Williams, George Ellis.
Neoproterozoic (Torridonian) alluvial fan succession, northwest Scotland, and its tectonic setting and provenance, *Geological Magazine*, 2001; 138 (2):161-184.

*Geol. Mag.* **138** (2), 2001, pp. 161–184. Printed in the United Kingdom © 2001 Cambridge University Press

# Neoproterozoic (Torridonian) alluvial fan succession, northwest Scotland, and its tectonic setting and provenance

GEORGE E. WILLIAMS

Department of Geology and Geophysics, Adelaide University, South Australia 5005, Australia* and
Postgraduate Research Institute for Sedimentology, University of Reading, Reading RG6 6AB, UK

(*Received* 16 March 2000; *accepted* 10 November 2000)

**Abstract** – The presence of alluvial fan deposits in the lower Neoproterozoic Torridon Group in northwest Scotland illuminates Torridonian basin development at the eastern Laurentian margin. The 450 m thick Cape Wrath Member of the Applecross Formation consists of alluvial fan conglomerate and arkose succeeded by more distal, braidplain feldspathic sandstone. Palaeocurrent data comprising > 2650 measurements on trough cross-bedding are of low variability and show overall eastward flow. The projection upcurrent of regionally divergent flow directions for the lower part of the member indicates a fan of *c*. 50 km radius with its apex 30 km to the west near a basement (pre-Caledonian) normal fault with downthrow to the east beneath the north Minch Basin. Extensional tectonics controlled deposition of the Applecross Formation. Regional uplift, causing erosion of a youthful topography on the Lewisian Gneiss, was followed by the development of the Applecross extensional basin in two main stages. Uplift of a western source area by movement on basin-bounding normal faults occurred first in the north and caused pediplanation and alluvial fan deposition in the Cape Wrath area, with subsequent uplift of the source area for the main body of the Applecross Formation occurring further to the west and south along the line of the Minch Fault. The bulk of the Applecross Formation was derived from a weathered terrain of felsic crystalline and related supracrustal rocks reaching from the Outer Hebrides region westward for up to *c*. 250 km onto what are now the continental margins of the North Atlantic. The tectonic events may mark an early phase in the crustal extension that led ultimately to the opening of the Iapetus ocean.

## 1. Introduction

The Torridonian continental deposits of northwest Scotland were deposited on the Lewisian Gneiss at the eastern margin of Laurentia during late Mesoproterozoic–early Neoproterozoic times. Torridonian deposition may have been connected with a prolonged phase of crustal extension that preceded the opening of the Iapetus ocean in latest Neoproterozoic times (Stewart, 1991).

The Applecross Formation, which is the thickest and most widespread formation of the lower Neoproterozoic Torridon Group, has been interpreted as alluvial fan and braided stream deposits that accumulated in a rift basin with a faulted basin margin near the Outer Hebrides (Williams, 1969*a*; Stewart, 1982) and a source area reaching westward onto what are now the continental margins of the North Atlantic (Williams, 1969*b*). Subsequently, however, it has been argued that all the Applecross Formation is of distal braidplain origin and was deposited in a thermal subsidence basin with no active basin-bounding faulted margins and a source area > 300 km distant (Nicholson, 1993).

Clearly, an understanding of Applecross palaeoenvironments is critical to interpreting the tectonic setting of the Torridonian and the location and extent of the Applecross source area. This is the aim of the present paper: detailed sedimentological study and palaeocurrent analysis of the Applecross Formation in the Cape Wrath area permits discrimination between the above conflicting views on Applecross palaeoenvironments, basin development and provenance and illuminates early Neoproterozoic tectonism at the eastern Laurentian margin.

## 2. Geological setting

The Torridonian succession is *c*. 11.5 km thick and comprises the Stoer, Sleat and Torridon groups (Stewart, 1991). The Torridon Group has a total thickness of *c*. 6 km and includes, in order, the Diabaig, Applecross, Aultbea and Cailleach Head formations, and is exposed from Rum to Cape Wrath (Fig. 1). The Stoer Group, which unconformably underlies the Torridon Group, gave a Pb–Pb calcite age of 1199 ± 70 Ma and Diabaig and Applecross siltstones gave Rb–Sr whole-rock ages of 994 ± 48 Ma and 977 ± 39 Ma, respectively (Turnbull, Whitehouse & Moorbath, 1996).

---

*Permanent address; e-mail: george.williams@adelaide.edu.au

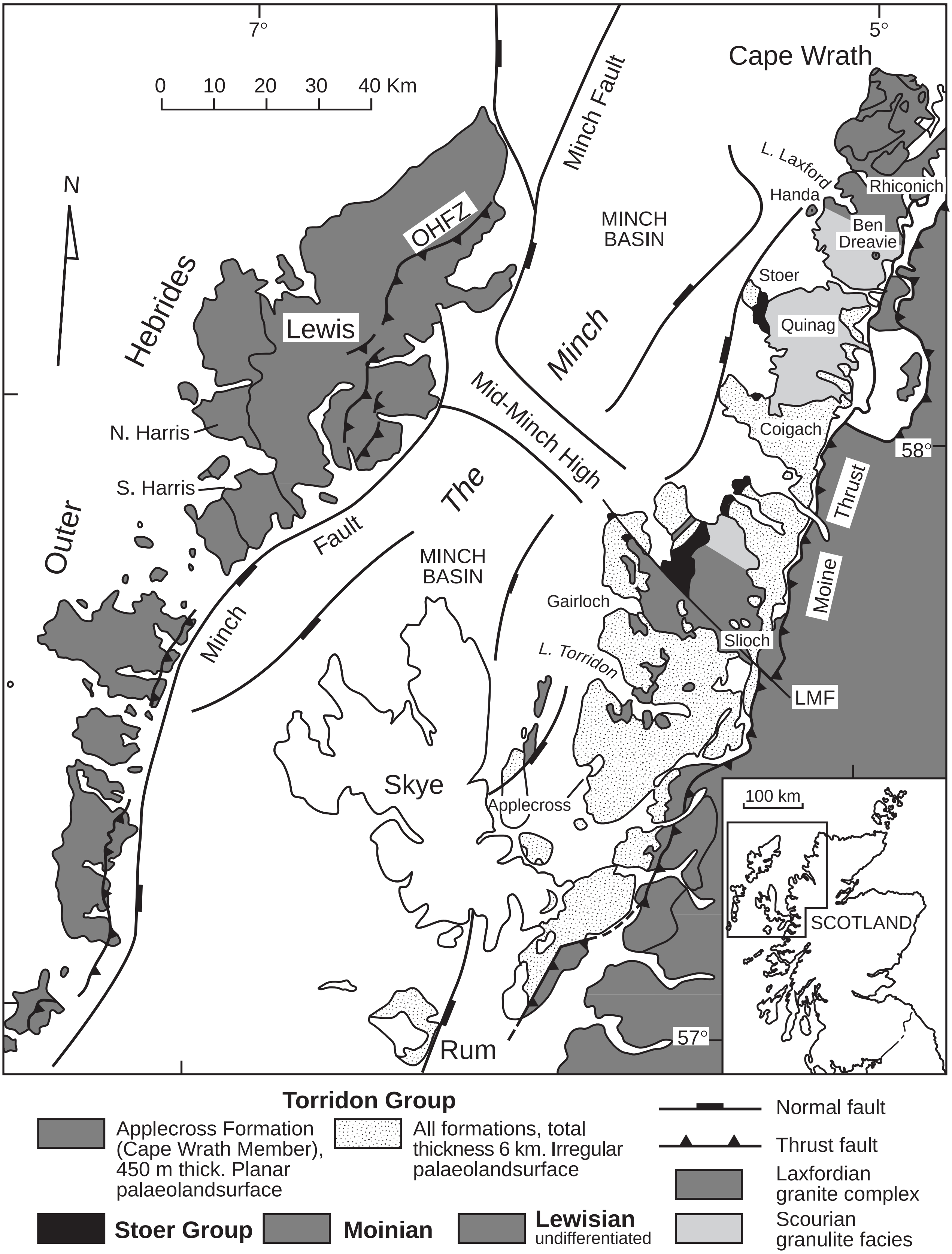

Figure 1. Geological map of northwest Scotland, showing the distribution of the Torridon and Stoer groups. The Scourian complex of the Lewisian Gneiss occupies the Central region of the mainland, and the Laxfordian complex occupies the Northern and Southern regions and is widespread in the Outer Hebrides. The tectonic elements for the Outer Hebrides and the late Palaeozoic–Mesozoic Minch Basin are from Fettes *et al.* (1992) and Stein (1988). OHFZ, Outer Hebrides Fault Zone; LMF, Loch Maree Fault.

Over most of the region the Torridon Group buries a Lewisian palaeolandsurface that is unweathered and irregular to mountainous, with relief as great as 600 m and deep valleys marking youthful palaeotopography (Fig. 2b,c; Penck, 1897; Peach *et al.* 1907). Near Cape Wrath, by contrast, a weathered palaeolandsurface of low relief is developed on the Lewisian and overlain by a 450-m-thick succession assignable solely to the Applecross Formation (Williams, 1969*a*). There, the general parallelism of the Lewisian unconformity and major bedding planes in the overlying strata is revealed in detailed exposures of the unconformity with relief of < 1 m and sections up to several kilometres long (Fig. 2a; Williams, 1968*a*, 1969*a*). Hence the Applecross Formation in the far north was deposited on a Lewisian landsurface that was mostly planar and near-horizontal. Geological, geochemical and mineralogical data indicate that the weathering profiles on the palaeoplain, which usually are 1–3 m thick with local tongues of alteration reaching depths of ≥ 6 m along joints, are palaeosols (Williams, 1968*a*; Retallack & Mindszenty, 1994; Young, 1999*a*). Palaeomagnetic data for the palaeosols and Applecross sandstones (Williams & Schmidt, 1997) indicate that the palaeosols formed penecontemporaneously with Applecross deposition in moderate palaeolatitudes (35–40°).

The Applecross Formation that occurs in an area of 150 km$^2$ between Cape Wrath and Rhiconich (Fig. 1; here referred to as the 'Cape Wrath area') comprises conglomerate and sandstone and is termed the Cape Wrath Member. From Quinag southward the Applecross Formation is mostly sandstone and up to 3.5 km thick (Stewart, 1991). The underlying Diabaig Formation includes fan breccias and lacustrine deposits that accumulated within palaeovalleys in the Lewisian palaeolandsurface. The Aultbea Formation

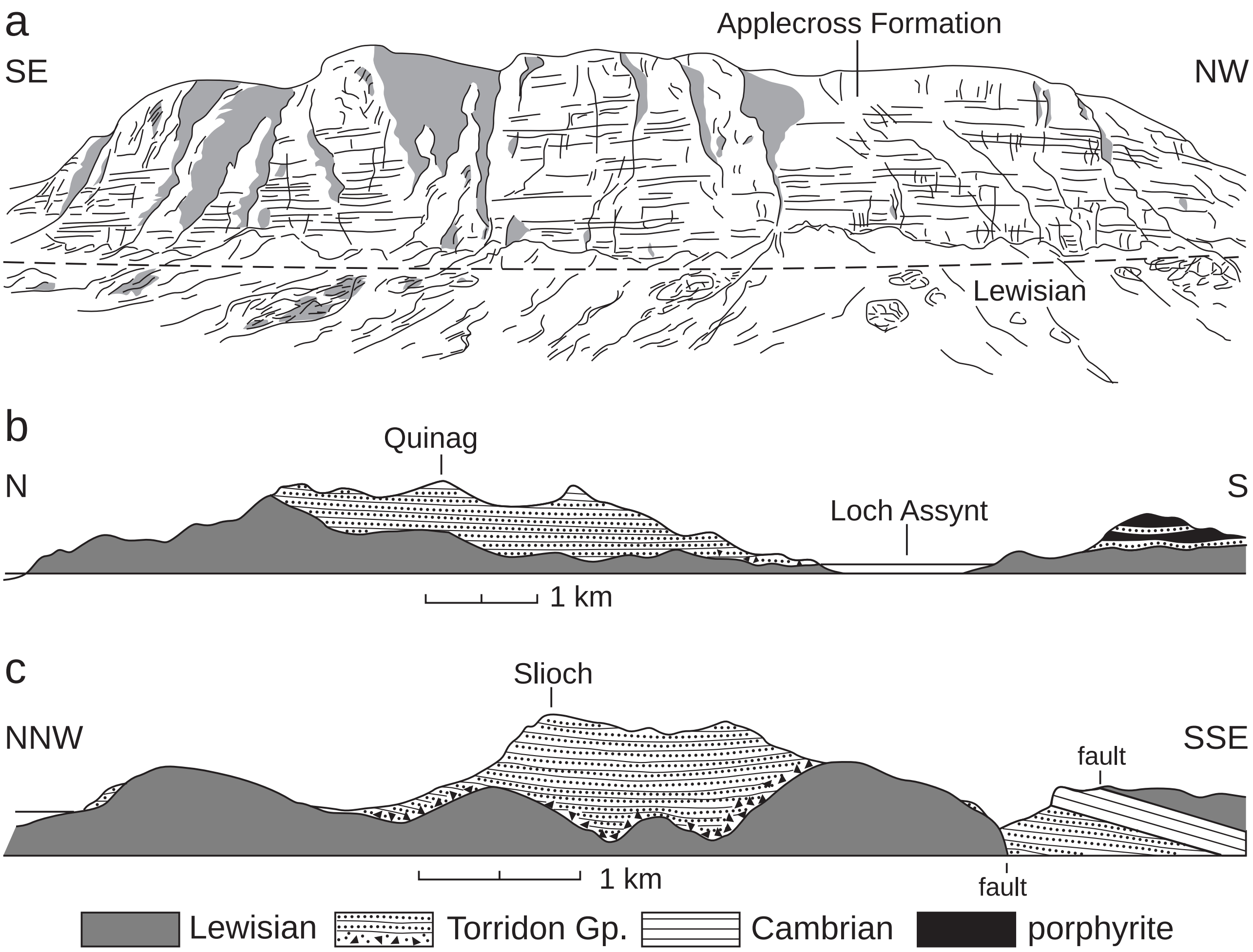

Figure 2. Sections showing the Lewisian palaeotopography buried by the Torridon Group; no vertical exaggeration of scale. (a) Cliff section at Creag Riabhach in the Cape Wrath area [NC 278643–285632], showing the sedimentary architecture of the lowest 150 m of the Cape Wrath Member (facies assemblage 2 and lower facies assemblage 3), with major bedding planes paralleling the unconformity with the Lewisian Gneiss (dashed line). Exposure 1.3 km long. (b) Section across Quinag, showing an irregular Lewisian palaeolandsurface overlain by the Applecross Formation and basal breccia of the Diabaig Formation (adapted from Peach *et al.* 1907, plate XXXI). (c) Section across Slioch, showing a mountainous Lewisian palaeolandsurface overlain by the Applecross Formation and Diabaig basal breccia (adapted from Peach *et al.* 1907, p. 277).

is a distal fluvial deposit genetically related to the Applecross Formation, which it conformably overlies. The Cailleach Head Formation may be deltaic.

