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Eclogite-facies shear zones—deep crustal reflectors?

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Abstract

Strongly foliated eclogite-facies rocks in 30–150 m thick shear zones of Caledonian age occur within a Grenvillian garnet granulite-facies gabbro–anorthosite terrain in the Bergen Arcs of Norway. The predominant eclogite-facies mineral assemblages in the shear zones are omphacite + garnet + zoisite + kyanite in gabbroic anorthosite and omphacite + garnet in gabbro. Eclogite-facies rocks in shear zones are generally fine-grained; alternating omphacite/garnet- and kyanite/clinozoisite-rich layers define gneissic layering. A strong shape preferred orientation of omphacite, kyanite, and white mica (phengitic muscovite and/or paragonite) define the foliation. The anorthositic eclogites show omphacite *b*-axis maxima approximately normal to the foliation and *c*-axis girdles within the foliation plane. P-wave velocities (V_P) determined at confining pressures to 600 MPa for samples from eclogite-facies shear zones range from 8.3 to 8.5 km s^{-1} and anisotropy ranges from 1 to 7%. The few samples with more pronounced anisotropy tend to be approximately transversely isotropic with minimum velocities for propagation directions normal to foliation and maximum velocities for propagation directions parallel to foliation. The fast propagation direction lies within the *c*-axis girdles (parallel to foliation) and the slow propagation direction is parallel to the *b*-axis concentration (normal to foliation) in samples for which omphacite crystallographic preferred orientation was determined. V_P for the granulite-facies protoliths average about 7.5 km s^{-1}. High calculated reflection coefficients for these shear zones, 0.04–0.14, indicate that they are excellent candidates for deep crustal reflectors in portions of crust that experienced high-pressure conditions but escaped thermal reactivation.

1. Introduction

Formation of eclogite near the base of thickened continental crust or within the subcontinental mantle is a process in which geologists and geophysicists seem to have an increasing interest (e.g. Snyder and Flack, 1990; Pearson et al., 1991; Hughes and Luetgert, 1992). Crustal material may be converted to eclogite when portions of the continental crust are forced into the mantle during collisional orogenies, such as the Alpine orogeny (e.g. Butler, 1986; Laubscher, 1990; Austrheim, 1991). Laubscher (1990) cited the well-known occurrences of eclogites with continental crustal affinities within Phanerozoic orogenic belts such as the Alps (e.g. Compagnoni et al., 1977) and Caledonides (e.g. Austrheim and Mørk, 1988). Others postulate that, for various reasons, lower continental crust and its underlying mantle lithosphere delaminate from the upper crust and sink into the mantle (e.g. Bird, 1979; Kay and Kay, 1986; Turcotte, 1989). In either case, eclogi-

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tization of the crustal material is likely. Austrheim (1990, 1991) proposed that a mixture of granulite and eclogite formed as a consequence of such processes may dominate the deepest continental crust in some areas. A different mode of eclogite formation was envisioned by Furlong and Fountain (1986) and Griffin and O'Reilly (1987). They postulated that mafic magmas that cool at appropriately deep subcontinental levels will convert to eclogite if the pressure and temperature conditions are right. Rough calculations (Furlong and Fountain, 1986; Griffin and O'Reilly, 1987) indicate that the eclogite product will have seismic velocities similar to surrounding mantle peridotites making distinction between eclogitic and peridotitic mantle impossible.

We know little about the geometrical relationship between eclogites and other rocks in the deep crust or upper mantle because eclogites usually occur as tectonic lenses within lower grade rocks or as xenoliths in kimberlites and volcanic pipes. This makes realistic geophysical modelling difficult. The situation is enlightened by an exceptional metamorphic terrain in the Bergen Arcs of southern Norway (Fig. 1) where Grenvillian granulite-facies rocks were partially converted to

