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Deformation in the mid to lower continental crust: analogues from Proterozoic shear zones in NW Scotland

TIMOTHY JAMES WYNN
Department. of Geology, Imperial College, Prince Consort Road, London SW7 2BP, UK
Present address: GeoScience Ltd, Falmouth Business Park, Bickland Water Road, Falmouth, Cornwall TR11 4SZ, UK

Abstract: A suite of 1.8–1.7 Ga (Laxfordian), E–W striking, amphibolite-facies shears in the Lewisian of NW Scotland have been mapped across a pre-existing rheological boundary in the Laxford–Scourie area. The rheological contrast is provided by the southern boundary of the NW striking Laxford shear zone (LSZ), which displays complex Proterozoic folding and shearing of relict Archaean granulites. The more competent Archaean gneisses lie to the SW of the LSZ which has been a focus of shearing, amphibolitization and metasomatism from 2.6 to 1.4 Ga. The shears of interest display a second-order, extensional shear band relationship to the LSZ and formed during a phase of sinistral transtensional shearing and granite intrusion at 1.8–1.7 Ga. The second-order shears show displacement variations consistent with their propagation from the strongly foliated, less competent gneisses in the LSZ into the Archaean granulites to the SW.

A map of the Laxford–Scourie area can be regarded as a section through the mid to lower crust as observed on some deep seismic lines such as the BIRPS DRUM line from the north of Scotland. In the Laxford area, the change in geometry of the second order shears from SW to NE, as they curve into the LSZ, is comparable to the curvature of extensional shears in the mid to lower continental crust. Displacement variations indicate that the shear zones may nucleate at the base of the upper/mid crust and propagate up and down dip. Alternatively the shears may nucleate in the mid/lower crust and propagate up dip, becoming steeper as they do so.

The Lewisian of NW Scotland provides an excellent opportunity to study a variety of rocks that have undergone several phases of deformation throughout the spectrum of crustal levels. Peach *et al.* (1907) established the principal features of the mainland Lewisian by excellent systematic mapping in the latter part of the last century. An important unit was recognized at Loch Laxford separating highly deformed gneisses to the north from the 'fundamental Complex' to the south.

This work was built upon by Sutton & Watson (1951) who recognized the importance of the Scourie dyke swarm as markers separating tectonic and metamorphic events throughout the Lewisian. The area between Loch Laxford and Badcall Bay was studied in some detail by Sutton & Watson (1951) and the principal metamorphic and structural changes recorded from S to N. These show a progressive increase in amphibolitization and shearing into the core of an intense zone of deformation occupied by syn-tectonic granites, later termed the Laxford Shear Zone (Beach *et al.* 1974). A variety of rock types exist in this area from ultramafic to trondhjemitic and from orthogneiss to paragneiss and provide a good opportunity to study the effects of progressive deformation and metamorphism on rocks of contrasting rheology.

Figure 1 shows the extent of Lewisian outcrop in NW Scotland and the variety of gneissic types mapped. This paper is concerned with the evolution of the northern part of the relict Archaean Central Region (CR) where it becomes progressively sheared into the Laxford shear zone (LSZ) (Sutton & Watson 1951; Beach *et al.* 1974; Davies 1978; Coward & Park 1987; Park 1991). This area is shown in more detail in Fig. 2 where a suite of subvertical sinistral shears can be seen to merge with the main LSZ; the northern boundary of the LSZ is marked by the zone of Laxford granite sheets which dip steeply southwest. This suite of sinistral shears has been mapped in detail and their geometrical, kinematic and petrological changes recorded from E to W as they merge with the LSZ. In this paper these changes are compared with the possible kinematic and geometric evolution of crustal scale extensional faults formed in the continental crust during extensional thinning.

From COWARD, M. P. & RIES, A. C. (eds), 1995, *Early Precambrian Processes*, Geological Society Special Publication No. 95, pp. 225–241.

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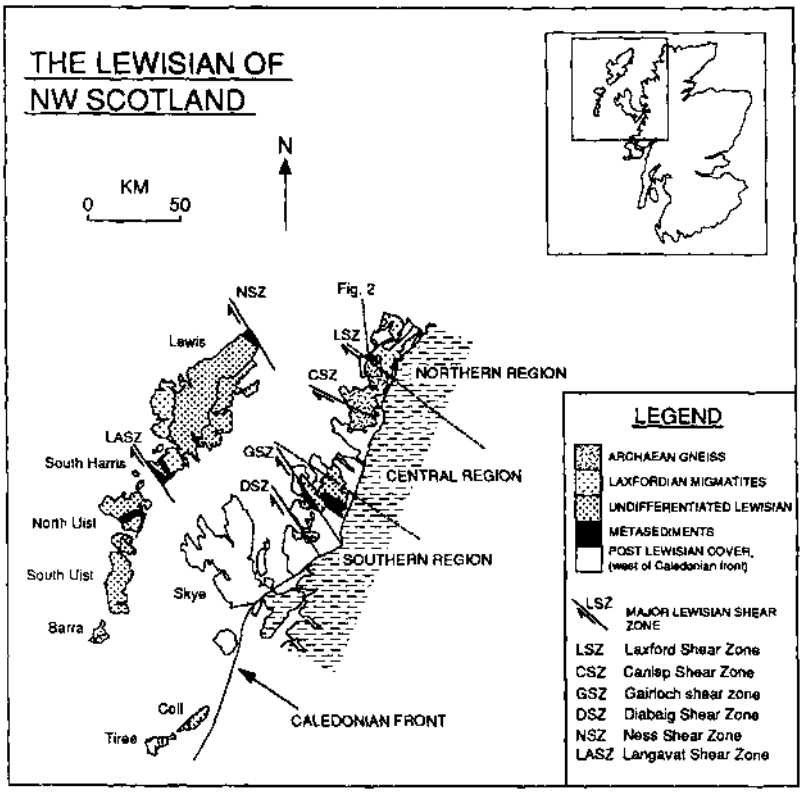

Fig. 1. Map of the Lewisian in NW Scotland. After Park & Tarney (1987).

Comparisons are provided by deep seismic lines and areas of exhumed crustal cross sections.

Geological history

Scourian

Table 1 shows the main chronological subdivisions of the Lewisian and the associated tectono-metamorphic events. One of the important events in the Lewisian was the late Scourian or Inverian tectonometamorphism which marked the change from predominantly pervasive, subhorizontal shearing (compression) to subvertical oblique-slip shearing and overthrusting on the LSZ and many of the other NW trending shears shown in Fig. 1 (Evans & Tarney 1964; Park 1964; Evans & Lambert 1974; Sheraton *et al.* 1973). On the southern margin of the LSZ this resulted in the formation of upright WNW-trending folds and stretching lineations, plunging 50–60° to the SE (Davies 1976, 1978). This predominantly dextral shearing was accompanied by extensive amphibolitization within both the shears and the surrounding gneisses (Evans & Lambert 1974; Beach & Tarney 1978).

Scourie dyke swarm

Scourian tectonometamorphism was followed by the emplacement of the Scourie dyke swarm which produced the volumetrically important basic and ultrabasic dykes used so effectively by Sutton & Watson (1951) in unravelling the chronology of the Lewisian. It is now recognized that the dykes were intruded between 2.4 and 2.0 Ga, probably in several pulses (Tarney 1973; Park & Tarney 1987; Heaman & Tarney 1989). However this concept of polyphase intrusion does not alter their fundamental importance as chronological markers. The dykes are predominantly quartz dolerites in the Laxford area and are mainly subvertical with NW trends. Most dykes were metamorphosed shortly after intrusion so that hornblende completely replaces the igneous pyroxene although some original ophitic textures are visible in places, (Beach & Tarney 1978; Park & Tarney 1987).

