The Čaravarri Formation of the Kautokeino Greenstone Belt, Finnmark, North Norway; a Palaeoproterozoic foreland basin succession

TORE TORSKE & STEFFEN G. BERGH

Torske, Tore & Bergh, Steffen G. 2004: The Čaravarri Formation of the Kautokeino Greenstone Belt, Finnmark, North Norway; a Palaeoproterozoic foreland basin succession. *Norges geologiske undersøkelse Bulletin, 442, 5–22.*

The Čaravarri Formation is the youngest supracrustal unit in the Palaeoproterozoic Kautokeino Greenstone Belt of western Finnmark. This greenstone belt and correlative units at Alta farther north were formerly ascribed to a continental rift setting near the northern margin of the Fennoscandian Shield. The Čaravarri Formation is metamorphosed at very low grade and made up of c. 4000 m-thick, upward-coarsening, siliciclastic sandstones and conglomerates deposited in a presently NNW-SSE-trending basin. It rests with depositional contact on the underlying argillitic Bik'kačåk'ka Formation. The eastern boundary is an east-dipping thrust, with a hangingwall block of older, more highly deformed and metamorphosed metasupracrustals.

The Čaravarri Formation is subdivided into three lateral segments, from south to north, the Njargavarri, Gæsvarri and Čaravarri segments, separated by inferred faults. The facies associations reveal: (i) an extensive lower conglomeratic member (unit 1), present in all segments and interpreted as subaqueous debris-flow deposits, (ii) pebbly sandstones and conglomerates interpreted as shallow-marine turbidites, debris flows or fan-delta deposits with locally derived carbonate clasts, which dominate in the Njargavarri segment and lower part of the Gæsvarri and Čaravarri segments, and (iii) conglomerates and channelled and cross-bedded sandstones interpreted as prograding alluvial fan and fluvial trunk-river deposits that characterise the northern, Čaravarri segment. Uniform transport directions from an easterly source can be demonstrated for the fluvial and alluvial successions. Distinct differences in clast maturity and various proportions of quartzite, jasper/chert, and carbonate clasts, suggest a younger age of clast deposition when moving northwards.

The Čaravarri Formation is reinterpreted as a molasse-type succession, deposited in a foreland basin that reflects changing provenance and depocenter migration through time. In the south, the basin was underfilled, with subsidence-induced, relatively high-gradient topography generating fan deltas with repeated mass flows and turbidity currents. In the central and northern segment, the basin became overfilled, i.e., fluvial trunk-rivers were overridden by large, westward-prograding alluvial fans with clasts derived from a more distal source, probably the uplifted, Svecokarelian thrust belt to the east. The lateral segmentation of the Čaravarri basin may have been controlled by frontal transverse faults of the thrust belt.

The Čaravarri Formation is tentatively correlated with the Kumpu Group (<1.88 Ga) of Central Finnish Lapland, which overlies the Palaeoproterozoic Kittilä Greenstone Complex. The Kumpu Group is now interpreted as a foreland basin, molasse-type unit deposited after the Svecokarelian orogeny. Such an interpretation and tectonic setting support derivation of the Čaravarri Formation from the Svecokarelian collisional orogen of eastern Finnmark, during which time the Lapland granulite belt and related allochthonous units were thrust westward onto the Karelian craton.

Tore Torske & Steffen G. Bergh: Department of Geology, University of Tromsø, N-9037 Tromsø, Norway. Tel. 77644464, e-mail: Steffen.Bergh@ig.uit.no

Introduction and geological setting

The Čaravarri Formation is the youngest supracrustal unit in the Kautokeino Greenstone Belt of western Finnmark, North Norway (Fig. 1; Holmsen et al.1957, Olsen & Nilsen 1985, Siedlecka et al. 1985, Solli 1983,1990, Olesen & Sandstad 1993). Together with the Karasjok Greenstone Belt, farther east, the Kautokeino belt forms the bulk of the Palaeoproterozoic cover rocks; major units of tholeiitic metabasalts, tuffaceous greenstones and amphibolites alternating with metasedimentary rocks. This pile of supracrustals rests upon crystalline basement within the Karelian craton near the northwestern margin of the Fennoscandian Shield. These greenstone belts are separated by a structural basement dome of Archaean to Palaeoproterozoic tonalitic gneisses and granites: the Jer'gul Gneiss Complex and its par-autochthonous sedimentary cover, the Skuvvanvarri and Masi Formations (Fig. 1; Siedlecka et al. 1985).

The rocks of the Čaravarri Formation compose a 3-6 km wide, NNW-SSE-trending, linear mountain ridge in the northwestern part of the Finnmarksvidda plateau. It rises to 300 m above the generally more subdued and heavily covered areas on both sides. The ridge is well exposed for about 25 km along strike, from lake Čarajav'ri in the north to the hill Lulli Lik'ča in the south (Fig. 2). The formation has a sharp, conformable contact with underlying units to the west, but is thrust-bounded to the east (Fig. 2). The ridge is also segmented by transverse faults. Lithologically, the formation consists of well preserved, very low-grade metamorphic, siliciclastic sedimentary rocks, with a stratigraphic thickness of up to about 4000 m; its bedding dips consistently eastward (Fig. 2).

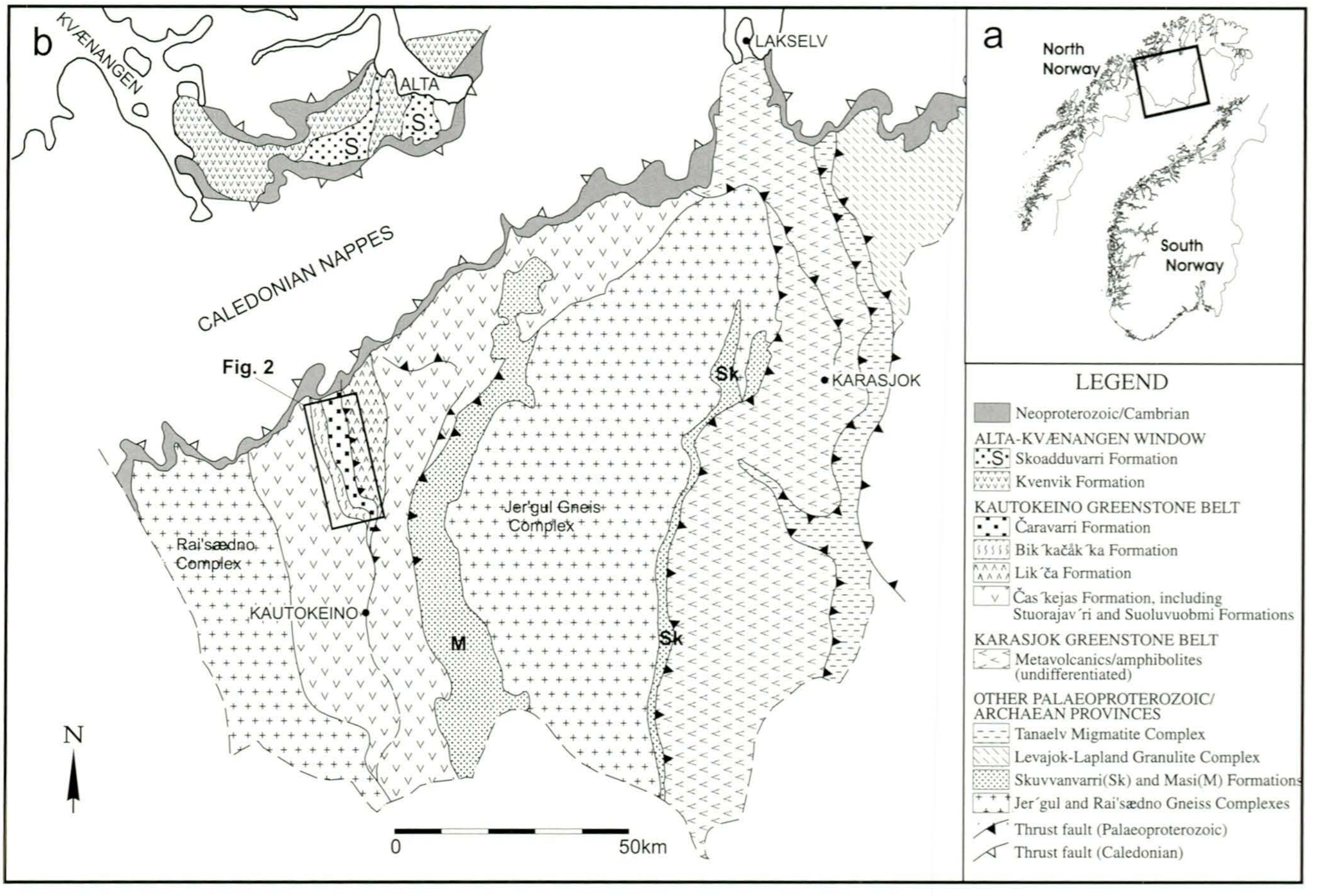

Fig. 1. Simplified geological map of western Finnmarksvidda and the Alta-Kvænangen tectonic window. Location of the Čaravarri Formation is shown, in the rectangle (from Siedlecka et al. 1985).

The units flanking the Čaravarri Formation all belong to the Kautokeino Greenstone Belt (Fig. 1); they include: (i) to the west, the conformably underlying, steeply east-dipping, 1000–1500 m thick, argillite-siltstone dominated, Bik'kačåk'ka Formation (Siedlecka et al. 1985). This formation is in turn underlain by mafic metavolcanic rocks of the Cas'kejas Formation, including the Stuorajav'ri and Suoluvuobmi Formations (Siedlecka et al. 1985); (ii) to the east, the Čaravarri Formation is bounded by a moderately to steeply east-dipping thrust system (Fig. 2) with hangingwall rocks of the Lik'ča Formation, comprising greenschist-facies metabasalts similar to those of the Čas'kejas Formation, and a few metasandstones. No thrusts are recorded west of the Čaravarri Formation.

Within a distance of about 500 m from the east-bounding fault zone, the Čaravarri sandstones are variably altered to light red and grey quartzite (Fig. 2). The fault zone itself is c. 1 km wide and made up of a heterogeneous, highly disrupted subzone of semi-ductilely deformed phyllitic and graphite-bearing schists interpreted as a thrust detachment. Locally, sheared and folded dolomite-marble lenses are observed which represent tectonised rocks of the hangingwall, the Lik'ča Formation (Siedlecka et al. 1985). Sheets of deformed mafic intrusive rocks and a lens-shaped lherzolite body are also present in this zone (Fig.2).

The Čaravarri Formation can be correlated with the Skoadduvarri Sandstone Formation at Alta (Holmsen et al. 1957, Barth & Reitan 1963, Zwaan & Gautier 1980, Torske & Bergh 1984a,b, Bergh & Torske 1986), the youngest unit of the Raipas Supergroup in the Alta-Kvænangen tectonic window within the Caledonian nappe area (Fig. 1).

Previous work and models

The rocks of the Čaravarri Formation had not been reported in the geological literature until Holmsen et al. (1957) introduced them under the designation 'Čaravarri Grit'. Oftedahl (1981) included the 'Grit' unit of Holmsen et al. (1957) and the Bik'kačåk'ka Formation in 'the Čaravarri Group', while Siedlecka et al. (1985) gave the grit-unit the formal name, 'Čaravarri Formation'.

The depositional character and tectonic significance of the Čaravarri Formation, relative to other supracrustal units of the Kautokeino and Karasjok Greenstone Belts, have not yet been addressed. However, Torske (1977) and Bergh & Torske (1986,1988) ascribed, tentatively, the Čaravarri and Čas'kejas Formations and the correlative Raipas Supergroup

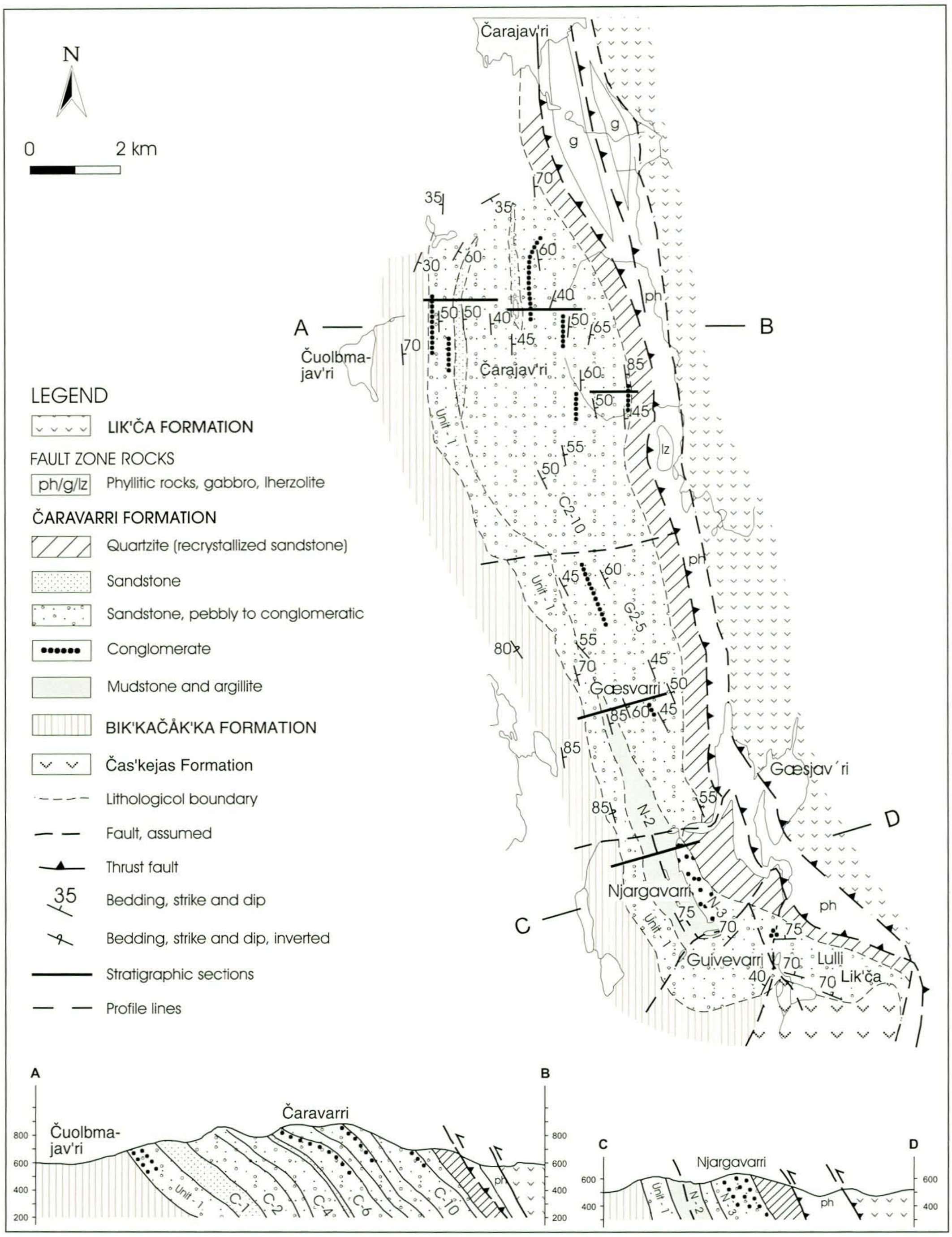

Fig. 2. Geological map, with cross-sections, of the Čaravarri Formation and adjoining units between lake Čarajav'ri and Lulli Lik'ča.