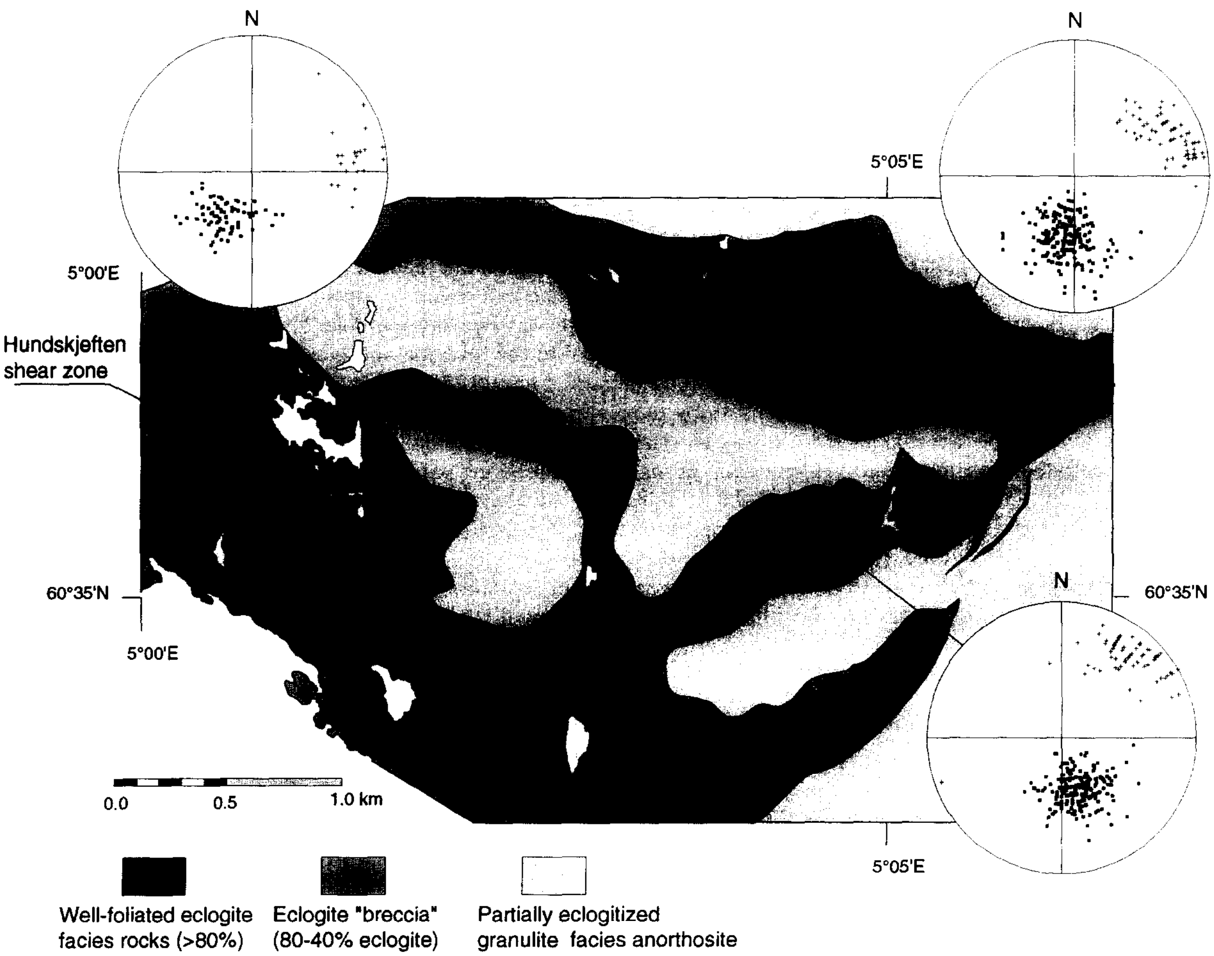

Fig. 1. Geologic map of a portion of Holsnøy (after Boundy et al., 1992b) showing eclogite-facies shear zones and key structural data on stereographic projections (● = poles to foliation; + = lineations). Structural data from Boundy et al. (1992b) were augmented by new data collected during the summer of 1992. Topography is not shown so the widths of outcrop are not related to true thickness because of possible topographic effects. A detailed version of this map is available in Boundy (1990).

eclogite-facies rocks during the Caledonian orogeny (Austrheim and Griffin, 1985). The eclogite-facies rocks occur 30–150 m thick shear zones surrounded by uneclogitized to partially eclogitized granulite-facies gabbroic anorthosites. The terrain, therefore, represents an example of deep continental crust partially converted to eclogite-facies during continental collision in the sense envisioned by Butler (1986) and Laubscher (1990) and affords an opportunity to study the variation of physical properties associated with this process.

Austrheim and Mørk (1988) suggested that eclogite shear zones could be deep crustal reflectors because of their juxtaposition against lower-velocity granulite-facies rocks. Austrheim (1990) further proposed that a partially eclogitic lower crust could explain transitional crust–mantle velocities. To evaluate these hypotheses we measured compressional wave velocities (V_P) and densities of rock samples collected along a traverse across the shear zones and flanking granulite-facies gabbroic anorthosites (Fig. 1). The data illustrate that, under certain circumstances, eclogitic shear zones could generate significant reflectors in the deep crust and significantly increase the average velocity of deep continental crust.

2. Geologic background

The Bergen Arcs of western Norway are situated within the Caledonian orogen, which is characterized by a variety of thrust-sheets displaced from west-to-east onto the Baltic Shield (Roberts and Gee, 1985). One of the arcuate Caledonian nappes of the Bergen Arcs consists of a Proterozoic granulite-facies anorthosite complex that originated as a layered intrusion composed of anorthosite to gabbroic anorthosite (Austrheim, 1987), which was intruded by jotunite (monzonorite) and mangerite (orthopyroxene monzonite) (Kolderup and Kolderup, 1940; Griffin, 1972; Austrheim and Griffin, 1985).

Two main phases of deformation and metamorphism recognized in the Bergen Arcs anorthosite complex are especially well-illustrated on NW Holsnøy (situated north of the city of Bergen) where granulite-facies anorthosite, gabbroic anorthosite and gabbro layers are intruded by mafic mangerite and jotunite (Austrheim and Mørk, 1988). Garnet–granulite-facies metamorphism and deformation of the complex occurred at mid- to lower-crustal levels during the Proterozoic Grenvillian (Sveconorwegian) orogeny (907 ± 9 Ma) at pressure and temperature conditions of 1000 MPa and 850°C (Austrheim and Griffin, 1985; Cohen et al., 1988). The second phase of deformation and metamorphism occurred in response to continental collision and crustal thickening during the early Paleozoic Caledonian orogeny that brought the granulite-facies rocks into eclogite-facies conditions (Austrheim and Griffin, 1985). The terrain experienced a localized Caledonian deformation and eclogite-facies metamorphism (Austrheim and Griffin, 1985; Austrheim, 1987; Austrheim and Mørk, 1988; Jamtveit et al., 1990). Recent $^{40}Ar/^{39}Ar$ dating of phengites from the eclogites (Boundy et al., 1992a) suggests the terrain cooled through 350°C between 460 and 440 Ma. Metamorphic pressure and temperature conditions were in the range 1600 to 2100 MPa and 700 to 750°C (Austrheim and Griffin, 1985; Jamtveit et al., 1990). However, not all the granulite-facies rocks converted to eclogite. The occurrence of hydrous eclogite-facies minerals along veins and within the eclogites indicates that eclogite-facies metamorphism along shear zones was promoted by fluid infiltration during metamorphism (Austrheim and Griffin, 1985; Austrheim, 1987; Andersen et al., 1990, 1991; Boundy et al., 1992b).