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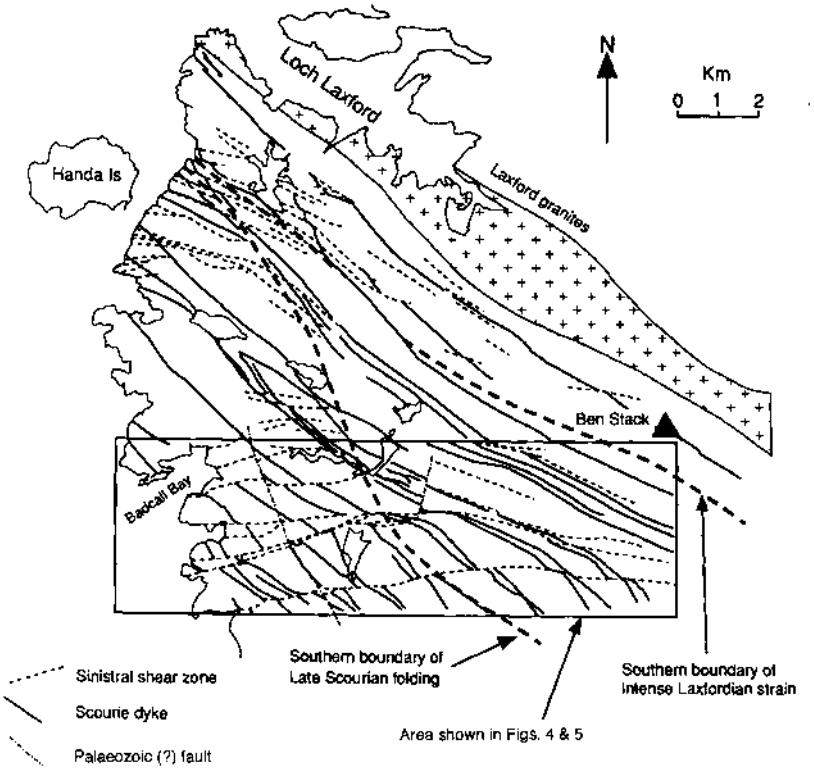

Fig. 2. Geological map of the Laxford–Badcall–Ben Stack area. After Beach *et al.* (1974).

Laxfordian

In the early Laxfordian at 1.9–1.8 Ga the Lewisian complex was subjected to the intense Laxfordian deformation and amphibolitization. Within the LSZ this resulted in continued dextral transpression along a movement vector plunging 30–40° SE. The late Scourian folds were tightened and overturned to the north and the Scourie dykes were also strongly sheared in places; (Beach *et al.* 1974). The Northern and Southern regions were intensely reworked at this time, largely obliterating the Scourian structures. Deformation in the Northern region resulted in strongly foliated gneisses dipping gently south, possibly as part of a shear zone 'flat' to the steeper LSZ which may have formed as a mid-crustal ramp (Beach 1974; Beach *et al.* 1974).

The sinistral shears of interest may have formed as second order antithetic shears to this early Laxfordian transpression (Davies 1978). However the sinistral shears at Tarbet appear to cut the late Scourian folds, so it is more likely they may have formed during a later phase of distinct sinistral transtension affecting the LSZ at 1.8–1.7 Ga (Beach 1974, 1976; Coward & Park 1987; Coward 1990). It is possible that the Laxford granites were intruded during the transtensional shearing. The Laxford granites may have formed by partial melting of the enriched Northern region which was emplaced beneath the Central region by early Laxfordian overthrusting (Beach 1974).

Late Laxfordian to Grenvillian deformation was restricted to greenschist-facies folding and re-activation of earlier foliations and shears, mainly in the Northern and Southern regions.

P–T–t history

Figure 3 shows the *P–t* and *T–t* paths for the Central Region based on geochronological, palaeothermal and palaeobaric studies by Giletti *et al.* (1961), Evans (1965), Evans & Tarney (1964), Moorbath (1969), Lambert & Holland (1972), Beach (1973), Tarney (1973), Wood (1977), O'Hara (1977), Rollinson (1980*a*, 1980*b*, 1981), Pride & Muecke (1980), Savage & Sills (1980), Humphries & Cliff (1982), Newton & Perkins (1982),

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Table 1. *Geological events in the Central Region of the Lewisian*

Archaean (Scourian)	
2900 Ma	Scourian tonalites and trondhjemites produced from partial melts of subducted oceanic crust. Subhorizontal gneissic layering produced.
2700 Ma	Badcallian grasnulites facies metamorphism. Isoclinal, recumbent folding of layering, fold axial surfaces dip gently NW.
2600 Ma	Late Scourian (Inverian) tectonometamorphism, widespread static amphibolitisation. Formation of steep NW striking lineaments at Laxford, Canisp and Gruinard Bay.
Proterozoic	
2400–2000 Ma	Intrusion of the Scourie dyke swarm, 4 suites from ultrabasic to basic derived from 2 separate mantle sources. The dykes mark the boundary between the Scourian and the Laxfordian (Sutton & Watson 1951).
(Laxfordian)	
1900–1800 Ma	Laxfordian amphibolite-facies deformation. Central region thrust over the northern region during dextral transpression on the Laxford Shear Zone.
1800–1700 Ma	Peak of Laxfordian metamorphism, intrusion of the Laxford granites. Sinistral transtension on the main Laxford Shear Zone with the minor sinistral shears formed on the SW margin.
1700–1400 Ma	Mid–Late Laxfordian retrogression to greenschist facies, associated with kilometre-scale folding in the Southern region. Development of some pseudotachylite belts throughout the Lewisian.
1150 Ma	Minor chlorite and muscovite growth and the formation of later pseudotachylite belts. Both possibly related to the Grenvillian–Sveconorwegian orogeny.
1100–1040 Ma	Exposure at the Torridonian land surface, deposition of the Stoer and Torridon groups.

After Park & Tarney (1987).

Barnicoat (1983), Sills (1983), Sills & Rollinson (1987), Cartwright & Barnicoat (1986, 1987) and Heaman & Tarney (1989).

The main features of the curves shown in Fig. 3 are relatively rapid cooling and uplift from peak Badcallian conditions at 2.7–2.6 Ga, fairly stable *P* and *T* conditions during dyke intrusion, a slight thermal peak during the early to mid-Laxfordian orogeny (and collapse) and slow cooling and uplift from 1.6–1.0 Ga when the Lewisian was exposed at the Torridonian land surface. This extremely slow exhumation history was a major factor in allowing

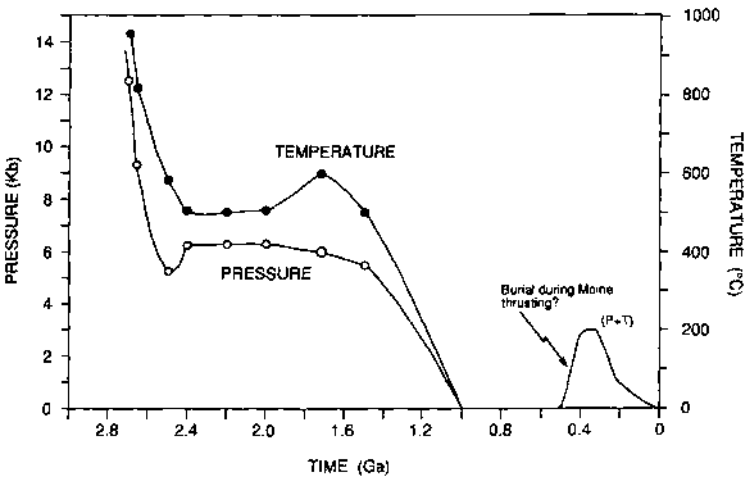

Fig. 3. Pressure–time and temperature–time plots for the Central Region (see text for references).