The mainland Lewisian is divisible into Northern, Central and Southern regions (Park, 1991). The Central region comprises the Scourian granulite complex that formed at 2.9–2.7 Ga. The Northern and Southern regions consist of the Laxfordian complex that was affected by Laxfordian deformation and metamorphism: (1) 1.86–1.63 Ga, main Laxfordian amphibolite-facies metamorphism, with the emplacement of granites and pegmatites at 1.75–1.65 Ga (Taylor, Jones & Moorbath, 1984); (2) 1.5 Ga, late Laxfordian metamorphism and deformation; (3) 1.4–1.1 Ga, late Laxfordian and/or Grenvillian brittle deformation and metamorphism (movement on crush belts). Laxfordian deformation and metamorphism are virtually ubiquitous in the Outer Hebrides, and Laxfordian granites, pegmatites and migmatites are widespread (Fettes *et al.* 1992). A Grenvillian event with uplift and cooling occurred in Harris and Lewis at 1.1 Ga (Cliff & Rex, 1989).

## 3. Cape Wrath Member in the Cape Wrath area

The Cape Wrath Member shows progressive upward fining and is divisible stratigraphically into five facies assemblages (FA1–FA5; Table 1 and Fig. 3) based on the proportion of conglomerate and sandstone, architecture of bedding, and sedimentary structures. Strata are low-dipping. The type area is in the southwest, with the base of the succession defined at Droman [NC 185594] and the top at An Grianan [NC 265627]. Measured sections are shown in Figure 4 and typical exposures in Figure 5a–e.

### 3.a. Cyclicity in FA2 and FA3

Coarsening-upward cycles usually 6–8 m thick occur in FA2 between Oldshoremore and Sandwood Loch and also at Creag Riabhach. The cycles (Fig. 4b) typically comprise three units, as follows:

(1) At the base, a unit 1.0–4.3 m thick of sandstone, usually trough cross-bedded and locally plane-bedded or rippled, that coarsens upward from coarse sand to granule grade.

(2) A middle unit 1.2–4.6 m thick of tabular conglomerate beds and trough cross-bedded, coarse-grained to granule sandstone. Some conglomerates contain thin (< 10 cm) lenses of sandstone. Most pebbles are < 40 mm in diameter. The unit tends to fine toward the top.

(3) At the top, a tabular conglomerate unit 0.6–1.5 m thick, commonly with pebbles 40–60 mm long and showing no obvious sign of pedogenesis. Sandy lenses are rare. The upper bounding surface (Fig. 5b) usually is sharp, planar and laterally extensive.

The sedimentary architecture of FA2 (Fig. 2a) shows that the bedding planes bounding the cycles persist for 0.5 to > 1 km and are classifiable as 5th-order bounding surfaces of Miall (1996, p. 82).

Cycles 6–9 m thick that occur low in FA3 (Fig. 4c.1) are like those of FA2 except that in the cycles of FA3 the middle unit contains a larger proportion of sandstone and the top unit is thinner; higher in FA3 this top conglomerate is absent (Fig. 4c.2). Hence, progressing up the succession the coarsening-upward cycles of FA2 are replaced by symmetrical cycles in FA3.

### 3.b. Red and grey striping in FA4 and FA5

Grey streaks in the upper 30 m of FA4 produce a red and grey striping that is best seen in the Clò Mòr cliffs. The grey streaks are guided by bedding but in detail are cross-cutting, showing that the striping occurred after deposition. The displacement of grey streaks by joints indicates that the striping preceded joint formation. Grey streaks are common also in the lowest 30 m of FA5 (Fig. 5f).

### 3.c. Soft-sediment deformation structures in FA5

FA5 displays abundant soft-sediment deformation, including: (1) structureless sandstone that may pass laterally and/or vertically into well-bedded sandstone; (2) convolutions; (3) steepened trough sides; (4) steepened and overturned cross-beds; (5) domed bedding involving several beds over a thickness of 2+ m (Fig. 5e); (6) diapiric structures 6 cm to 2 m high (Fig. 5f). Many of the structures are truncated, showing that the deformation occurred during deposition. Their origin is discussed below in Sections 5.d and 8.b.

## 4. Cape Wrath Member south of the Cape Wrath area

### 4.a. Handa and Ben Dreavie

Handa consists of *c.* 350 m of low-dipping strata assignable to FA1–4, with thicknesses comparable to those in the Cape Wrath area. FA1 is exposed in the northeastern cliffs, where the presence of cobbles > 120 mm long indicates proximity to the Lewisian unconformity. Coarsening-upward cycles occur in FA2 in the northern cliffs, and FA3 and FA4 occur in the northwest and west.

A small (400 × 240 m) outlier of conglomerate and cross-bedded sandstone 15 m thick occurs at Ben Dreavie [NC 261392]. Clasts > 100 mm in diameter are common and the strata are assignable to FA1. Reddened gneiss crops out within 10 m, with no sign of major relief in the unconformity.

### 4.b. Quinag and Stoer

The low-dipping succession at Quinag (Table 2) is 520 m thick and overlies an unweathered Lewisian palaeolandsurface with a relief of *c.* 500 m (Fig. 2b). The

Table 1. Succession of facies assemblages for the Cape Wrath Member of the Applecross Formation, Cape Wrath area

| Facies assemblage | Palaeoenvironment |
|---|---|
| FA5. 90+ m thick. Type section at An Grianan (Fig. 4e) and well exposed also at Beinn a'Chraisg. Pale red, coarse-grained sandstone ($\geq 98$ % of facies assemblage). Trough cross-bedding and tabular cross-bedding equally common. Troughs mostly 0.5–2.0 m wide and 10–20 cm thick, forming cosets. Tabular cross-bedded sets 10–40 cm thick, rarely 1 m thick. Minor plane-bedded sandstone. Local channel forms $\leq 30$ m wide and 1.6 m deep. Rare pebble bands, pebbles $< 30$ mm in diameter. Abundant soft-sediment deformation structures (Fig. 5e, f). Grey streaks and patches (Fig. 5f) common in the lowest 30 m. | Distal braidplain setting, with deposition mainly on dunes and bars in shallow channels. Rare pebble lags. Poorly drained tract with high water-table, causing soft-sediment deformation structures. |
| FA4. 90 m thick in southwest, 120 m in north. Type section at Port Mòr (Figs 4d, 5d). Red, mostly coarse-grained sandstone ($\geq 90$ %). Trough cross-bedding and tabular cross-bedding equally common. Troughs typically 2–3 m wide and $\leq 40$ cm thick, forming cosets. Tabular cross-bedded sets 0.1–3.0 m thick. Channel forms $\leq 30$ m wide and 1.6 m deep. Minor plane-bedded sandstone and rare beds of red mudstone. Local ripple mark and ripple cross-lamination. Tabular pebble-conglomerate beds $\leq 20$ cm thick ($\leq 10$ %), pebbles $<30$ mm in diameter. Bands of polygonal mudstone intraclasts. Rare fining-upward cycles 2–3 m thick of sandstone and mudstone. Grey streaks and patches common in the uppermost 30 m. | Braidplain setting beyond the alluvial fan, with deposition mainly on dunes and bars in channels up to 3+ m deep. Rare distal sheet gravels. Local desiccation and erosion of muds. Laterally migrating streams producing fining-upward cycles. |
| FA3. 90 m thick. Type section at Port Beag (Fig. 4c.1). Tabular pebble-conglomerate beds $\leq 1$ m thick (35–10 %); most pebbles $< 40$ mm in diameter, supported by a sandstone matrix; sandy lenses $<10$ cm thick. Trough cross-bedded, pale red, coarse-grained sandstone (65–90 %); cosets $\leq 6$ m thick, troughs 0.5–2.0 m wide and $\leq 40$ cm thick (Fig. 5c,g). Minor tabular cross-bedded sandstone sets $\leq 1$ m thick. Rare plane-bedded and rippled sandstones. Symmetrical cycles 6–9 m thick of sandstone and conglomerate. | Distal alluvial fan setting, with deposition by sheetfloods and on dunes and bars in shallow channels. Symmetrical cycles record deposition without abandonment of fan lobes. |
| FA2. 150 m thick in southwest, 100 m in northeast. Type section at Sheigra (Fig. 4b). Tabular pebble-conglomerate beds 0.1–2.5 m thick (70–35 %) that maintain their thickness laterally over hundreds of metres; many pebbles $>40$ mm in diameter; clast-supported or supported by a sandstone matrix; usually ungraded, but some normal and inverse grading; major axes of clasts usually oriented roughly parallel to the bedding, with some imbrication and orientation along the local flow direction. Trough cross-bedded, pale red, very coarse-grained sandstone (30–65 %); troughs 0.5–2.0 m wide and 10–30 cm thick. Isolated tabular cross-bedded sandstone sets $\leq 3.0$ m thick (Fig. 5h). Laterally persistent, coarsening-upward cycles 6–8 m thick of sandstone and conglomerate, with tops of cycles sharp and planar (Fig. 5b). | Mid-alluvial fan setting, with deposition by sheetfloods and on dunes and bars in channels up to 3+ m deep. Coarsening-upward cycles record the repeated progradation and abrupt abandonment of fan lobes. |
| FA1. 30 m thick in southwest, absent in northeast. Type section at Droman (Fig. 4a), and the Lewisian unconformity and palaeosols are exposed at Poll a'Mhurain (Fig. 5a) and Sheigra [NC 183608]. Trough and tabular cross-bedded and lenticular pebble- and cobble-conglomerates $\leq 4$ m thick (70 %); erosional contacts and channelling; clasts $\leq 180$ mm in diameter, supported by a sandstone matrix. Trough and tabular cross-bedded, pale red, very coarse-grained sandstone (30 %). Red mudstone with desiccation cracks at Sheigra. | Proximal to mid-alluvial fan setting, with deposition by flashy discharges on gravel and sand dunes and bars in channels up to 4+ m deep. Local deposition of overbank muds. |

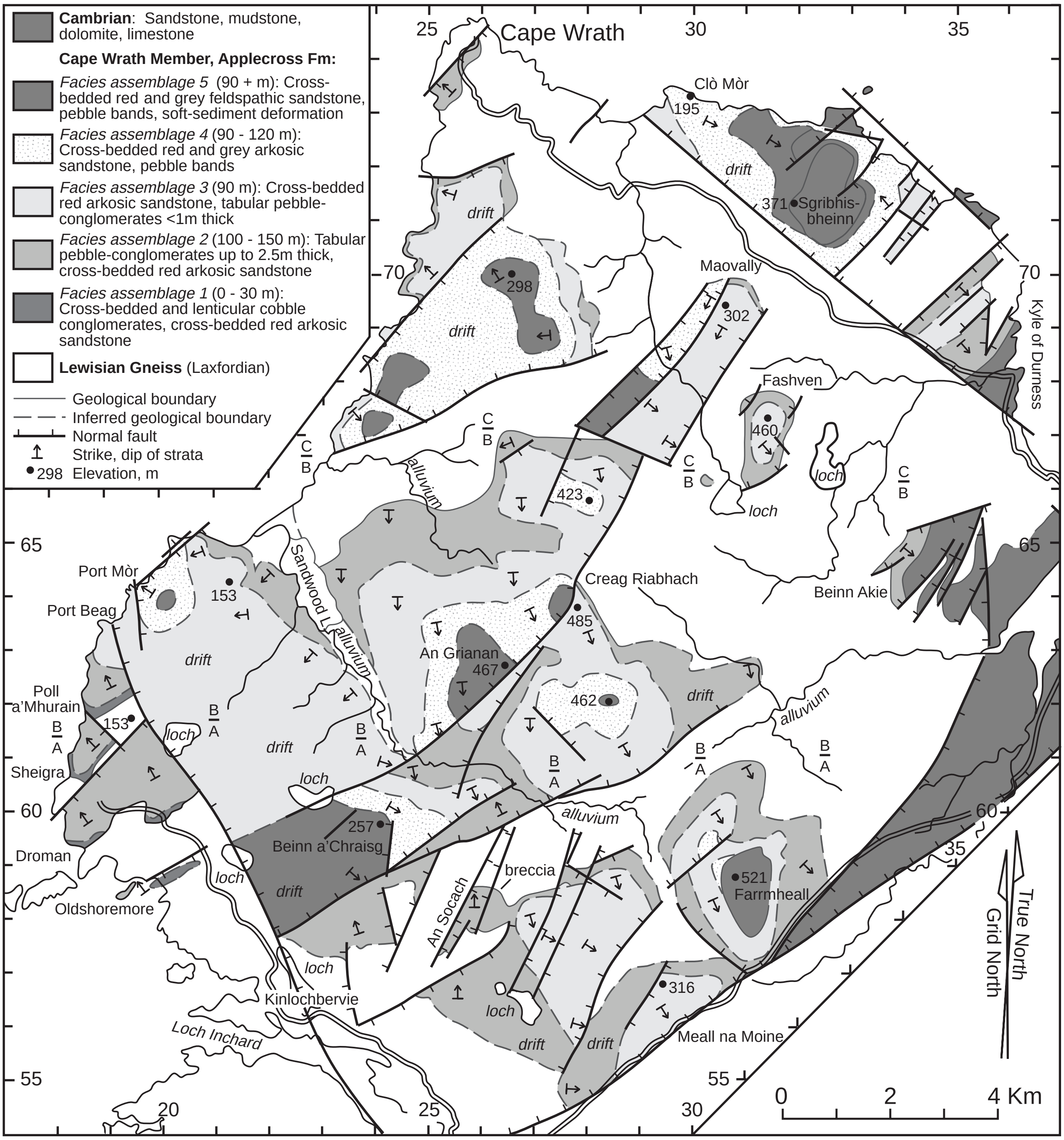

Figure 3. Map of the Applecross Formation (solid geology) in the Cape Wrath area, showing the distribution of facies assemblages 1–5 of the Cape Wrath Member. Dips are ≤ 10–15° except near faults where they may reach 20–25°. $\frac{B}{A}$ and $\frac{C}{B}$ mark east–west boundaries between subareas A (south), B and C (north) for palaeocurrent analysis. The Lewisian palaeolandsurface is of low relief except near An Socach, where a gneiss hill is mantled by locally-derived breccia (Williams, 1969*a*). The member is unconformably overlain by Cambrian quartzites. Numbers along the border refer to the 1 km National Grid (NC series). See also Cape Wrath, Scotland, Sheet 113, Solid and Drift Geology, 1:50 000, Provisional Series (Keyworth, Nottingham: British Geological Survey, 1997).

lack of weathering suggests stripping of weathered material from Lewisian slopes prior to burial.

The Torridon Group rests on the Stoer Group by the Bay of Culkein (Williams, 1966*a*) and commences with 20 m of sandstone–boulder conglomerate, sandstone and siltstone of the Diabaig Formation. These beds are overlain by 50 m of conglomerate and sandstone like upper FA3 which dip westward to a meridional fault that runs through the bay. About 100 m of south-dipping sandstone like FA4 occur between the fault and Rubh'an Dunain on the western side of the bay. Applecross strata east of Rubh'an Dunain are

Table 2. Stratigraphy of the Torridon Group at Quinag

| Member or formation | Lithology | Thickness (m) |
|---|---|---|
| FA4 | Trough and tabular cross-bedded, pale red, coarse-grained sandstone, with local bands of mudstone or pebbles. Grey streaks and patches and soft-sediment deformation structures near the top of the succession. | 250+ |
| FA3 | Mostly trough cross-bedded, pale red, coarse-grained sandstone; minor pebble-conglomerate. | 150 |
| FA2 | Tabular pebble-conglomerate beds ≤ 1.7 m thick; clasts ≤ 60–80 mm in diameter and like those in the Cape Wrath area. Mostly trough cross-bedded, pale red, coarse-grained sandstone. No cyclicity. | 30 |
| cf. FA3 | Mostly trough cross-bedded, pale red, coarse-grained sandstone. Thin pebble-conglomerate beds; pebbles < 20 mm in diameter and like those in the Cape Wrath area. | 75 |
| Diabaig Formation | Locally-derived basal breccia ≤ 40 cm thick, at Loch Assynt overlain by tabular, greyish red, poorly sorted sandstone with thin mudstone and conglomerate beds. | 15 |
| | | Total 520+ |

FA2–4 refer to facies assemblages 2–4 of the Cape Wrath Member of the Applecross Formation.

assigned to the Cape Wrath Member. The succession west of Rubh'an Dunain is *c*. 780 m thick and has affinity with the Applecross Formation to the south.