of Alta, including the Kvenvik and Skoadduvarri Formations, to a continental rift setting near the northern margin of the Fennoscandian Shield. In this scenario, the Skoadduvarri Formation at Alta was interpreted as fan-delta deposits from rivers running along the postulated rift basin, from a source of prograding, trough-filling, alluvial fans in the Čaravarri Formation to the south (Bergh & Torske 1986). A possible strike-slip rift origin for the Čaravarri Formation was suggested by Torske & Bergh (1984 a, b) on the basis of differences in clast types and tentative, but elusive, source areas along strike.

Recently, a rift basin setting for the Alta-Kautokeino greenstone belt has been questioned by us because no synsedimentary, rift-marginal, master normal faults have been demonstrated (see Fig. 2). Instead, the Čaravarri Formation has been proposed as a foreland basin deposit of molasse type (Torske 1997), preserved in front of remnants of a Palaeoproterozoic (Svecokarelian) fold and thrust belt (e.g., Krill 1985) to the east.

Purpose and methods of the present study

The purpose of this study is to establish a stratigraphic

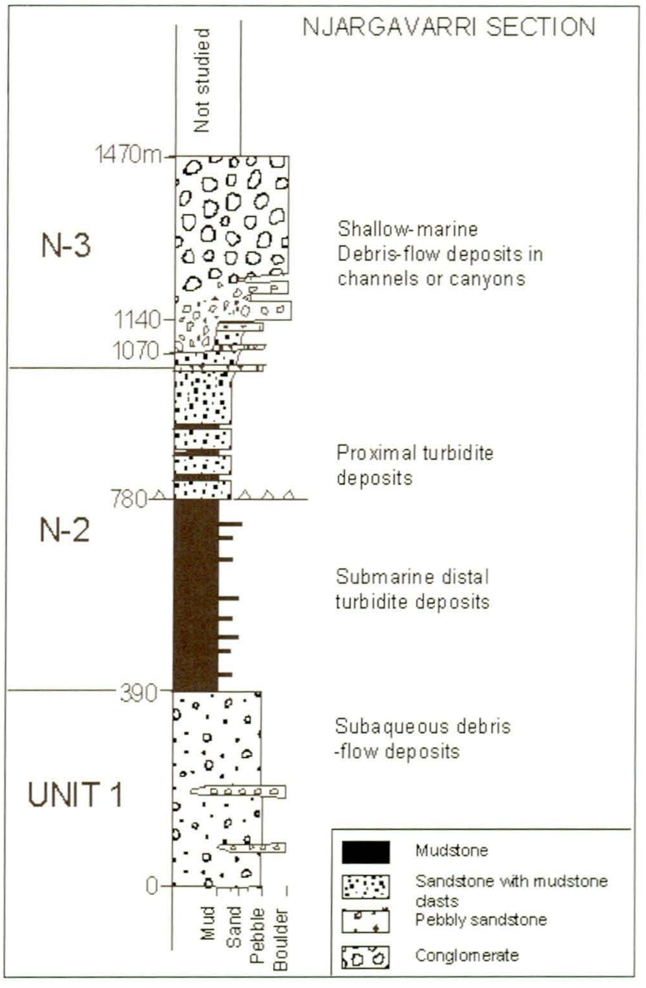

Fig. 3. Generalised stratigraphic section of the *Njargavarri segment* showing the main facies associations and interpretations. See Fig. 2 for location of the section.

framework for the Čaravarri Formation and to determine its depositional environment on the basis of facies associations, sediment transport directions, and clast provenance. This is done with a view to determine the tectonic relations between this thick, coarse clastic sedimentary unit and other Palaeoproterozoic units in the northern part of the Fennoscandian Shield.

No complete section through the Čaravarri Formation is found, due to an extensive Quaternary cover that obscures vertical and lateral continuity of the strata. However, a stratigraphy has been established on the basis of separate, spot, or continuous observations of more or less isolated outcrops. Lithofacies boundaries and transitions are commonly hidden, which hampers the detailed analysis of genetically related facies associations. The uncertain and ambiguous associative aspect of the rock successions affected our field methods, and the present study is therefore of a reconnaisance character.

Stratigraphic levels and intervals above the formational datum line, i.e., the western, lower boundary of the Čaravarri Formation, were measured and/or computed from positions plotted on topographic maps and air photos. Thicknesses of major sections and subsections were calculated as the across-strike horizontal distance of the location plots, multiplied by the sine of an average dip angle for inferred local bedding along the sections. As these calculations ignored topographic distortion, the method gives only approximate values.

Our study shows that facies characters, changing source areas, and depositional environments of the Čaravarri Formation are consistent with deposition in a foreland basin, most likely from erosion of an uplifted fold and thrust belt to the east. This belt includes thrust units of the Kautokeino and Karasjok Greenstone Belts, as well as erosional material from the adjacent gneiss, migmatite and granulite complexes (see Fig. 1; Krill 1985, Marker 1985, Braathen 1991, Andreassen 1993, Braathen & Davidsen 2000).

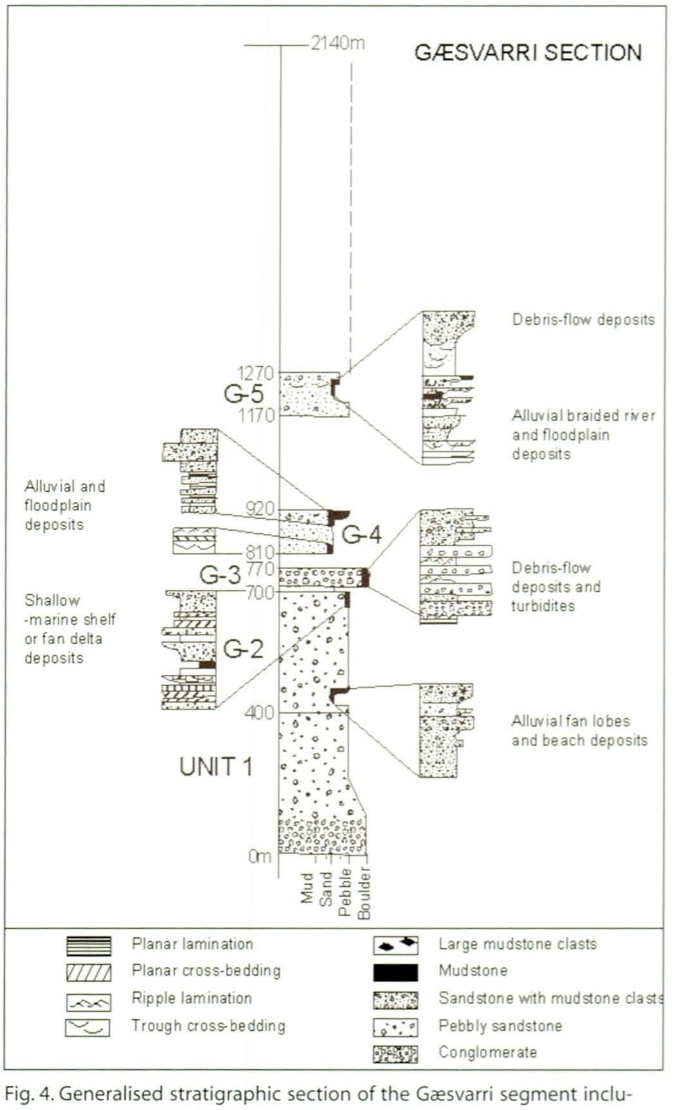

Fig. 4. Generalised stratigraphic section of the Gæsvarri segment including detailed sedimentological logs of parts of the units, main facies associations, sedimentary structures and depositional interpretations.

General stratigraphy and facies interpretations

The Čaravarri Formation constitutes a c. 4000 m-thick assemblage of siliciclastic sandstones, pebbly sandstones, conglomerates, quartzites, and some intercalated mudstones and siltstones (Fig. 2). The lower boundary of the formation is defined as the first occurrence of coarse, pebbly sandstone to conglomerate conformably overlying the argillite- and siltstone- dominated Bik'kačåk'ka Formation. This boundary marks the formational datum line and is, together with the basal unit 1, used for correlation of the reconnaissance sections (Fig. 2). Along strike, the Čaravarri Formation is split into three main segments by distinct transverse passes, possibly concealing faults (Fig. 2). Separate stratigraphic sections were established for the three main segments, Njar'gavarri, Gæsvarri and Čaravarri (Figs. 3, 4 and 5). Individual units characteristic of the separate sections of the formation have been designated by capital letters ('N' for Njargavarri, 'G' for Gæsvarri, and 'C' for the Čaravarri section), and numbered successively. Additional detailed logs are shown for the Gæsvarri section (Fig. 4). For the Čaravarri section, facies associations are described and interpreted from between calculated, metric-scale, stratigraphic levels.

The c. 400 m-thick basal unit 1, which contains conglomerates with distinct felsic porphyritic clasts, is the only stratigraphic unit that can be correlated along the three segments (Fig. 6). Most other units appear to be confined to each of the separate sections, with the exception of unit N-2 in the Njargavarri section which extends nearly 4 km northward into the neighbouring Gæsvarri segment, and the units G-2, G-3 and C2 (Fig. 6). The description of facies associations below starts with the underlying Bik'kačåk'ka Formation, since the contact is inferred to be sedimentary, then follows the unit 1 conglomerate, and finally, the three main sections of the Čaravarri Formation itself.

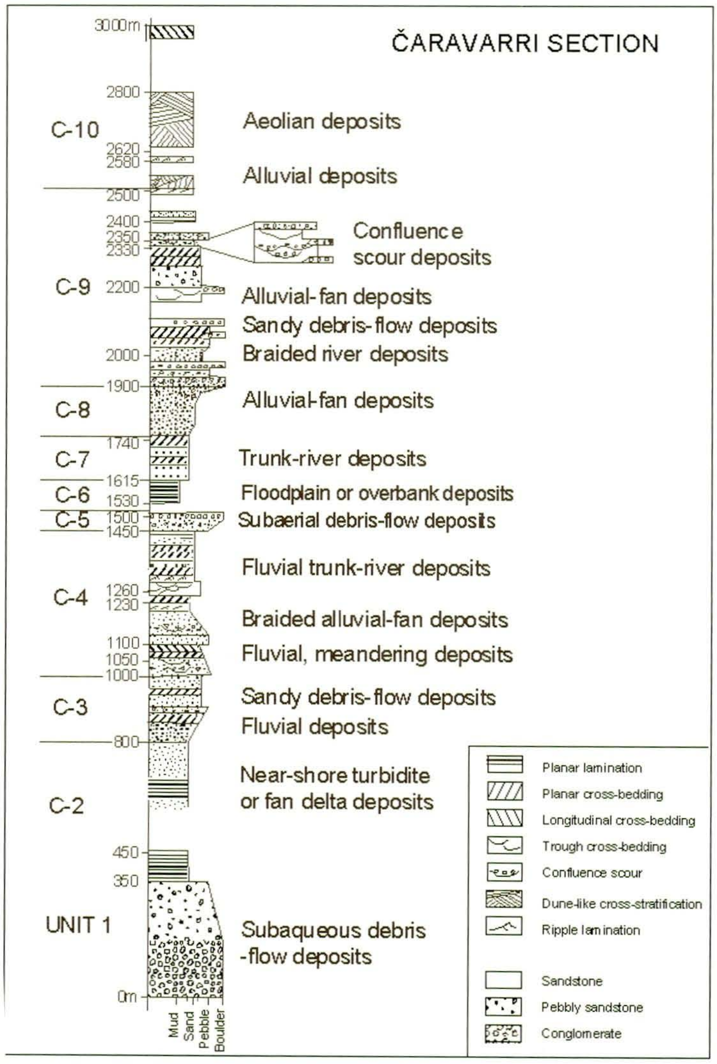

Fig. 5. Generalised stratigraphic section of the Čaravarri segment, with main facies associations, sedimentary structures and depositional interpretations.

Bik'kačåk'ka Formation

The very low to low-grade metamorphic Bik'kačåkka Formation (defined formally by Siedlecka et al. 1985) is c. 1000 to 1500 m thick and composed mostly of dark grey, laminated mudstones with thin intercalations of grey and green siltstone and sandstone. Near the upper boundary, Bik'kačåk'ka sandstones have desiccation cracks and symmetrical and transverse ripple marks indicating deposition in shallow water, with instances of emergence. The lower boundary of the formation is not observed in the study area but has been described as transitional, locally interdigitating with metavolcanic rocks of the Časkejas Formation farther west (Siedlecka et al. 1985). Its upper boundary is sharp to transitional towards the Čaravarri Formation (Siedlecka et al. 1985).