The major exposures of eclogites on Holsnøy are within anastomosing, subparallel, 30–150 m thick shear zones that are laterally continuous along strike over distances up to several kilometers (Austrheim, 1987; Austrheim and Mørk, 1988; Boundy, 1990; Klaper, 1990, 1991; Boundy et al., 1992b). Boundy (1990) mapped three major eclogite shear zones on NW Holsnøy (Fig. 1): the Hundskjeften shear zone (HSZ), the lower Eldsfjell shear zone (LESZ), and the upper Eldsfjell shear zone (UESZ). The zones are composed of more than 80% eclogite and contain a few blocks of granulite-facies rocks. They have strong folia-

Table 1
Structural data for eclogite shear zones

Structure	UESZ	LESZ	HSZ
Foliation pole trend	185°	162°	217°
Foliation pole plunge	52°	58°	56°
Foliation strike	275°	252°	307°
Foliation dip [a]	38°	32°	34°
Lineation trend	77°	39°	84°
Lineation plunge	11°	21°	26°

[a] Right-hand dip rule.

tions that have a general N–NE dip (Fig. 1 and Table 1). The upper boundaries are defined by marked increases in mylonitic foliation intensity and the bases of the shear zones are gradients where shear-zone fabric loses intensity downward over 10 m intervals.

The layering within the shear zones is defined by alternating omphacite/garnet- and kyanite/clinozoisite-rich layers at outcrop scale and foliation, which is parallel to layering, is defined by the shape preferred orientation of omphacite. The eclogites are generally fine-grained (0.1–0.5 mm). Lineations consist of rod-shaped mineral aggregates, elongate relict corona structures, and mineral lineations that trend to the NE or ENE with gentle plunges (Fig. 1 and Table 1). Analysis of kinematic indicators (e.g. sigmoid-type porphyroblasts, S–C′ features, etc.) demonstrate that the hanging wall of the shear zones moved to the NE or ENE with a normal sense of movement as postulated by Boundy et al. (1992b). Universal stage measurements of omphacite crystallographic axes (Boundy et al., 1992b) show that omphacite *b*-axis maxima are approximately normal to the foliation and *c*-axis girdles within the foliation plane. Weak *c*-axis maxima within the girdles are approximately parallel to the lineation.

The shear zones are surrounded by an eclogite breccia (Fig. 1) that consists of granulite- and eclogite-facies rocks in about equal proportion. The breccia is characterized by anastomosing zones of strongly foliated eclogite that enclose and wrap around areas of granulite-facies rocks, which occur as angular, lensoidal blocks typically less than 5 m across. These breccia zones are in turn flanked by anorthositic to gabbroic anorthosite granulite-facies rocks (Fig. 1). The meta-anorthositic rocks are massive to well-layered and are generally coarse grained (1–3 mm) with equigranular and granoblastic textures. The granulite-facies rocks show partial conversion to eclogite, but there are areas in excess of several hundred square meters where the granulite is preserved with no eclogite-facies overprint.

3. Sample description

We collected oriented samples along a traverse of the major eclogite-facies shear zones and surrounding granulites (Fig. 1, Tables 2–4). Major element chemical analyses were determined on powders from each sample and standard point count methods were used to estimate modes. Microprobe analyses were performed for a few samples to determine mineral composition (Boundy et al., 1992b). Petrofabric analyses of the crystallographic preferred orientation of omphacite (see above) for a few selected samples are also reported in Boundy et al. (1992b).

Granulite-facies gabbroic anorthosite, the presumed protolith for the eclogites in the shear zones, typically consists of plagioclase, diopside, garnet ± hypersthene ± scapolite ± green spinel ± hornblende. In thin section, the plagioclase may contain zoisite needles and white mica, and pyroxene shows fine-grained rims of undetermined minerals, probably omphacite. These textures are interpreted to result from incipient reactions under eclogite-facies conditions. The predominant mineral assemblage in gabbroic anorthosite in the eclogite-facies shear zones consists of omphacite (Jd_{44-49}), garnet (predominantly $Pyr_{38}Alm_{40}Gross_{21}$), zoisite, kyanite with minor Na-rich phengitic muscovite ± rutile ± quartz ± amphibole. Minor gabbroic eclogites consist of predominantly of omphacite (Jd_{39}), garnet ($Pyr_{26}Alm_{48}Gross_{26}$), with minor phengitic muscovite, rutile, quartz ± carbonate.

The eclogite-facies samples show some geochemical diversity that may be related to variable protolith composition and/or reflect compositional changes associated with the eclogite conversion process. To illustrate this, we compare samples used in this study to a reference sample (BA-31). Sample BA-31 was chosen as a reference because of its proximity to the shear zones, its lack of textures indicative of incipient reactions, and its compositional similarity to a gabbroic anorthosite collected at Indre Arna, 30 km southeast of Holsnøy. When compared to BA-31 (Fig. 2), the other gabbroic anorthosite granulites separate into three groups: those very similar to BA-31 (BA-24, BA-32, BA-38), two samples

Table 2
Compressional wave velocity data for granulite facies gabbroic anorthosites