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repeated phases of movement on the LSZ which resulted in the complex, medium to high grade deformation visible in the Lewisian today.

Sinistral shear zones

Figure 2 shows the positions of the sinistral shears with respect to the main LSZ and the boundaries of late Scourian and early Laxfordian strain. These boundaries indicate a progressive northeastwards migration of deformation on the LSZ, corresponding to an increase in strain localization. All the fabrics appear to be amphibolite facies, so the decrease in the width of the zone of penetrative shearing is unlikely to be due to cooling and is probably a result of the availability of fluids for metasomatism and subsequent weakening of the gneisses (Beach 1976; Beach & Tarney 1978).

On the basis of the similarity in structural, metamorphic and kinematic style, the Laxfordian sinistral shears in the Tarbet area (Beach 1974; Coward 1990) are regarded as coeval with the E–W sinistral shears studied in this project. The Tarbet shears occur within, and partly define, the main Laxford shear zone and the foliation and lineation orientations have been used as reference orientations for the LSZ shear plane and the LSZ movement vector, respectively during structural analysis of the E–W sinistral shears. The geometrical and kinematic relationships between the two sets of sinistral shears indicates that the E–W set has a second order, extensional, synthetic relationship to the Tarbet shears (cf. Beach 1974; Coward & Park 1987).

Figure 4 shows the E–W sinistral shears in detail. The area mapped has been subdivided into four zones, where the zone boundaries join up inflection points on the shears. The overall change is a clockwise rotation in strike from W to E. The zone boundaries are approximately parallel to the main LSZ. Thin section analysis of the sinistral shear fabrics has shown that they were formed in amphibolite-facies conditions in all the zones

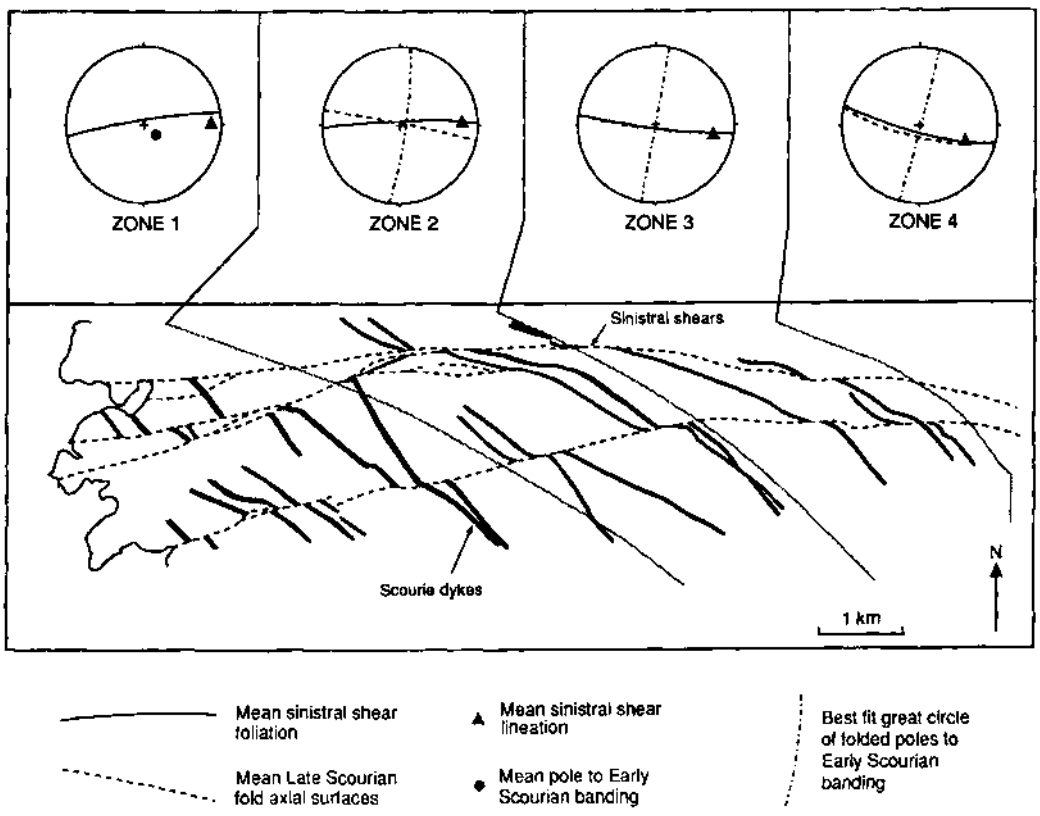

Fig. 4. Map of sinistral shears in zones 1 to 4 of the Badcall–Ben Stack–Kylesku area. Stereonets of the main fabrics in each zone are also shown.

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although later, localized greenschist facies retrogression has occurred. This indicates that the kinematic and geometrical changes of the sinistral shears from W to E is not a direct function of a lateral change in metamorphic grade.

Characteristics of the gneisses surrounding the sinistral shears

Structural analysis of the older fold fabrics in each zone from the gneisses surrounding the shear zones shows progressive changes from W to E, as shown in Fig. 4. In zone 1 the Scourie gneiss banding dips gently NW, in the centre of zone 2 tight, WNW-striking folds occur which become progressively tighter and overturned to the NE in zones 3 and 4. The changes in structures from SW to NE in the gneisses surrounding the shears (as seen in Figs 2 and 4) can be summarized as follows:

(i) increase in intensity of steep, penetrative (axial planar) foliations;
(ii) clockwise rotation of axial planar foliations and tightening of folds;
(iii) increase in retrogression of the Scourie gneisses, amphibolitization and biotite growth in the foliation zones, probably as a result of increased metasomatic activity;
(iv) Increase in effective anisotropy of the gneisses, i.e. change from sub-horizontal layering perpendicular to sinistral shears, to sub-vertical foliations, sub-parallel to shears.

These features indicate that the development of the sinistral shears in zones 1 to 4 was closely linked to changes in the structure of the surrounding gneisses. The general trend is one of decreasing obliquity between the sinistral shear fabrics and the late Scourian axial surfaces (shown in zones 2 and 3). In zone 4, all fabrics including the Scourie dykes are sub-parallel.

It is possible that these changes are equivalent to an increase in ductility of the gneisses from SW to NE, although this will depend to an extent on the orientation of the stress and strain fields during any subsequent deformation. Beach (1973, 1976) indicated that the growth of the sinistral shears was controlled by the availability of fluids to propagate fractures and metasomatize the gneisses, allowing the growth of biotite. These factors were believed to have contributed to the strain localization of these shears. However the shears only appear to be very localized in zones 1 and 2 and in zones 3 and 4 the shearing appears to be more diffuse.

This characteristic is highlighted by Fig. 5 which shows the positions of correlatable Scourie dykes across the two largest shears and the approximate positions of diffuse zones of deformation either side of the discrete parts of the shears in zones 3 and 4. These diffuse zones can be recognized by patchy fabric development within the Scourie dykes and their gradual anti-clockwise deflection as the discrete parts of the sinistral shears are approached. It is suggested that these diffuse zones of shear have formed by the reactivation of late Scourian fold axial surfaces during sinistral shearing in zones 2 and 3 (see Fig. 4). Penetrative shearing probably occurred throughout zone 4, coeval with the more discrete shearing in zones 1 to 3.