The Cape Wrath Member wedges out south of the Coigach district and lies stratigraphically below the main body of the Applecross Formation (Williams, 1969*a*; Stewart, 1991). The southern Applecross Formation is finer grained than FA1–2 and much is comparable to FA5.

## 5. Sedimentary features and their environmental significance

### 5.a. Trough cross-bedding

The trough cross-bedding in the Cape Wrath Member (Table 1) typifies that formed by the migration down-current of subaqueous dunes. Hence the down-current azimuth of trough cross-bedded sets seen on bed surfaces (Fig. 5g) or in three dimensions provides a reliable indicator of local flow direction. Such azimuths are of low variability (see Section 7), which is consistent with fluvial deposition. The trough cross-bedded and lenticular conglomerates in FA1 are comparable to the deposits of fluvial channels that display three-dimensional gravel dunes and are marked by flashy discharges (Brierley, Liu & Crook, 1993; Khadkikar, 1999).

### 5.b. Tabular cross-bedding

Tabular cross-bedded sandstone sets (Fig. 5h) are ≤ 3 m thick and typically have a planar lower surface and planar top. Sets accreted laterally for up to 60 m, with most wedging out and a few passing laterally into trough cross-bedding. Mudstone and conglomerate cross-beds and reactivation structures occur within some sets.

The dip direction of cross-beds in tabular sets commonly deviates widely from the local flow direction indicated by adjacent trough cross-bedding. The deviation for 18 typical sets is 16–102° and averages 55° (Williams, 1966*b*). At least 80 % of the tabular sets in FA1–3 in the Cape Wrath area have cross-bed dip directions that deviate to the north relative to typical eastward local flow directions, whereas higher in the succession and for all facies assemblages from Handa to Quinag the cross-beds in tabular sets have no obvious preferred direction of dip.

The tabular cross-bedded sets are best compared to the deposits of fluvial channel bars (e.g. Collinson, 1970; Williams, 1971). Isolated mudstone cross-beds record the temporary ponding of a channel. The thickness of a channel-bar deposit provides a minimum estimate of channel depth, hence channels attained depths of ≥ 3–4 m during deposition of the Cape Wrath Member.

### 5.c. Tabular conglomerates

The main features of the tabular conglomerates in FA2 (Table 1) agree with those of sheetflood deposits in fluvial settings and on alluvial fans (e.g. Harvey, 1984; Nemec & Steel, 1984; Brierley, Liu & Crook, 1993), particularly on alluvial fans with catchments underlain by crystalline rocks such as granite and andesite (Blair, 1999). A fluvial environment marked by an abundant supply of sand, pebbles and cobbles and flashy sheetflood discharge is indicated.

### 5.d. Soft-sediment deformation structures

Soft-sediment deformation structures are important only in FA5 and are similar to those characterizing the

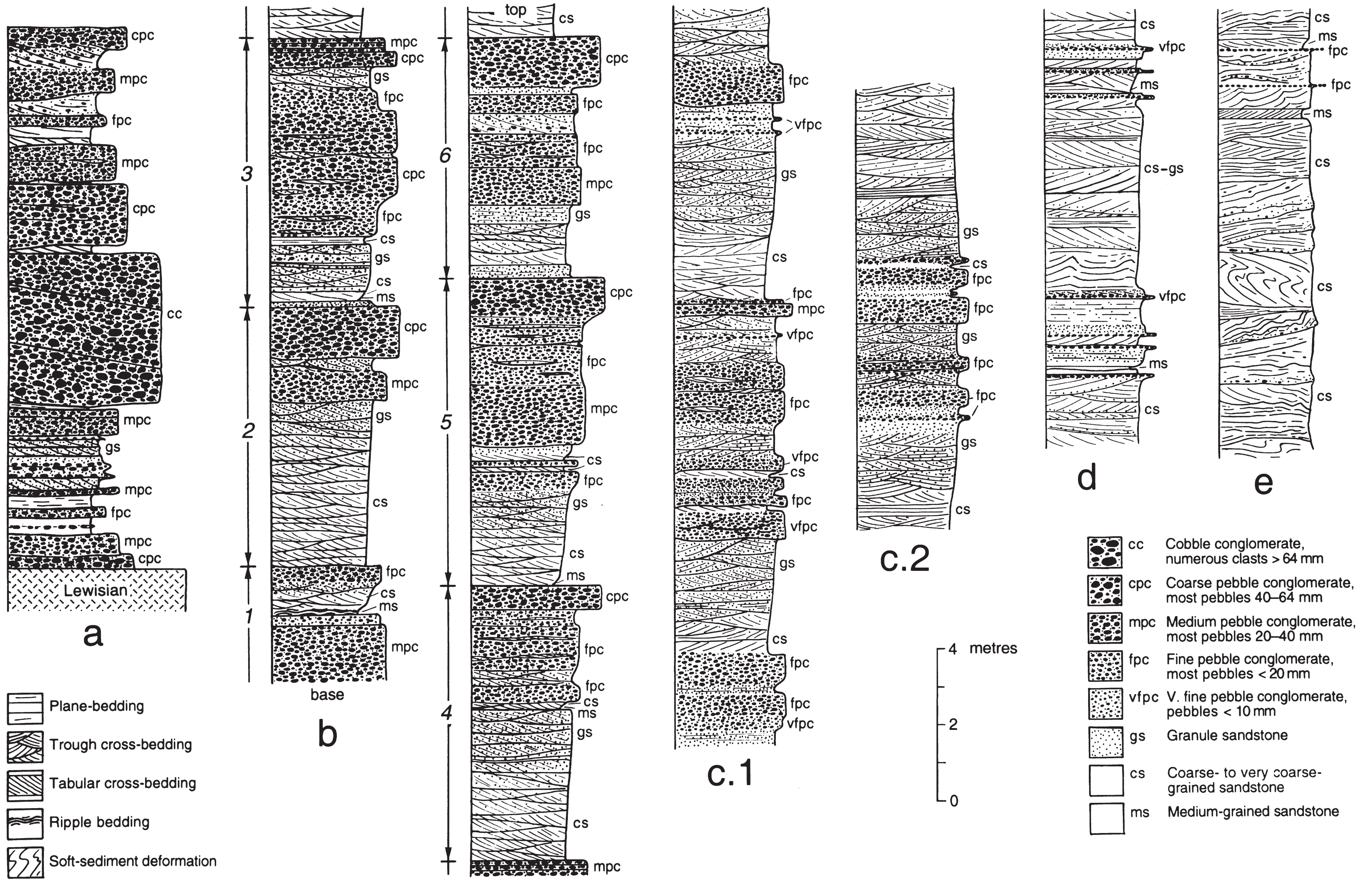

Figure 4. For legend see opposite.

Applecross Formation in the south (e.g. Selley *et al.* 1963; Owen, 1995). The flexed bedding indicates that the sands were partly or wholly liquidized when deformed. The complex structures indicate upward movement of water near the surface, in places the flow being sufficiently strong to disrupt the bedding. The Cape Wrath Member does not show structures that are diagnostic of earthquake shock (e.g. Mohindra & Bagati, 1996), and some other mechanism seems required for the liquidization of much of FA5 (see Section 8.b).

## 6. Composition of the Cape Wrath Member

### 6.a. Petrography of sandstones

The main clastic constituents of the sandstones as determined from modal analyses are as follows:

(1) Quartz and quartzose rock fragments form 53–57 % of specimens from FA1–4 and 72 % from FA5. Quartz occurs as single and composite grains that commonly show undulose extinction.

(2) Feldspar forms 27–32 % of specimens from FA1–4 and 19 % from FA5. Microcline commonly forms > 50 % and plagioclase (albite) forms ≤ 7–11 % of the feldspar suite, with some orthoclase and antiperthite. The microcline is mostly unaltered but other feldspars show variable alteration to mica.

(3) Grains of schist, chert, cataclasite, mylonite, and volcanic rocks form ≤ 6 % of the specimens.

(4) Detrital muscovite and biotite are rare, with the biotite usually containing secondary hematite.

Most specimens from FA1–4 are classifiable as arkose and a few specimens from FA1–4 and all specimens from FA5 as feldspathic sandstone (Williams, 1969*a*). The abundance of feldspar and the angularity of many grains indicate that the sandstones are immature, with the bulk being first cycle. Van de Kamp & Leake (1997) similarly concluded that Applecross sandstones from the Gairloch–Applecross area are first cycle. Upward increase in maturity of the Cape Wrath succession is indicated by the higher quartz/feldspar ratio, greater roundness of grains, and better sorting for sandstones from FA5 compared to those from FA1. Upward increase in quartz/feldspar ratio also is shown by the Applecross succession at Quinag and Coigach (Stewart & Donnellan, 1992).

### 6.b. Geochemistry

Analyses for sandstones from FA1–5 and for mudstone are given in Table 3. The sandstones show the following upward changes, which are greatest between FA3 and FA4: (1) decrease in $Na_2O$ and $Na_2O/K_2O$, and to a lesser degree in CaO, reflecting progressive loss of plagioclase relative to K-feldspar; (2) increase in $K_2O$ and K/Rb. Upward decrease in $Na_2O/K_2O$ also is displayed by the Applecross Formation at Coigach (Stewart & Donnellan, 1992) and north of the Loch Maree Fault (Stewart, 1991), and was ascribed to the progressive disappearance of plagioclase relative to K-feldspar. These upward changes for the Cape Wrath Member and the Applecross Formation further south accord with the petrographic evidence for upward increase in maturity of the sandstones.

Red sandstones assayed 0.94–4.81 % $Fe_2O_3$ (mean = 2.20 %) and red mudstone 9.33 % $Fe_2O_3$, whereas FeO values are low. Most of the Fe is contained in hematitic pigment. In addition to the results for FA5 in Table 3, eight samples of red sandstone assayed 1.3–2.3 % total Fe as $Fe_2O_3$ (mean = 1.85 %) and six samples from associated grey sandstone assayed 0.3–0.5 % total Fe as $Fe_2O_3$ (mean = 0.38 %). Hence much Fe has been removed from the grey beds in the striped interval.

### 6.c. Pebbles and cobbles

Up to 70 % of clasts in the Cape Wrath Member can be matched with rock types in the Lewisian as now exposed (Williams, 1969*b*). Such rock types include quartz, felsic gneiss, granulite, quartz-mica schist, cataclasite, pegmatite, feldspar, and several types of metaquartzite. The member also contains clasts that are unknown in the Lewisian, including quartz-tourmaline rock, tourmaline-aplite, quartz-fuchsite schist, ferruginous metaquartzite, felsic volcanics, chert, jasper and orthoquartzite.

### 6.d. Implications for the source area

Peach *et al.* (1907, p. 273) noted that microcline, ‘in a wonderfully fresh condition’, is the dominant feldspar in the Applecross sandstones. Because microcline is abundant in the mainland Lewisian only between Cape Wrath and Loch Laxford, they concluded that the Applecross Formation as a whole could not have been derived from the bulk of the Lewisian Gneiss as exposed on the mainland.

In the weathering profiles on the Lewisian palaeoplain, microcline is little altered, whereas much of the plagioclase and hornblende was altered to clay minerals, and hematite partly replaced biotite. The presence of pedogenic carbonate and the degree of weathering indicate a temperate to warm, subhumid palaeo-

Figure 4. Measured sections for FA1–5. (a) FA1, lower part of the type section at Droman [NC 185594]. (b) FA2, part of the type section at Sheigra [NC 184614]. Nos 1–6 mark coarsening-upward cycles. (c) FA3. 1, lower part of the type section at Port Beag [NC 190634]. 2, middle part of the section at Creag Riabhach [NC 285630]. (d) FA4, middle part of the type section at Port Mòr [NC 196642]. (e) FA5, part of the type section at An Grianan [NC 265627].

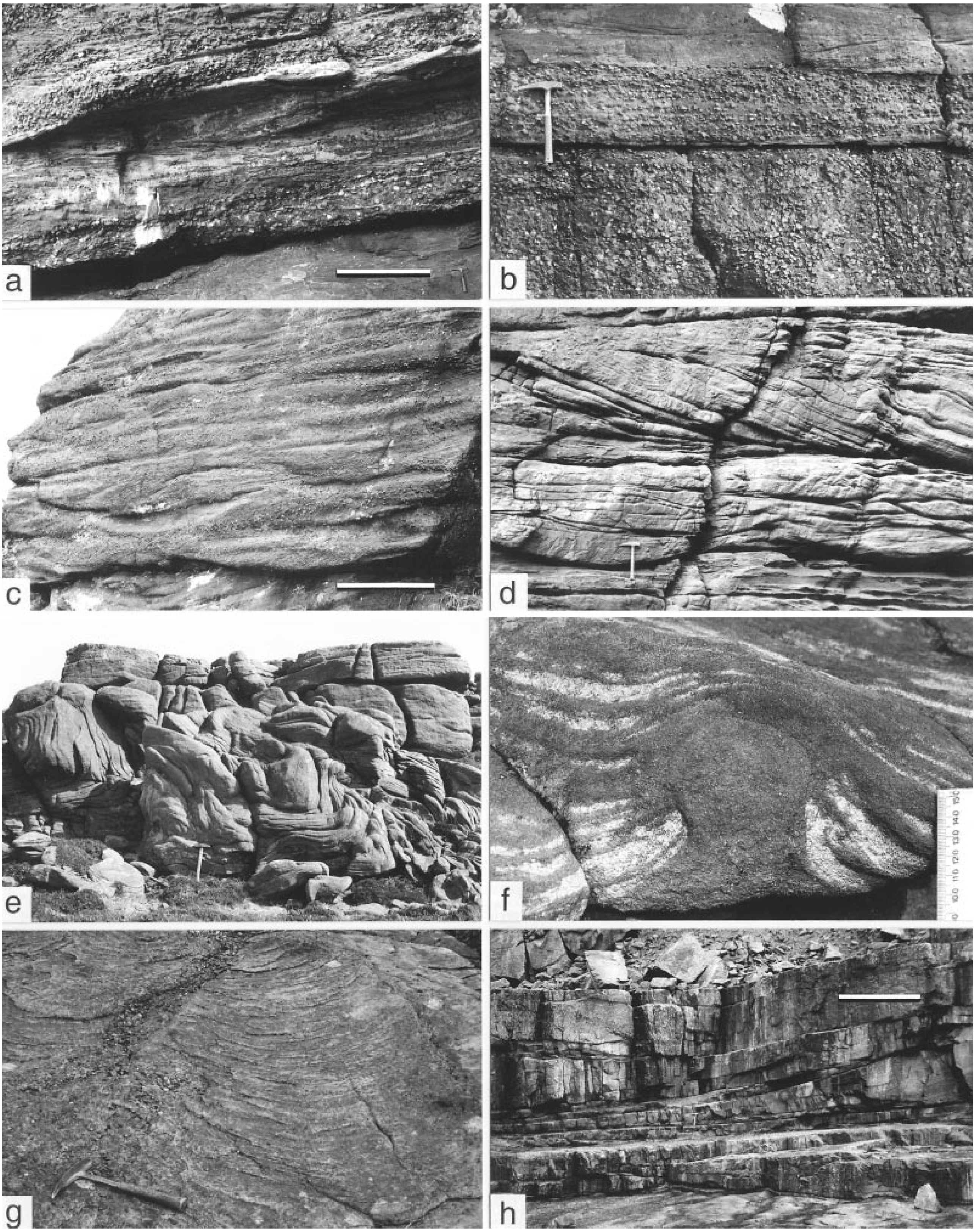

Figure 5. (a) Weathered gneiss overlain by pebble- and cobble-conglomerate and sandstone of FA1, Poll a'Mhurain [NC 185619]. Looking east; scale bar 1 m. (b) Top of a coarsening-upward cycle, FA2 [NC 185613], with tabular pebble-conglomerate overlain by sandstone. Looking southeast; hammer 33 cm long. (c) Trough cross-bedded, pebbly sandstone, FA3 [NC 317583]. Looking south; scale bar 50 cm. (d) Trough and tabular cross-bedded sandstone, FA4 [NC 196642]. Looking west; hammer 33 cm long. (e) Soft-sediment deformation structure with overturning of beds, FA5 [NC 238598]. Looking east; hammer 30 cm long. (f) Diapiric structure in red and grey sandstone, FA5 [NC 315725]. Scale in millimetres. (g) Trough cross-bedding shown on a sandstone bed surface, FA3 [NC 206648]. Looking east–northeast down-current; hammer 33 cm long. (h) Tabular cross-bedded sandstone, FA2 [NC 184614]. Looking southeast; scale bar 1 m.