Sample No.	ρ ($g\ cm^{-3}$)	V_p 0 MPa ($km\ s^{-1}$)	$\partial V_p/\partial P$ ($\times 10^{-4}$ km $s^{-1}\ MPa^{-1}$)	V_p 600 MPa ($km\ s^{-1}$)	V_p 1000 MPa ($km\ s^{-1}$)
BA-23A[1]	3.032	7.314	3.840	7.544	7.698
BA-23B[1]	3.042	7.497	2.692	7.659	7.766
BA-23C[1]	3.057	7.409	4.283	7.666	7.838
BA-23M[1]	3.044	7.407	3.605	7.623	7.767
BA-24A	2.999	7.457	1.678	7.648	7.715
BA-24B	2.950	7.344	3.403	7.548	7.684
BA-24C	2.895	7.302	2.418	7.447	7.544
BA-24M	2.948	7.368	2.500	7.548	7.648
BA-26A	2.793	7.155	2.587	7.310	7.414
BA-26B	2.788	7.073	1.624	7.170	7.235
BA-26C	2.788	7.381	2.934	7.557	7.674
BA-26M	2.789	7.203	2.382	7.346	7.441
BA-31A	2.942	7.350	2.541	7.502	7.604
BA-31B	2.943	7.661	3.358	7.863	7.997
BA-31C	2.889	7.550	4.705	7.837	8.025
BA-31M	2.925	7.520	3.535	7.734	7.875
BA-32A	2.998	7.263	3.162	7.452	7.579
BA-32B	3.021	7.350	3.279	7.547	7.678
BA-32C	3.013	7.349	4.554	7.623	7.805
BA-32M	3.011	7.321	3.665	7.541	7.687
BA-33A	3.068	7.542	3.425	7.748	7.885
BA-33B	3.078	7.886	4.500	8.156	8.336
BA-33C	3.115	7.594	4.310	7.853	8.025
BA-33M	3.087	7.674	4.078	7.919	8.082
BA-38A	2.864	6.995	3.513	7.206	7.347
BA-38B	2.878	7.182	4.302	7.440	7.612
BA-38C	2.876	7.375	1.379	7.458	7.513
BA-38M	2.873	7.184	3.065	7.368	7.491
BA-42A	2.806	7.017	3.055	7.201	7.323
BA-42B	2.842	7.061	3.517	7.278	7.419
BA-42C	2.814	7.047	4.094	7.292	7.456
BA-42M	2.821	7.042	3.555	7.257	7.399

[1] Abbreviations: *A* = propagation normal to foliation; *B* = propagation parallel to foliation and perpendicular to lineation; *C* = propagation parallel to foliation and parallel to lineation; *M* = mean.

poorer in Fe_2O_3 and MgO (BA-26 and BA-42), and two samples higher in Fe_2O_3 and MgO (BA-23 and BA-33). The samples less mafic than BA-31 are more feldspathic (see Boundy et al., 1992b). Mean atomic weight for the granulite samples ranges from 21.4 to 21.7, with the higher mean atomic weight corresponding to the more mafic samples. Comparing the LESZ eclogite-facies samples to BA-31 (Fig. 2), we observe that most of the samples, as well as a few samples from the UESZ, are similar to the latter group. Three samples from the UESZ (BA-2, BA-6, and BA-12), however, are much more enriched in Fe_2O_3 and MgO than other shear-zone eclogites. Mean atomic weight for LESZ samples ranges from 21.6 to 21.8 but sample BA-22, a gabbroic composition eclogite, has a mean atomic weight of 22.9. UESZ samples have mean atomic weights between 21.5 and 22.1; the more ferromagnesian samples have the higher values.

Table 3
Compressional wave velocities for Lower Eldsfjell Shear Zone eclogites

Sample No.	ρ ($g\ cm^{-3}$)	V_p 0 MPa ($km\ s^{-1}$)	$\partial V_p/\partial P$ ($\times 10^{-4}$ km s^{-1} MPa^{-1})	V_p 600 MPa ($km\ s^{-1}$)	V_p 1000 MPa ($km\ s^{-1}$)
BA-13A[1]	3.238	7.758	3.236	7.953	8.082
BA-13B[1]	3.310	8.224	3.120	8.411	8.536
BA-13C[1]	3.289	8.105	3.977	8.343	8.502
BA-13M[1]	3.279	8.029	3.444	8.236	8.373
BA-20A	3.332	7.995	1.127	8.063	8.108
BA-20B	3.352	8.379	1.272	8.455	8.506
BA-20C	3.290	8.203	2.618	8.360	8.465
BA-20M	3.324	8.192	1.672	8.293	8.360
BA-21A	3.340	7.675	6.109	8.041	8.286
BA-21B	3.333	8.148	7.223	8.582	8.871
BA-21C	3.338	8.389	3.384	8.592	8.727
BA-21M	3.337	8.071	5.572	8.405	8.628
BA-22A	3.580	8.144	2.260	8.279	8.370
BA-22B	3.557	8.080	2.882	8.253	8.369
BA-22C	3.561	8.161	3.288	8.359	8.491
BA-22M	3.566	8.128	2.810	8.297	8.410
BA-39A	3.392	8.016	6.244	8.390	8.640
BA-39B	3.371	8.164	5.790	8.511	8.743
BA-39C	3.364	8.095	5.538	8.427	8.649
BA-39M	3.376	8.092	5.857	8.443	8.677
BA-40A	3.358	7.973	4.751	8.258	8.448
BA-40B	3.351	8.317	3.816	8.546	8.699
BA-40C	3.295	8.095	5.538	8.427	8.649
BA-40M	3.335	8.128	4.702	8.410	8.599
BA-41A	3.421	8.126	2.283	8.263	8.354
BA-41B	3.325	8.091	5.012	8.392	8.593
BA-41C	3.341	8.184	4.658	8.464	8.650
BA-41M	3.362	8.134	3.984	8.373	8.532

[1] For legend see Table 2.

