.92	2.75	0.13	16.1	23.1	1.92	0.38	<0.025	16800	98600	167000	1850	33.1	0.70	0.59	56	7.7	14.5
RP_21	12.0	2.94	2.33	0.09	16.3	25.7	1.58	0.10	<0.025	15000	107000	189000	1660	35.6	0.63	0.57	47	7.5	14.6
RP_20	8.48	1.58	1.92	0.07	16.5	28.1	0.70	0.25	<0.025	12300	112000	211000	1440	44.4	0.59	0.53	32	7.3	15.1
Profiles C-D																			
RP_15	22.1	5.73	1.83	0.18	12.8	21.5	3.33	0.09	<0.025	11000	86000	160000	1300	33.3	0.60	0.54	39	8.1	14.5
RP_16	17.3	3.84	1.89	0.14	14.8	24.2	2.22	0.04	<0.025	12900	98300	179000	1350	34.7	0.61	0.55	39	8.2	14.7
RP_17	19.4	3.58	2.80	0.11	14.2	23.3	2.04	0.04	<0.025	13400	94600	171000	1690	36.5	0.61	0.55	46	8.2	14.4
RP_18	22.0	4.20	2.61	0.14	13.0	22.6	2.08	0.27	<0.025	11200	86400	168000	1400	37.0	0.58	0.51	38	8.3	14.3
RP_19	22.6	4.46	2.95	0.15	13.5	20.9	2.58	0.05	<0.025	12700	89700	158000	1700	33.9	0.65	0.57	50	8.2	14.2
Bergmark, Upper dolostone, Section 3																			
Bedded dolostone lithofacies																			
JS1337	16.0	2.25	1.50	0.09	13.8	27.9	<0.1	0.66	<0.025	9200	81200	195000	668	32.3	0.49	0.42	21	7.9	16.9
JS1338	11.5	2.46	1.18	0.10	12.7	31.8	<0.1	0.70	<0.025	7020	73700	219000	667	35.8	0.40	0.34	19	8.0	17.1
JS1339	28.0	8.46	1.70	0.32	13.3	17.4	<0.1	2.19	0.054	6930	66100	122000	498	17.5	0.76	0.54	28	8.2	17.6
JS1340	31.4	8.01	2.01	0.38	12.1	17.2	<0.1	1.90	0.050	6390	54700	120000	507	18.0	0.70	0.46	28	8.2	17.4
JS1341	12.8	2.35	2.95	0.08	13.0	30.1	<0.1	0.44	<0.025	12600	68600	203000	1050	33.4	0.43	0.34	31	7.5	16.1
JS1342	25.3	5.80	2.15	0.21	13.7	19.6	<0.1	1.61	0.048	10900	74100	135000	690	21.5	0.70	0.55	32	7.4	16.3

Table 4 (continued).