Displacement plots

Figure 6 shows the geometrical construction used to calculate the true displacements on the two

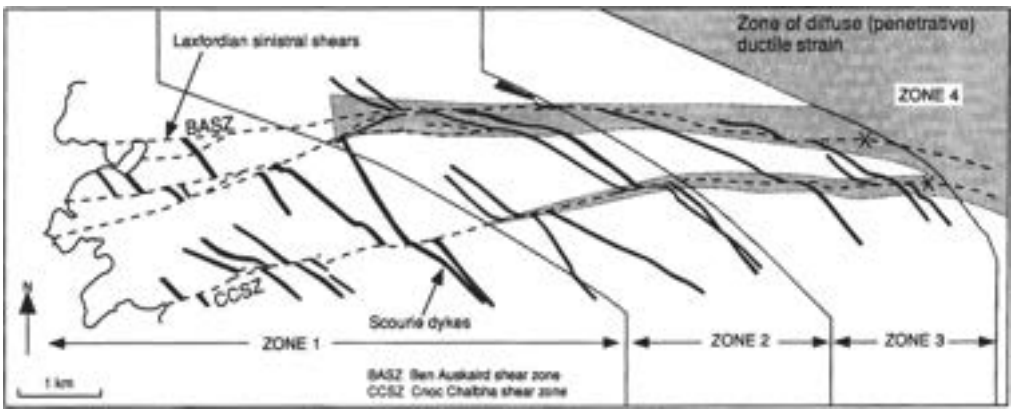

Fig. 5. Map of the sinistral shears in zones 1 to 4. The zones of diffuse deformation on the margins of the shears are also shown.

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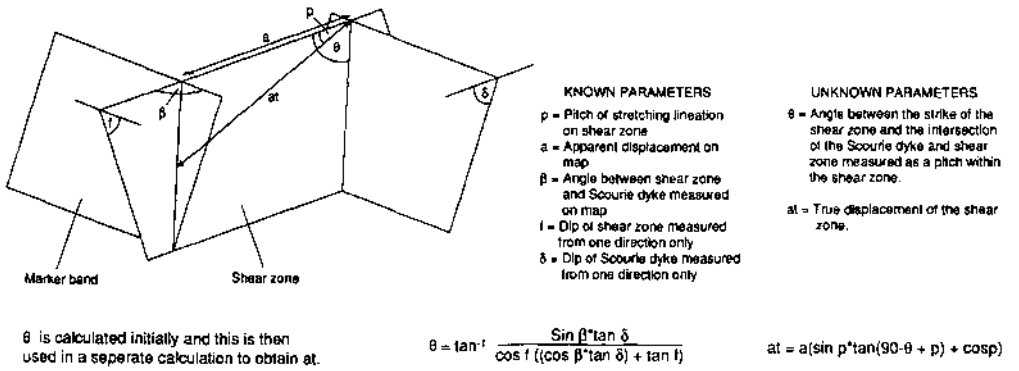

Fig. 6. Geometrical construction showing how the true displacements on the displacement plots in Fig. 7 were calculated.

sinistral shears shown in Fig. 5. This construction takes into account the dip of the shear zone, the dip of the markers (Scourie dykes) and the plunge of the movement vector. Because of the large strains accumulated on the sinistral shears, the movement vector was taken to be represented by the stretching lineations in the shears. The true displacement values obtained were then used in the displacement plots shown in Fig. 7, these displacement plots are for the Ben Auskaird and Cnoc Chalbha shear zones (BASZ and CCSZ, respectively, see Fig. 5).

Ben Auskaird shear zone

For the BASZ it can be seen that the displacements are relatively high in zone 1 (1400 m) and the western half of zone 2, between 4000 and 5000 m from the coast (1900 m displacement); this rapidly decreases in 1000 m of strike length at the boundary of zones 2 and 3 to 500 m. The rapid decrease in displacement can only be realistically explained by the partitioning of strain into the surrounding gneisses in the eastern part of zone 2 and all of zone 3, where the favourably oriented late Scourian axial surfaces could take up the deformation. This idea is further supported by the two dykes on Ben Auskaird in the centre of zone 2. Here the displacement can be separated into a total component, indicated by the deflection of the dykes across the Ben Auskaird shear zone and a discrete component, where the dykes are actually cut by a discrete shear zone. The total component is similar to the displacement values for zone 1 and the discrete component similar to the values for zone 3 (see linked points in Fig. 7a).

The large total displacement on the BASZ has shifted the southern boundary of late Scourian folding approximately 2 km to the east, so that the CCSZ cuts it at the boundary of zones 2 and 3.

Cnoc Chalbha shear zone

The displacements for the CCSZ can also be subdivided into diffuse and discrete shearing. The plot shown in Fig. 7b contains the total displacement estimates and the discrete component in the eastern part of zone 2 and in zone 3. The total displacement systematically increases from zone 1 to zone 3, although the best fit curve changes in gradient in zone 3. The discrete component decreases in value eastwards from equal to the total displacement at the southern boundary of late Scourian folding, to virtually zero at the eastern margin of zone 3. Again this seems to indicate a partitioning of strain into the late Scourian fold axial surfaces. However, the strain partitioning is less pronounced in the CCSZ compared with the BASZ where the total displacements are much higher.

The total displacement decreases from E to W which can possibly be attributed to propagation of the CCSZ from zone 3 into zones 2 and 1. It is possible that this may be due to the relative weakness of the Scourie gneisses in zone 3 compared to the Scourie gneisses in zones 1 and 2 which contain relict granulite-facies minerals with granoblastic textures and where the compositional banding is sub-horizontal and unavailable for reactivation. This would therefore mean that the shears had to grow as new fractures in zones 1 and 2, possibly driven by fluids emanating from the Laxford granites, northeast of zone 4 (Beach 1976).

Orientations of the sinistral shears with respect to the LSZ

Figure 8 shows an equal area stereonet showing the change in orientation of the sinistral shear segments from zone 1 to zone 4 with respect to the LSZ (Tarbet) mean foliation and lineation. It is clear

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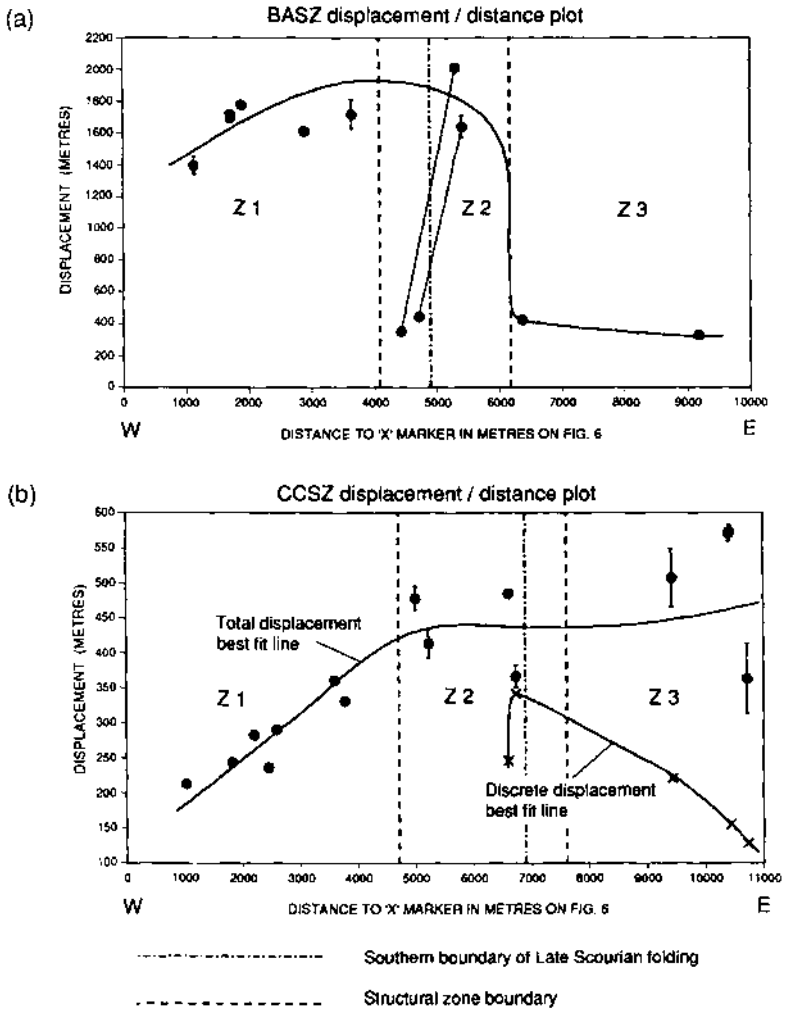