Čaravarri Formation: Unit 1

Facies associations: This basal unit of the Čaravarri Formation is up to 400 m thick, and is traceable in scattered outcrops along the whole length of the formation (Fig. 2). It consists of coarse-grained, light red sandstones and grey, pebbly to conglomeratic sandstone intervals and beds of polymictic conglomerate (Figs. 3–5). Well rounded granules and pebbles, averaging five centimetres in diameter (Fig. 7), of red-coloured felsic porphyries form a distinctive type of clast. Similar porphyries have not been reported from other Precambrian supracrustal rocks in Finnmark. Another clast type in the unit 1 conglomerate that distinguishes it from other Čaravarri Formation subunits is that of pebbles of vein quartz. Other clast types include microcline, plagioclase, lithic sandstone, quartzite, chert, metabasalt, silicified metabasalt, quartz-feldspar porphyry, vitroclastic feldspathic tuff and aphyric felsite, and there are also mudflake clasts and rare pebbles of red-oxidised granite.

The conglomeratic character decreases upward, through pebbly sandstones and into pale reddish-grey, coarse- to

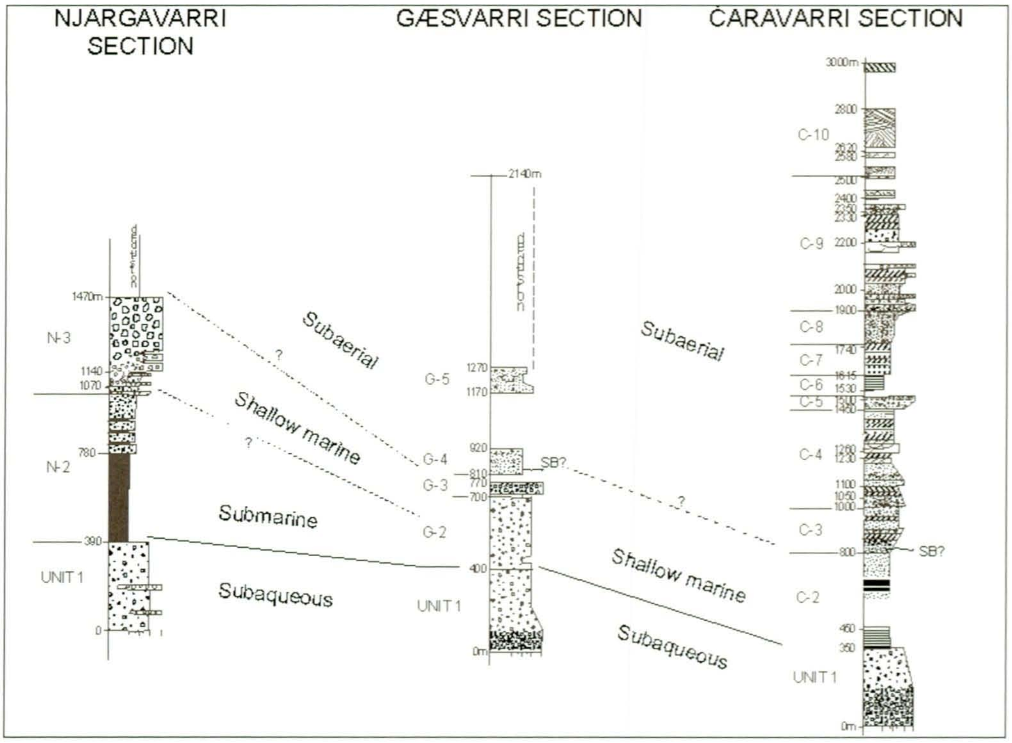

Fig. 6. Compilation of the three main sections of the Čaravarri Formation with overall depositional regimes and proposed correlation of units. Note the presence of an inferred sequence boundary (SB) above units N-2, G-3 and C-2.

medium-grained sandstone (Figs. 3 and 5). In outcrop, the sandstone facies types are mostly structureless, while the conglomeratic types are disorganised to crudely stratified (Fig. 7). As current-induced sedimentary structures were not found, no transport directions or provenance could be estimated. Only one outcrop locality (Fig. 7) in the northern, Čaravarri segment was found to yield information about the local mode of deposition. It shows an inversely graded conglomerate bed, about 10 cm thick, with clasts protruding into overlying massive and structureless sandstone. This conglomerate bed is underlain by a 5 cm-thick, coarse-grained sandstone, beneath which there is another, crudely bedded and inversely graded conglomerate.

Facies interpretation: The conglomerate rocks of unit 1 are interpreted as debris-flow deposits, with the associated sandstones representing sandy debris-flow deposits (cf. Shanmugam 1996). The generally structureless and disorganised appearance throughout most of the unit would indicate sediment gravity-flow deposition as the prevalent sedimentation mechanism. A subaquatic environment of deposition seems most likely for this extensive, thick unit. The provenance of the conglomerates, however, is unknown.

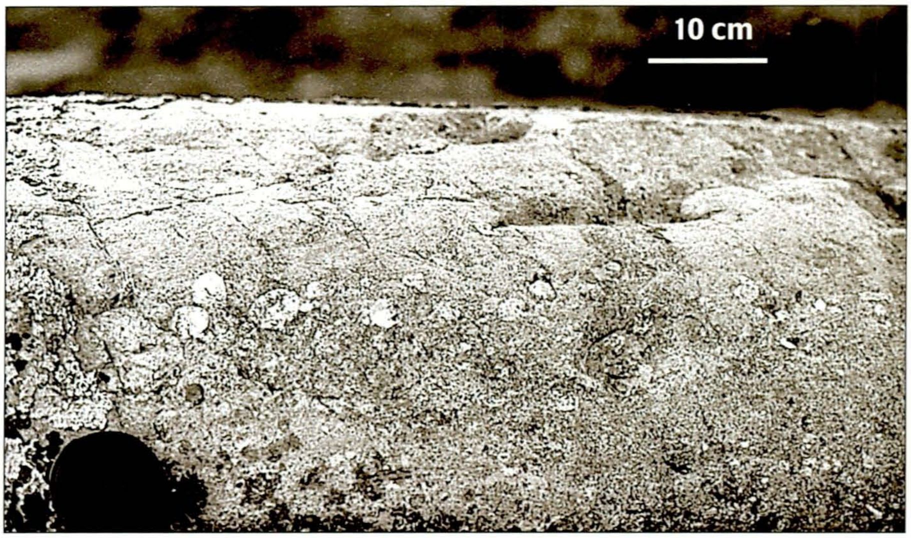

Fig. 7. Polymict conglomerate of unit 1 at Njargavarri. The conglomerate is inversely graded and includes well-rounded quartz and porphyritic pebbles protruding into overlying, massive sandstone, interpreted as a debris-flow deposit.

The Njargavarri section

Facies associations: This section is compiled from outcrops in the corresponding segment, with additional information from the adjoining hills Guivevarri and Lulli Lik'ča (Fig. 2). Above the unit 1 conglomerates (Fig. 3), ***unit N-2*** follows conformably; it comprises a dark grey, thinly bedded, fine-grained and flaggy mudstone succession in its lower part, and a greywacke sandstone succession in the upper part, with a total thickness of about 850 m. Individual mudstone laminae of the lower succession are commonly graded, and sandstone beds show loading into subjacent mudstone layers. The unit extends northward c. 4 km into the Gæsvarri section, where it wedges out. Higher up in the section, unit N-2 appears to be tectonically imbricated or partially excised by a bedding-parallel thrust. To the east of this fault, greywacke-like pebbly sandstones with thin conglomerate beds predominate. The latter may form single one-pebble layers or conglomerate beds up to 1 m thick in sandstone, as well as coarsening-upward beds with clasts increasing in size from 3 to 6 cm in diameter. Some of the conglomerate beds have a carbonate-rich matrix.

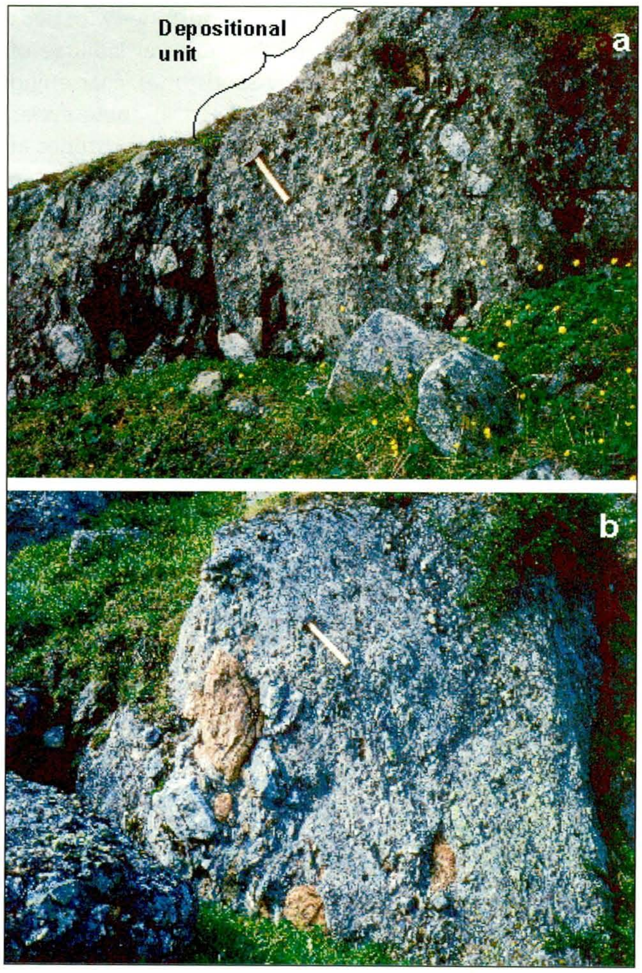

Fig. 8. a) Conglomerate of unit N-3 (Njargavarri segment), in part crudely bedded (parallel to hammer shaft) and with inverse grading. A possible single depositional unit is marked. Stratigraphic way up is to the right. View is toward NNW. b) Massive conglomerate of unit N-3. Note outsize dolostone boulder, 90 cm long (tan-coloured) in boulder cluster of smaller, subangular quartzite clasts. View is toward NNW and stratigraphic way up is to the right.

The Guivevarri and Lulli Lik'ča sub-segments, next to the bounding thrust zone (Fig.2), comprise a narrow belt of dark grey sandstones with intercalated thin beds of conglomerate with dolostone clasts that may correspond to unit N-2 at Njargavarri.

Unit N-3 is up to 350 m thick and composed entirely of conglomerates. Part of its lower boundary is erosional, apparently cutting c. 70 m downward into the underlying unit, N-2 (Fig. 3). The conglomerates are polymictic and disorganised to crudely stratified, with poor sorting and mostly subrounded to subangular clasts. The matrix, mostly of lithic wacke, consists of grey silty sandstone and, locally, carbonate-cemented quartz wacke and chlorite schist. Crudely bedded intervals may show normal as well as inverse grading of clasts (Fig. 8a). Disorganised conglomerates have scattered clasts oriented vertically relative to the crude stratification. The predominant clast material is quartzite, and subordinate greenstone. In higher stratigraphic parts of unit N-3, conglomerates also have conspicuous clasts of fine-grained, brown dolostone. Otherwise, conglomerate clasts in unit N-3 range in size from granules to boulders, with examples of dolostone reaching 90 cm in their longest dimension (Fig. 8b). Quartzite and dolostone represent the largest clast types. Carbonate clasts show by far the greatest size variation. Fragments of the underlying and adjacent, fine-grained, grey sandstone of N-2 type are also found. One single, bedded sandstone slab, about 1.5 m in length, is incorporated as an exotic clast (Fig. 9).

Interpretations: As the boundary between unit N-2 and the overlying conglomerate deposits of unit N-3 mostly is conformable, although locally erosive, our facies interpretation is that these highly contrasting subunits were deposited without interruption, and within closely related depositional regimes. The most abundant facies associations of ***unit N-2*** are uniform, thinly bedded to laminated and normally graded sandstones and mudstones; they are here interpreted as submarine, distal turbidites (cf. Hampton 1972, Walker 1992). The pebbly character of greywackes in the upper half of unit N-2 probably represents a more proximal turbidite facies.

The overlying, massive conglomerates of ***unit N-3*** are interpreted as cohesive to non-cohesive debris-flow deposits. This concentration of stacked, largely disorganised conglomerates of limited width (2 km preserved along strike) indicates that they may have been deposited within a near-shore constrictional channel or a submarine canyon (see Nemec 1990, Reading & Richards 1994). By comparison, the more regular, thin conglomerate beds and greywacke sandstones of unit N-2 may represent sediment gravity-flow material which had emerged from the canyon and been deposited on an unconstricted slope or fan.