Sample #	SiO ₂	Al ₂ O ₃	Fe ₂ O ₃	TiO ₂	MgO	CdO	Na ₂ O	K ₂ O	BaO	Fe	Mg	Ca	Mn	Sr	MgO/CaO	Mg/Ca	Mn/Sr	δ ¹³ C	δ ¹⁸ O
																			‰
Bergmark, Upper dolostone, Section 3																			
<i>Impure, laminated, dolostone lithofacies</i>																			
JS1333	37.9	8.79	5.33	0.28	12.3	13.0	0.36	1.00	0.034	9940	37600	90600	643	25.1	0.95	0.42	26	7.7	16.1
JS1334	27.3	5.44	2.67	0.21	12.2	19.8	<0.1	1.60	0.055	12100	66800	139000	1010	30.9	0.62	0.48	33	7.7	15.7
JS1335	16.2	2.79	2.65	0.11	13.6	27.3	0.22	0.54	<0.025	13500	75800	190000	1180	43.9	0.50	0.40	27	7.6	15.7
JS1336	30.6	6.38	2.62	0.19	11.2	19.0	<0.1	1.87	0.037	11300	56700	130000	949	31.8	0.59	0.44	30	7.6	15.3
Bergmark, Upper dolostone, Section 2																			
<i>Bedded dolostone lithofacies</i>																			
JS1316	13.1	2.10	3.14	0.06	15.5	25.5	1.19	0.03	<0.025	19300	98300	184000	1030	37.8	0.61	0.53	27	7.8	15.0
JS1317	13.4	2.19	3.23	0.06	16.0	26.2	1.24	0.04	<0.025	12100	85500	195000	731	28.8	0.61	0.44	25	7.7	15.6
JS1318	13.8	2.39	1.64	0.08	14.5	27.7	1.20	0.11	<0.025	10500	89100	194000	537	27.0	0.52	0.46	20	8.1	16.7
JS1319	32.4	7.27	3.22	0.26	11.2	16.5	2.74	0.71	<0.025	16600	63000	121000	567	19.5	0.68	0.52	29	7.8	15.4
JS1320	27.1	5.95	1.82	0.21	12.8	19.3	0.90	1.52	<0.025	11100	77100	138000	504	19.7	0.66	0.56	26	7.0	15.8
JS1321	25.6	5.18	2.40	0.15	12.9	20.1	2.49	0.29	<0.025	13800	74800	138000	571	22.4	0.64	0.54	25	7.1	15.9
Impure, laminated, dolostone lithofacies																			
JS1311	57.7	15.4	2.48	0.50	5.99	3.03	8.30	1.14	<0.025	7350	14800	21800	158	7.8	1.98	0.68	20	7.4	13.5
JS1312	56.5	14.8	2.40	0.52	9.98	2.11	7.00	0.91	<0.025	4020	12400	15000	141	4.0	4.73	0.83	35	5.5	13.6
JS1313	35.0	4.49	4.48	0.15	9.84	17.0	2.54	0.10	<0.025	23300	61400	124000	947	23.0	0.58	0.50	41	6.1	13.4
JS1314	46.9	11.6	2.81	0.43	10.9	7.63	5.50	1.05	<0.025	9170	30300	54500	217	8.4	1.43	0.56	26	6.3	13.7
JS1315	39.2	9.43	2.73	0.34	9.39	13.4	4.04	0.59	<0.025	13100	49800	95600	477	16.6	0.70	0.52	29	7.7	14.9
JS1332	44.0	14.7	10.6	1.0	9.35	3.72	4.81	3.87	<0.025	15400	19500	26800	624	8.7	2.51	0.73	72	1.7	14.1
Bergmark, Lower dolostone, Section 1																			
<i>Pale-yellow, laminated, dolostone lithofacies</i>																			
JS1322	23.6	6.28	1.62	0.22	13.4	19.7	3.50	0.32	<0.025	10300	79000	137000	938	32.1	0.68	0.58	29	7.3	15.0
JS1323	29.0	8.46	2.64	0.35	11.7	16.7	4.42	0.75	<0.025	7520	66200	116000	1080	26.4	0.70	0.57	41	5.4	16.5
JS1326	26.3	7.56	1.68	0.24	12.6	18.3	4.11	0.45	<0.025	9350	73200	128000	567	35.3	0.69	0.57	16	7.2	17.3
JS1327	21.7	5.99	1.81	0.23	14.1	20.5	3.24	0.36	<0.025	11200	82800	145000	576	38.5	0.69	0.57	15	7.2	17.0

Table 4 (continued).

Sample ##	SiO ₂	Al ₂ O ₃	Fe ₂ O ₃	TiO ₂	MgO	wt. %	CaO	Na ₂ O	K ₂ O	BaO	Fe	Mg	Ca	Mn	Sr	MgO/CaO	Mg/Ca	Mn/Sr	δ ¹³ C	δ ¹⁸ O
Bergmark, Lower dolostone, Section 1																				
<i>Pale-yellow, laminated, dolostone lithofacies</i>																				
JS1328	23.8	3.91	2.44	0.17	2.43		33.3	0.29	0.29	3.35	551	2320	232000	335	264	0.07	0.01	1.3	7.6	17.0
JS1329	29.2	4.86	2.83	0.24	1.38		32.7	1.84	0.31	<0.025	539	2150	228000	465	164	0.04	0.009	2.8	8.5	16.9
<i>Red, massive, dolostone lithofacies</i>																				
JS1331	28.1	3.80	1.63	0.14	14.5		19.6	0.35	0.09	<0.025	5780	76800	135000	1230	48.0	0.74	0.57	26	7.2	23.3
JS1330	24.4	2.03	1.56	0.09	3.40		37.2	<0.1	0.14	<0.025	563	2780	259000	366	446	0.09	0.01	0.8	7.0	17.0
JS1325	7.77	0.47	1.79	0.02	19.3		27.5	<0.1	0.16	<0.025	7380	112000	191000	2210	72.1	0.70	0.59	31	6.2	21.8
JS1324	21.4	2.03	2.29	0.08	17.0		21.7	<0.1	0.62	<0.025	5120	82700	151000	2620	50.9	0.78	0.55	51	5.7	22.4

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