Fig. 7. Displacement plots for the Ben Auskaird shear zone (**a**) and the Cnoc Chaibha shear zone (**b**).

from Fig. 8 that the shears become closer in orientation to the LSZ as the LSZ is approached from SW to NE. Figure 9 is a 3D block diagram of the sinistral shears and the LSZ; Fig. 9 also shows how the true thicknesses of the structural zones were calculated, these thicknesses are used on the vertical axes of the plots in Fig. 10.

Figure 10a shows the angles of the mean sinistral shear foliations and the late Scourian axial surfaces with respect to the LSZ, as measured in the profile plane of the LSZ. The BASZ segments in zones 3 to 4 and the CCSZ segment in zone 3, have similar orientations to the late Scourian axial surfaces in these zones. However the BASZ segment in zone 2 and the CCSZ segment in zone 3 are oriented at slightly higher angles than the axial surfaces. The larger obliquity between these shear segments and the late Scourian axial surfaces may be a function of the lower strains (displacements) for the western parts of the shears. Therefore, the late Scourian axial surfaces have been rotated by a smaller amount and still retain an appreciable obliquity to the discrete shear segments.

In addition to the changes in angle of the shear zone segments, the movement vectors (lineations) in the sinistral shears in zones 1 to 4 also change orientation. The best fit great circle for these lineations, shown in Fig. 8, does not have an orientation that is consistent with the sinistral shears forming as plane strain, simple shear splays. It is more likely that the sinistral shears contain a finite component of rotational movement, defined by an

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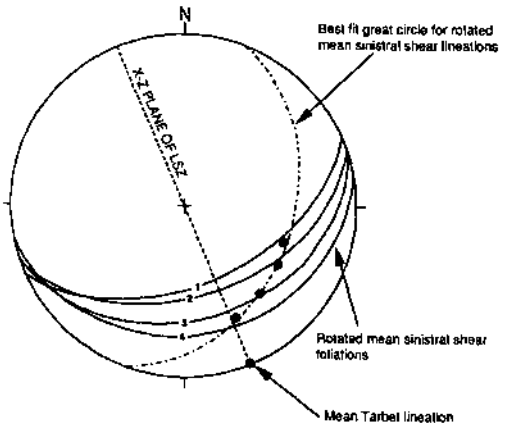

Fig. 8. Stereonet of rotated mean foliations and lineations from zones 1 to 4 of the Badcall–Ben Stack–Kylesku area.

arc contained within a plane at an angle to the profile plane to the LSZ.

Figure 10b shows the angles of the CCSZ and BASZ with respect to the LSZ as measured within the best fit great circle to the mean lineations from zones 1 to 4. The angles were obtained by measuring the plunges of the intersections of the best fit great circle to the sinistral shear lineations and the mean foliations from each zone (see Fig. 8). The angles between the shear segments and the LSZ are more consistent than those obtained in the profile plane (X–Z section) of the LSZ (see Fig. 10a). In addition, there appears to be smaller obliquity between the sinistral shear segments and the late Scourian axial surfaces in zone 2 for the BASZ and zone 3 for the CCSZ (see Fig. 10b).

If the shear zone segments propagated in a similar direction as the movement vectors of the shears, then this feature indicates that the shears propagated in a direction that had a smaller change in curvature from zone 4 to zone 1 than would be the case if they had propagated in a direction contained within the profile plane of the LSZ. The reason for this is not clear, although it is probably a function of local stress field variations and the strong control of pre-existing anisotropies in the Scourian gneisses.

Model for the formation of the sinistral shears

Figure 11 shows two models for the evolution of the sinistral shears in the Scourie–Laxford area. The main feature of the models is that the Central Region was being displaced down to the SE during sinistral transtension on the LSZ, whilst the Central

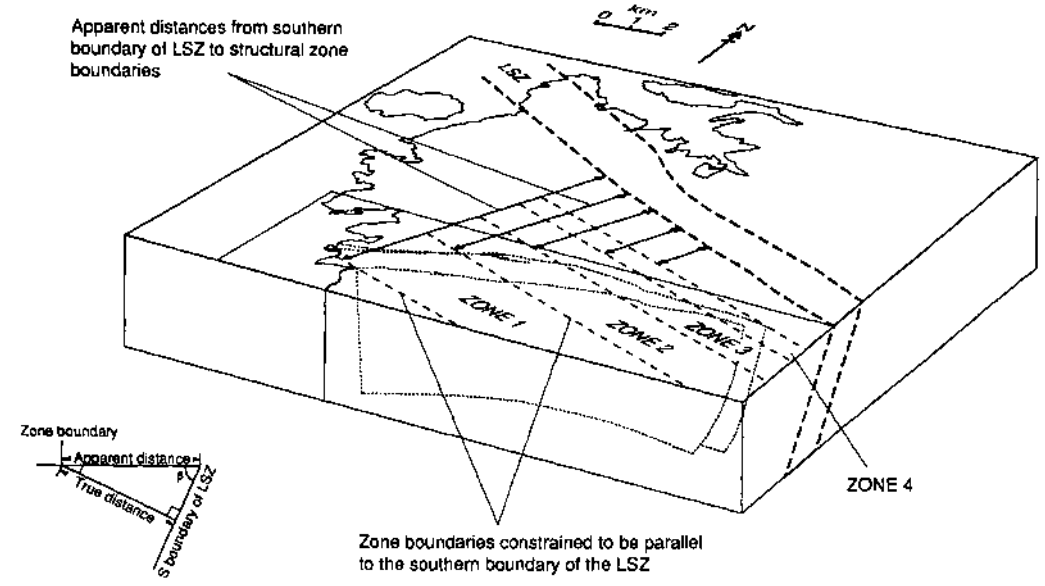

Fig. 9. Block diagram of structural zones 1 to 4 to show how the true thicknesses of the zones were calculated. These thicknesses are used on the vertical axes of the plots in Fig. 10.