Furthermore, the partly erosive basal surface of unit N-3, and inverse grading lowermost in the conglomerate, suggest that debris flows may have occupied a channel eroded

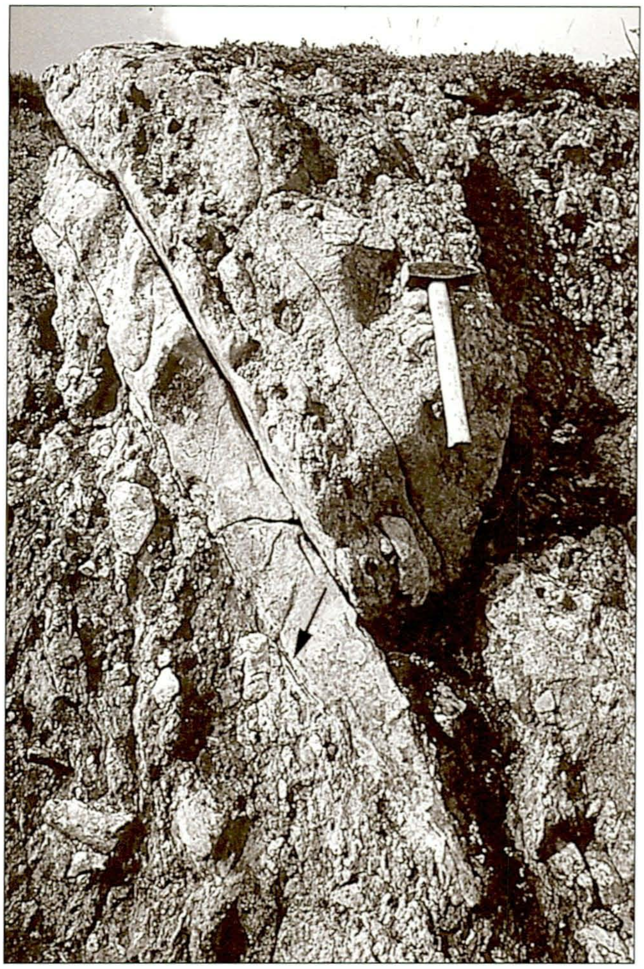

Fig. 9. Crudely bedded dolostone conglomerate bed of unit N-3 (Njargavarri segment) with exotic, short-distance transported sandstone slide, about 1.5 m long. Note primary shear fabric (arrow) and apparent drag structure in the underlying conglomerate. Stratigraphic way up is to the right.

earlier, probably by turbulent flows. The non-erosive character of this boundary in other parts of its course would be consistent with laminar flow of the debris. The occurrence of local stratification indicates that the gravel flows may have been variably liquefied and sporadically turbulent.

The character of the N-3 conglomerates allows an assessment to be made of clast provenance and local sediment transport. The labile, carbonate fragments were most likely eroded from sources considerably closer to the sedimentary basin than were the quartzite clasts. They may have been derived from a local carbonate shelf, where some of the largest dolomite boulders, present at the southern end of the conglomerate (Fig. 8b), may have been incompletely lithified during their redeposition. A rapid, northward reduction in clast size and local transition from conglomerate to pebbly sandstones within the inferred same stratigraphic level (unit N-3), are suggestive of a local southerly source area for the carbonate clasts. Thus, the largest dolostone clasts may have been supplied to passing debris-flows by breaking off the walls of E-W-trending channels or canyons. A similarly close provenance is inferred for the slide of carbonate pebble conglomerate-bearing sandstone with shear-induced cleavage along the sole, and with drag along the contact of the underlying conglomerate (Fig. 9).

The Gæsvarri section

Facies associations: The lowermost exposed part of the Gæsvarri segment (Fig. 4) starts with about 400 m of red and grey, disorganised conglomerates and pebbly to conglomeratic sandstones of unit 1. The conglomeratic character of this unit decreases upward, ending with an uppermost part of pale grey and red, medium-grained sandstone. Stratigraphic equivalents to units N-2 and N-3 of the Njargavarri section are absent in the Gæsvarri section, as these units likely pinch out northward (Fig. 6).

Unit G-2 is approximately 300 m thick in the area east of unit 1; to the south of the section, it overlies unit N-2 (Figs. 2 and 6). Characteristic facies associations in unit G-2 are shown in separate sedimentological logs for its lower and upper parts (Fig. 4). In the lower part of unit G-2, there is a recurrent pattern of medium-grained sandstones with beds of planar lamination, including heavy-mineral laminae of hematite and structureless pebbly sandstones, coarsening upward within 2 to 10 m-thick cycles (Fig. 4). These cycles start with fine-grained, massive to laminated sandstones at the base, grading into pebbly sandstone at the top. Within the middle part of unit G-2, the fine-grained parts are lacking, leaving a monotonous succession of amalgamated, structureless, pebbly sandstones higher up. The most common clast types are rounded quartzite in the lower part, and shale and argillite in the middle part. Angular mudstone flakes may be up to 25 cm in length. Locally, diffuse stratification is perceptible through variations in clast content.

In the upper c. 50 m of unit G-2, the pebbly sandstones are interrupted by 1 to 2 m-thick intercalations of planar cross-bedded, carbonate-rich, pebbly sandstones with mud clasts, and homogeneous, dark grey, faintly laminated siltstones and coarsening-upward, graded sandstones. Coarse to pebbly sandstone occur at the top, at the c. 700 m stratigraphic level. The latter rock has a number of irregular to lens-shaped, patchily distributed zones of coarse-grained carbonate, enclosing grains of the sandstone framework. The carbonate was identified in thin-section as secondary, poikilotopic (Friedman 1965) calcite cement.

After a covered interval of twenty metres, unit G-3 follows (Fig. 4). It is c. 50 m thick and made up of polymictic conglomerates. From the base upwards follows a succession of conglomerate beds, 1 to 7 m thick. Conspicuous clasts are quartzite, dolostone, greenstone and red jasper (Fig. 10). With the exception of the dolostone pebbles, these clast types are mineralogically more mature than similar clast types in the Njargavarri segment (where jasper clasts are unknown). Proportion variations among the G-3 clast types indicate the existence of two to five metres thick subunits within some of the conglomerate beds. Some of the conglomerates are matrix-supported (Fig. 10b). Less conspicuous, smaller clasts, identified in thin-section, include: silty

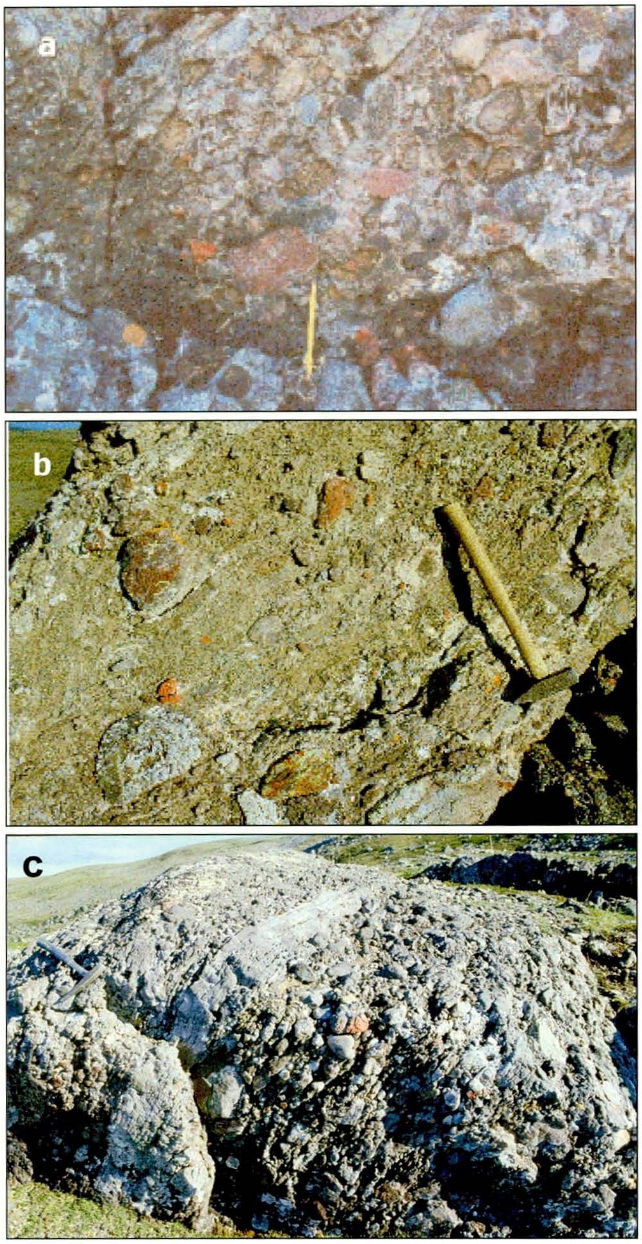

Fig. 10. Diverse conglomerate facies of unit G-3 of the Gæsvarri segment viewed toward the SSE: a) Disorganised, polymict debris-flow conglomerate, with subangular to rounded clasts of greenstone, diverse varieties of red jasper, and scattered clasts of carbonate. At its base (lower half), the conglomerate clasts penetrate into the underlying fine sandstone, probably by loading. b) Structureless, matrix-supported debris-flow conglomerate. c) Conglomerate composed of clasts of quartzite, red jasper, greenstone and sandstone. Note the bed of medium-grained sandstone with erosional base. Stratigraphic way up is to the left.

carbonate, fine-grained granoblastic quartzite with dust-sized carbonate inclusions, vein-quartz aggregates, metasiltstone, tourmaline-rich carbonate sandstone, subarkose, recrystallised jasper with remnants of colloform hematite and quartz intergrowths, fine-grained carbonate-cemented metasandstone, carbonate claystone, leucogabbro and marble. The lower conglomerate contact is irregular and characterised by loading of nearly vertical clasts into underlying, fine- to medium-grained sandstone. Some conglomerate beds are individually separated by up to 0.5 m-thick intervals and lenses of medium-grained sandstones with erosional bases (Fig. 10c). One of the conglomerates is capped by a 10 cm-thick graded, pebbly sandstone to mudstone deposit, with wave ripples on its upper surface (Fig. 11). Overlying this is 20 m of medium-grained, rippled sandstone and a pebbly to conglomeratic sandstone with dolostone clasts near the top.

Unit G-4 is 110 m thick, and made up of variously cross-bedded and thin-laminated, red and grey sandstones (levels 810 to 920 m). Medium-grained, feldspathic sandstones with large-scale, trough cross-stratification predominate in the lower, c. 20 m. This succession is capped by upward-coarsening, medium- to coarse-grained sandstone sheets with internal, planar and rippled laminations. A number of similar, recurrent fine-grained sandstone sheets occur in the middle part. The upper c. 30 m of unit G-4 consists of interbedded, 1-2 m-thick, coarse- and fine-grained, plane-laminated sandstones and grey mudstones, and red-coloured, sheeted pebbly sandstones at the top (to level 920 m).

Unit G-5 is the uppermost stratigraphic interval studied (from level 1170 m to 1270 m) in the Gæsvarri segment (Fig. 4). This unit, and the remaining part of the section (above 1270 m), largely consists of recrystallised, red and grey, rather nondescript quartzite. Unit G-5 starts with individual sandstone deposits that coarsen upward from fine-grained to pebbly sandstone within sets up to 5 m thick, with erosive bases. Above these, the pebbly sandstones contain large-scale trough cross-beds, with troughs up to 4 m wide and 0.75 m deep. The next c. 20 m include repeated, 1 to 5 m-thick, massive to trough cross-bedded sandstones, many with internal upward coarsening. At this level, the sandstones are separated by distinctive, conformable, red mudstones (<0.5 m thick). A 30 m-thick package of dominantly trough cross-bedded, coarse-grained sandstones, with troughs up to 10 m wide and 2 m deep, overlies the mudstone-sandstone intervals. The upper 20 m of unit G-5 consists of massive, pebbly sandstone, with scattered quartzite and mudstone clasts.

Interpretations: The lowermost part of *unit G-2* represents an upward-coarsening succession, in which the laminated and planar-bedded facies record upper flow regime conditions and the pebbly sandstones formed by deposition from sediment gravity flows (e.g., Hampton 1972). We interpret, tentatively, the sandstone, with its heavy-mineral laminae, as a foreshore sediment with beach lamination (cf. Elliot 1978). The upward-coarsening, pebbly sandstones overlying the beach deposits may consist of redeposited material from alluvial fan lobes along steep coastlines that have migrated above beach sediments (cf. Nemec & Postma 1993). Massive pebbly sandstone, transported as mass gravity flows, probably inherited a restricted clast size distribution, mostly 2 to 3 cm, from the source deposit, perhaps an older, elevated beach.

In the middle and upper parts of *unit G-2*, debris flows

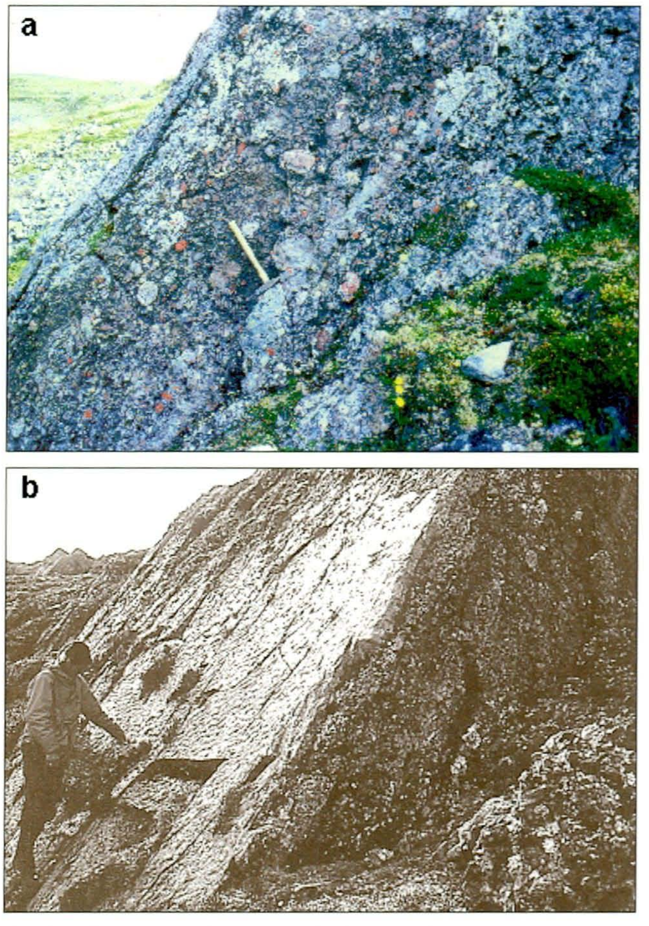

Fig. 11. Individual conglomerate bed of unit G-3 (Gæsvarri segment) viewed toward the SSE: a) Crudely bedded debris-flow conglomerate, c. 2 m thick, with lower half (to the right) rich in angular, brown, recessively weathered, carbonate clasts. The upper half (above hammer) lacks carbonate clasts, but contains scattered, rounded, larger siliciclastic clasts. At the top, the conglomerate is capped by a c. 10 cm-thick turbidite deposit that shows grading from sandstone to rippled mudstone. b) View of the upper surface of the turbidite deposit of figure 11a, showing rippled mudstone.

were predominant, with deposition of thick, massive and structureless, pebbly sandstones. Here, the presence of large, angular, argillitic clasts suggests erosion and resedimentation after equally short transport of the alluvial fans as envisioned above. In addition, angular argillite slabs up to 25 cm in length would indicate that these source beds were partly lithified at this time. Carbonate-rich, planar cross-bedded sandstones and the dark siltstone bed represent new rock types in the unit; they may indicate that the foreshore or beach sandstone below may have introduced a shallow-marine, littoral depositional regime (cf. Mueller & Dimroth 1987).