4. Laboratory methods and results

Compressional-wave velocities were measured to confining pressures of 600 MPa for three mutually orthogonal cores (normal to foliation, perpendicular to lineation, and parallel to lineation) cut from each sample using the equipment and methods described in Fountain et al. (1990). Densities (ρ) were determined for each core based on its mass and its volume (determined with a helium pycnometer). Results are reported here for granulite-facies gabbroic anorthosites (Table 2), LESZ eclogites (Table 3), and UESZ eclogites (Table 4). Table 2 also includes data for gabbroic anorthosites collected from other areas on the island and from a distant site at Indre Arna southeast of Holsnøy. In these tables we report the density data for each core and the parameters of a line fit to the V_P-pressure data between 400 and 600 MPa. Pressure derivatives are typically in the range $1\text{–}3 \times 10^{-4}$ km s^{-1} MPa^{-1} and are comparable to derivatives determined for similar rocks between 400 and 1000 MPa (Christensen, 1974; Manghnani et al., 1974). Extrapolated values for 1000 MPa are also listed in these tables.

Shear-zone eclogites are variably anisotropic. P-wave anisotropy ranges from 1 to 6.5% and averages about 3% for all samples studied (Tables 3 and 4). Such low values seem to be typical for eclogites (Fountain and Christensen, 1989).

Table 4
Compressional wave velocities for Upper Eldsfjell Shear Zone eclogites

Sample No.	ρ (g cm^{-3})	V_P 0 MPa (km s^{-1})	$\partial V_P/\partial P$ (×10^{-4} km s^{-1} MPa^{-1})	V_P 600 MPa (km s^{-1})	V_P 1000 MPa (km s^{-1})
BA-02A[1]	3.400	8.306	2.053	8.429	8.511
BA-02B[1]	3.405	8.713	2.570	8.867	8.970
BA-02C[1]	3.379	8.758	2.846	8.929	9.043
BA-02M[1]	3.395	8.592	2.490	8.742	8.841
BA-03A	3.252	7.945	2.247	8.080	8.170
BA-03B	3.269	8.192	1.878	8.304	8.379
BA-03C	3.211	7.996	1.863	8.108	8.182
BA-03M	3.244	8.044	1.996	8.164	8.244
BA-04A	3.238	8.037	2.383	8.180	8.275
BA-04B	3.183	8.100	2.226	8.234	8.326
BA-04C	3.317	7.814	5.915	8.168	8.405
BA-04M	3.246	7.984	3.508	8.194	8.335
BA-06A	3.420	8.166	2.879	8.339	8.454
BA-06B	3.432	8.330	2.526	8.482	8.583
BA-06C	3.397	8.260	2.264	8.396	8.487
BA-06M	3.416	8.252	2.556	8.406	8.508
BA-08A	3.277	8.115	1.781	8.222	8.293
BA-08B	3.237	8.006	1.589	8.101	8.165
BA-08C	3.256	8.035	1.436	8.121	8.178
BA-08M	3.256	8.052	1.602	8.148	8.212
BA-12A	3.348	8.102	2.498	8.167	8.269
BA-12B	3.306	8.135	1.932	8.251	8.328
BA-12C	3.334	8.201	2.304	8.339	8.432
BA-12M	3.329	8.146	2.245	8.252	8.343

[1] For legend see Table 2.

We expected greater anisotropy for these eclogites in light of the strong macroscopic fabric, but there is no significant difference in anisotropy between the eclogites and the gabbroic anorthosite protolith (Table 2). Those eclogite samples with more pronounced anisotropy tend to be approximately transversely isotropic with minimum velocities for propagation directions normal to foliation and maximum velocities for propagation directions parallel to foliation. The fast propagation direction lies within the *c*-axis girdles (parallel to foliation) and the slow propagation direction is parallel to the *b*-axis concentration (normal to foliation) in samples for which omphacite crystallographic preferred orientation was determined. V_P is typically high parallel to the *c*-axes and slow parallel to the *b*-axes in the diopside–omphacite–jadeite series pyroxenes (Levien et al., 1979; Kandelin and Weidner, 1988; Bhagat et al., 1992). This pattern would reinforce any anisotropy pattern induced by the micas, which also have a strong shape preferred orientation.

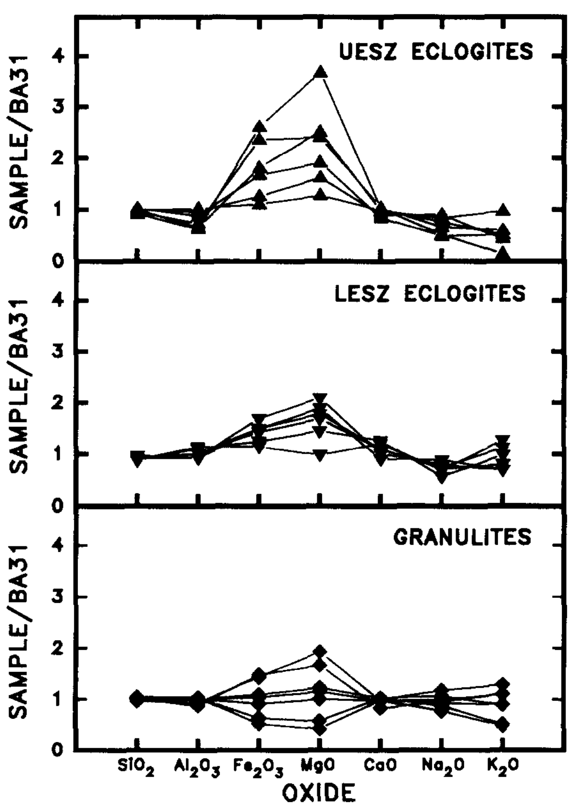

Fig. 2. Variation of major oxide compositions, normalized to the composition of BA-31, for gabbroic anorthosite granulites (♦), UESZ eclogites (▲) and LESZ eclogites (▼).

