# Seismic reflection coefficients from mantle fault zones

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**Summary.**
Several bright reflections from structures within the mantle can be seen on BIRPS' deep seismic reflection profiles. We have calculated apparent reflection coefficients for the brightest of these events and obtain values around 0.1. It is not possible to produce such large reflections by either compositional layering or seismic anisotropy if olivine and pyroxene are the only significant minerals in the mantle. These large reflections can be produced by a mafic layer or a partially hydrated layer within normal peridotite. The brightest reflections seem to be best explained as major faults or shear zones within the mantle.

## 1. Mantle reflections

Several of the British Institutions Reflection Profiling Syndicate (BIRPS) deep seismic reflection profiles show reflections originating from within the upper mantle. The DRUM line (Fig. 1) was shot in 1984 to 30 s two-way-time (about 110 km depth) to examine mantle reflections. On this profile two significant sub-crustal reflectors can be seen.

The first appears as a dipping sequence of reflections running from near the base of the crust down to a two-way-time (TWT) of at least 28 s. This event was labelled the "Flannan Thrust" by Smythe *et al.* (1982) when it was first observed on the MOIST line which runs sub-parallel to DRUM. It has since been recorded on several deep reflection lines in this area (Brewer *et al.* 1983). On these adjacent lines it can be seen to cut and possibly offset the Moho. It subsequently flattens out in the lower crust and does not appear to cut through the middle and upper crust.

The second event (the W-reflector) consists of a packet of sub-horizontal reflections lying between 13 and 15 s TWT, running east from the Flannan Thrust for about 100 km to the edge of the profile. Both events are described in more detail by McGeary & Warner (1985).

Cross-lines in this area show that the Flannan Thrust is indeed located within the mantle and cannot be caused by side-swipe from a feature within the crust. As yet we have no cross-line control for the W-reflector. Amplitude modelling of the raw records shows that it is very unlikely that either event is due to a converted S-wave reflection from, for example, the Moho.

## 2. Reflection coefficients

In an attempt to limit the range of acceptable geological explanations for these bright mantle reflections we have tried to determine their absolute reflection strength. The most useful procedure for limiting possible explanations is to estimate the minimum equivalent reflection coefficient for the brightest reflections. In order to calculate reflection coefficients
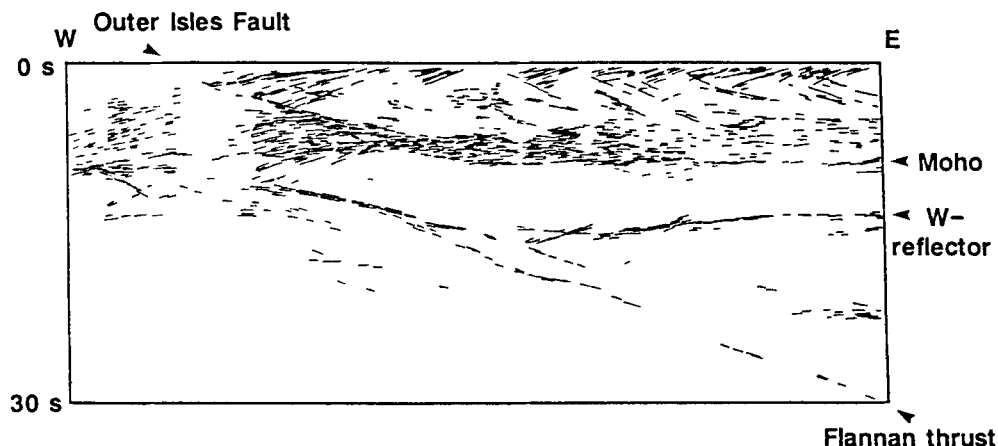
**Figure 1.** Unmigrated line drawing of the DRUM deep reflection line.

one must work with unprocessed, unstacked raw data, calibrate the acquisition system, determine the velocity structure, and estimate the cumulative losses due to anelastic absorbtion, scattering and multiple reflections.

We have calibrated the acquisition system by using a profile recorded in deep water (~4 km) which used a similar source and identical recording system to that used on the DRUM line. In deep water the ratio of the amplitude of the primary sea bottom reflection to that of twice the first multiple gives an accurate direct estimate of the reflectivity of the sea floor and hence calibrates the system. This method is offset-independent as the multiple used is recorded at twice the offset of the primary. It is also a multichannel method which allows many estimates to be made for the same point on the sea bed. The errors introduced during calibration in this way will produce no more than about a 5% error in final estimates of mantle reflection coefficients.