Unit G-3 is dominated by conglomerates with some rippled sandstone intervals. The irregular, loaded, lower conglomerate contact may be an indication of rapid deposition, perhaps as a result of aquaplaning of the debris flow down a steep gradient (cf. Postma et al. 1988, Falk & Dorsey 1998). The capping, thin, turbidite bed with a veneer of rippled mudstone at the top (Fig. 11b) is interpreted as formed by a turbidity current entrained from the faster-flowing debris flow below it.

The red and grey-coloured, trough cross-bedded sandstones in the lower part of unit G-4 may represent alluvial deposits (as described by Miall 1992), marking a change from the marine, near-shore depositional regime of the underlying unit G-3. The overlying, repeated successions of planar- and ripple-laminated, fine-grained sandstones may then represent fluvial or floodplain deposits.

Unit G-5 reveals that the constellations of red and grey, recurrently upward-coarsening pebbly sandstones with intervening mudstones, and large-scale trough cross-bedded sandstones appear consistent with an alluvial setting. The trough cross-bedding may then represent in-channel bedforms, but the largest ones are probably better explained as deep, braid-channel, confluence scours (cf. Siegenthaler & Huggenberger 1993). The mudstones could be overbank or floodplain deposits. The upper part of unit G-5 reflects sedimentation of successively coarser grained, hyperconcentrated flow to debris-flow deposits.

The Čaravarri section

The stratigraphic column of the Čaravarri segment (Fig. 5) is not a complete, exposed section but forms a reconnaissance traverse across the lower c. 3000 m of the about 4000 m-thick segment. The upper, c. 1000 m of it has been more or less pervasively recrystallised to quartzite in the proximity of the eastern, fault boundary of the Čaravarri Formation (Fig. 2). As these recrystallised rocks were largely covered and provided little stratigraphic and sedimentological information, they were identified but not studied more closely.

The lowermost 350 m of the section (***unit 1***) is made up of pebbly sandstones and porphyritic conglomerates. Above unit 1 follows a thick, monotonous succession of fine-grained sandstones, with local small-scale cross-laminations, and dark grey, laminated mudstones (***unit C-2***, levels 350-800 m).

Interpretation: The stratigraphic position of these fine-grained sediments (unit C-2) corresponds to that of *unit N-2* in the Njargavarri section, and probably also to parts of unit G-2, and they may be interpreted similarly, as near-shore turbidites or fan deltas prograding into a deeper-water basin (Figs. 3 and 4).

Unit C-3 (stratigraphic succession between levels 800 m and 1000 m) coarsens upward from massive siltstone and fine sandstone with mudclasts and clastic mica grains, to reddish-grey, medium- to coarse-grained, slightly pebbly sandstone, with tabular and planar cross-bedding.

Interpretation: Mudclasts occur within massive and structureless sandstones, which may be interpreted as prograding, fluvial, sandy debris-flow deposits.

Unit C-4 (between levels 1000 m and 1450 m) comprises pebbly sandstones and light grey to white, medium- to coarse-grained, non-pebbly sandstones. The pebbly sand-

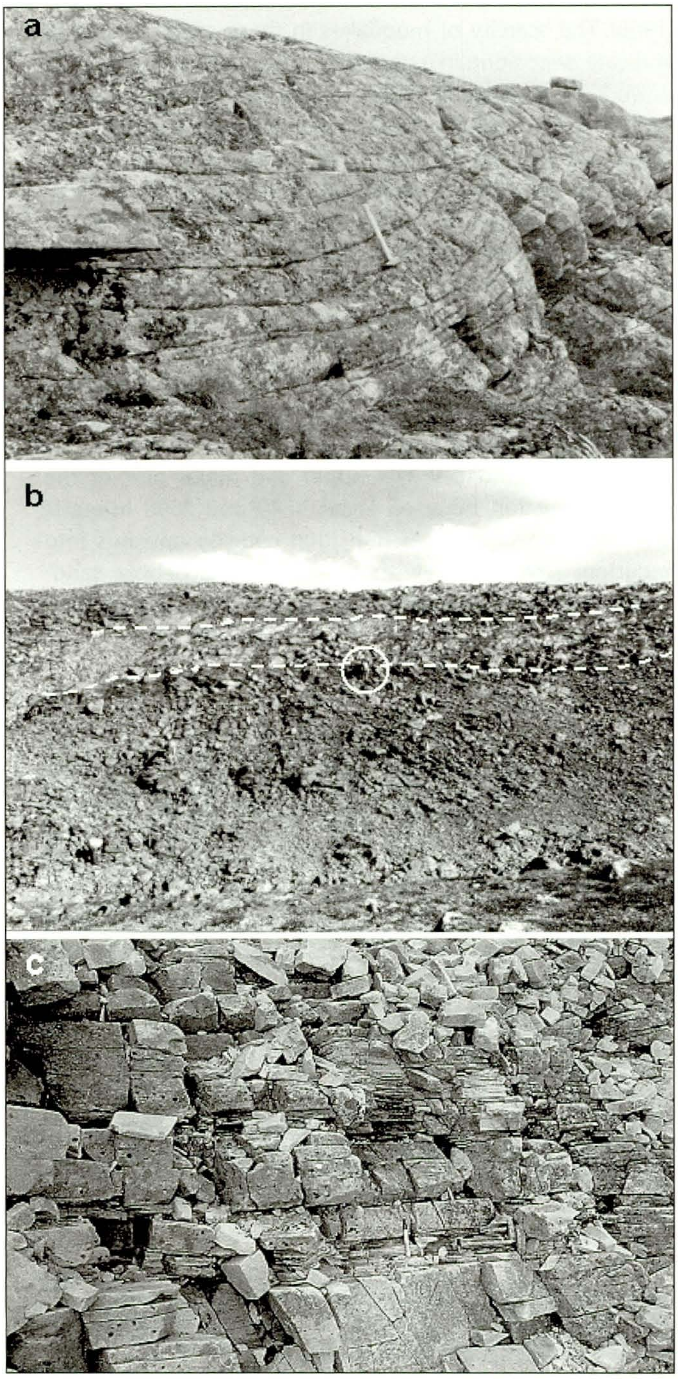

Fig. 12. a) Large-scale trough cross-bedding in pebbly sandstone of unit C4 (Čaravarri segment). View is toward the SE. b) Longitudinal cross-bedding in trunk-river sandstone of unit C-4 (between dashed lines), probably representing point bar deposits of a meandering river. Person ringed for scale. The cross-bedded sandstone is overlain by beds of mudstone and fine sandstone. View is toward the east. c) Parallel-laminated (horizontal), pale grey mudstone and bedded fine sandstone of unit C-6 interpreted as floodplain deposit associated with trunk-river sandstones.

stones show large-scale trough cross-bedding (Fig. 12a), with trough axes indicating deposition towards the west; i.e., transverse to the general strike of the Čaravarri Formation. Within this c. 400 m-thick succession, at least three distinct, pale grey-coloured and cross-bedded, non-pebbly sandstone intervals were identified. With thicknesses of approximately 50 metres each, they are separated by trough cross-bedded sandstone intervals of similar thickness (Fig. 5). The non-pebbly sandstone units show parallel bedding and tabular, planar cross-bedding. One typical unit starts with tabular, 5-20 cm-thick sets of planar cross-beds making up a c. 3 m-thick coset of medium-grained sandstone. This is followed by 3.5 m of fine-grained sandstone with subhorizontal beds, and then by 10 m of sandstone with a number of shale lenses, 0.2 m thick and 1.5 m long. The uppermost part represents massive sandstone that contains thin beds and lenses of mudstone. Mudstone clasts are, however, far less common here than in other sandstone types of the Čaravarri Formation. About 2 km farther north, along strike, we recognised longitudinal cross-bedding inclined toward the north and occurring in sets 1.5 and 2 m thick in a similar type of sandstone (Fig. 12b). Faint, smaller scale, sedimentary structures have been nearly wiped out by recrystallisation in the quartz rich sandstone.

Interpretation: The pebbly sandstones with large-scale trough cross-beds of *unit C-4* can be interpreted as braided alluvial fan deposits. The pale grey, plane-tabular, cross- bedded sandstones most likely reflect trunk-river deposits in front of transverse alluvial fan deposits (cf. Miall 1992). From the observed facies relations, at least three trunk-river intervals, separated by possible alluvial fan toes invading the trunk-river system from the east, exist in this part of the section. The longitudinal cross-bedded sandstone may then represent fluvial intercalations in the form of laterally accreted point-bar deposits in a sinuous, perhaps meandering, river (cf. Sweet 1988).

Unit C-5 (between levels 1450 and 1500 m) contains beds of red-brown, structureless, pebbly to cobbly sandstone and disorganised, matrix-supported conglomerate with subrounded to rounded clasts exposed over a thickness of at least 25 m. In some intervals, cm-thick, red siltstone to very fine sandstone beds subdivide these conglomerates into a number of depositional units.

Interpretation: These rocks are interpreted as a series of debris flows on the basis of their disorganised nature, lack of sorting, and the presence of vertically positioned cobbles and boulders. The red-brown coloration and the red coloration of intercalated fine deposits distinguish this particular debris-flow deposit from the majority of mass-flow conglomerates elsewhere in the Čaravarri segment, and suggest a subaerial environment of deposition.

Units C-6 and C-7: Between levels 1500 and 1615 m (*unit C-6*) there is a succession of parallel-bedded and laminated, pale grey and greenish-grey mudstones and argillites, with thin interbeds of fine-grained quartz, feldspar and mica-bearing sandstones (Fig. 12c). The finest types have millimetre-thin laminae, grading varve-like from silt to clay. Conformably overlying the *C-6* argillites is a succession of medium- to coarse-grained, pale grey sandstones with large-scale planar cross-beds that occur in tabular sets (*unit C-7*: levels 1615 to 1740 m). Parallel-stratified, fine-grained sandstone beds and lenses of argillite are present at several intervals.

Interpretation: The pale-grey colour of the argillites in *unit C-6* distinguishes them from the dark bluish-grey colours of other argillites, of inferred marine origin, elsewhere in the Čaravarri Formation; the pale colour may suggest deposition under freshwater conditions. The finely-graded silt and clay lamination and lack of current-induced structures indicate deposition by vertical accretion in stagnant water, possibly in a floodplain and/or fluvial overbank environment (see Wright & Marriott 1993, Sønderholm & Tirsgaard 1998). The sandstones of *unit C-7* are interpreted as fluvial trunk-river deposits similar to those between levels 1000 and 1400 metres. The association of fluvial, cross-bedded sandstones with fine suspension, argillitic deposits may indicate sedimentation in a mixed-load, channellised, leveed and probably perennial, river system with cohesive and well-defined banks, accompanied by a floodplain fed by overbank fines from the possibly meandering river (cf. Sweet 1988). The scarcity of mudflakes in these sandstones may indicate deposition in a perennial stream without pervasive reworking; the shale lenses, on the other hand, could possibly reflect deposition in abandoned channels. These associations, taken together, would be non-typical for coarse bed-load, braided rivers, as envisaged for the trough cross-bedded, pebbly to gravelly sandstones making up most of the Čaravarri section. The associations of alternating fluvial sandstones and inferred floodplain deposits are consistent with a river course migrating back and forth across a floodplain and building a multistorey sandstone architecture (cf. Martin & Turner 1998); with units *C-4, -6,* and *-7* constituting one facies association, punctuated by the transverse debris-flow deposit of *unit C-5* (Fig. 5).