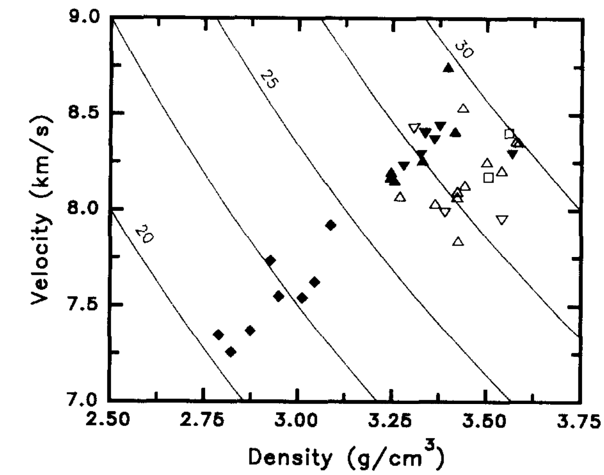

Fig. 3. Compressional-wave velocity measured at 600 MPa vs. density for granulite-facies gabbroic anorthosites (♦), UESZ eclogites (▲) and LESZ eclogites (▼). Also shown are 600 MPa data for eclogites from Christensen (1974) (□), Manghnani et al. (1974) (△) and Kumazawa et al. (1971) (▽). Curves are contours of constant acoustic impedance ($\times 10^{-6}$ kg m^{-2} s^{-1}).

The granulite to eclogite transition, as represented by these samples, exhibits a linear velocity–density trend (Figs. 3 and 4) that can be related to mineralogical changes associated with the transition. The granulites define a linear array between 2.75 and 3.1 g cm^{-3}; the lower density samples are more feldspathic and the higher density samples are more garnetiferous. The mineralogical controls on this V_P trend can be understood in terms of component mineral properties (Fig. 4). Plagioclase-rich samples have V_P and ρ similar to plagioclase and V_P and ρ systematically increase as clinopyroxene and garnet become more abundant (Fig. 4).

The eclogite-facies rocks, with few exceptions, cluster in a tight linear array where the differences in density and velocity are related to the volume of omphacite and garnet relative to other phases (Figs. 3 and 4). Velocities are lower than anticipated for an ideal garnet–omphacite mixture because of the influence of other phases, in

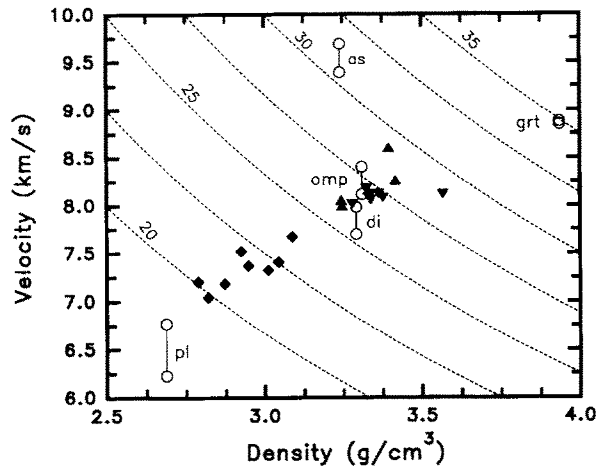

Fig. 4. Compressional-wave velocity extrapolated to 0 MPa vs. density for granulite-facies gabbroic anorthosites (◆), UESZ eclogites (▲) and LESZ eclogites (▼) compared to single crystal Voigt (high) and Reuss (low) averages (○) for the primary constituent minerals of plagioclase (*pl*), diopside (*di*), omphacite (*omp*), garnet (*grt*) and aluminosilicate (*as*). Appropriate values for plagioclase, garnet and omphacite solid solutions were linearly interpolated from the stiffness tensors of the end-member components. Sillimanite was used as a substitute for kyanite because single crystal data for kyanite are not available and the close similarity of the Voigt and Reuss averages for sillimanite and andalusite suggests that kyanite properties could be reasonably approximated by the other aluminosilicates. Single crystal data were based on room temperature, atmospheric pressure measurements of Ryzhova (1964), Vaughan and Weidner (1978), Kandelin and Weidner (1988), Bass (1989), O'Neill et al. (1989) and Bhagat et al. (1992). Curves are the same as in Fig. 3.

particular quartz, white mica, and zoisite (see Boundy et al., 1992b, table 1). One sample, BA-2, has a much higher V_P than the other samples owing to abundant kyanite. Although no single crystal elastic data for kyanite have been reported in the literature, both andalusite and sillimanite have similar velocities that are much higher than other silicates and thus we would expect that kyanite has a similarly high V_P. The velocity–density trend for this group of eclogites does not follow trends observed for most other eclogites reported in the literature due to the anorthositic composition of this particular suite. A few eclogites from other studies that overlap with this suite also have gabbroic anorthositic compositions. Sample BA-22 deviates significantly from this trend because it has a more mafic (gabbroic) composition and a higher mean atomic weight. It follows a trend defined by most of the eclogites reported in the literature, which tend to be mafic with mean atomic weights between 22 and 23.