Table 3. Analyses of sandstones and mudstone from the Cape Wrath Member in the Cape Wrath area

| | (1) | (2) | (3) | (4) | (5) | (6) | (7) |
|---|---|---|---|---|---|---|---|
| *Major elements (wt %)* | | | | | | | |
| $SiO_2$ | 83.21 | 81.21 | 77.60 | 83.91 | 87.26 | 86.48 | 49.39 |
| $TiO_2$ | 0.19 | 0.37 | 0.86 | 0.13 | 0.20 | 0.10 | 1.46 |
| $Al_2O_3$ | 7.90 | 8.42 | 8.32 | 7.78 | 5.78 | 6.93 | 21.08 |
| $Fe_2O_3$ | 1.69 | 2.32 | 4.81 | 0.94 | 1.24 | 0.33 | 9.33 |
| FeO | 0.05 | 0.13 | 0.37 | 0.09 | 0.05 | 0.04 | 0.88 |
| MnO | 0.03 | 0.01 | 0.02 | 0.01 | 0.00 | 0.00 | 0.05 |
| MgO | 0.47 | 0.76 | 0.76 | 0.30 | 0.35 | 0.46 | 4.11 |
| CaO | 0.27 | 0.10 | 0.32 | 0.05 | 0.02 | 0.02 | 0.40 |
| $Na_2O$ | 1.74 | 1.65 | 1.64 | 0.89 | 0.08 | 0.07 | 0.56 |
| $K_2O$ | 3.02 | 3.41 | 3.52 | 4.09 | 3.77 | 4.68 | 7.38 |
| $P_2O_5$ | 0.03 | 0.05 | 0.04 | 0.03 | 0.02 | 0.02 | 0.26 |
| $SO_3$ | 0.03 | 0.00 | 0.02 | 0.01 | 0.00 | 0.00 | 0.01 |
| LOI | 0.95 | 1.07 | 1.37 | 1.02 | 0.82 | 0.75 | 4.68 |
| Total | 99.58 | 99.50 | 99.65 | 99.25 | 99.59 | 99.88 | 99.59 |
| *Trace elements and rare earth elements (ppm)* | | | | | | | |
| Zr | 88.1 | 150.5 | 345.1 | 65.4 | 94.3 | 60.1 | 314.8 |
| Nb | 2.3 | 5.1 | 10.1 | 2.4 | 2.8 | 0.6 | 22.5 |
| Y | 6.8 | 7.3 | 10.0 | 5.4 | 7.1 | 6.9 | 74.7 |
| Sr | 66.6 | 71.8 | 101.3 | 76.3 | 46.2 | 55.0 | 173.5 |
| Rb | 75.6 | 80.6 | 86.9 | 84.2 | 74.7 | 80.1 | 238.8 |
| U | 2.1 | 2.0 | 3.2 | 0.2 | 2.0 | 1.4 | 5.6 |
| Th | 4.2 | 3.8 | 8.1 | 2.4 | 2.9 | 3.1 | 16.9 |
| Pb | 6.1 | 8.2 | 8.9 | 9.4 | 9.9 | 8.0 | 21.6 |
| Ga | 6.7 | 8.0 | 10.2 | 6.3 | 5.2 | 6.4 | 26.3 |
| Cu | 6 | 11 | 7 | 9 | 4 | 5 | 5 |
| Zn | 15 | 20 | 20 | 8 | 7 | 8 | 103 |
| Ni | 8 | 12 | 11 | 5 | 5 | 6 | 68 |
| Ba | 605 | 702 | 840 | 602 | 502 | 553 | 329 |
| Sc | 2.7 | 4.1 | 3.7 | 2.6 | 1.0 | 1.5 | 22.2 |
| Co | 46 | 43 | 50 | 63 | 68 | 82 | 28 |
| V | 32 | 34 | 87 | 16 | 19 | 13 | 80 |
| Ce | 19 | 31 | 36 | 24 | 31 | 51 | 178 |
| Nd | 5 | 10 | 12 | 8 | 10 | 15 | 69 |
| La | 6 | 13 | 15 | 9 | 12 | 20 | 107 |
| Cr | 18 | 30 | 47 | 12 | 12 | 2 | 137 |
| $SiO_2/Al_2O_3$ | 10.5 | 9.6 | 9.3 | 10.8 | 15.1 | 12.5 | 2.35 |
| $Na_2O/K_2O$ | 0.58 | 0.48 | 0.47 | 0.22 | 0.02 | 0.01 | 0.08 |
| $CaO/Na_2O$ | 0.16 | 0.06 | 0.20 | 0.06 | 0.25 | 0.29 | 0.71 |
| K/Rb | 332 | 351 | 336 | 404 | 419 | 485 | 257 |
| Rb/Sr | 1.14 | 1.12 | 0.86 | 1.10 | 1.62 | 1.46 | 1.38 |

(1) Medium- to coarse-grained, pale red sandstone (two samples), FA1, Sheigra [NC 183608]. (2) Medium- to coarse-grained, pale red sandstone (two samples), FA2, Sheigra [NC 184614]. (3) Medium- to coarse-grained, pale red sandstone (two samples), FA3, Port Mòr [NC 195642]. (4) Medium- to coarse-grained, pale red sandstone (three samples), FA4, Port Mòr [NC 195642]. (5) Medium- to coarse-grained, pale red sandstone (four samples), FA5, Beinn a'Chraisg [NC 233594]. (6) Medium- to coarse-grained, grey sandstone, FA5, Beinn a'Chraisg [NC 240595]. (7) Red mudstone, FA2, Sheigra [NC 183614]. Determinations by X-ray fluorescence and redox titration, Department of Geology and Geophysics, Adelaide University.

climate with dry intervals (Williams, 1968*a*; Retallack & Mindszenty, 1994), which is similar to the temperate– humid palaeoclimate inferred from the composition of Applecross sandstones (van de Kamp & Leake, 1997). Williams (1968*a*) concluded that similar weathering of a Lewisian terrain could, in principle, have supplied the microcline-rich sandstones of the Cape Wrath Member. Van de Kamp & Leake (1997) suggested a source terrain comprising mostly sub-granulite facies felsic gneisses, granitic rocks and supracrustal rocks for the Applecross Formation.

The conclusion of Young (1999*b*) that a Laxfordian source is not appropriate for Torridon Group mudstones he studied, based on a mean value of 14.1 ppm Th for the mudstones compared to a mean value of 8.4 ppm Th for Laxfordian gneisses at Rhiconich (Weaver & Tarney, 1981), must be queried. His value of 14.1 ppm Th for the mudstones is typical of Mesoproterozoic and Neoproterozoic fine-grained sediments (13 ± 1 ppm Th: Taylor & McLennan, 1985). Furthermore, the potential input of the widespread Laxfordian granite–migmatite complexes in the Outer Hebrides, which may have contributed to Applecross detritus (Stewart & Donnellan, 1992), should not be overlooked. Th values for samples of Laxfordian granites from 13 localities in Lewis and South and North Harris average 38.9 ppm (range = 21.2–59.0 ppm) (G. E. Williams, unpub. data). Evidence for a Laxfordian contribution to Applecross detritus is discussed in Section 9.c.

It is concluded that much of the Cape Wrath Member could have been derived from the weathering of a terrain consisting mostly of felsic gneisses and granites, with some supracrustal rocks. The degree of weathering shown by the palaeosols in the Cape Wrath area would be appropriate. The immature nature of the sandstones implies that their source area for was not far distant.

## 7. Palaeocurrent analysis of the Cape Wrath Member

Palaeocurrent data comprise the down-current azimuths of individual trough axes exposed on bed surfaces (Fig. 5g) and azimuths obtained from the three-dimensional structure of trough cross-bedding as seen on vertical joints where bed surfaces were poorly exposed. Because of the low dips, structural correction of the data was unnecessary. The dataset of 2667 measurements greatly exceeds previous palaeocurrent datasets for the Cape Wrath Member: Williams (1969*a*) recorded 800 trough azimuths between Cape Wrath and Quinag including 489 for the Cape Wrath area, and Nicholson (1993) made 607 measurements between Sheigra and Quinag including 159 near Sheigra.

Palaeocurrent data were arranged by individual bed surfaces, outcrops of more than one bed surface, and facies assemblages. Typically the data are of low variability and unimodal at all hierarchical levels. Rose diagrams are shown in Figure 6.

### 7.a. Cape Wrath area

A total of 2118 measurements was recorded in 1996–97 for virtually all exposures of FA1–5 in the Cape Wrath area. To ascertain whether lateral changes occur in mean palaeocurrent directions, the data were grouped by an arbitrary, east–west subdivision of the Cape Wrath area into subareas A (south), B and C (north) (Fig. 3). The vector means and vector magnitudes (Potter & Pettijohn, 1977) of 2085 palaeocurrent measurements, grouped by FA1–2, FA3 and FA4–5 in each of the three subareas, are given in Table 4 and rose diagrams are shown in Figure 6a–i.

Features of the data are as follows: (1) mean directions for subarea A and for FA4–5 in all subareas are eastward; (2) mean directions for FA1–2 and FA3 swing to the northeast in moving from subarea A to subarea C, with the swing being greatest for FA1–2; (3) mean directions for subareas B and C change upward from east–northeastward to eastward; (4) vector magnitudes are consistently large, indicating a strong central tendency of the data. This comprehensive palaeocurrent study supports earlier findings obtained with fewer data and using a sampling grid over the Cape Wrath area (Williams, 1969*a*). The consistency of results using two different sampling methods argues strongly for the validity of the identified mean directions and their systematic changes.

The mean palaeocurrent direction of 91° ($n = 159$) obtained by Nicholson (1993) for strata at Sheigra (stratigraphy unspecified, but presumably FA1–2) agrees with the data of Williams (1969*a*) and for subarea A in the present study. However, the single result for Sheigra given by Nicholson (1993) cannot be taken as representative of directions for the entire Cape Wrath succession.

Thirty-three additional trough-axis azimuths deviate by 109° on average (range = 91–127°) from other flow directions given in Table 4, with 72 % occurring in FA1–3. The bimodal nature of datasets with such anomalous directions (Fig. 6) justifies the separate analysis of the anomalous data. The possible cause of the anomalous flow directions is discussed below in Section 8.a.

### 7.b. Handa and Ben Dreavie

Data are given in Table 5 and rose diagrams for Handa shown in Figure 6j–l. The overall mean direction at Handa is 100° ($n = 228$), excluding four anomalous

Table 4. Trough-azimuth palaeocurrent data for the Cape Wrath Member in the Cape Wrath area

| | Facies assemblages 1–2 | | | Facies assemblage 3 | | | Facies assemblages 4–5 | | | Anomalous data | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | $V_{mean}$ (°) | $V_{mag}$ (%) | $n$ | $V_{mean}$ (°) | $V_{mag}$ (%) | $n$ | $V_{mean}$ (°) | $V_{mag}$ (%) | $n$ | $V_{mean}$ (°) | $V_{mag}$ (%) | $n$ |
| Subarea C | 62 | 92.4 | 60 | 71 | 96.2 | 378 | 84 | 94.5 | 146 | 338 | 97.3 | 11 |
| Subarea B | 76 | 96.1 | 105 | 78 | 96.9 | 265 | 92 | 93.0 | 151 | 339 | 99.1 | 8 |
| Subarea A | 87 | 95.3 | 336 | 86 | 98.0 | 433 | 82 | 96.4 | 211 | 328 | 89.0 | 14 |

$V_{mean}$, vector mean; $V_{mag}$, vector magnitude; $n$, number of measurements.

Table 5. Trough-azimuth palaeocurrent data for the Cape Wrath Member south of the Cape Wrath area

| Area | Facies assemblage | $V_{mean}$ (°) | $V_{mag}$ (%) | *n* |
|---|---|---|---|---|
| Handa | FA4 | 83 | 89.1 | 71 |
| | FA3 | 104 | 91.3 | 69 |
| | FA1–2 | 110 | 95.8 | 88 |
| | All data | 100 | 90.5 | 228 |
| Ben Dreavie | FA1 | 110 | 98.7 | 9 |
| Quinag | FA4 | 105 | 93.2 | 63 |
| | FA3 | 125 | 94.1 | 42 |
| | FA2 | 135 | 95.2 | 31 |
| | cf. FA3 | 130 | 96.3 | 139 |
| | All data | 124 | 93.5 | 275 |
| Stoer | cf. FA3–4 | 161 | 89.5 | 21 |

Data do not include rare anomalous directions (see text, Section 7.b,c). Abbreviations as for Table 4.

azimuths of 316–340° for FA2–3. Mean directions show an upward change from east–southeastward for FA1–2 to eastward for FA4, which is accompanied by an upward increase in variability. The mean palaeocurrent direction of 110° (*n* = 325) determined by Nicholson (1993) for Handa agrees broadly with results given here and by Williams (1969*a*), although Nicholson did not relate his data to stratigraphic position. The sparse data for Ben Dreavie indicate mean flow to the east–southeast (Table 5).

### 7.c. Quinag and Stoer

The great majority of troughs at Quinag are directed to the southeast or east–southeast (Table 5 and Figure 6m–o). Mean directions show an upward change from southeastward (130°, *n* = 212) for the lowest three facies assemblages to 105° (*n* = 63) for FA4. There is also an upward increase in variability of the data. The mean palaeocurrent direction of 129° (*n* = 123) obtained by Nicholson (1993) for Quinag is consistent with results given here and by Williams (1969*a*) but Nicholson did not relate his data to stratigraphic position.