Units C-8 and ***C-9***: The upper and major part of the Čaravarri section between levels 1740 and 2500 m starts with massive sandstones (*C-8*) that coarsen upwards into recurrent reddish-grey to yellowish-grey, pebbly sandstones and conglomerates (*C-9*) with well-rounded, nearly

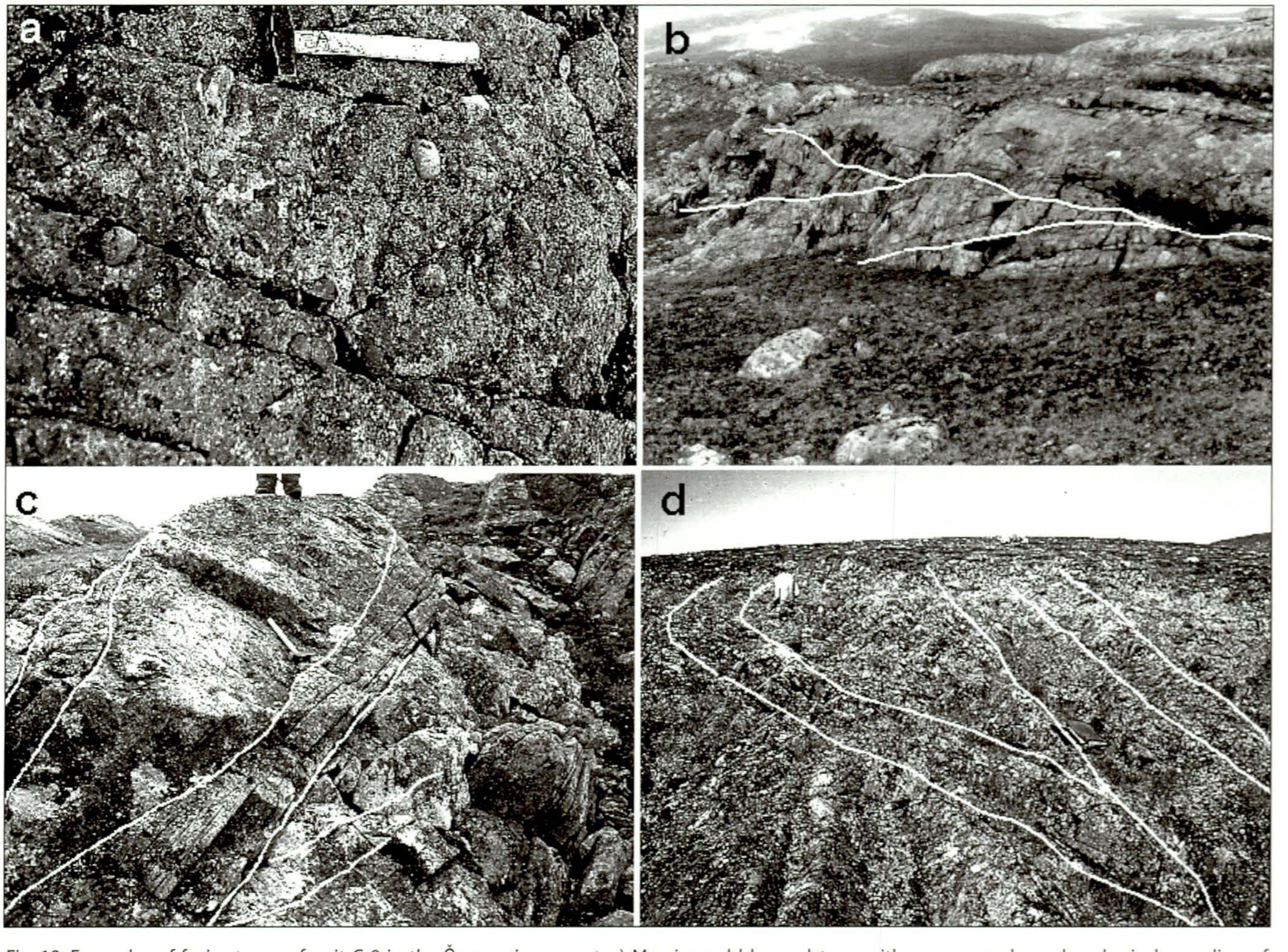

Fig. 13. Examples of facies types of unit C-9 in the Čaravarri segment: a) Massive pebbly sandstone with pronounced, nearly spherical rounding of quartzite and chert clasts. Stratigraphic way up is to the left. b) Mega-lens of massive, pebbly to conglomeratic sandstone (above) that truncates medium-grained sandstones with large-scale channel scours (below). View is toward the north. c) Close-up view of a single, parallel-laminated sandstone bed (below hammer head to the right) and an overlying pebbly sandstone unit with erosive base. View is toward the NNW. d) Giant channel in medium-grained sandstone. View is toward the SSE.

spherical quartzite and jasper pebbles (Fig. 13a). Some of the lowermost *C-9* conglomerates seem to be laterally extensive and can be traced for a few kilometres along strike. Farther up in the section, better exposed conglomerates can be easily delimited individually as lens-shaped units, up to a hundred metres in length, interbedded with, and sometimes truncating, medium- to coarse-grained sandstones with large-scale channel scours (Figs. 13b,c). Up to level 2350 m, erosionally based conglomerate lenses, as well as large-scale channel scours and a few trough cross-beds in the sandstones, are recurrent. Some conspicuous scours exceed 40 m in width and up to 8 m in depth (Fig. 13d). More commonly, channel scours are between 7 and 12 m wide and may have thin lag conglomerates along parts of their bottom surfaces. Pebbles and cobbles are generally well rounded and consist of quartzite, jasper, and chert.

Interpretation: The high mineralogical and textural maturity of the clasts contrasts with that of the host rock sandstones, which are mostly feldspar-bearing to feldspathic. This indicates that the clasts may record a significantly longer, polycyclic sedimentary history than the host sandstones with which they are intermixed. This upper part of the Čaravarri section is, in combination, interpreted as alluvial fan or slope deposits, intercalated with braided sandy debris-flow deposits (cf. Nemec & Postman 1993). The characteristic lens-shaped conglomerate bodies may represent gravelly braid bar deposits. Large-scale channel scours may form in a number of ways, and are difficult to interpret in detail from the Čaravarri outcrops. Some of the trough cross-beds observed may be remnants of in-channel bedforms. The larger channel scour structures, however, are better explained as confluence scours and fills that formed by erosion from channels establishing themselves on an alluvial fan or alluvial plain (cf. Bristow & Best 1993, Siegenthaler & Huggenberger 1993). Because they occupied the lowest topographic levels within the braid system, these deep scours had a high preservation potential; they were not as easily eroded, as were the shallower parts of the stream bed. Braided river morphology, however, has not been directly proven, but the coarse and pebbly sand to gravel type of sediment, and general lack of fining-upward grain sizes in the beds would seem to indicate braiding.

Unit C-10 (between levels 2500 and 3000 m) makes up the remaining part of the studied Čaravarri section. It includes large-scale, dune-like cross-bedded, pebbly sandstones frequently interlayered with thinly laminated and ripple-laminated sandstones with characteristic bimodally sorted laminae of coarse, well-rounded quartz grains, alternating with laminae with subrounded feldspar grains (Fig.14).

Interpretation: In isolation, the characteristic laminae and textures of these sandstones are interpreted as subcritically translatent, climbing wind ripple lamination (cf. Glennie 1970, Hunter 1977). These structures are common within a thickness of several hundred metres and are indicative of recurrent aeolian reworking of the sand during deposition. Bimodally sorted laminae typical of aeolian reworking (e.g.,

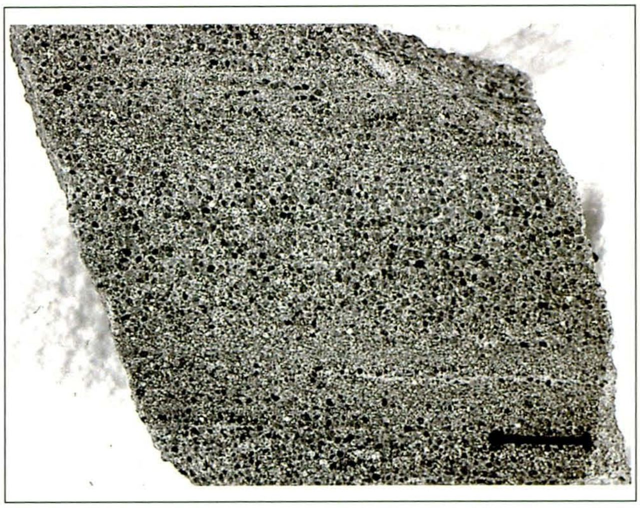

Fig. 14. Hand specimen of laminated, medium- to coarse-grained sandstone of unit C-10 (Čaravarri segment). Bimodal texture and well-rounded quartz grains indicate an aeolian origin.

Schenk 1983) were recognised among the sampled material only after the end of the field work. Some samples include poorly exposed, possible aeolian cross-bedding.

Stratigraphically above level 3000 m of the Čaravarri section, recrystallisation appears to have been pervasive, and the resulting quartzitic rocks were difficult or impossible to decipher, and therefore not included in the column (Fig. 5).

Depositional setting of the Čaravarri Formation: a summary

The base of the Čaravarri Formation is here defined as the lower, western boundary of the porphyritic conglomerates (*unit 1*), which is transitional towards the underlying Bik'kačåk'ka Formation mudstones. Unit 1 is the most extensive, coarsely clastic and most easily identifiable rock unit in the lower part of the Čaravarri stratigraphy, and the only unit that can be correlated in all sections (Fig. 6).

The *unit 1* conglomerate constitutes an exotic part of the lower Čaravarri sedimentary succession. Its deposition interrupted the steady accumulation of shallow-water mudstone and argillite in the underlying stratigraphic succession.

During or after deposition of *unit 1*, subsidence deepened the basin and affected the depositional regime, as compared to the shallow-water sedimentation in the upper parts of the Bik'kačåk'ka Formation. Above *unit 1*, in the Njargavarri segment of the basin (Fig. 6), dark grey, probably marine mudstones and shales (lower part of *unit N-2*) were deposited as distal type turbidites. *Unit N-2* coarsens upward, and has thin conglomerate beds in its upper part, indicating deposition as proximal turbidites. The upper part, *unit N-3*, of the Njargavarri segment, records proximal debris-flows in shallow-marine channels or canyons; thus the Njargavarri sediments, including the probably subaqueous unit 1 conglomerate, are marine over a thickness of 1470 metres (Fig. 6).

In the Gæsvarri segment, *unit G-2*, overlying *unit 1*, contains heavy-mineral bedded and laminated intervals and carbonates in otherwise massive sandstones; they may represent beach lamination formed in a near-shore environment. *Unit G-3* has polymict conglomerate beds deposited on a loaded fine substratum, attesting to deposition from sediment gravity flows in stagnant water, in a shallow-marine environment. The upper *units G-4* and *G-5*, characterised by red and grey, trough cross-bedded sandstones, with a number of thin, red- coloured, oxidised mudstones, reflect a marked change to largely terrestrial, fluvial and alluvial depositional conditions. The surface above *unit G-3* (and also above *unit C-2* of the Čaravarri segment; Fig. 6) may then be a sequence *boundary* of potential regional importance (cf. Posamentier & Allen 1993). This further indicates about 360 m of shallow-marine sediments, followed by a c. 460 m thickness of terrestrial sandstones in the Gæsvarri section, separated by a sequence boundary.

In the Čaravarri segment, the interval from 350 m to about the 800 m level (*unit C-2*) is characterised by dark, fine-grained sandstones, interpreted as marine, near-shore turbidite and fan delta deposits, that may be correlatable with *units G-2* through *G-3* of the Gæsvarri segment. Separated by a possible sequence boundary, the upper c. 2200 m-thick pile comprises terrestrial, coarse-clastic fluvial (braided and trunk-rivers) and alluvial channel deposition, with suspected, conspicuous aeolian sandstones in the highest part (Fig. 6).

Basin character and clast provenance

The three lateral segments of the Čaravarri Formation cannot be fully correlated along strike (Fig. 6); they differ markedly with respect to overall stratigraphy, facies associations, clast composition and maturity of coarse-clastic deposits, and may thus be considered as individual sedimentary entities separated by transverse faults. However, the proposed facies interpretations and transport directions of the individual entities made above, allow some assessment to be made of overall basin character and sediment provenance.

Firstly, the established stratigraphy of the Čaravarri Formation indicates deposition of a minimum 4000 m thickness of coarse-clastic sediments in a restricted, N-S-trending basin. The vertical and lateral disposition of facies and facies interpretations (Fig. 6) further indicate that the Čaravarri basin may have evolved as a shallow-marine shelf or near-shore depression, locally with carbonate platforms in the south that created repeated mass flow/debris flows and turbidity currents down the submarine canyons of the basin. This may have occurred through a restricted, terrestrial, fluvial-alluvial slope/plain setting in the northern part of the basin, including large progradational alluvial fans and eolian deposition.

Secondly, sediment transport directions for the shallow-marine and submarine facies seem uncertain. However, uniform transport directions from an easterly source can be demonstrated for the channelled trough cross-beds of the fluvial and alluvial successions.

Thirdly, there are distinct differences in composition, shape, size and maturity of clasts in the three segments. The southern segment is dominated by angular, locally derived carbonate, mudstone, greenstone and siltstone/sandstone clasts, the central segment contains subangular to better rounded quartz, jasper and a few greenstone-dolomite phenoclasts, and the northern segment is composed entirely of well-rounded quartzite and jasper phenoclasts. These observations suggest a younger age of clast deposition when moving northwards from segment to segment. Accordingly, the oldest, least mature clasts are preserved in the Njargavarri segment in the south, the Gæsvarri segment represents the intermediate situation, and the Čaravarri segment in the north has the youngest, most evolved clast population.

Fourthly, possible source rocks of the detrital clast material can be inferred. Most important, the abundant, well-rounded *quartzites* may be erosional products of the Lik'ča, Masi and Skuvanvarri Formations to the east (Fig. 2): The red- and grey-coloured *jasper/chert* and *porphyritic* clasts are believed to be derived from Palaeoproterozoic, porphyritic igneous rocks and banded iron formations, such as the Kittilä Greenstone Complex of north-central Finland (see Fig. 16). The angular *carbonate, mudstone, greenstone* and *sandstone/siltstone* clasts, on the other hand, may have been derived from a local source.

Foreland basin model

The Čaravarri basin is proposed to have developed in front of the Svecokarelian thrust belt as a N-S-trending foreland basin (Torske 1997). The main reason is that no master-bounding normal faults have been recognised; all bounding faults appear to be thrust generated. Despite this fact, the Čaravarri basin still remains structurally cryptic.

Foreland basins are commonly asymmetric and define linear to arcuate depressions situated on continental crust between a contractional fold and thrust belt and the adjacent craton (DeCelles & Giles 1996). They result from downward flexuring of the lithosphere (Beaumont 1981) possibly because of dynamic processes related to subduction, and are commonly filled longitudinally by sediments derived from the thrust belt (e.g., Eisenbacher et al. 1974, Cant & Stockmal 1989, Ricci Lucci 1986, Cowan 1993). By similarity, the Čaravarri sediments were deposited transversely from an eastern source within a narrow, N-S-trending, axially drained depression, while correlative marine-deltaic and fluvial sediments of the Skoadduvarri Sandstone at Alta farther north (Fig. 1) were deposited from a southerly source (Bergh & Torske 1986). Together with the notable change in clast maturity of the Čaravarri basin from south to north, these data suggest a drainage pattern known from many idealised foreland basins (e.g., Ricci Lucci 1986).