5. Discussion

The data presented here can be used to assess Austrheim's hypotheses that eclogite-facies shear zones in the deep continental crust could cause seismic reflections and high crust–mantle transition velocities. We constructed four simple models of these zones (Figs. 5 and 6) using measured thicknesses along the traverse across the UESZ and LESZ and by rotating the zones to horizontal. In two cases (Fig. 5A and B) we assumed that the properties of the bounding breccia could be approximated by a 50% eclogite–50% granulite mixture. In one of these cases we used the average V_P and ρ for all the granulites (Fig. 5A) and for the other we used the lowest granulite V_P and ρ. The other two cases (Fig. 6A and B) use the same granulite properties but assume the breccia is all granulite and thus represent the extreme case of granulite juxtaposed directly against eclogite.

In the multi-layer case (Fig. 5A and B) the layer thickness ranges from 30 to 250 m and averages 106 m. This average layer thickness is slightly larger than Hurich and Smithson (1987) argued is necessary for reflection amplitude enhancement due to constructive interference, but the thinner layers in the sequence are similar to this optimal thickness. Normal-incidence reflection coefficients (absolute value) calculated using standard equations given in Telford et al. (1982) range from 0.04 to 0.07, depending on the nature of the bounding granulite. The more extreme cases (Fig. 6A and B) exhibit reflection coefficients in the range of 0.09 to 0.14 with an average layer thickness of 88 m, a value closer to the optimal conditions outlined by Hurich and Smithson (1987).

Although the combination of the relatively high reflection coefficients and layer thicknesses may be conducive to generation of deep crustal reflectors, the dimensions of the shear zones may be a mitigating factor. Many shear zones are only 1–3 km long, but others must be longer as they are truncated by the coast. Assuming shear zones such as these exist at 50 km depth in the crust, we estimate Fresnel zone dimensions (see Fowler, 1990) of 4–8 km for frequencies between 10 and 40 Hz and a velocity of 7 km s^{-1}. Thus it appears

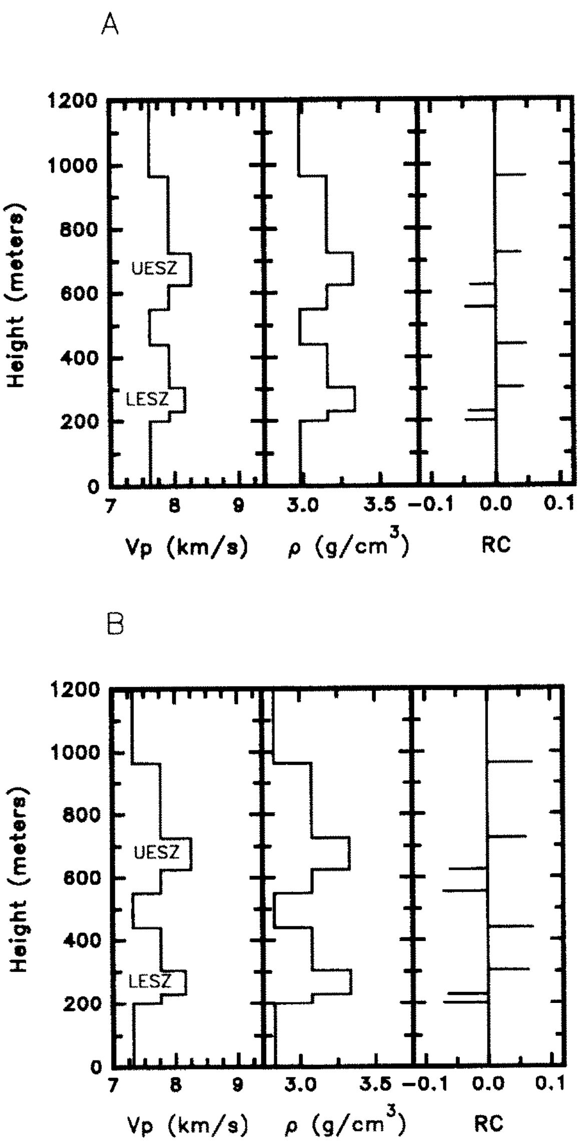

Fig. 5. (A) Variation of V_P, ρ, and reflection coefficients through the UESZ and LESZ assuming the breccia zone is a 50% eclogite–50% granulite mixture and the granulite V_P is approximated by the average of all the granulites. (B) Same as (A) except the granulite V_P is represented by the lowest V_P granulite.