The velocity structure in the area of the DRUM line is well determined. The near surface and sedimentary velocity structure is well controlled by using the move-out on the reflection data; deep crustal and upper mantle velocities have been determined by several refraction and wide angle reflection lines in the area (Smith & Bott 1975; Bamford *et al.* 1978; Jones *et al.* 1984). Errors in the velocity model will introduce about a 15% uncertainty in reflection coefficient estimates. Determining the cumulative losses as the signal propagates to and from the deep reflectors is difficult. The model we have assumed is that the effective Q is 100 down to a depth of 5 km and is infinite, i.e. no further losses, below this depth. This model is most likely to be an overestimate of the value of Q, that is an underestimate of the losses, and we are therefore calculating minimum estimates of deep reflection coefficients.

We have estimated the apparent reflection coefficient for the brightest parts of both the Flannan Thrust and W-reflector. Fig. 2 shows a close up of the stacked data for both these events. The reflection coefficients for each event are rather similar and range from about 0.08 to 0.14. For the remainder of this paper we will assume that any acceptable geological model must be capable of producing reflection coefficients of at least 0.1.

Table 1

| Medium 1 | Medium 2 | Maximum reflection coefficient |
|---|---|---|
| Compositional variation | | |
| lherzolite | harzburgite | 0.02 |
| dunite | peridotite | 0.03 |
| dunite | pyroxenite | 0.04 |
| fayalite | forsterite | 0.02 |
| | | |
| Anisotropy | | |
| isotropic peridotite | anisotropic peridotite | 0.01 |
| isotropic dunite | anisotropic dunite | 0.02 |
| anisotropic dunite | anisotropic dunite | 0.04 |
| (maximum velocity vertical) | (minimum velocity vertical) | |
| | | |
| Phase change | | |
| spinel-peridotite (10% spinel) | garnet-peridotite | 0.02 |
| | | |
| Mafic layer | | |
| peridotite | gabbro | 0.13 |
| peridotite | eclogite | 0.11 |
| | | |
| Partially hydrated peridotite | | |
| peridotite | peridotite + ~ 20% serpentine | 0.1 |
| peridotite | peridotite + ~ 30% mica | 0.1 |
| peridotite | peridotite + ~ 65% amphibole | 0.1 |

## 3. Geological models

In our geological model we have estimated the normal incidence p-wave reflection coefficient obtained at a plane infinite single interface between two dissimilar media. We are therefore ignoring both interference due to thin layers and focusing due to reflector curvature or velocity structure. For realistic geological models neither of these processes can significantly increase the apparent reflection coefficient (i.e. by not more than a few tens of percent) above that of a single plane interface (Raynard 1986). Although we are strictly using a single interface as our model, most of our examples would in fact consist of a layer of one medium contained within the other. We have used estimates of p-wave velocity, anisotropy and density from published laboratory measurements at high pressure (usually 10 Kbars) performed on real rocks, synthetic crystal aggregates and single crystals (Christensen 1965, 1966, 1974; Birch 1960, 1961, 1972; Chung 1971; Babuska 1972, 1984; Baker & Carter 1972).

**Flannan thrust**

a)

13.5 s

15.5 s

9 km

**W-reflector**

b)