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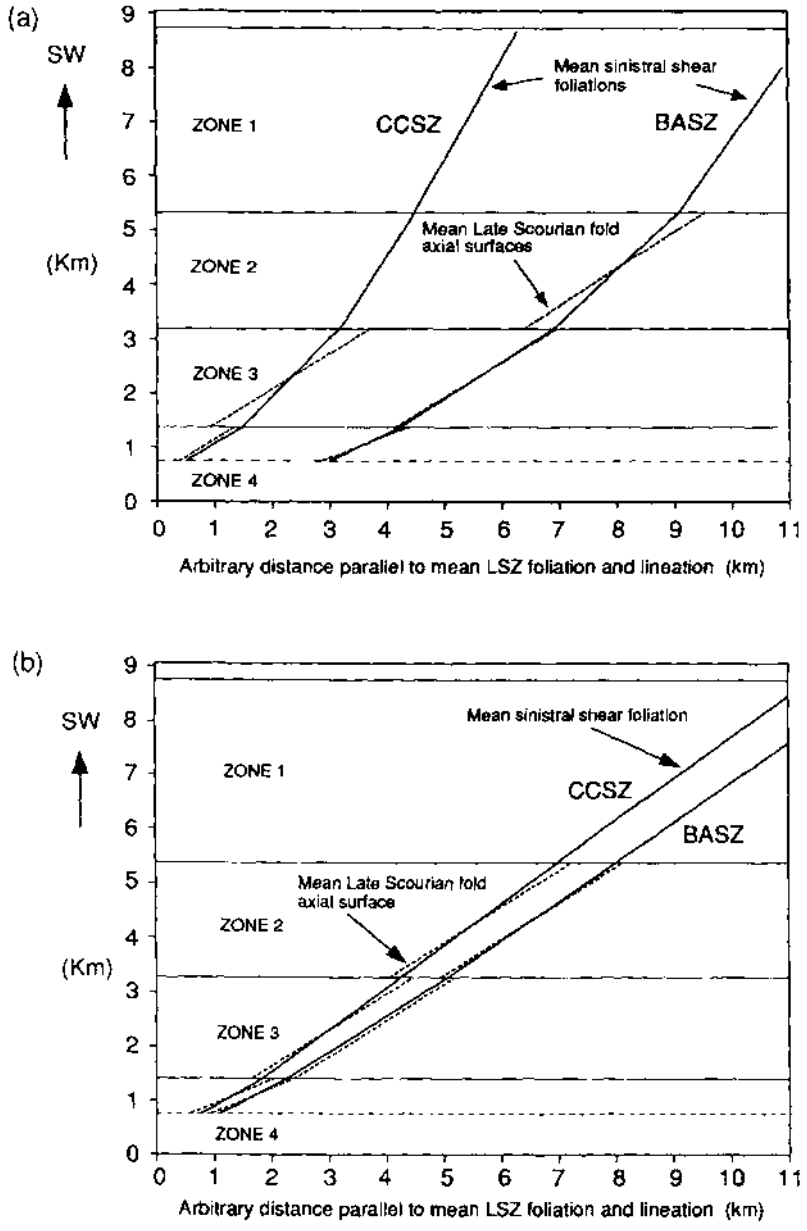

Fig. 10. Plots of angles between mean fabrics in zones 1 to 4 and the mean LSZ fabrics. (a) Plot of angles measured within the profile plane of the LSZ. (b) Plot of angles measured within the best fit great circle to the rotated mean lineations from zones 1 to 4.

Region was also extended parallel to the movement vector on the LSZ.

Figure 11a shows a model of the sinistral shears initiated as accommodation structures by a combination of simple shear and block rotation on the northern margin of the Central Region within the more penetratively deformed gneisses (Fig. 11a(i)). Continued deformation propagated the shears into the more competent Scourie gneisses and also allowed at least one discrete shear strand (CCSZ) to propagate from the southern boundary of Scourian folding eastwards back along the more diffuse zone of shearing (Fig. 11a(ii)).

Figure 11b shows an alternative model where the

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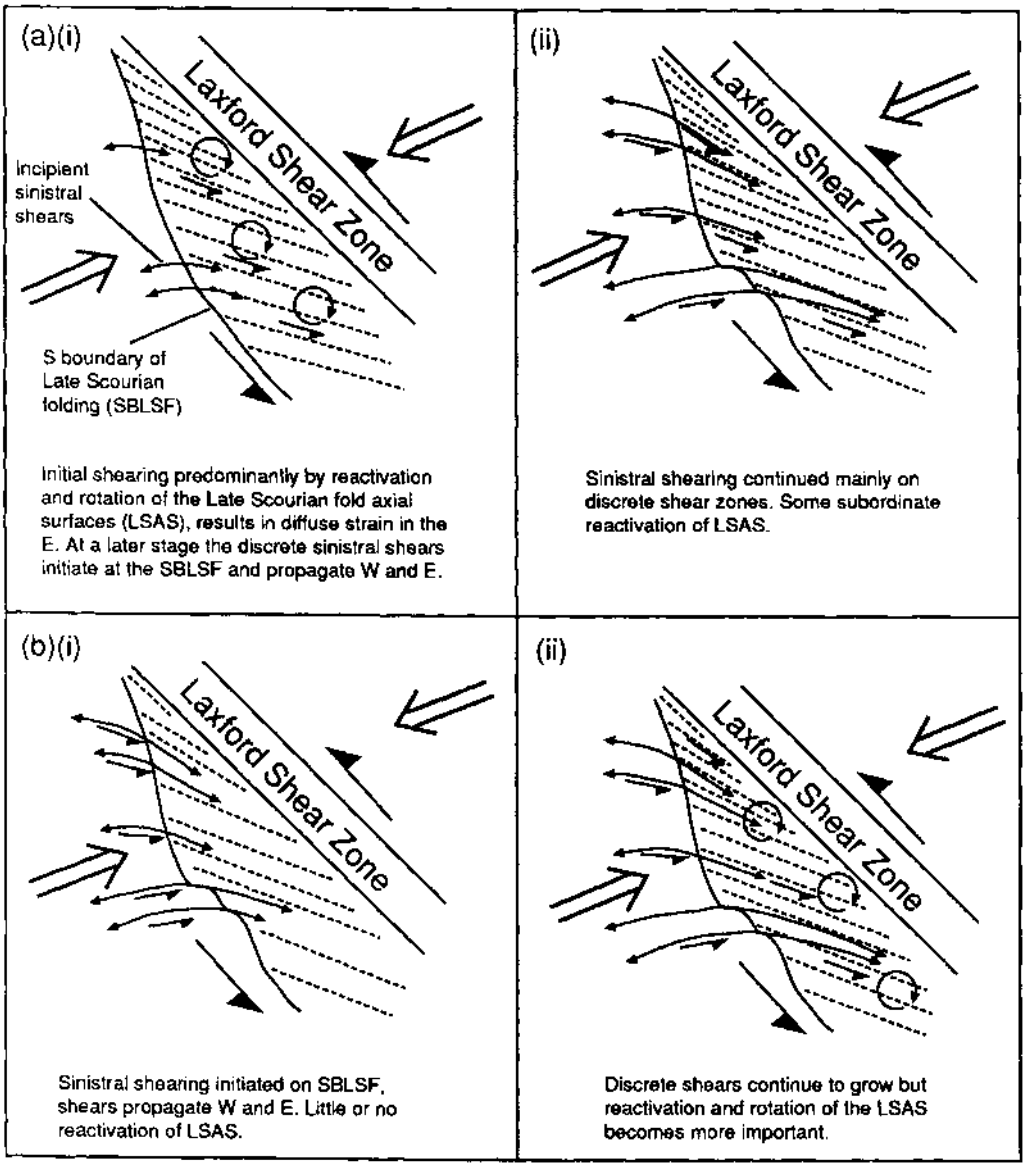

Fig. 11. Models for the formation of the sinistral shears. (**a**) Diffuse deformation followed by discrete shearing. (b) Discrete shearing followed by diffuse deformation.

sinistral shears nucleate at the boundary between the late Scourian folds and the early Scourian granulites and propagate WSW and ENE coevally as discrete shear zones (Fig. 11b(i)). Either during or after this discrete shearing, the gneisses affected by the late Scourian folding deformed by more diffuse shearing and rotation which produced the higher total displacements in zone 3 for the CCSZ (Fig. 11b(ii)).

