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Letter to the Editor

Probable Relations Between Seismic Anisotropy and a Fine Structure of the Lithosphere

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P-wave anisotropy in horizontal directions has been observed in various parts of the world in the upper mantle, for instance in some regions of the Pacific Ocean (Hess, 1964; Morris et al., 1969; Raitt et al., 1969; Shimamura and Asada, 1978) and in some continental areas (Nevskiy et al., 1974; Bamford, 1977; Hirn, 1977). The observations on continents are not yet substantiated by the same amount of data and by the same level of confidence as are the oceanic ones. From laboratory data peridotitic and metamorphic rocks are known to exhibit a strong seismic anisotropy (Peselnick et al. 1974; Kern and Fakhimi, 1975; Meissner and Fakhimi, 1977; Kern, 1978) which has been attributed to the preferred lattice orientation of olivine in peridotites, the main rock type of the upper mantle. In fact, the preferred orientation of olivine seems to be the main reason for the observed large anisotropy in oceanic areas.

Parallel to the observations of anisotropy, mostly performed on medium range deep seismic sounding (=DSS) profiles, of 200 to 300 km in length, some so called 'Lithospheric profiles' of more than 500 km in length have been observed in oceanic and continental regions (Hirn et al., 1973; Bamford et al., 1976; Orcutt and Dorman, 1977; Nagumo et al., 1978; among others). Unfortunately, only very few lithospheric profiles exist so far, and none of them are crossing each other. From measurements along lithospheric profiles various interpreters derive a fine structure of *P*-wave velocity for the upper mantle. This is an alteration of layers of high and low velocities ($\Delta v \approx 0.5$ km/s) showing individual thicknesses of some tens of kilometers. In some cases, such a fine structure can be followed over adjacent and very long parts of a profile (Hirn et al., 1973). Recently such a fine structure has been derived for the western part of the Pacific based on surface wave studies and long range profiles (Nagumo et al., 1978). Independently, anisotropy has been observed in the same region along normal DSS-profiles (Shimamura and Asada, 1978). It seems possible, hence, that lithospheric fine structure is mainly an effect of a crystallographic

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anisotropy and not related to a fine structure in density and composition. Some arguments for this hypothesis can be summarized as follows:

(i) The range of anisotropy values from 7.4 to 8.7 km/s as derived from laboratory measurements as well as from field experiments in the Pacific (Raitt et al., 1969) is only slightly larger than that of the reported fine structure of the upper mantle.

(ii) Although up to now no intersecting lithospheric profiles (>500 km in length) exist which could definitely prove or disprove the existence of an azimuthal (and hence anisotropically generated) fine structure, there are two regions where anisotropy (along several crustal profiles with different azimuths) and fine structure (along one lithospheric profile) do co-exist. These observations are from the Western Pacific region, as mentioned before, and from parts of France (Hirn et al., 1973; Hirn, 1977). No other areas are known to us where both kinds of observations exist, i.e., several crustal profiles with different azimuths and a lithospheric profile. In the southern and southwestern part of Germany, where Bamford (1977) reported an anisotropy which is however much debated, a fine structure for a large area was reported based on body waves from earthquakes (Baer et al., 1979) providing another, though weaker, indication for a correlation of fine structure and anisotropy.

(iii) If the reported fine structure in *P*-velocity would be caused by a related density structure, for instance by the correlation of Woollard (1959), densities as low as $\rho = 3.3$ g/cm^3 ($v_p = 7.8$ km/s) would underly layers of $\rho = 3.5$ g/cm^3 ($v_p = 8.4$ km/s) in some places. Large areas would be unstable. Calculations indicate that such a layering would be very short-lived, especially near the oceanic ridges where new lithosphere is created and viscosities are near 10^{17} poise (Vetter and Meissner, 1977). Even if viscosity were 10^{19} poise and the density differences only 0.01 g/cm^3 a spherical inhomogeneity of 30 km radius would need less than 100,000 years to intrude the denser layer. (From a set of master curves, using Maxwell rheology, Inst. f. Geophys., Kiel, unpubl.). As the oceanic lithosphere is definitely created at oceanic ridges in a high temperature and low viscosity environment, the assumption of a density structure in the oceanic upper mantle which corresponds to the *P*-velocity structure seems impossible from genetic and stability reasons.

(iv) The assumption that the velocity structure represents different layers with different anisotropy, caused by a preferred orientation of olivine crystals in peridotite, seems plausible from genetic reasons: During the generation of new oceanic lithosphere at the oceanic ridges, which is certainly a high temperature process, a preferred orientation of olivine crystals is easily obtained due to stress fields. Such a process of creating an anisotropy by weak or medium stresses and elevated temperatures is much debated. Plastic flow and/or recrystallization may be involved in these processes (Francis, 1969; Ave'Lallemant and Carter, 1970).

Various stresses have to be considered for creating a preferred orientation of olivine crystals near ridges: The largest stresses seem to originate from bending of flow lines. Due to Francis (1969), *a*-axes of olivines adjust to flow lines. Other stresses, such as those of thermal origin, overburden pressure, and perhaps solidification stresses seem to play an additional role. All stresses may really be different at different depth levels due to flow inhomogeneities and discontinuous flow patterns. Later tectonic events such as the origin of fracture zones, dyke injections, re-heating and plastic flow may destroy (or raise) the anisotropy. However, basically anisotropy – like magnetization – is assumed to be 'frozen in' during the period of bending and cooling near the ridges. According to possible differences in stress at different depth levels, also the anisotropy may be depth-dependent, i.e., show a layered structure. In general, of course, a-axes of olivines and the direction of maximum P_n-velocity should be directed perpendicular to the ridges, as found for the uppermost part of the mantle in many marine in situ-experiments. The greater the depth down from the M-discontinuity, the smaller is the bending stress, the longer the cooling time, and the greater the possibility of a contribution of other stresses. A density layering would certainly not be created during a gravitational differentiation of oceanic lithosphere at ridges.

All four points mentioned above might well explain a correlation between anisotropy and fine structure under oceans and continents. Points (iii) and (iv), which primarily seem to explain anisotropy and fine structure in the *oceanic* lithosphere, might well be extended to continental areas if one assumes a general underplating of oceanic lithosphere below continents, which was especially strong and shallow in the Precambrian (Meissner, 1979). Old continental shields, on the other hand, have often undergone various orogenic cycles with a re-heating and metamorphism, increasing the probability of destroying large scale uniform orientation of olivines in peridotitic layers. Additionally, shear stresses acting on the base of the continental lithosphere as suggested by Fuchs, 1977, may generate a new orientation of the olivine crystals and thereby of maximum velocity orientation. Long range (lithospheric) observations of different orientations are required in suspected areas to establish a solid basis for the suggested relation between anisotropy and fine structure. They should show, then, an azimuthal variation of the fine structure.

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