Anomalous trough azimuths at Quinag are confined to the lowest facies assemblage (cf. FA3). One anomalously oriented trough (310°) was observed in the south. Troughs directed west–southwest (vector mean = 247°, vector magnitude = 97%, *n* = 11) occur in the north [NC 206302] and may reflect an irregularity in the palaeolandsurface. The anomalous directions are shown in Figure 6 but are treated separately from other directional data for Quinag.

Data for beds assigned to the Cape Wrath Member at Stoer are sparse, with the mean direction for 21 troughs being south–southeastward (Table 5). Strata that evidently are stratigraphically above the Cape Wrath Member exhibit eastward to northeastward flow (Williams, 1969*a*). Nicholson (1993) also found an eastward mean flow direction for the upper Applecross Formation at Stoer.

### 7.d. Palaeocurrent pattern for the Cape Wrath Member

Figures 6 and 7 show the systematic upward changes in mean palaeocurrent directions for subareas B and C and at Handa and Quinag; for subareas B and C the upward change is clockwise, whereas for Handa and Quinag the upward change is anticlockwise. No important upward change is identified for subarea A, where the switch in the sense of upward change occurs.

Systematic lateral changes in mean directions for FA1–2 and FA3 also are shown by Figures 6 and 7. To better show these changes, pooled data for FA1–3 are plotted in Figure 8, using mean directions of 70° (*n* = 438, vector magnitude = 95.6%) for subarea C, 77° (370, 96.6%) for subarea B, 86° (769, 96.8%) for subarea A, 107° (157, 93.7%) for Handa, and 130° (212, 95.6%) for Quinag. Mean directions for FA1–3 (for $n > 150$) projected upcurrent from the centre of respective sampling areas indicate a divergent palaeocurrent pattern with a spread of 60°. By contrast, palaeocurrent directions for FA4–5 show no systematic lateral change. Hence the upward changes in mean palaeocurrent directions result in the disappearance of the divergent flow pattern for FA1–3 and the establishment of eastward flow for FA4–5.

Maximum clast size in conglomerates at the base of the Cape Wrath Member declines eastward in the Cape Wrath area and southeastward to Quinag (Fig. 8), consistent with the palaeocurrent data.

## 8. Deposition of the Cape Wrath Member

### 8.a. Alluvial fan deposition

FA1–3 display the following features of alluvial fan deposits (e.g. Heward, 1978; Nilsen, 1982; Brierley, Liu & Crook, 1993; Blair & McPherson, 1994): (1) a fan pattern indicated by palaeocurrent data of low variability; (2) coarse-grained, poorly sorted deposits; (3) tabular, laterally-extensive conglomerates with sand-grade matrix, ascribable to sheetflood deposition; (4) cross-bedded channel sandstones; (5) coarsening-upward cycles; (6) down-current decrease in clast size at the base of the succession; (7) immature, largely first-cycle sandstones.

The upcurrent projection of mean flow directions for FA1–3 shows that the fan pattern has a radius of *c.* 50 km and its apex 30 km to the west (Fig. 8). Such projection of flow directions back to a fan apex is justified by the low variability of the large palaeocurrent dataset, and has been demonstrated for mean flow directions in other ancient fluvial distributary systems (Hirst & Nichols, 1986; Jupp *et al.* 1987). The fan

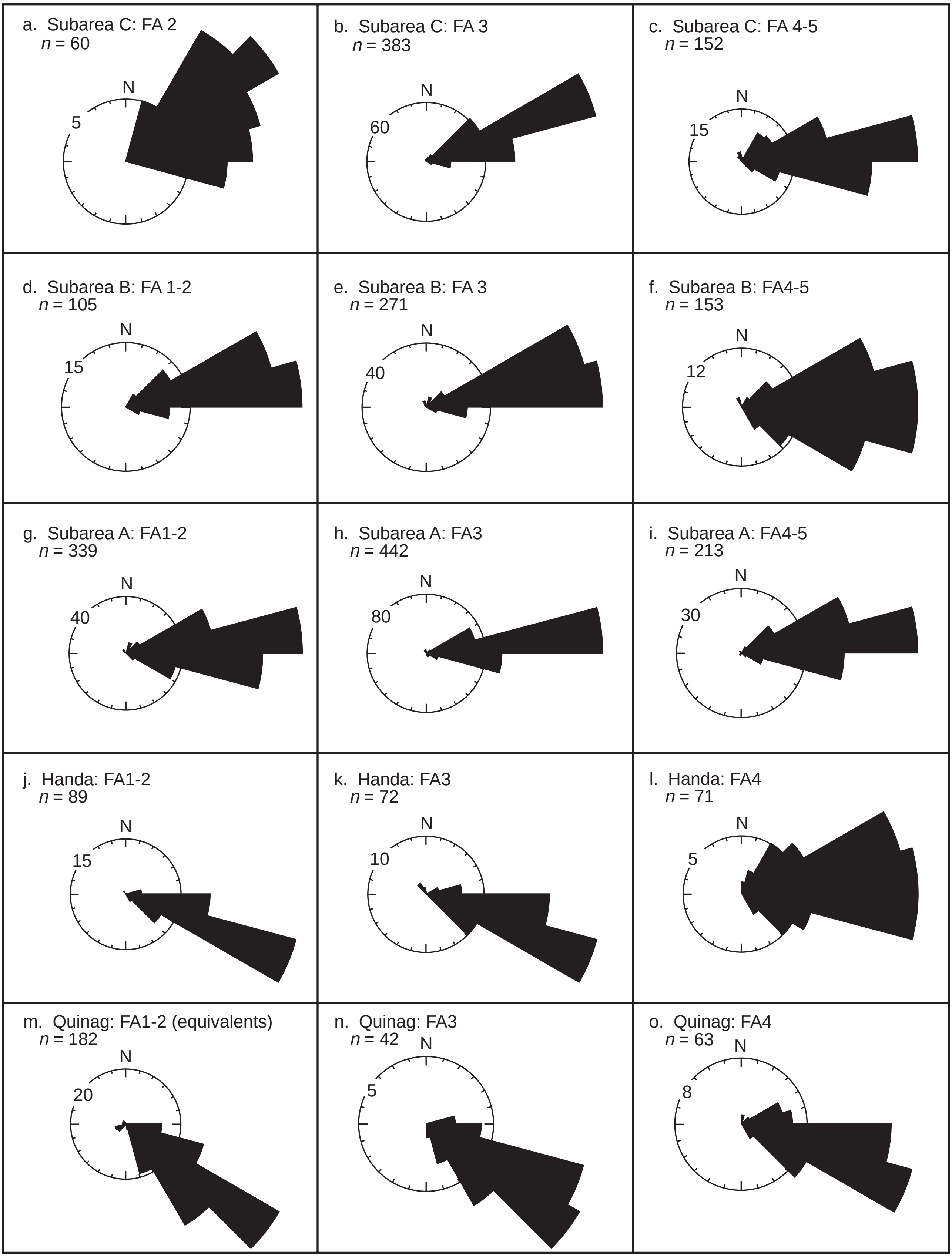

Figure 6. Rose diagrams of palaeocurrent directions for FA1–2, FA3 and FA4–5 of the Cape Wrath Member for subareas A, B and C in the Cape Wrath area, and for equivalent facies assemblages at Handa and Quinag. Rare anomalous directions are included (see Section 7). Class interval = 15°, *n* = number of measurements.

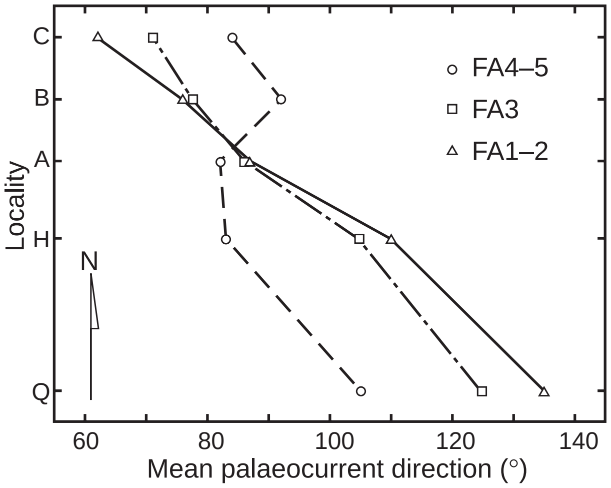

Figure 7. Plot of mean palaeocurrent directions for FA1–2, FA3 and FA4–5 of the Cape Wrath Member (Tables 4 and 5). A, B, C, subareas of the Cape Wrath area; H, Handa; Q, Quinag. The plot shows the upward change in mean directions for subareas B and C and the upward change at Handa and Quinag in the opposite sense, with the switch in sense occurring near subarea A.

median has a bearing of *c*. 100° and is near subarea A where mean flow directions are consistently eastward. The fan area (*c*. $1.3\times10^3$ km$^2$) falls within the range of modern 'fluvial megafans' ($10^3$–$10^5$ km$^2$: Gohain & Parkash, 1990; DeCelles & Cavazza, 1999), but the approach of Miall (1996, p. 245) is followed and the term 'alluvial fan' used here.

Importantly, the presence of coarsening-upward cycles in FA2 argues strongly for an alluvial fan setting. Coarsening-upward cycles are common in alluvial fan deposits (Steel *et al.* 1977; Steel & Aasheim, 1978; Heward, 1978) and range from cycles of a few metres to megasequences hundreds of metres thick. Such cyclicity indicates the repeated progradation of fan lobes mainly in the mid-fan area, punctuated by their abrupt abandonment, and has been ascribed to climate change (Garner, 1979), autogenic cyclicity generated by intrabasinal sedimentary processes (I. A. Nyambe, unpub. Ph.D. thesis, Univ. Ottawa, 1993), and repeated tectonic uplift of the source area (Steel *et al.* 1977; Steel & Aasheim, 1978). The scale of the cyclicity may indicate which mechanism was dominant.

The coarsening-upward cycles of FA2 indicate the aggradation of fan lobes in mid-fan areas, first by cross-bedded channel sands then by tabular gravels spread by sheetfloods, followed by fan-lobe abandonment. The lack of obvious pedogenesis at the top of cycles indicates that the cycles were of relatively short duration, suggesting autogenic processes. The cycles are classifiable as 'parasequences' or 'facies successions' (Miall, 1997, p. 330) representing *c*. $10^3$–$10^4$ years. The symmetrical cycles of FA3 mark distal fan areas (Heward, 1978; Steel & Aasheim, 1978) and record the alternation of channel sands and thinner sheetflood gravels without abandonment of the fan lobe.

Williams (1969*a*) interpreted the Cape Wrath Member as a diachronous succession whose upward changes reflect downstream changes at the alluvial surface. Accordingly, FA1–3 record the change from channel gravels in proximal to mid-fan areas to the domain of sheetfloods and channel sands in the mid-fan to distal fan areas (Table 1). The cross-bedded sandstone at the base of the member at Quinag may have accumulated before a large catchment developed. The upward changes in mean palaeocurrent directions for FA1–3 (Figs 6, 7) are consistent with the pattern expected for an onlapping, diachronous alluvial fan succession.

As observed above (Section 5.b), the dip directions of tabular cross-bedding in FA1–3 for the Cape Wrath area indicate a preferential northward accretion of channel bars, suggesting a northward migration of channels across the fan (Williams, 1966*b*). Viseras & Fernandez (1994) confirmed that channel displacement in a constant direction across an alluvial fan can occur by the preferential accumulation of bars on one of the channel banks. Furthermore, since 1731 the active river channel on the Kosi alluvial fan in the Ganges Valley has migrated westward across the fan through avulsive steps and gradual filling and shifting of the channel (Gohain & Parkash, 1990).

The anomalous northwestward flow directions in the Cape Wrath area (Table 4) occur mostly in FA1–3 and may be related to the preferential northward accretion of channel bars on the alluvial fan. Locally reversed current directions and upstream-dipping cross-bedding can result from flow separation on point bars at sharp bends in rivers (Taylor, Crook & Woodyer, 1971; Hiller & Stavrakis, 1982), and the anomalous flow directions in FA1–3 may record such flow separation and the local northward deflection of flow around north-facing channel bars.

### 8.b. Braidplain deposition

FA4–5 display features indicating a less energetic, more distal fluvial setting, as follows: (1) tabular cross-bedding and trough cross-bedding are equally common; (2) the palaeocurrent data show no regional fan pattern and in some areas display a greater variability than do data for FA1–3; (3) conglomerate beds are thin and have only small pebbles; (4) thin mudstone beds are rare. These features agree with braided river deposits (South Saskatchewan and Platte types: Miall, 1977, 1978).

Hence, FA4–5 were deposited on a braidplain beyond the alluvial fan (Table 1) while maintaining eastward drainage. This setting accords with the greater maturity of sandstones from FA4–5. Comparable downstream passage from alluvial fans to

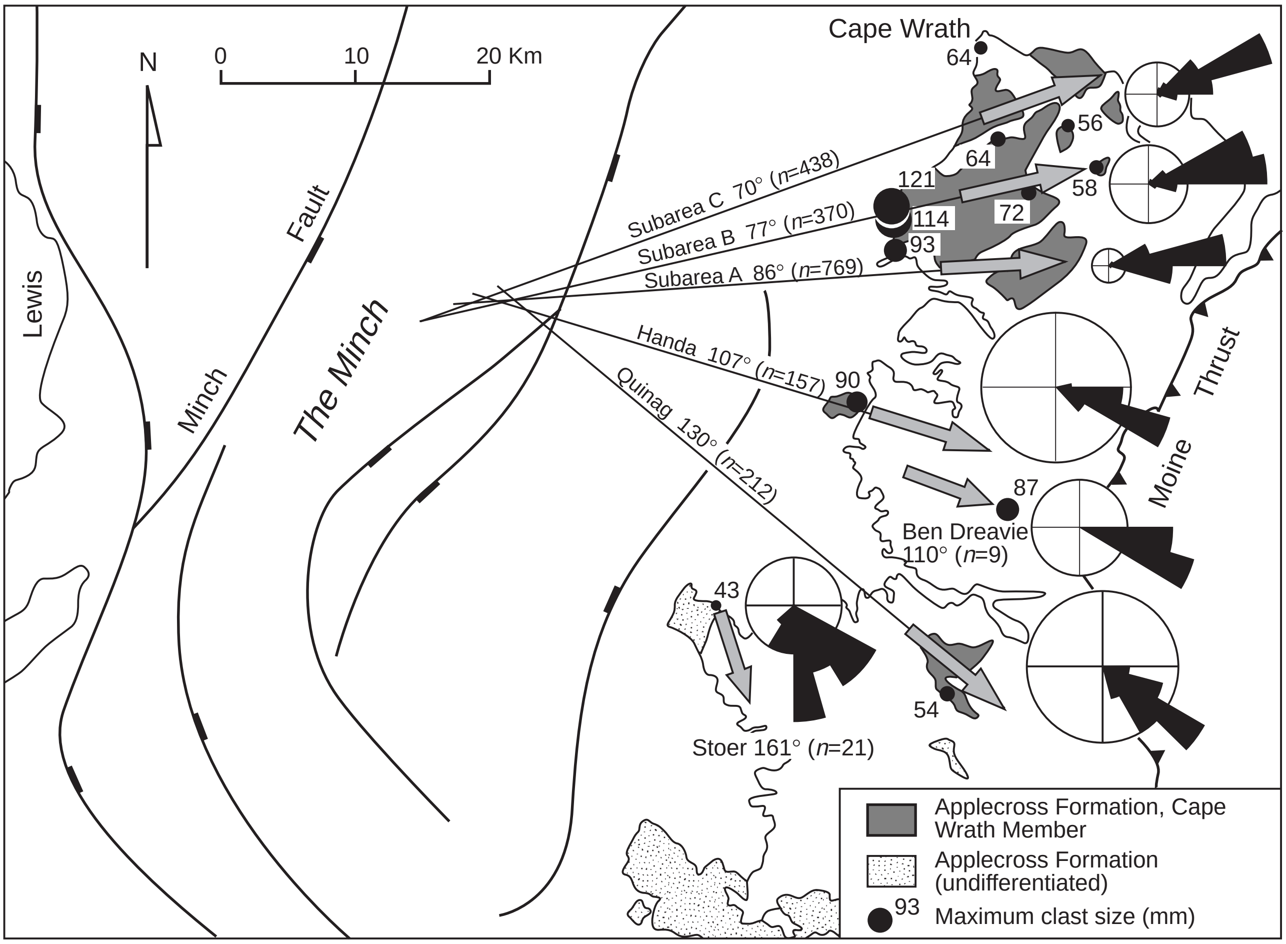

Figure 8. Vector mean palaeocurrent directions (arrows) for FA1–3 in subareas A, B and C of the Cape Wrath area and for equivalent facies assemblages at Handa, Ben Dreavie, Stoer and Quinag (see Section 7.d for details). Rose diagrams are for FA1–3; circles represent 50 measurements except for Ben Dreavie and Stoer (two measurements). Mean directions (for $n > 150$) projected upcurrent from the centre of respective sampling areas intersect near a major basement (pre-Caledonian) normal fault with downthrow to the east that runs north–northeast beneath the north Minch Basin (basement faults shown from Stein, 1988; Stein & Blundell, 1990). Maximum clast size for conglomerate beds at or near the base of the Cape Wrath Member is taken as the mean maximum dimension in millimetres of the five largest clasts within an area of 50 × 50 cm on bed surfaces.

braidplain occurs across the 40-km-wide Aurès piedmont plain in Algeria (Williams, 1970*b*).