The overall deformation of discrete shearing and block rotation is approximately equivalent to pure shear in a horizontal plane in this part of the Central Region (Jackson 1987). This deformation can also be equated to the 'stretching faults' described by Means (1989) but instead of the wall rocks to the shears deforming by penetrative pure shear, the strain is partitioned predominantly into simple shear and rotation. This is further complicated by

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the partitioning in time and space from bulk pure shear in zone 4 to very localized simple shearing with rotation in zone 1. Because there is also a decrease in displacement from zone 4 to zone 1, there is an equivalent decrease in the wall rock extension parallel to the LSZ from zone 4 to zone 1.

It is proposed that this zone of simple shear and block rotation and penetrative pure shear on the northern margin of the Central Region is equivalent to the change from simple shear and block rotation to penetrative pure shear in the middle to lower continental crust in areas of continental extension.

Mid to lower continental crust

Before a direct comparison can be made between NW Scotland and areas of continental extension it is necessary to describe the various features of the mid to lower crust via the study of exhumed orogens and extensional terranes (Fountain & Salisbury 1981; Percival *et al.* 1992) or deep seismic data from areas of continental extension. A large amount of deep seismic data from the continental crust has been collected in the last ten years, mainly from the Northern hemisphere and marine, rifted passive margin situations (e.g. BIRPS) but also onshore orogenic belts (e.g. DEKORP, ECORS and COCORP) (McGeary *et al.* 1985; Meissner *et al.* 1990; Pfiffner *et al.* 1990; Hauser *et al.* 1987).

Many of the surveys from Northwest Europe and North America have traversed extensional terranes (Mooney & Meissner 1992) which may be a result of the extensional collapse of orogenic belts or true continental rifting (McGeary *et al.* 1987; Meissner & Tanner 1993; Rey 1993). A number of features are present on many of the lines which are believed to be a direct result of the extensional deformation (Serpa & de Voogt 1987; Warner 1990; Reston 1988, 1993; Hyndmann *et al.* 1991; Mooney & Meissner 1992). These are listed below:

(i) there is generally a zone of seismically transparent upper to mid crust, also with upper crustal half-grabens with sediment infill;
(ii) there is a strongly reflective mid to lower crust where reflectors vary in length;
(iii) there is a reflection Moho at the base of the lower crustal reflectors which is usually subhorizontal and laterally continuous;
(iv) there is a transparent upper mantle, with occasional dipping reflectors.

The most consistent feature of many areas is that of lower crustal reflectivity which appears to be a feature of approximately 50% of the continental crust imaged to date, (McGeary *et al.* 1987). The possible origins of these reflectors are uncertain. Warner (1990) outlined the three most likely probabilities:

(i) basaltic intrusions;
(ii) shear zones;
(iii) Free aqueous fluids.

In some areas the upper and lower boundaries of the zone or layer of mid to lower crustal reflectors do not correspond to changes of velocity measured by refraction studies, i.e. the midcrustal Conrad discontinuity and the Moho, respectively. These differences indicate that the reflectivity is not always restricted to layers with particular velocities caused solely by compositional variations and indicates that the discontinuous reflectivity results in part from shearing and/or variations in metamorphic grade (Holbrook *et al.* 1991).

From studies of exposed 'lower crustal sections' e.g. the Ivrea zone, North Italy (Rutter & Brodie 1990; Zingg *et al.* 1990; Handy & Zingg 1991), it is generally agreed that the percentages of mafic and ultramafic gneisses increase with depth as does the metamorphic grade (Rutter & Brodie 1990; Percival *et al.* 1992). These mafic gneisses are generally interbanded with more tonalitic, granodioritic and rarely metasedimentary gneisses which would provide a strong velocity contrast and hence possible reflection surfaces in the lower crust. If this subhorizontal layering has been enhanced by pervasive pure shearing from extension or subhorizontal thrusting, then the resultant mineral alignment anisotropies can also produce strong acoustic contrasts and hence reflectivity (Mainprice & Nicolas 1989; Reston 1993; Mooney & Meissner 1992). This lithological banding may be further enhanced during extension by the intrusion of mafic sills produced by adiabatic melting of the upper mantle lithosphere. In areas of relatively recent extension, such as the Basin and Range Province, Western USA, some of these sills may still be liquid magmas (Hauser *et al.* 1987; Parsons *et al.* 1992).

Models of crustal extension from deep seismic reflection data

A number of authors have suggested models to explain the various features seen on deep seismic lines from extended terranes, particularly from BIRPS data (Reston 1990, 1993 Stein & Blundell 1990; Klemperer & Hurich 1991), but also from COCORP onshore data from the Basin and Range province (Allmendinger *et al.* 1987; Hauser *et al.* 1987). Essentially these models fall in a spectrum between lithospheric pure shear (McKenzie 1978)

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and simple shear (Wernicke 1985; Coward 1986). It is generally accepted that lithospheric pure shear is the most applicable model to the Mesozoic basins of the North Sea (cf. Klemperer & Hurich 1991).

Models from Stein & Blundell (1990), Klemperer & Hurich (1991) and Reston (1993) are basically similar and indicate that the lower crust acts as a ductile zone deforming by pure and/or simple shear to accommodate discrete faulting and block rotation in the upper crust and upper mantle. However the deformation averaged over the lithosphere is that of pure shear. Figure 12 shows a composite cartoon model of this idea. There may be variable amounts and types of shearing in the lower crust, as listed below:

(i) large scale, broad low angle zones of simple shear linking offset loci of upper crustal and upper mantle faulting;
(ii) distributed pure shear from anastomosing simple shears bounding deformable pods;
(iii) distributed pure shear extension forming non-rotational subhorizontal foliations.

Comparison of NW Scotland to deep reflection data

Figure 13 shows a simplified version of the map in Fig. 2 reflected about a N–S axis and rotated 135° anticlockwise to orientate the zone of Laxford granites to a horizontal position. This reorientation allows a comparison between Fig. 13 and the DRUM deep seismic line shown in Fig. 14 which shows upper crustal half-grabens, a reflective lower crust and a 6 km thick, bright upper mantle feature termed the Flannan reflector (McGeary & Warner 1985). The Flannan reflector exhibits strong positive reflection coefficients, which indicate that the composition is eclogitic (M. Warner pers. comm. 1993). Given that it is has an appreciable thickness, it may represent a relict subduction zone (M. Warner pers. comm. 1993). However the Flannan structure may have subsequently acted as a thrust during the Caledonian orogeny and then as an extensional fault during formation of the upper crustal half-grabens in Devonian to Triassic times (Coward *et al.* 1987).