Thus, the major facies associations of the Čaravarri Formation can be tentatively applied to deposition in a foreland basin (Fig. 15). Following the deposition of the subaqueous Bik'kačåk'ka Formation mudstones, in the south, shallow-marine conditions prevailed throughout the depositional history, with subsidence-induced, relatively high gradients of shelf topography and instability generating fine-grained turbidite or fan delta sediments that spread across the basin (Fig. 15a). This may be due to the basin being partly underfilled (Burbank et al. 1988, Flemings & Jordan 1989) receiving sediments from more local sources, and resembling a carbonate-influenced, turbiditic foreland basin (Barnolas & Teixel 1994). In the central and northern segments, restricted shallow-marine and fluvial floodplain and trunk-rivers of the basin were overridden by large, westward progradational alluvial fans and fan deltas reflecting the end stage of coarse-clastic infilling of the basin (Fig. 15b), i.e., an *overfilled* foreland basin (Flemings & Jordan 1989, Burbank & Beck 1991). In this stage of foreland basin evolution the mature clasts reflect a more distal source; they were probably derived from the uplifted thrust belt to the east.

A foreland basin model may explain the observed vertical and lateral differences in stratigraphy and clast composition as due to changing provenance (local or distal), depocenter character and migration, and/or subsidence of the basin. For example, clasts of the lower stratigraphic sections of the Gæsvarre and Njargavarre segments may reflect locally derived material (e.g., Čas'kejas, Lik'ča and Bik'kačåk'ka formations), whereas mature quartzite and jasper clasts of the Čaravarri segment were most likely derived from a more easterly source, i.e., the uplifted thrust belt in the east. In addition, the lateral segmentation and sedimentation patterns of the Čaravarri basin may have been controlled by the thrust belt geometry, e.g., the presence of transverse faults in front of the thrust belt (Fig. 15b).

On the other hand, and in disfavour of a foreland model, most of the sedimentological features mentioned above can as well be applied to an extensional basin when looked upon in isolation. For example, rift basins normally start by being underfilled (the marine stage), passing up to be overfilled (the alluvial stage), and the sediments may be transported both transversely and axially (e.g., Gawthorpe & Leeder 2000). The great range of sedimentary infill patterns of foreland basins in general (e.g., Einsele 2000) may also question our model. Finally, a foreland basin that is underlain by thick metavolcanic successions (i.e., the Čas'kejas Formation) is rather unusual, at least in terms of modern foreland basins. If there is a primary contact between the Čaravarri Formation and underlying successions, it implies that the Čas'kejas Formation metabasalts may have formed in the same foreland basin. Thus, the mafic magmatism of the Čas'kejas Formation was a precursor to clastic sediment infilling of the basin, a feature which is common in many Precambrian foreland basins of the Canadian Shield (see Hoffman 1987). On the other hand, if there is a very large hiatus between the metavolcanics and the sedimentary successions, one may consider the Čas'kejas greenstones as a terrane that became accreted to the Fennoscandian craton, thus forming the cratonic crust below the Čaravarri basin sediments.

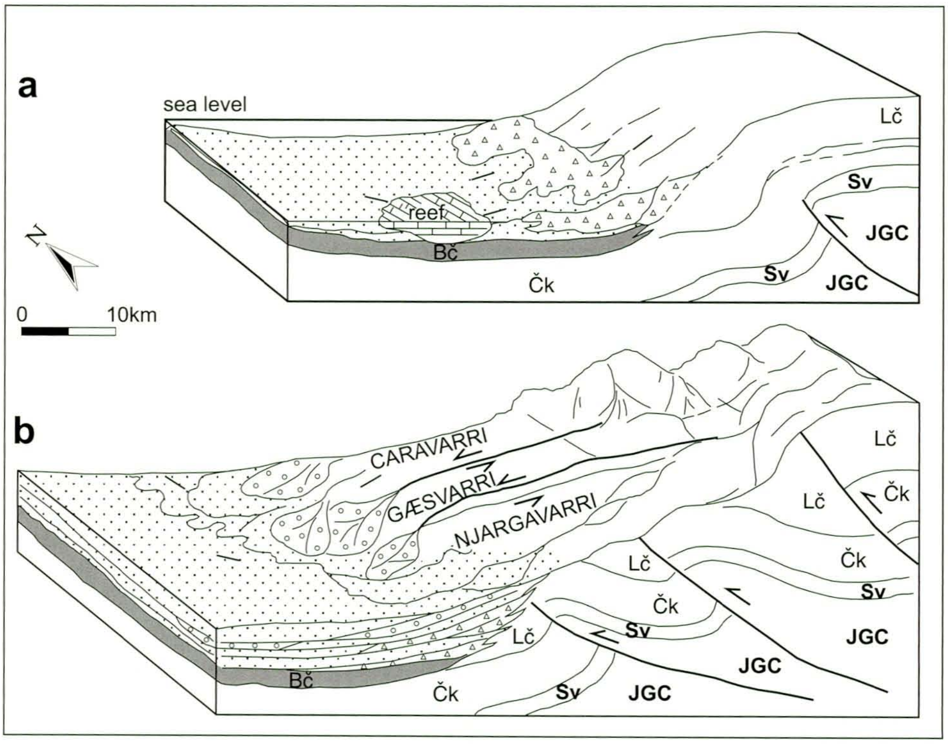

Fig. 15. Generalised block diagrams summarising the tentative evolution of the Čaravarri foreland basin in front of the uplifted Svecokarelian thrust belt to the east. a) Deposition of shallow-marine debris flows (unit 1 conglomerate), shelf turbidites and fan-delta sediments (units N-2, N-3, G-2, G-3 and C-2) in an underfilled foreland basin. b) Deposition of fluvial floodplain, trunk-river and westward progradational alluvial fan deposits (remaining parts of the sections) in an overfilled foreland basin. Arrows indicate local palaeocurrent direction. Note transverse faults that laterally segment the basin. Abbreviations are: JGC - Jer'gul Gneiss Complex, Sv - Skoganvarri Formation, Čk - Čas'kejas Formation, Lč - Lik'ča Formation, Bč - Bik'kačåk'ka Formation.

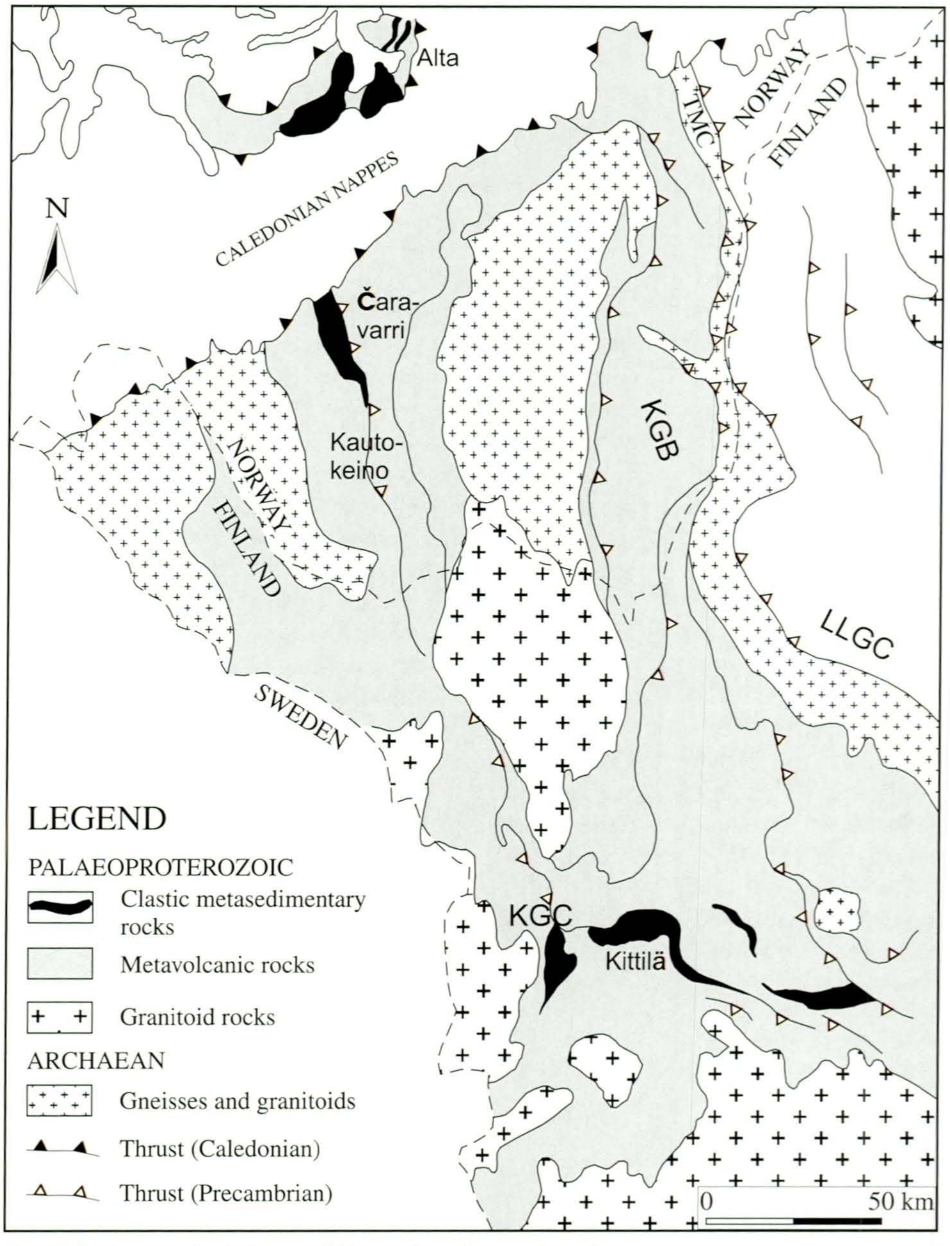

Fig. 16. Regional geological map of Finnmark and parts of northern Finland showing (in black) the location of Palaeoproterozoic molasse-type metasedimentary rocks, e.g. the Čaravarri Formation and correlative Kumpu and Laino Groups of the Kittilä area and Skoadduvarri Formation near Alta. Map is modified from Siedlecka et al. (1985) and Korsman et al. (1997). Abbreviations are as follows: KGB - Karasjok Greenstone Belt, KGC - Kittilä Greenstone Complex, LLGC - Levajok Lapland Granulite Complex, TMC - Tanaelv Migmatite Complex.

Palaeoproterozoic tectonic relations of the Čaravarri Formation; discussion and tentative conclusions

The interpretation of the Čaravarri Formation as a molasse-type foreland basin succession places important constraints on regional tectonics and correlation of Palaeoproterozoic units in the northernmost Fennoscandian Shield. For example, a correlation is suggested between the Kautokeino Greenstone Belt and parts of the Kittilä Greenstone Complex (Hanski et al. 1997) of northern Finland (Fig. 16). This complex includes several lower units of highly deformed and metamorphosed mafic and ultramafic, volcanogenic rocks (Sodankylä, Savukoski and Kittilä Groups), interpreted as remnants of 2.0 Ga oceanic lithosphere (Hanski et al. 1998, Rastas et al. 2001). They are unconformably overlain by the Lainio and Kumpu Groups which are composed solely of coarse-clastic, molasse-type sediments with clasts dated at 1.9 Ga (Räsänen et al. 1995, Hanski et al. 1997, 2000, Lehtonen et al. 1998). The Kumpu Group has recently ben interpreted as a foreland basin succession deposited after the Svecokarelian obduction, metamorphism and uplift of the underlying, deformed and metamorphosed lower units of the Kittilä Greenstone Complex.

Regarding the Finnmark portion of the Svecokarelian provinces, proposed plate-models also favour such a correlation and, besides, yield important time-constraints on the developmental history of the Čaravarri Formation relative to thrust provinces farther east (Fig. 16). These models suggest an early stage of crustal extension or rifting (2.1 Ga), producing various mafic volcanic units, e.g., the Kautokeino and Karasjok Greenstone Belts, followed by a subduction stage (2.0 Ga) with formation of arc-related igneous and volcaniclastic units, e.g., the metaflysch of the Levajok Granulite Complex and Tanaelv Migmatite Complex (Barbey et al. 1980, Krill 1985, Marker 1985, Berthelsen & Marker 1986). During the final continent collision between the Karelian craton and the arc-related oceanic provinces (1.9 Ga), the Levajok granulites, Tanaelv migmatites and Karasjok Greenstone Belt were thrusted toward the west (Krill 1985, Andreassen 1993, Braathen & Davidsen 2000). During this collision, imbricate, west-vergent thrusting and orogenic uplift also affected the Jer'gul Gneiss Complex and adjacent parts of the Kautokeino Greenstone Belt, i.e., the Čas'kejas and Lik'ča formations to the east of the Čaravarri Formation

(Holmsen et al. 1957, Vennervirta 1969, Krill 1985). On the other hand, thrusted units have not been observed west of the Čaravarri Formation.

A similar collisional evolution model has been advocated for the granulite belt and related Kittilä Greenstone Complex of northern Finland (Fig. 16; Hanski et al. 1996).

Acknowledgements

This project was supported by grants from the Norwegian Research Council and the University of Tromsø. Constructive reviews by Johan Petter Nystuen, Kauko Laajoki and Reidulv Bøe are gratefully acknowledged. We thank student field assistants Arild Seim, Kjell Bugge Johansen, Arnfinn Johansen, Sigvart Sørensen, Tor M. Broks and technician Hans Vasseng.

References

Andreassen, T.O. 1993: *Strukturelle og tektoniske undersøkelser i de nordlige deler av Karasjok grønnsteinsbeltet, Finnmark fylke.* Cand. Scient thesis, University of Tromsø, 146 pp.