On modern alluvial fans, water infiltrates the coarse alluvium of fanheads and commonly emerges in lower fan areas (Rachocki, 1981). Hence alluvial piedmont plains can comprise an upper, well-drained tract and a lower marshy tract with high water-table, exemplified by the Aurès piedmont plain in Algeria (Williams, 1970*b*). The ubiquity of soft-sediment deformation structures in FA5 and the southern Applecross Formation indicates deposition in such poorly drained, distal alluvial tracts. Saturated sands at and near the surface could have been liquidized by high pore pressure and rendered susceptible to soft-sediment deformation. Complex soft-sediment deformation structures may record upwelling groundwater (Williams, 1970*a*; Owen, 1995). The grey streaks and patches at the transition from FA4 to FA5 may record the reduction and removal of much ferric oxide pigment within a near-surface zone of fluctuating water table.

The upward decrease in $Na_2O/K_2O$ for sandstones from the Cape Wrath Member, ascribed to severe loss of plagioclase relative to K-feldspar (see Section 6.b above), may have occurred mainly through weathering during alluvial storage with transport across the sedimentary basin. The relatively large decrease in $Na_2O/K_2O$ and CaO between FA3 and FA4 suggests increase in such weathering with the passage from a well-drained alluvial-fan setting to a poorly drained, braidplain setting.

### 8.c. Relationship between weathered palaeoplain and alluviation

A sedimentary model for the Cape Wrath Member should account for the following: (1) the Lewisian palaeoplain in the Cape Wrath area with its thin palaeosols that formed penecontemporaneously with Applecross deposition; (2) the eastward slope of the palaeoplain, indicated by the eastward flow directions for the immediately overlying sediments; (3) the nature of the Cape Wrath Member, with its upward change from alluvial fan to more distal, braidplain deposits; (4) evidence that the member could have been derived largely from a weathered terrain that included felsic gneisses and granitic rocks such as occur in the Lewisian; (5) the greater thickness of FA4 at Quinag and the presence there of an unweathered, mountainous Lewisian palaeolandsurface.

The Lewisian palaeoplain shows no evidence for the former presence of a thick regolith like the deep (120+ m) weathering on granite in peneplaned regions (Ollier, 1965). However, analogues of the palaeoplain may occur in mountainous, now semi-arid regions, where uplands commonly are bordered by piedmont plains whose upper parts comprise an eroded bedrock surface of low relief, or pediment, that slopes gently away from the uplands (Tuan, 1959; Oberlander, 1974, 1997). The basinward part of a pediment that is buried by piedmont alluvium is termed a suballuvial bench. Pediplanation has been ascribed to several processes including backwearing of the mountain front, sheetwash or sheetflood erosion, and lateral planation. Granitic pediments typically are moderately weathered and a regime of chemical weathering seems required for active pediplanation on granitic or crystalline bedrock. Indeed, Oberlander (1974) argued that the granitic pediment landforms of the Mojave Desert in California are inherited features that formed through rapid chemical breakdown of granitic rock and backwearing under a more humid climate during the Tertiary.

The above features of the Lewisian palaeoplain and Cape Wrath Member may be integrated by the model of Lawson (1915), by which a pediment forming at the foot of a backwearing mountain front or scarp is buried by onlapping piedmont alluvium (Fig. 9a). Lawson concluded that the succession above the suballuvial bench would fine upward, reflecting downstream changes at the alluvial surface, and so would be diachronous. Hence, the Lewisian weathered palaeoplain may be viewed as an eastward-sloping suballuvial bench that formed at the foot of a westward-retreating scarp and was buried by the onlapping Cape Wrath Member. The progressive upward changes in the member would reflect backwearing of the scarp and probably also reduction in relief of the source area. The upward decrease in divergence of palaeocurrent directions for FA1–3 implies retreat of the scarp and/or widening of the fan bay, mechanisms that could cause pediplanation by backwearing and lateral planation. Quinag lay basinward of the pediplaned area. This scenario requires extensive scarp retreat to produce a suballuvial bench for a fan whose apex was 30 km to the west.

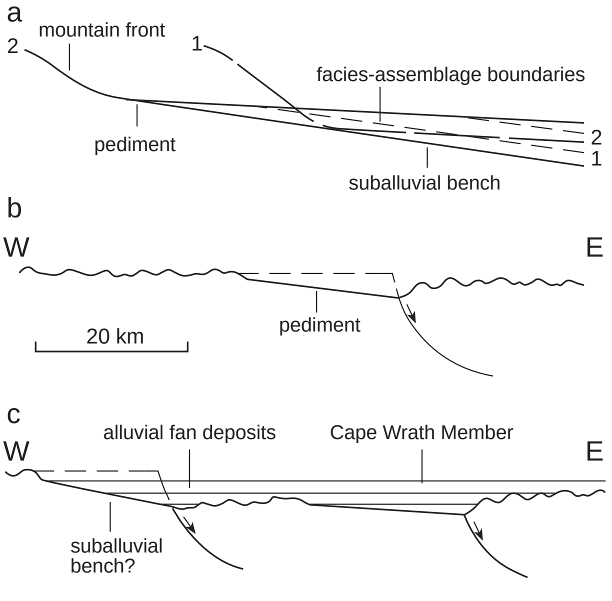

Figure 9. (a) Schematic cross-section of a piedmont zone showing the postulated relationships among retreating mountain front or scarp, weathering pediment, suballuvial bench, and onlapping alluvial fan deposits. Facies assemblages are diachronous and their boundaries are parallel to the suballuvial bench. Based on Lawson (1915). (b) Schematic cross-section showing an eastward-sloping pediment that may have formed in the Cape Wrath area by scarp retreat during an initial cycle of erosion following uplift of the area along a fault to the east. (c) Schematic cross-section showing how a pediment in the Cape Wrath area may have been buried by alluvial fan deposits during a second cycle of erosion following uplift along a fault to the west. A suballuvial bench may have formed through retreat of the western fault scarp. The horizontal scale is for (b) and (c), vertical exaggeration *c.* × 8.

Alternatively, the Lewisian palaeoplain may have formed during an initial cycle of erosion, with the Cape Wrath area being uplifted by faulting and an eastward-sloping pediment forming by westward retreat of the fault scarp (Fig. 9b). Northeast-trending normal faults occur immediately east of the area (Mendum, Evans & Hitchen, 1989). An initial alluvial cover may have been stripped from the pediment, possibly by uplift as proposed for the exhumed 600 km$^2$ Tortilla–Black Mountain granitic pediment in Arizona (Tuan, 1959). Later uplift along a fault further to the west may have led to a second cycle of erosion and the burial of the pediment by the Cape Wrath Member (Fig. 9c).

## 9. Applecross basin development

### 9.a. Palaeogeography of the Applecross Formation

Alluvial fan deposits can provide high-quality information on the former existence, proximity and history of basin-margin faults, basin orientation, and the location of the source area (Blair & McPherson, 1994). The recognition of alluvial fan deposits in the Cape Wrath Member therefore provides a key to interpreting the architecture and tectonic development of the Applecross basin.

Williams (1966*a*, 1969*a*) suggested that the Applecross Formation was deposited on at least two large alluvial fans, including a northern fan between Cape Wrath and Quinag and one or more fans to the south. Differences in pebble suite for the Cape Wrath Member and the Applecross Formation further south (Williams, 1969*b*) suggested different depocentres, but Williams (1969*a*) stressed that the idea of a large southern fan may be oversimplified because of the paucity of stratigraphic and palaeocurrent data. Later studies that included the acquisition of more palaeocurrent data permit modification of this scenario. The present study confirms the former existence of a northern alluvial fan. However, Nicholson (1993) showed that the Applecross Formation south of Quinag is a braidplain succession draining to the east–southeast. Hence the Applecross Formation comprises two principal alluvial components that were at least partly coeval: (1) alluvial fan deposits, best represented by FA1–3 which record overall eastward drainage; (2) more distal, braidplain deposits comprising FA4–5 and much of the Applecross Formation south of Quinag, which maintained eastward to southeastward flow.

The reconstruction of Nicholson (1993) showing all Applecross environments from Rum to Cape Wrath as distal-fluvial cannot be sustained. Apart from palaeocurrent measurements at Sheigra, Nicholson ignored the Cape Wrath area with its weathered palaeoplain and the sedimentological evidence for an alluvial fan origin for FA1–3, and hence his Applecross palaeogeography is incomplete and misleading. Moreover, the palaeohydrological reconstruction of Applecross rivers by Nicholson (1993), in which palaeorivers estimated as 2–7 m deep and 200–1400 m wide were believed to be up to 500+ km long and > 300 km from the source area, is invalidated by observations of rivers in sparsely-vegetated regions that most closely resemble the Precambrian landscape. Sand-bed rivers in central Australia attain flood depths of > 2 m and widths of several hundred metres and deposit cross-bedded sets like those of the Applecross Formation in reaches as close as 25–35 km to the headwaters (Williams, 1968*b*, 1971). Rivers crossing alluvial fans in northern Algeria attain flood depths of 2 m and widths of 800 m, yet the catchment in the Aurès Mountains is only a few kilometres distant (Williams, 1970*b*). Topography and intensity and amount of rainfall are major determinants of flood discharge and channel cross-section in such regions.

### 9.b. The Applecross extensional basin

The Applecross Formation accumulated in a region dominated by uplift, crustal extension and syndepositional faulting during early Neoproterozoic time, for the following reasons:

(1) The presence of an irregular, unweathered Lewisian palaeolandsurface from Quinag southward, with relief as great as 600 m, can be explained by kilometre-scale regional uplift resulting in deep erosion and the development of a youthful topography on the Lewisian prior to Torridon Group deposition. The Sleat Group, which occurs in a restricted rift basin conformably below the Torridon Group (Stewart, 1991), may have formed during crustal extension accompanying the regional uplift.

(2) The alluvial fan deposits and fan flow pattern of FA1–3 indicate proximity to the northwestern margin of the basin. Importantly, Figures 8 and 10a show that the fan apex plots near a major, basement (pre-Caledonian) normal fault with downthrow to the east that runs north–northeast beneath the north Minch Basin (see Stein, 1988; Stein & Blundell, 1990). This fault may mark the margin of the Applecross basin during deposition of the Cape Wrath Member.

(3) As the Cape Wrath Member underlies Applecross strata further south, faulting at the margin of the Applecross basin evidently occurred first in the north (Fig. 10a). Laubach & Marshak (1987) and Stein (1988) observed that normal faults, including a dominant northeast–southwest set, are common in the Laxfordian basement of the Cape Wrath area but are absent in the Scourian to the south. They concluded that basement structure influenced the preferential development of normal faults in the Laxfordian of the Cape Wrath area and inhibited their southward propagation into the Scourian. The nature of the Lewisian thus may have led to early normal faulting and pediplanation at the northern margin of the Applecross basin.

(4) Following deposition of the Cape Wrath Member (Fig. 10a), uplift of the source area for the main body of the Applecross Formation to the south occurred along a second normal fault located further to the west (Fig. 10b). This fault evidently ran as far north as the Cape Wrath area, because the Minch Basin is floored by a Proterozoic section > 6 km thick (Stein, 1992), much of which may belong to the Torridon Group. The development of a second basin-bounding normal fault cutting back toward the undeformed footwall agrees with the sequence of listric faulting at extensional basin margins (Gibbs, 1984). The line of principal uplift was likely controlled by the Outer Hebrides Fault, which probably originated at *c.* 2.5 Ga as an eastward-inclined ramp and was reactivated as an extensional fault along the line of the Minch Fault in Torridonian times (Stewart, 1982; Stein, 1988; Lailey, Stein & Reston, 1989; Stein & Blundell, 1990). Pediments and alluvial fan deposits may have formed in the piedmont zone of the fault scarp in an area now beneath the Minch, while sedimentary filling of the Applecross basin and subsidence