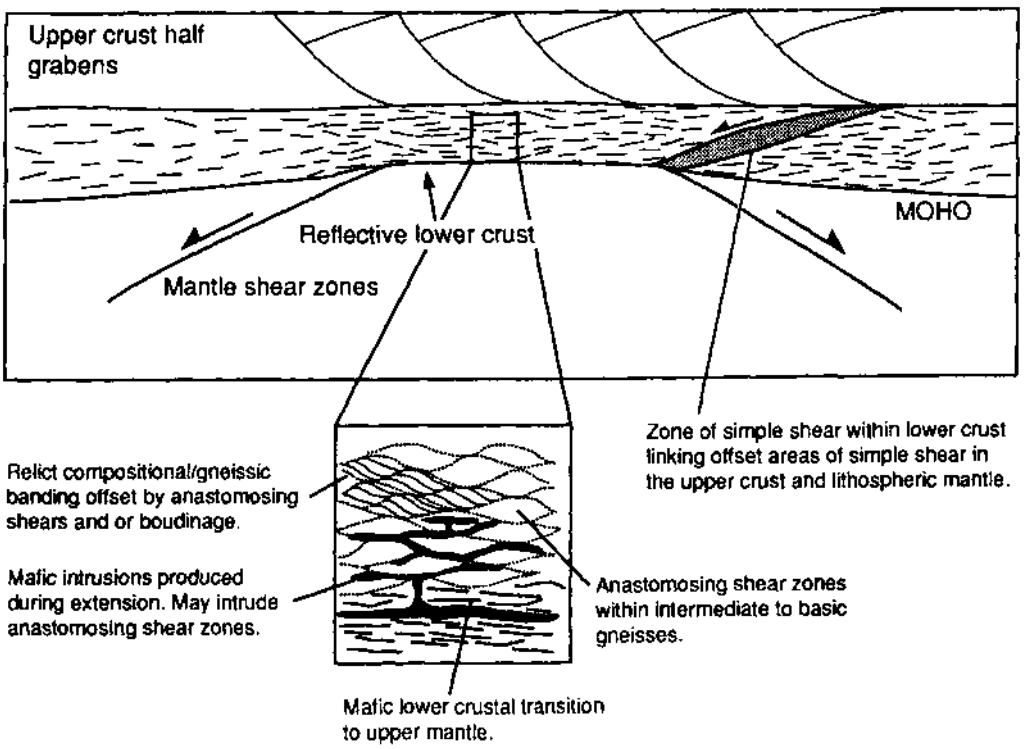

Fig. 12. Model of lower crustal shear fabrics based on BIRPS seismic reflection data. After Stein & Blundell (1990); Warner (1990); Klemperer & Hurich (1991); Reston (1993).

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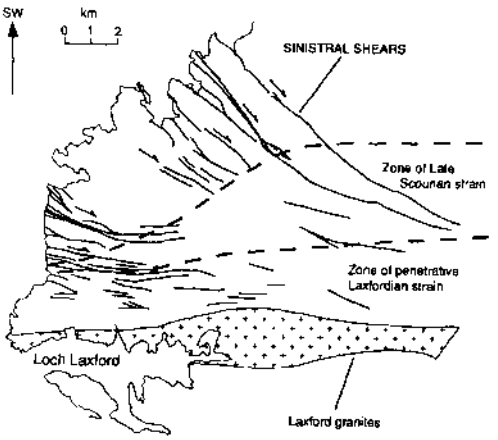

Fig. 13. Rotated and reflected version of the map in Fig. 2 to compare with Figs 12 and 14.

It is proposed that features of the sinistral shears as shown in Fig. 13 are kinematically and geometrically equivalent to the main features of the mid to lower crust in the DRUM line in Fig. 14 and the general model in Fig. 12. These features are discussed below.

(i) lower crustal anastomosing shears represented by the zone of penetratively foliated Scourie gneisses, Scourie dykes and Laxford granites on the southern margin of the LSZ to zone 4;
(ii) listric upper crustal half-graben bounding faults represented by the discrete shear zones (BASZ and CCSZ) in zones 1 and 2;
(iii) 'a brittle–ductile' transition represented by the gradual change from discrete shearing in zones 1 and 2 to more penetrative, ductile strain in zones 3 and 4.

There has been considerable strain concentration into the sinistral shears in zones 1 and 2 of the Scourie–Laxford area and it is possible that this is a function of the higher strength of the relict granulite-facies rocks causing localization of deformation, compared with the more diffuse strain in the ductile gneisses in zones 3 and 4. The higher strength relict granulite-facies rocks in zones 1 and 2 can be equated with the upper continental crust and the penetratively sheared amphibolite-facies rocks in zones 3 and 4 with the mid to lower continental crust. If the model in Fig. 11b is regarded as a cross section through the upper to lower crustal transition, then it corresponds to a situation where the shears nucleate at the brittle-plastic transition and propagate into the upper crust and mid/lower crust (cf. Jackson 1987; Jackson & White 1989; Kusznir & Park 1987). However the model in Fig. 11b indicates that later deformation is concentrated into the mid/lower crust which would result in an increase in extension with depth (Coward 1986).

An alternative model for the development of the shears is provided by Fig. 11a which can also be regarded as a cross section through the upper to lower crustal transition. In this case, the shears nucleate within the lower crust as relatively diffuse zones of shearing and propagate into the upper crust as more discrete zones. However when the brittle–plastic transition is reached the shears begin to propagate back into the lower crust as more discrete features. The later formation of discrete shears in the lower crust may be equivalent to the effective downward migration of the brittle-plastic transition during extension (England & Jackson 1987; LePichon & Chamot-Rooke 1991).

The model in Fig. 11a, as described above, implies that the extension nucleated in the lower crust is at odds with accepted models of lithospheric failure. In such models, the faults or shears nucleate at, or just above, the brittle ductile transition which is believed to be the region of highest strength and stress concentration (Kusznir & Park 1987). Several factors may be invoked to explain this possibility.

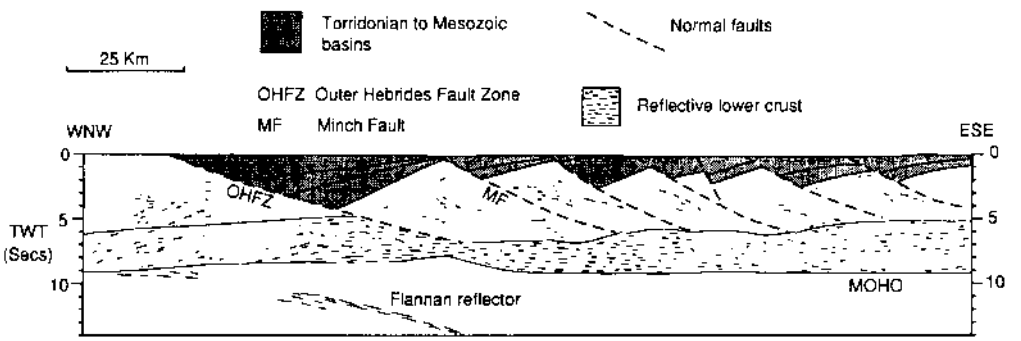

Fig. 14. Line interpretation of the DRUM deep seismic reflection profile. After McGeary *et al.* (1987).

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(i) The brittle upper crust does not equate to Byerlees friction law with depth and may be weaker than previously thought. Sub-critical (sub acoustic) crack growth may allow the initiation and growth of faults whilst also supporting geologically significant stresses (Atkinson 1987). That is, stable frictional sliding controls the deformation rather than pressure sensitive cataclasis (acoustic failure).

(ii) The plastic regime of the mid to lower crust can support higher stresses than previously thought, therefore allowing stress and strain concentration and propagation of shears upwards.

(iii) Lithospheric extension is not equal with depth and the lower crust may extend locally by a larger amount compared with the upper crust especially during the initial stages of extension (Ziegler 1983; Coward 1986). If this lower crustal extension initiates before the upper crustal extension then the shear zones may nucleate within the lower crust and propagate upwards.

Conclusions

The suite of shears studied from NW Scotland formed during flattening of a relatively competent block against the LSZ to produced shears that have cut across and modified a pre-existing rheological transition. The shears may have propagated from a region of macroscopic ductile behaviour near the LSZ into the more competent, relict Scourie granulites to the southwest. These features are kinematically and geometrically equivalent to lithospheric scale extensional shears as imaged on some deep seismic reflection profiles. It is possible that large extensional faults may propagate from the plastic lower crust up into the more competent, brittle upper crust. At present, there is no evidence to suggest that the plastic regime of the lower crust can support higher stresses than the mid to upper crust. However, shearing may initiate within the lower crust if a combination of a weak upper crust and increased extension with depth can be invoked.

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