Barbey, P., Convert, J., Martin, H., Capdevila, R. & Hameurt, J. 1980: Relationships between granite-gneiss terranes, greenstone belts and granulite belts in the Archean crust of Lappland (Fennoscandia). *Geologische Rundschau 69*, 648-658.

Barnolas, A. & Teixel, A. 1994: Platform sedimentation and collapse in a carbonate-dominated margin of a foreland basin (Jaca basin, Eocene, southern Pyrenees). *Geology* 22, 1107-1110.

Barth, T.F.W. & Reitan, P.H. 1963: The Precambrian of Norway. In Rankama, K. (ed.): *The Precambrian, I.* Wiley Interscience, New York, 27-88.

Beaumont, C. 1981: Foreland basins. *Geophysical Journal of the Royal Astronomical Society 65*, 291-329.

Bergh, S.G. & Torske, T. 1986: The Proterozoic Skoadduvarri Sandstone Formation, Alta, Northern Norway: a tectonic fan delta complex. *Sedimentary Geology 47*, 1-25.

Bergh, S.G. & Torske, T. 1988: Palaeovolcanology and tectonic setting of a Proterozoic metatholeiitic sequence near the Baltic Shield margin, northern Norway. *Precambrian Research 39*, 227-246.

Berthelsen, A. & Marker, M. 1986: Tectonics of the Kola collision suture and adjacent Archean and Early Proterozoic terrains in the northeastern region of the Baltic Shield, *Tectonophysics 126*, 31-55.

Braathen, A. 1991: *Stratigrafi og strukturgeologi sentralt i Karasjok grønnsteinsbelte, Finnmark*, Cand. Scient thesis, University of Tromsø, 130 pp.

Braathen, A. & Davidsen, B. 2000: Structure and stratigraphy of the Palaeoproterozoic Karasjok Greenstone Belt, north Norway – regional implications. *Norsk Geologisk Tidsskrift 80*, 33-50.

Bristow, C.S. & Best, J.L. 1993: Braided rivers: perspectives and problems. In Best, J.L. & Bristow, C.S (eds): *Braided rivers*, Geological Society Special Publication 75, 1-11.

Burbank, D.W. & Beck, R.A. 1991: Models of aggradation versus progradation in the Himalayan Foreland. *Geologische Rundschau 80*, 623-638.

Burbank, D.W., Beck, R.A., Raynolds, R.G.H., Hobbs, R. & Tahirkheli, R.A.K. 1988: Thrusting and gravel progradation in foreland basins: A test of post-thrusting gravel dispersal. *Geology 16*, 1143-1146.

Cant, D.J. & Stockmal, G.S. 1989: The Alberta foreland basin: relationship between stratigraphy and Cordilleran terrane-accretion events. *Canadian Journal of Earth Sciences 26*, 1964-1975.

Cowan, E.J. 1993: Longitudinal fluvial drainage pattern within a foreland basin-fill: Permo-Triassic Sydney Basin, Australia. *Sedimentary Geology 85*, 557-577.

DeCelles, P.G. & Giles, K.A. 1996: Foreland basin systems. *Basin Research 8*, 105-123.

Einsele, G. 2000: *Sedimentary basins – Evolution, facies and sediment budget.* Springer Verlag, Berlin. 792 pp

Eisenbacher, G.H., Carrigy, M.A. & Campbell, R.B. 1974: Paleodrainage pattern and late-orogenic basins of the Canadian Cordillera. In Dickinson, W.R. (ed.): *Tectonics and sedimentation.* Society Economic Paleontological Mineralogical Special Publications, 22, 143-166.

Elliot, T. 1978: Clastic shorelines. In Reading, H.G. (ed): *Sedimentary Environments and Facies.* Blackwell, Oxford, 143-177.

Falk, P.D. & Dorsey, R.J. 1998: Rapid development of gravelly high-density turbidity currents in marine Gilbert-type fan deltas, Loreto Basin, Baja California Sur, Mexico. *Sedimentology 45*, 331-349.

Flemings, P.B. & Jordan, T.E. 1989: A synthetic stratigraphic model of foreland basin development. *Journal of Geophysical Research 94*, 3851-3866.

Friedman, G.M. 1965: Occurrence and stability relationships of the aragonite, high-magnesian calcite and low-magnesian calcite under deep-sea conditions. *Bulletin of the Geological Society of America 76*, 1191-1196.

Gawthorpe, R.L. & Leeder, M.R. 2000: *Basin Research 12*, 195-218.

Glennie, K.W. 1970: *Desert sedimentary environments.* Amsterdam, Elsevier, 222 pp.

Hampton, M.A. 1972: The role of subaqueous debris flow in generating turbidity currents. *Journal of Sedimentary Petrology 42*, 775-793.

Hanski, E.J., Huhma, H., Lehtonen, M.I. & Rastas, P. 1997: Isotopc (Sm-Nd, U-Pb) and geochemical evidence for an oceanic crust to molasse evolution of the Palaeoproterozoic Kittila greenstone complex, northern Finland. *Norges geologiske undersøkelse Report 97*.131, COPENA conference, Abstracts and Proceedings.

Hanski, E.J, Huhma, H., Lehtonen, M.I., & Rastas, P. 1998: 2.0 Ga old oceanic crust in northern Finland. In, Hanski, E.J. & Vuollo, J. (eds.). International ophiolite symposium and field excursion: generation and emplacement of ophiolites through time. Abstract. *Geological Survey of Finland. Special Paper 26*, 24.

Hanski, E.J., Manttari, I., Huhma, H., & Rastas, P. 2000: Post-1.88 Ga deposits of the Kumpu and Laino group molasse-type sediments in northern Finland: evidence from conventional NORDSIM zircon dating. Abstract, 24. Nordiske geologiske vintermøte, Trondheim 6.-9. janur 2000. *Geonytt 1*, 75.

Hoffmann, P. 1987: Early Proterozoic foredeep, foredeep magmatism, and Superior-type iron formations of the Canadian Shield. In Krøn, A. (ed.): *Proterozoic lithospheric evolution.* American Geophysical Union, Geodynamics Series, 17, 85-89.

Holmsen, P., Padget, P. & Pehkonen, E. 1957: The Precambrian geology of Vest-Finnmark, northern Norway, *Norges geologiske undersøkelse 201*, 106 pp.

Holtedahl, O. 1918: Bidrag til Finnmarkens geologi. *Norges geologiske undersøkelse 84*, 1-34.

Hunter, R.E. 1977: Basic types of stratification in small eolian dunes. *Sedimentology 24*, 361-387.

Korsman, K., Koistinen, T., Kohonen, J., Wennerstöm, M., Ekdahl, E., Honkamo, M., Idman, H. & Pekkala, Y. (eds.) 1997: Suomen kallioperäkartta – Berggrundskarta over Finland – Bedrock map of Finland 1: 1000000. Geological Survey of Finland.

Krill, A. 1985: Svecokarelian thrusting with thermal inversion in the Karasjok-Levajok area of the northern Baltic Shield. *Norges geologiske undersøkelse 403*, 89-101.

Lehtonen, M., Airo, M-L., Eilu, P., Hanski, E., Kortelainen, V., Lanne, E., Manninen, T., Rastas, P., Räsänen, J. & Virransalo, P. 1998: The stratigraphy, petrology and geochemistry of the Kittilä greenstone area, northern Finland. *Geological Survey of Finland, Report of Investigations 140*, 144 pp.

Marker, M. 1985: Early Proterozoic (c. 2000-1900 Ma) crustal structure of northeastern Baltic Shield: tectonic division and tectogenesis, *Norges geologiske undersøkelse 403*, 55-74.

Martin, C.A.L. & Turner, B.R. 1998: Origins of massive-type sandstones in braided river system. *Earth-Science Reviews 44*, 15-38.

Miall, A.D. 1992: Alluvial deposits. In Walker, R.G. & James, N.P. (eds.): *Facies Models, Response to sea level change*, Geological Association of Canada, 119-142.

Mueller, W. & Dimroth, E. 1987: A terrestrial-shallow marine transition in the Archaean Opemica Group east of Chapais, Quebec. *Precambrian Research 37*, 29-55.

Nemec, W. 1990: Aspects of sediment movement on steep delta slopes. *Special Publications International Associations of Sedimentology 10*, 29-73.

Nemec, W. & Postman, G. 1993: Quaternary alluvial fans in southwestern Crete; Sedimentation processes and geomorphologic evolution. *In*: Marzo, M. & Puigdefabregas, C. (eds.), *Alluvial Sedimentation*, IAS Special Publication, 17, 235-276.

Olesen, O. & Sandstad, J.S. 1993: Interpretation of the Proterozoic Kautokeino greenstone belt, Finnmark, Norway , from combined geophysical and geological data. *Norges geologiske undersøkelse 425*, 43-64.

Olsen, K.I., & Nilsen, K.S. 1985: Geology of the southern part of the Kautokeino Greenstone Belt, West-Finnmark; Rb-Sr geochronology and geochemistry of associated gneisses and late intrusions. *Norges geologiske undersøkelse 403*, 131-160.

Oftedhal, C. 1981: *Norges geologi*, Tapir, Trondheim, 169 pp.

Posamentier, H.W. & Allen, G.P. 1993: Siliciclastic sequence stratigraphic patterns in foreland ramp-type basins. *Geology 21*, 455-458.

Postma, G., Nemec, W. & Kleinspehn, K. 1988: Large floating clasts in turbidites: a mechanism for their emplacement. *Sedimentary Geology 58*, 47-61.

Reading, H. & Richards, M. 1994: Turbidite systems in deep-water basin margins classified by grain size and feeder system. *American Association of Petroleum Geologists Bulletin 78*, 792-822.

Rastas, P., Huhma, H., Hanski, E.J., Lehtonen, M.I., Härkönen, I., Kortelainen, V., Mänttäri, I. & Paakkola, J. 2001. U-Pb isotopic studies on the Kittila greenstone area, central Finland. In, Vaasjoki, M. (ed.). *Radiometric age determinations from Finish Lapland and their bearing on the timing of Precambrian volcano-sedimentary sequences.* Geological Survey of Finland. Special Paper 33, 95-141.

Räsänen, Hanski, E., Juopperi, H., Kortelainen, V., Lanne, E., Lehtonen, M.I., Manninen, T., Rastas, P., Väänänen, J. 1995: New stratigraphical map of Central Finnish Lappland, 22nd Nordic Geologic Vintermeeting, Turku, 1996, Abstracts, p.182.

Ricci Lucci, F. 1986: The Oligocene to recent foreland basins of the Northern Apennines. *International Associations of Sedimentology Special Publications 8*, 105-139.

Schenk, C. 1983: Textural and structural characteristics of some experimentally formed eolian strata. In Brookfield, M.E. & Albrandt, T.S. (eds.): *Eolian sediments and processes: Development in Sedimentology 38*, 41-49.

Shanmugam, G. 1996: High-density turbidity currents: are they sandy debris flows? *Journal of Sedimentary Research 66*, 2-10.

Siedlecka, A., Iversen, E., Krill, A.G., Lieungh, B., Often, M., Sandstad, J.S. & Solli, A. 1985: Lithostratigraphy and correlation of the Archean and Early Proterozoic rocks of Finnmarksvidda and the Sørvaranger district. *Norges geologiske undersøkelse 403*, 7-36.

Siegenthaler, C. & Huggenberger, P. 1993: Pleistocene Rhine gravel: deposits of a braided river system with dominant pool preservation. *In* Best, J.L. & Bristow, C.S (eds.): *Braided rivers*, Geological Society Special Publication 75, 147-162.

Solli, A. 1983: Precambrian stratigraphy in the Masi area, southwestern Finnmark, Norway. *Norges geologiske undersøkelse 380*, 97-105.

Solli, A. 1990: Čarajavri, 1933 I – berggrunnsgeologisk kart – M 1: 50000. *Norges geologiske undersøkelse.*

Sweet, I.P. 1988: Early Proterozoic stream deposits: braided or meandering – evidence from central Australia. *Sedimentary Geology 58*, 277-293.

Sønderholm, M. & Tirsgaard, H. 1998: Proterozoic fluvial styles: response to changes in accomodation space (Riverdale sandstones, eastern North Greenland). *Sedimentary Geology 120*, 257-274.

Torske, T. 1977: A paleorift in the Precambrian of western Finnmark? Abstract, *Nytt fra Oslofeltgruppen 6*, 63.

Torske, T. 1997: The Čaravarri Formation of the Kautokeino Greenstone Belt, Northern Norway; Foreland basin sediments in front of the Lapland - Kola thrust belt? *Norges geologiske undersøkelse Report 97*.131.

Torske, T., & Bergh, S.G. 1984a: Relasjonen mellom Alta-Kautokeinoriften og Raipas-bergartene i Komagfjord-Repparfjordvinduet. In Often (ed.), *Norges geologiske undersøkelse Report 84.095*, 34-36.

Torske, T., & Bergh, S.G. 1984b: Carravarregruppen på Finnmarksvidda; Alluviale vifter og massestrømsavsetninger i et særpreget riftbasseng. *Geolognytt 20*, 49.

Vennervirta, H. 1969: Karasjok-områdets geologi. *Norges geologiske undersøkelse 258*, 131-184.

Walker, R.G. 1992: Turbidites and submarine fans. In Walker, R.G. & James, N.P. (eds.): *Facies Models, Response to sea level change,* Geological Association of Canada, 239-263.

Wright, V.P. & Marriott, S.B. 1993: The sequence stratigraphy of fluvial depositional systems: the role of floodplain sediment storage. *Sedimentary Geology 86*, 203-210.

Zwaan, K.B. & Gautier, A.M. 1980: Alta og Gargia. Beskrivelse til de berggrunnsgeologiske kart 1834 I og 1934 IV, 1:50,000. *Norges geologiske undersøkelse 357*, 1-47.