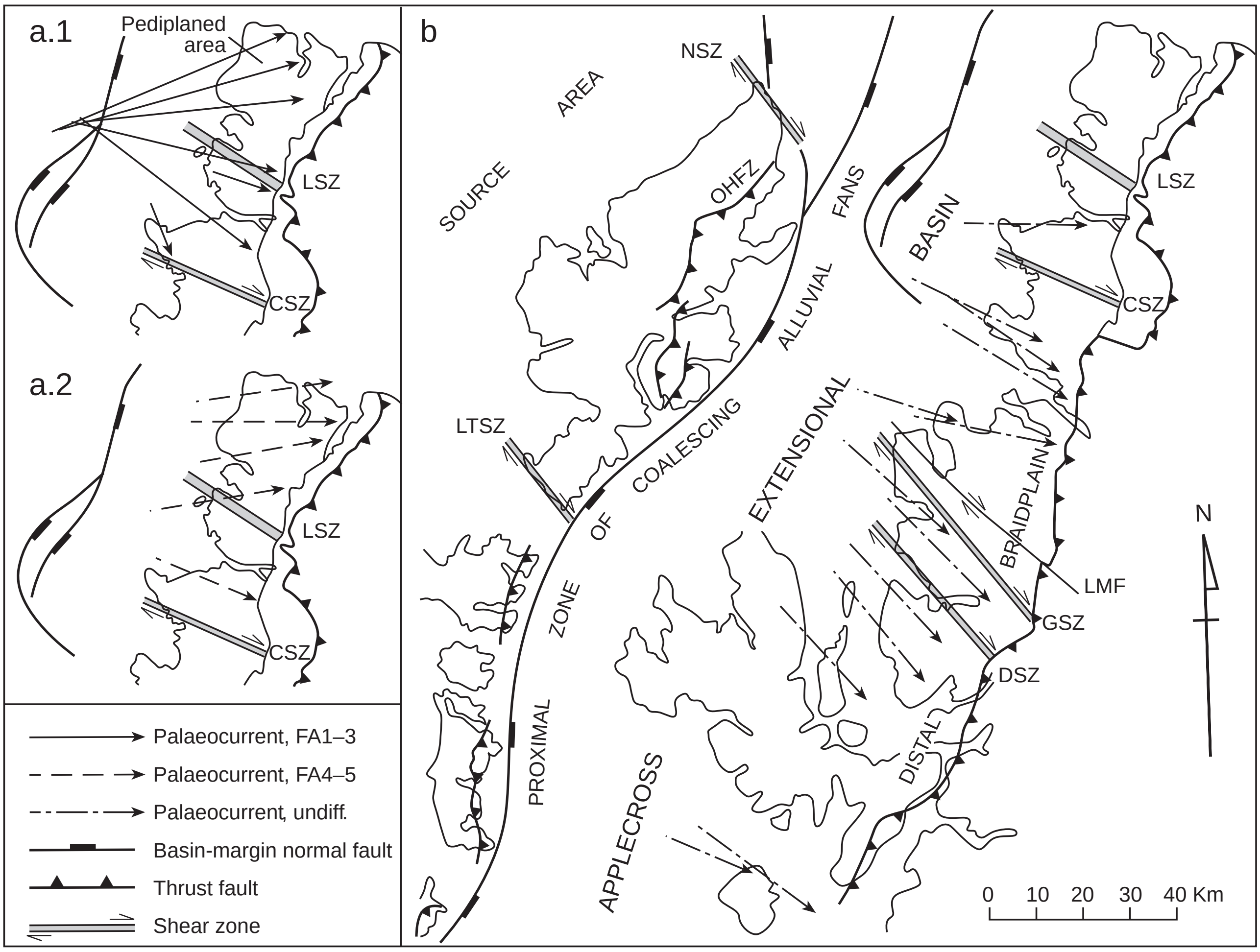

Figure 10. Proposed two main stages in the development of the Applecross extensional basin and deposition of the Applecross Formation. (a.1) Deposition of FA1–3 of the Cape Wrath Member on an alluvial fan to the east of a normal fault at the northern basin margin; the fault shown is a major basement (pre-Caledonian) fault beneath the north Minch Basin (see Stein, 1988; Stein & Blundell, 1990). The Lewisian pediplaned landsurface buried by the fan deposits formed during this, or an immediately preceding, cycle of erosion. (a.2) Deposition of FA4–5 of the Cape Wrath Member in a more distal, braidplain setting. (b) Deposition of the Applecross Formation stratigraphically above and south of the Cape Wrath Member and east of the main basin-margin normal fault along the line of the Minch Fault. The exposed Applecross Formation accumulated on a braidplain draining to the southeast (flow directions from Nicholson, 1993), with alluvial fan deposits postulated to occur in a proximal zone now beneath the Minch. Lewisian shear zones are from Coward & Park (1987): LSZ, Laxford Shear Zone; CSZ, Canisp Shear Zone; GSZ, Gairloch Shear Zone; DSZ, Diabaig Shear Zone; NSZ, Ness Shear Zone; LTSZ, Langavat Shear Zone; OHFZ, Outer Hebrides Fault Zone. LMF, Loch Maree Fault.

caused the burial of the youthful topography on the Lewisian in more distal settings. This proposed arrangement and sequence of faulting at the basin margin explains the distribution of Applecross environments and palaeotopography, from alluvial fan above a weathered pediment in the Cape Wrath area to distal braidplain burying a mountainous landsurface to the south.

(5) Extensional basins typically are segmented by transverse strike-slip transfer or compartmental faults (Gibbs, 1984, 1987) that can be important in controlling sediment movement. Northwest–southeast Lewisian shear zones within the Applecross basin (Fig. 10b) have histories of strike-slip movement (Coward & Park, 1987) and were active at *c.* 1.1 Ga or later (Moorbath & Park, 1971; Park, 1991), and Blundell, Reston & Stein (1989) suggested they were reactivated as transfer faults which compartmentalized the Proterozoic and Phanerozoic extensional basins of northwest Scotland. South of the Canisp Shear Zone most Applecross flow directions are parallel to the nearest shear zone (Fig. 10b), suggesting that the shear zones acted as transfer faults during Applecross deposition and influenced flow directions in distal settings. The northwest–southeast Loch Maree Fault, which had a dextral displacement of 12 km in Neoproterozoic times and was active during Torridon Group deposition (Stewart, 1991), also may have acted as a transfer fault within the Applecross basin.

The proposal of Nicholson (1993) that all the Applecross Formation is of distal braidplain origin and was deposited in a thermal subsidence basin showing no evidence of active basin-bounding faulted margins or syndepositional tectonism, with a source area >300 km distant, is refuted by the following: (1) the presence of alluvial fan deposits in the Cape Wrath Member; (2) palaeocurrent data indicating that alluvial fan deposition was controlled by a nearby basement fault to the west; (3) the variation in the Lewisian palaeolandsurface, from weathered pediment near the basin margin to mountainous in distal settings; (4) the immature character of Applecross sandstones, indicating a relatively nearby source; (5) the occurrence of syndepositional faults within the Applecross basin.

The sequence of extensional tectonic events prior to and during Applecross deposition as proposed here may mark the upwelling of a mantle plume (see Hill, 1991) and is comparable to the main stages in the evolution of the Neoproterozoic basin in western Baltoscandia that Kumpulainen & Nystuen (1985) ascribed to uplift and crustal extension above a rising mantle plume. Although Torridonian volcanism is known only from the Stoer Group (Lawson, 1972), much of the Applecross basin including its faulted margins is not exposed (Fig. 10) and further volcanism cannot be ruled out. Interestingly, the indicated WNW–ESE extension during Applecross deposition accords with the palaeostress regime at 1.2 Ga during deposition of the Stoer Group (Beacom, Anderson & Holdsworth, 1999).

Applecross deposition at 980 Ma followed the Grenvillian orogeny at 1.3–1.0 Ga, but the extensional character of the Applecross basin indicates that the events were genetically unrelated. The Applecross Formation may mark a phase of crustal extension that led ultimately to the birth of Iapetus.

### 9.c. Provenance of the Applecross Formation

Geochronological studies of detritus from the Applecross Formation (Moorbath *et al.* 1967; Allen, Sutton & Watson, 1974; Allen, 1991; Rainbird, Hamilton & Young, 1998, 2001) indicate four main age groups: *c.* 3.09–3.04, 2.9–2.5, 1.8–1.55 and 1.2–1.03 Ga. The latter three groups agree with Scourian, Laxfordian and Grenvillian ages as recorded in the Lewisian (see Section 2 above).

Rb–Sr ages of *c.* 1.55 Ga for detrital microcline from the Applecross Formation (Moorbath *et al.* 1967) indicate that granites or pegmatites of late Laxfordian age occurred in the source area. Pebbles of porphyry and metasedimentary schist gave K–Ar ages of 1.8–1.67 Ga and evidently were derived from now-vanished Laxfordian supracrustal rocks. The absence of Laxfordian supracrustal rocks in Scotland is to be expected, given the major uplift experienced by the Outer Hebrides since the early Neoproterozoic (Steel & Wilson, 1975; Stewart, 1991). Stewart & Donnellan (1992) concluded that Lewisian rocks like those now exposed in the Outer Hebrides and related but now-vanished supracrustal rocks could have provided suitable detritus for the bulk of the Applecross Formation.

U–Pb dating of detrital zircons from Applecross and Aultbea sandstones by Rainbird, Hamilton & Young (1998, 2001) identified weak clusters of ages at 3.09–3.04 and 2.86–2.65 Ga, dominant modes at 1.80 and 1.66 Ga and a lesser mode at 1.10 Ga. They thought the three latter modes indicate Ketilidian, Labradorian and Grenvillian sources, respectively, in eastern Laurentia. However, these ages also agree with main Laxfordian, late Laxfordian and Grenvillian events recorded in the Lewisian. Indeed, the Applecross–Aultbea zircon ages accord with geochronological data from most continents that define terrestrial rock-forming events at *c.* 2.9–2.6, 1.9–1.6 and 1.2–0.9 Ga (see Moorbath, 1977).

Stone *et al.* (1999) argued that major differences in composition between Applecross sandstones and the Lewisian Gneiss, including higher K and lower Ca, Sr and basic elements in the sandstones, indicate a distant source, possibly the Laurentian basement of Labrador. However, these differences may be ascribed to weathering of a Lewisian complex then at a higher crustal level with now-vanished supracrustal rocks, and further alteration of derived material during transport. Hence, detrital geochronology and geochemistry by themselves do not locate unambiguously the Applecross source area; additional arguments must be considered.

The marked upward loss of detrital plagioclase relative to K-feldspar displayed by the Cape Wrath Member (see Sections 6.b and 8.b above) is consistent with a nearby basin margin. Assuming this loss occurred mainly during transport across the sedimentary basin from alluvial fan to braidplain settings, such vulnerability of plagioclase during transport implies that at least the feldspar component of sands reaching the Cape Wrath fan was derived from a relatively nearby source.

One method of estimating the extent of the Applecross source area employs the non-linear relation between alluvial fan area and catchment area

$$F = 0.72A^{0.82} \qquad (1)$$

where $F$ and $A$ are fan area and catchment area, respectively, measured in square kilometres (determined from data for 68 Spanish fans up to 200 km$^2$: Harvey, 1997). This regression equation lies midway among other data for alluvial fans in Spain and the Basin and Range Province in the United States. Taking $F = 1.3 \times 10^3$ km$^2$ for the Cape Wrath fan gives $A = 9.4 \times 10^3$ km$^2$. $A/F$ for six modern fluvial megafans (DeCelles & Cavazza, 1999) averages 13.5 (range = 0.75–44.0), which gives $A \approx 1.8 \times 10^4$ km$^2$ and possibly more for the Cape Wrath fan.

These results support the suggestion that the Applecross source area reached from the Outer Hebrides region westward for up to *c*. 250 km onto what are now the continental margins west of Scotland and east of Greenland (Williams, 1969*b*). Granulites and granites from the Rockall Bank gave a Laxfordian age of 1625 Ma (Morton & Taylor, 1991) with evidence of Grenvillian reworking at 987 Ma (Miller, Matthews & Roberts, 1973), findings that are consistent with the evidence for a varied Applecross source terrain now situated on the North Atlantic continental margins. The suggestion of Rainbird, Hamilton & Young (2001) that the Torridon Group was deposited by an intermontane river system flowing northeastward from Labrador, parallel to a proposed Grenvillian orogenic front, is not supported by Applecross fluvial geology, composition and palaeocurrent data. Although recycled material, which forms only a small fraction of the total Applecross detritus (Stewart & Donnellan, 1992), may have had an ultimate source in that part of Laurentia, such a distant provenance is not appropriate for the alluvial-fan conglomerates of the Cape Wrath Member and the immature arkoses and feldspathic sandstones that form the bulk of the Applecross Formation.

## 10. Conclusions

The following conclusions are reached concerning the palaeoenvironment, basin development and provenance of the lower Neoproterozoic Applecross Formation in northwest Scotland.

(1) The Cape Wrath Member of the Applecross Formation comprises alluvial fan and braidplain deposits. Regionally divergent flow directions for the lower part of the member reveal a fan of *c*. 50 km radius with its apex 30 km to the west. More distal, braidplain deposits derived from the west form the upper part of the member and the bulk of the Applecross Formation to the south.

(2) Extensional tectonics controlled Applecross deposition. Regional uplift causing deep erosion of the Lewisian was followed by maximum crustal extension with the development of the Applecross extensional basin in two main stages. Applecross deposition was controlled by episodic uplift of a western source area by movement on basin-bounding normal faults. Initial faulting and uplift in the north caused pediplanation near the northern basin margin and deposition of the Cape Wrath Member. Subsequent uplift of the source area for the main body of the Applecross Formation occurred further to the west and south along the line of the Minch Fault.

(3) A source terrain of moderately weathered, felsic crystalline rocks of Scourian, Laxfordian and Grenvillian age and Laxfordian supracrustals, with further weathering of detritus during transport, can account for the bulk of the Applecross Formation. The Applecross source area reached westward for up to *c*. 250 km onto what are now the continental margins of the North Atlantic. A more distant provenance is not appropriate for the alluvial-fan conglomerates and immature, feldspar-rich sandstones of the Applecross Formation.

(4) The sequence of tectonic events is consistent with the upwelling of a mantle plume and may mark an early phase in the crustal extension that led ultimately to the opening of Iapetus.

**Acknowledgements.** This paper is based on Ph.D. research in 1963–66 at the University of Reading and further work in 1996–99. I am indebted to Prof. P. Allen for suggesting the Torridonian of the Cape Wrath area as a research topic and for his continued interest. Dr P. Friend and an anonymous referee provided helpful reviews. S. Proferes, J. Stanley and K. Turnbull gave technical assistance.

## References

ALLEN, P. 1991. Provenance research: Torridonian and Wealden. In *Developments in Sedimentary Provenance Studies* (eds A. C. Morton, S. P. Todd and P. D. W. Haughton), pp. 13–21. Geological Society of London, Special Publication no. 57.

ALLEN, P., SUTTON, J. & WATSON, J. V. 1974. Torridonian tourmaline–quartz pebbles and the Precambrian crust northwest of Britain. *Journal of the Geological Society, London* **130**, 85–91.

BEACOM, L. E., ANDERSON, T. B. & HOLDSWORTH, R. E. 1999. Using basement-hosted clastic dykes as syn-rifting palaeostress indicators: an example from the basal Stoer Group, northwest Scotland. *Geological Magazine* **136**, 301–10.

BLAIR, T. C. 1999. Cause of dominance by sheetflood vs. debris-flow processes on two adjoining alluvial fans, Death Valley, California. *Sedimentology* **46**, 1015–28.

BLAIR, T. C. & MCPHERSON, J. G. 1994. Alluvial fans and their natural distinction from rivers based on morphology, hydraulic processes, sedimentary processes, and facies assemblages. *Journal of Sedimentary Research* **A64**, 450–89.

BLUNDELL, D. J., RESTON, T. J. & STEIN, A. M. 1989. Deep crustal structural controls on sedimentary basin geometry. In *Origin and Evolution of Sedimentary Basins and their Energy and Mineral Resources* (ed. R. A. Price), pp. 57–64. American Geophysical Union Geophysical Monograph no. 48.

BRIERLEY, G. J., LIU, K. & CROOK, K. A. W. 1993. Sedimentology of coarse-grained alluvial fans in the Markham Valley, Papua New Guinea. *Sedimentary Geology* **86**, 297–324.

CLIFF, R. A. & REX, D. C. 1989. Evidence for a ‘Grenville’ event in the Lewisian of the northern Outer Hebrides. *Journal of the Geological Society, London* **146**, 921–4.

COLLINSON, J. D. 1970. Bedforms of the Tana River, Norway. *Geografiska Annaler* **52A**, 31–56.

COWARD, M. P. & PARK, R. G. 1987. The role of mid-crustal shear zones in the Early Proterozoic evolution of the Lewisian. In *Evolution of the Lewisian and Comparable Precambrian High Grade Terrains* (eds R. G. Park and J. Tarney), pp. 127–38. Geological Society of London, Special Publication no. 27.