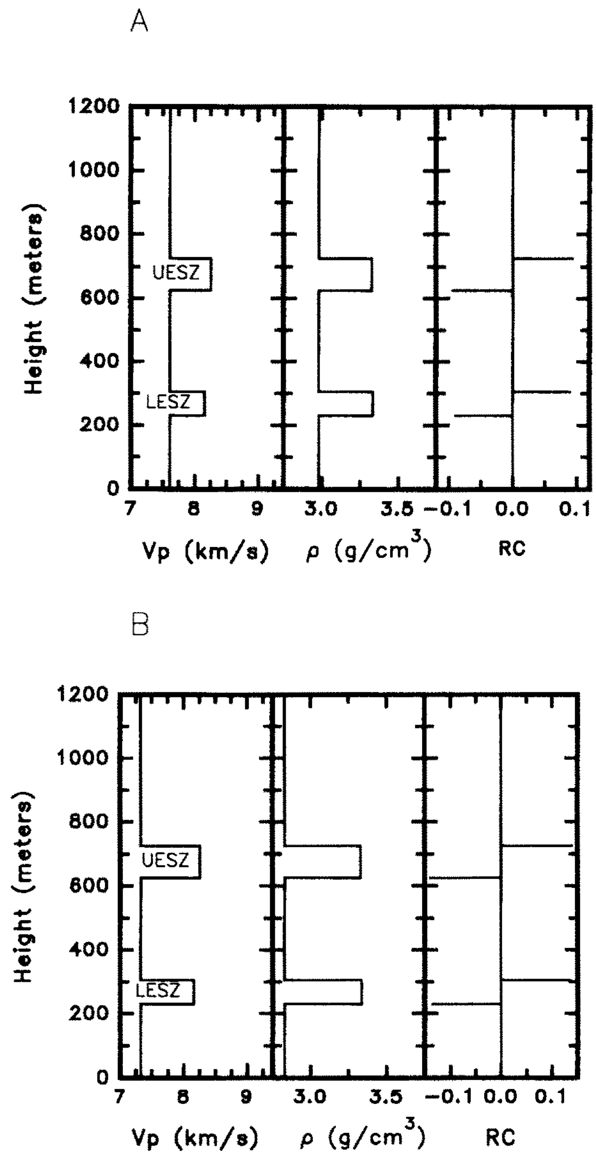

Fig. 6. (A) Variation of V_P, ρ, and reflection coefficients through the UESZ and LESZ assuming the breccia zone horizon is a 100% granulite and the granulite V_P is approximated by the average of all the granulites. (B) Same as (A) except that the granulite V_P is represented by the lowest V_P granulite studied.

that individual shear zones may be too short to image in near-vertical incidence seismic profiles. However, these zones form an anastomosing network across the island that, taken as a whole, has an effective length greater than the Fresnel zone range for the deep crust.

To assess the average properties of this terrain it is useful to examine the eclogite breccia unit as a scale model of a lower crust dominated by geology of the type mapped on Holsnøy. The zone is made up of eclogitic shear zones bounding blocks of granulite-facies rocks that are in various stages of conversion to eclogite. By estimating that the breccia zone is about half eclogite and half granulite we calculate an average V_P of 7.7–7.9 km s^{-1} at 600 MPa, depending on which granulite we use as an end-member. Assuming the range of pressure and temperature gradients summarized in Fountain and Christensen (1989), we expect these same velocities at the base of the crust after metamorphism. These numbers are clearly in the range of transitional velocities reported from deep crustal wide-angle surveys (e.g. Luosto and Korhonen, 1986; Boland and Ellis, 1989) and we therefore argue that terrains like Holsnøy may exist in the deep crust.

6. Conclusions

The objective of this study was to evaluate the ideas put forward by Austrheim and Mørk (1988) and Austrheim (1991) that eclogite-facies shear zones formed in the deep crust of a collisional orogenic belt might be seismic reflectors and might constitute zones characterized by velocities transitional between crustal and mantle velocities. The shear zones are bounded by granulite-facies rocks that have an average V_P of about 7.5 km s^{-1} at 600 MPa confining pressure. P-wave velocities determined at confining pressures to 600 MPa for samples from eclogite-facies shear zones range from 8.3 to 8.5 km s^{-1} and anisotropy is low to modest (1–7%). Average layer thicknesses of the zones are close to the optimal conditions for enhanced reflectivity due to constructive interference (Hurich and Smithson, 1987) and the zones are typified by high reflection coefficients (0.04–0.14). Individual shear zones, however, have length scales less than the Fresnel zone for the lower crust suggesting that they may not be imaged on near-vertical incidence seismic profiles. However, these zones form an anastomosing network across the island that, taken as a whole, has an effective length greater than the Fresnel zone range for the deep crust. If considered on a larger scale, the terrain approximates a mixture of 50% eclogite with 50% granulite and should give an average of 7.7–7.9 km s^{-1}. This velocity would be sensed by standard refraction surveys and wide-angle experiments and is close to values transitional between the crust and mantle recorded in some regions.

We propose that terrains similar to Holsnøy may be prevalent components of the deep crust or uppermost mantle in regions that experienced crustal shortening and fluid infiltration in their evolution provided that there has been no significant thermal overprint following high pressure metamorphism. Thermal models predict that the $P–T$ history experienced by terrains such as the one discussed here experience heating after convergence (e.g. Thompson and England, 1984). This implies that the eclogites should be transient features. Although this may hold for many orogenic belts, we envision that this thermal relaxation does not apply to all contractional orogenic systems thus favoring preservation of rocks like those exposed on Holsnøy. For example, the high pressure event related to the Caledonian convergence was closely followed by extension in the early Devonian (e.g. Lux, 1985; Kullerud et al., 1986; Chauvet and Dallmeyer, 1992). Exhumed eclogitic rocks in the Western Gneiss Region experienced decompression into the amphibolite-facies without significant temperature change (Andersen and Jamtveit, 1990; Chauvet et al., 1992; Cuthbert and Wilks, 1992). This indicates that the deep crust resided under eclogite-facies conditions when extension began and the consequent extension occurred at relatively low temperatures. The absence of thermal relaxation in the Scandinavian Caledonides is confirmed by the fact that the exhumed terrains preserve Precambrian structures (e.g. Powell et al., 1988) and by the near absence of synextensional migmatitic rocks and crustally-derived granitoid plutons (Stephens, 1988). Our argument receives further support from the recognition of eclogite-facies overprints on deep crustal rocks in other regions and increased documentation that fluid infiltration and deformation play an important role in the kinetics of the gabbro–eclogite transition (e.g. Blattner, 1976; Compagnoni et al., 1977; Koons et al., 1987; Philippot, 1987; Sanders, 1988; Wilkerson et al., 1988; Biino and Pognante, 1989; Philippot and Kienast, 1989; Lardeaux and Spalla, 1991; Philippot and Selverstone, 1991; Ellis and Maboko, 1992; Selverstone et al., 1992; Indares, 1993).

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