13 s

15 s

9 km
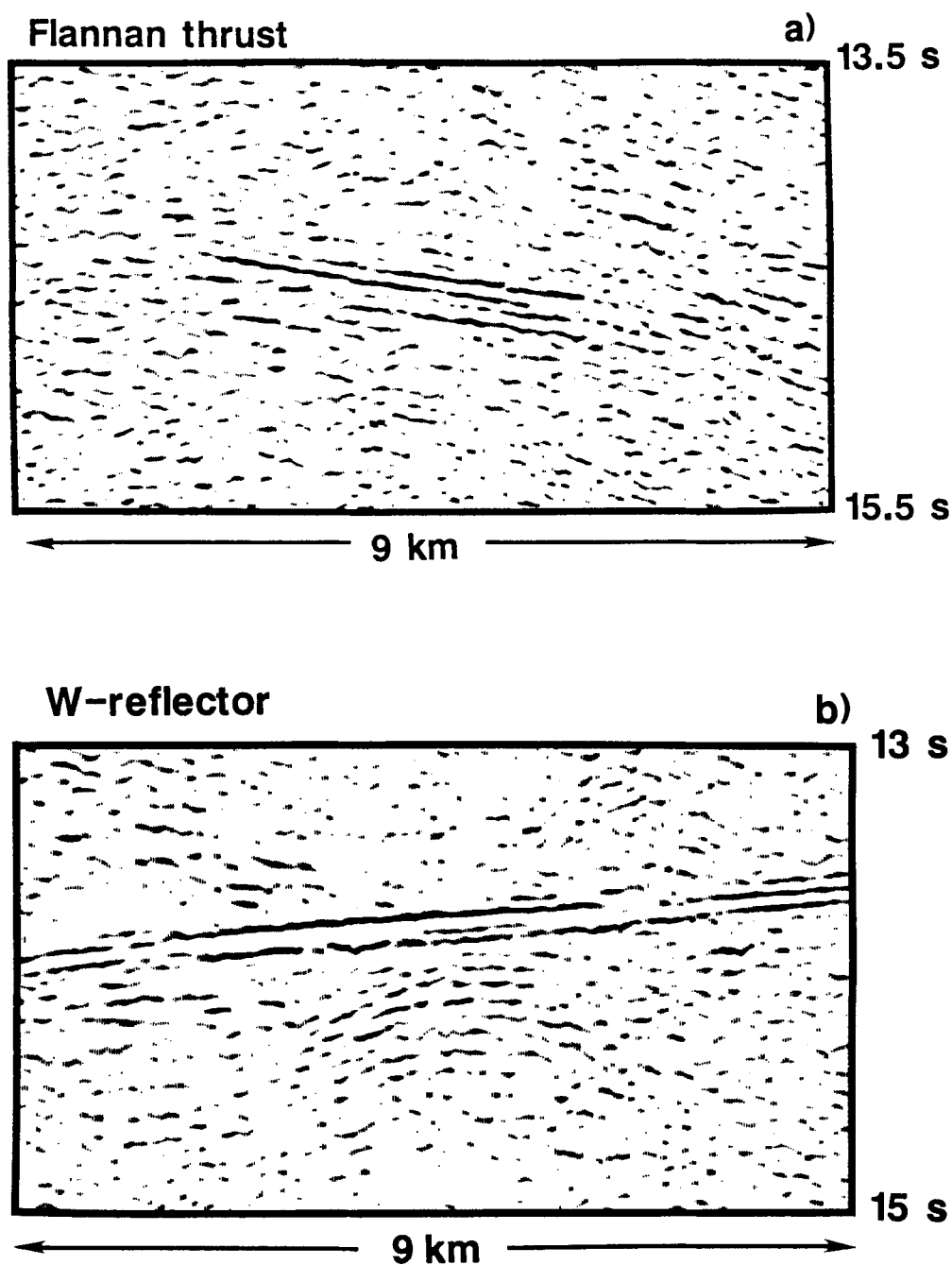
**Figure 2. (a).** Detail from the stacked DRUM line showing part of the Flannan Thrust. **(b).** Detail from the stacked DRUM line showing part of the W-reflector.
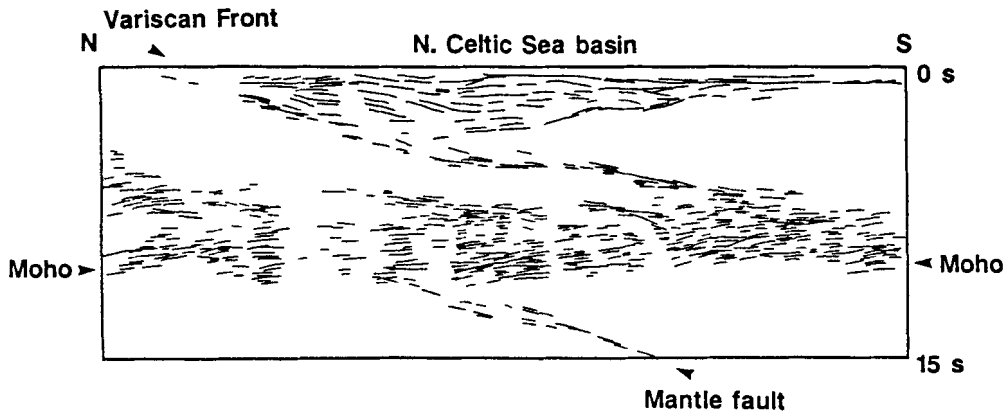
**Variscan Front**