DECELLES, P. G. & CAVAZZA, W. 1999. A comparison of fluvial megafans in the Cordilleran (Upper Cretaceous) and modern Himalayan foreland basin systems. *Geological Society of America Bulletin* **111**, 1315–34.

Fettes, D. J., Mendum, J. R., Smith, D. I. & Watson, J. V. 1992. *Geology of the Outer Hebrides.* Memoir of the British Geological Survey, Sheets (solid edition) Lewis and Harris, Uist and Barra (Scotland). London: HMSO, 197 pp.

Garner, H. F. 1979. Coarsening-upward cycles in the alluvium of Hornelen Basin (Devonian), Norway: Sedimentary response to tectonic events: Discussion. *Geological Society of America Bulletin* **90**, 121–4.

Gibbs, A. D. 1984. Structural evolution of extensional basin margins. *Journal of the Geological Society, London* **141**, 609–20.

Gibbs, A. 1987. Development of extension and mixed-mode sedimentary basins. In *Continental Extensional Tectonics* (eds M. P. Coward, J. F. Dewey and P. L. Hancock), pp. 19–33. Geological Society of London, Special Publication no. 28.

Gohain, K. & Parkash, B. 1990. Morphology of the Kosi megafan. In *Alluvial Fans: A Field Approach* (eds A. H. Rachocki and M. Church), pp. 151–78. Chichester: Wiley.

Harvey, A. M. 1984. Debris flows and fluvial deposits in Spanish Quaternary alluvial fans: implications for fan morphology. In *Sedimentology of Gravels and Conglomerates* (eds E. H. Koster and R. J. Steel), pp. 123–32. Canadian Society of Petroleum Geologists Memoir no. 10.

Harvey, A. M. 1997. The role of alluvial fans in arid zone fluvial systems. In *Arid Zone Geomorphology: Process, Form and Change in Drylands* (ed. D. S. G. Thomas), pp. 231–59. Chichester: Wiley.

Heward, A. P. 1978. Alluvial fan sequence and megasequence models: with examples from the Westphalian D–Stephanian B coalfields, Northern Spain. In *Fluvial Sedimentology* (ed. A. D. Miall), pp. 669–702. Canadian Society of Petroleum Geologists Memoir no. 5.

Hill, R. I. 1991. Starting plumes and continental break-up. *Earth and Planetary Science Letters* **104**, 398–416.

Hiller, N. & Stavrakis, N. 1982. Reversed currents over a point bar on the Great Fish River. *Transactions of the Geological Society of South Africa* **85**, 215–19.

Hirst, J. P. P. & Nichols, G. J. 1986. Thrust tectonic controls on Miocene alluvial distribution patterns, southern Pyrenees. *Special Publication of the International Association of Sedimentologists* **8**, 247–58.

Jupp, P. E., Spurr, B. D., Nichols, G. J. & Hirst, J. P. P. 1987. Statistical estimation of the apex of a sediment distribution system from paleocurrent data. *Mathematical Geology* **19**, 319–33.

Khadkikar, A. S. 1999. Trough cross-bedded conglomerate facies. *Sedimentary Geology* **128**, 39–49.

Kumpulainen, R. & Nystuen, J. P. 1985. Late Proterozoic basin evolution and sedimentation in the westernmost part of Baltoscandia. In *The Caledonide Orogen—Scandinavia and Related Areas* (eds D. G. Gee and B. A. Sturt), pp. 213–32. Chichester: Wiley.

Lailey, M., Stein, A. M. & Reston, T. J. 1989. The Outer Hebrides fault: a major Proterozoic structure in NW Britain. *Journal of the Geological Society, London* **146**, 253–9.

Laubach, S. E. & Marshak, S. 1987. Fault patterns generated during extensional deformation of crystalline basement, NW Scotland. In *Continental Extensional Tectonics* (eds M. P. Coward, J. F. Dewey and P. L. Hancock), pp. 495–9. Geological Society of London, Special Publication no. 28.

Lawson, A. C. 1915. The epigene profiles of the desert. *University of California Publications, Bulletin of the Department of Geology* **9**, 23–48.

Lawson, D. E. 1972. Torridonian volcanic sediments. *Scottish Journal of Geology* **8**, 345–62.

Mendum, J. R., Evans, D. & Hitchen, K. 1989. Sutherland sheet 58°N–06°W, solid geology. *British Geological Survey 1:250,000 Series*. Natural Environment Research Council.

Miall, A. D. 1977. A review of the braided-river depositional environment. *Earth-Science Reviews* **13**, 1–62.

Miall, A. D. 1978. Lithofacies types and vertical profile models in braided river deposits: a summary. In *Fluvial Sedimentology* (ed. A. D. Miall), pp. 597–604. Canadian Society of Petroleum Geologists Memoir no. 5.

Miall, A. D. 1996. *The Geology of Fluvial Deposits: Sedimentary Facies, Basin Analysis, and Petroleum Geology*. Berlin: Springer, 582 pp.

Miall, A. D. 1997. *The Geology of Stratigraphic Sequences*. Berlin: Springer, 433 pp.

Miller, J. A., Matthews, D. H. & Roberts, D. G. 1973. Rocks of Grenville age from Rockall Bank. *Nature Physical Science* **246**, 61.

Mohindra, R. & Bagati, T. N. 1996. Seismically induced soft-sediment deformation structures (seismites) around Sumdo in the lower Spiti valley (Tethys Himalaya). *Sedimentary Geology* **101**, 69–83.

Moorbath, S. 1977. Ages, isotopes and evolution of Precambrian continental crust. *Chemical Geology* **20**, 151–87.

Moorbath, S. & Park, R. G. 1971. The Lewisian chronology of the southern region of the Scottish mainland. *Scottish Journal of Geology* **8**, 51–74.

Moorbath, S., Stewart, A. D., Lawson, D. E. & Williams, G. E. 1967. Geochronological studies on the Torridonian sediments of north-west Scotland. *Scottish Journal of Geology* **3**, 389–412.

Morton, A. C. & Taylor, P. N. 1991. Geochemical and isotopic constraints on the nature and age of basement rocks from Rockall Bank, NE Atlantic. *Journal of the Geological Society, London* **148**, 631–4.

Nemec, W. & Steel, R. J. 1984. Alluvial and coastal conglomerates: their significant features and some comments on gravelly mass-flow deposits. In *Sedimentology of Gravels and Conglomerates* (eds E. H. Koster and R. J. Steel), pp. 1–31. Canadian Society of Petroleum Geologists Memoir no. 10.

Nicholson, P. G. 1993. A basin reappraisal of the Proterozoic Torridon Group, northwest Scotland. In *Tectonic Controls and Signatures in Sedimentary Successions* (eds L. E. Frostick and R. J. Steel), pp. 183–202. Special Publication of the International Association of Sedimentologists no. 20.

Nilsen, T. H. 1982. Alluvial fan deposits. In *Sandstone Depositional Environments* (eds P. A. Scholle and D. Spearing), pp. 49–86. Tulsa: American Association of Petroleum Geologists.

Oberlander, T. M. 1974. Landscape inheritance and the pediment problem in the Mojave Desert of southern California. *American Journal of Science* **274**, 849–75.

Oberlander, T. M. 1997. Slope and pediment systems. In *Arid Zone Geomorphology: Process, Form and Change in Drylands*, 2nd ed. (ed. D. S. G. Thomas), pp. 135–63. Chichester: Wiley.

OLLIER, C. D. 1965. Some features of granite weathering in Australia. *Zeitschrift für Geomorphologie* **9**, 285–304.

OWEN, G. 1995. Soft-sediment deformation in upper Proterozoic Torridonian sandstones (Applecross Formation) at Torridon, northwest Scotland. *Journal of Sedimentary Research* **A65**, 495–504.

PARK, R. G. 1991. The Lewisian complex. In *Geology of Scotland*, 3rd ed. (ed. G. Y. Craig), pp. 25–64. London: Geological Society.

PEACH, B. N., HORNE, J., GUNN, W., CLOUGH, C. T., HINXMAN, L. W. & TEALL, J. J. H. 1907. The geological structure of the north-west highlands of Scotland. *Great Britain Geological Survey Memoir*, 668 pp.

PENCK, A. 1897. Geomorphologische Probleme aus Nordwest-Schottland. *Gesellschaft für Erdkunde zu Berlin Zeitschrift* **32**, 146–91.

POTTER, P. E. & PETTIJOHN, F. J. 1977. *Paleocurrents and Basin Analysis*, 2nd ed. Berlin: Springer, 425 pp.

RACHOCKI, A. 1981. *Alluvial Fans*. Chichester: Wiley, 161 pp.

RAINBIRD, R. H., HAMILTON, M. A. & YOUNG, G. M. 1998. U–Pb detrital zircon geochronology of the Neoproterozoic Torridonian succession of NW Scotland. *Geological Society of America Abstracts with Programs* **30**, A-291.

RAINBIRD, R. H., HAMILTON, M. A. & YOUNG, G. M. 2001. Detrital zircon geochronology and provenance of the Torridonian, NW Scotland. *Journal of the Geological Society, London* **158**, 15–27.

RETALLACK, G. J. & MINDSZENTY, A. 1994. Well preserved late Precambrian paleosols from northwest Scotland. *Journal of Sedimentary Research* **A64**, 264–81.

SELLEY, R. C., SHEARMAN, D. J., SUTTON, J. & WATSON, J. 1963. Some underwater disturbances in the Torridonian of Skye and Raasay. *Geological Magazine* **100**, 224–43.

STEEL, R. & AASHEIM, S. M. 1978. Alluvial sand deposition in a rapidly subsiding basin (Devonian, Norway). In *Fluvial Sedimentology* (ed. A. D. Miall), pp. 385–412. Canadian Society of Petroleum Geologists Memoir no. 5.

STEEL, R. J., MÆHLE, S., NILSEN, H., RØE, S. L. & SPINNANGR, Å. 1977. Coarsening-upward cycles in the alluvium of Hornelen Basin (Devonian) Norway: Sedimentary response to tectonic events. *Geological Society of America Bulletin* **88**, 1124–34.

STEEL, R. J. & WILSON, A. C. 1975. Sedimentation and tectonism (?Permo-Triassic) on the margin of the North Minch Basin, Lewis. *Journal of the Geological Society, London* **131**, 183–202.

STEIN, A. M. 1988. Basement controls upon basin development in the Caledonian foreland, NW Scotland. *Basin Research* **1**, 107–19.

STEIN, A. M. 1992. Basin development and petroleum potential in The Minches and Sea of the Hebrides Basins. In *Basins of the Atlantic Seaboard: Petroleum Geology, Sedimentology and Basin Evolution* (ed. J. Parnell), pp. 17–20. Geological Society of London, Special Publication no. 62.

STEIN, A. M. & BLUNDELL, D. J. 1990. Geological inheritance and crustal dynamics of the northwest Scottish continental shelf. *Tectonophysics* **173**, 455–67.

STEWART, A. D. 1982. Late Proterozoic rifting in NW Scotland: the genesis of the 'Torridonian'. *Journal of the Geological Society, London* **139**, 413–20.

STEWART, A. D. 1991. Torridonian. In *Geology of Scotland*, 3rd ed. (ed. G. Y. Craig), pp. 65–85. London: Geological Society.

STEWART, A. D. & DONNELLAN, N. C. B. 1992. Geochemistry and provenance of red sandstones in the Upper Proterozoic Torridon Group of Scotland. *Scottish Journal of Geology* **28**, 143–53.

STONE, P., PLANT, J. A., MENDUM, J. R. & GREEN, P. M. 1999. A regional geochemical assessment of some terrane relationships in the British Caledonides. *Scottish Journal of Geology* **35**, 145–56.

TAYLOR, G., CROOK, K. A. W. & WOODYER, K. D. 1971. Upstream-dipping foreset cross-stratification: origin and implications for paleoslope analyses. *Journal of Sedimentary Petrology* **41**, 578–81.

TAYLOR, P. N., JONES, N. W. & MOORBATH, S. 1984. Isotopic assessment of relative contributions from crust and mantle sources to the magma genesis of Precambrian granitoid rocks. *Philosophical Transactions of the Royal Society, London* **A310**, 605–25.

TAYLOR, S. R. & MCLENNAN, S. M. 1985. *The Continental Crust: its Composition and Evolution*. Oxford: Blackwell, 312 pp.

TUAN, YI-FU. 1959. Pediments in southeastern Arizona. *University of California Publications in Geography* **13**, 1–163.

TURNBULL, M. J. M., WHITEHOUSE, M. J. & MOORBATH, S. 1996. New isotopic age determinations for the Torridonian, NW Scotland. *Journal of the Geological Society, London* **153**, 955–64.

VAN DE KAMP, P. C. & LEAKE, B. E. 1997. Mineralogy, geochemistry, provenance and sodium metasomatism of Torridonian rift basin clastic rocks, NW Scotland. *Scottish Journal of Geology* **33**, 105–24.

VISERAS, C. & FERNANDEZ, J. 1994. Channel migration patterns and related sequences in some alluvial fan systems. *Sedimentary Geology* **88**, 201–17.

WEAVER, B. L. & TARNEY, J. 1981. Lewisian gneiss geochemistry and Archaean crustal development models. *Earth and Planetary Science Letters* **55**, 171–80.

WILLIAMS, G. E. 1966*a*. Palaeogeography of the Torridonian Applecross Group. *Nature* **209**, 1303–6.

WILLIAMS, G. E. 1966*b*. Planar cross-stratification formed by the lateral migration of shallow streams. *Journal of Sedimentary Petrology* **36**, 742–6.

WILLIAMS, G. E. 1968*a*. Torridonian weathering, and its bearing on Torridonian palaeoclimate and source. *Scottish Journal of Geology* **4**, 164–84.

WILLIAMS, G. E. 1968*b*. Formation of large-scale trough cross-stratification in a fluvial environment. *Journal of Sedimentary Petrology* **38**, 136–40.

WILLIAMS, G. E. 1969*a*. Characteristics and origin of a Precambrian pediment. *Journal of Geology* **77**, 183–207.

WILLIAMS, G. E. 1969*b*. Petrography and origin of pebbles from Torridonian strata (late Precambrian), northwest Scotland. In *North Atlantic—Geology and Continental Drift* (ed. M. Kay), pp. 609–29. American Association of Petroleum Geologists Memoir no. 12.

WILLIAMS, G. E. 1970*a*. Origin of disturbed bedding in Torridon Group sandstones. *Scottish Journal of Geology* **6**, 409–11.

WILLIAMS, G. E. 1970*b*. Piedmont sedimentation and late Quaternary chronology in the Biskra region of the northern Sahara. *Zeitschrift für Geomorphologie Supplementband* **10**, 40–63.

WILLIAMS, G. E. 1971. Flood deposits of the sand-bed ephemeral streams of central Australia. *Sedimentology* **17**, 1–40.

WILLIAMS, G. E. & SCHMIDT, P. W. 1997. Palaeomagnetic dating of sub-Torridon Group weathering profiles, NW Scotland: verification of Neoproterozoic palaeosols. *Journal of the Geological Society, London* **154**, 987–97.

Young, G. M. 1999*a*. A geochemical investigation of palaeosols developed on Lewisian rocks beneath the Torridonian Applecross Formation, NW Scotland. *Scottish Journal of Geology* **35**, 107–18.

Young, G. M. 1999*b*. Some aspects of the geochemistry, provenance and palaeoclimatology of the Torridonian of NW Scotland. *Journal of the Geological Society, London* **156**, 1097–1111.