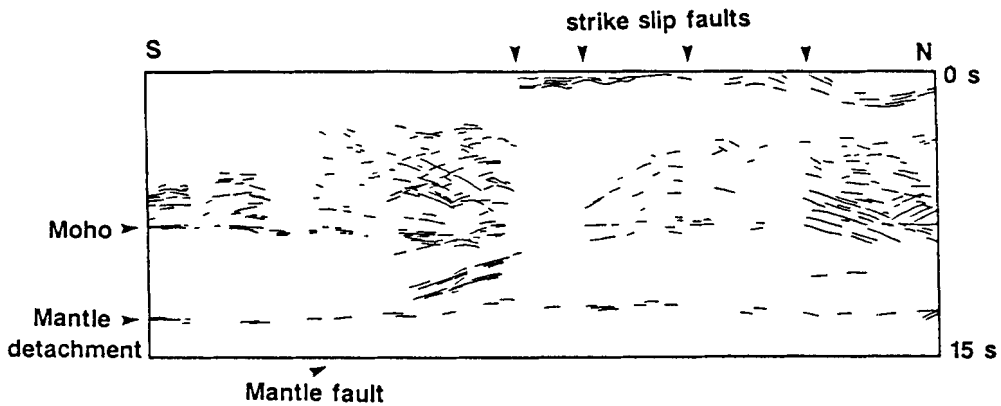Figure 3. Unmigrated line drawing of part of the SWAT survey.



Figure 4. Unmigrated line drawing of part of the WINCH survey.

Table 1 summarizes our results. It is clear that reflection coefficients large enough to explain the reflections we see can not be produced by compositional changes or seismic anisotropy within peridotites, or by a phase change from a spinel-peridotite to a garnet-peridotite.

There are two models which can explain the data. The first is to introduce a layer of mafic rock into a mantle composed of peridotite. Such a layer, whether it were gabbroic or eclogitic, could produce reflection coefficients in excess of 0.1. This mafic layer could be derived from the mantle by partial melting or it could have been originally in the crust and have found its way into the mantle during subduction of oceanic lithosphere.

The second possibility is that the mantle contains some layers of partially hydrated peridotite. Hydrated minerals which might be present in the mantle include serpentine, a mica such as phlogopite and an amphibole such as hornblende. Each of these minerals, if present in sufficient quantity, can produce a reflection coefficient of 0.1. The concentration of serpentine ($\sim 20\%$) and mica ($\sim 30\%$) are probably realistic for a thin layer within the

mantle. An amphibole rich layer requires rather higher concentrations of the hydrated mineral (~65%).

Both hornblende and phlogopite should be stable in the uppermost continental Serpentine however will not be stable above about 500°C at a depth of 50 km. This is a rather low temperature for the mantle at this depth, but our data are from a stable area with low heat flow and such temperatures are perfectly possible. The reflection coefficient we have calculated for phlogopite assumes that the mica is randomly orientated. Phlogopite is extremely anisotropic, and if, as we suggest below, the hydrated layer was formed as a shear zone then any anisotropy induced would be orientated so as to significantly increase this reflection coefficient.

## 4. Discussion

Figs. 3 and 4 show line drawings of two other profiles from around the U.K. which also show mantle reflections. Both the DRUM line and these two profiles have features in common. All three show major crustal-penetrating faults in the middle and upper crust: the Outer Isles Fault in Fig. 1; the Variscan Front reactivated as a normal fault in Fig. 3; and the Great Glen strike slip fault in Fig. 4. In each case, dipping reflectors in the upper mantle lie beneath these crustal faults. Both the DRUM line and the profile in Fig. 4 show sub-horizontal events below the Moho at a depth of 40 - 50 km. This apparent spatial association of major crustal faults with mantle reflectors suggests that the events in the mantle may also be zones of high localised strain. That is, the dipping reflectors are faults or shear zones within the mantle and the sub-horizontal events are detachment faults or decollement surfaces.

Faults and shear zones within the crust often show evidence for the passage of large amounts of water (Beach 1976), and water is necessary for significant movement to occur on low angle faults (Hubbert & Rubey 1959). It seems likely therefore that these analogous structures within the mantle will also have been wet when they were active and thus contain the hydrated minerals required to make them visible on deep reflection profiles. The second possibility is that the dipping events in the mantle are relic subduction zones rather than intra-plate faults. In this instance we have the added possibility that we may see reflections because crustal rocks have been carried down the subduction zone into the mantle.

## 5. Conclusion

Seismic reflection data have revealed major dipping and horizontal events below the continental Moho within the uppermost mantle. The best explanation for these events appears to be that they are mantle faults or shear zones. They are visible on seismic profiles because they contain hydrated minerals and/or mafic rocks.

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