FORMATION OF ANISOTROPY IN UPPER MANTLE PERIDOTITES - A REVIEW

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Abstract. In this review of the rapidly growing literature on seismic anisotropy in the upper mantle, the accent is put on the origin of this phenomenon. The predominant role of olivine is recalled. It is shown that the anisotropy is induced by plastic strain and directly related to it. There is no direct relation between anisotropy and stress. Finally, the nature and geometry of asthenosphere flow at oceanic ridges are discussed in view of their bearing on seismic anisotropy.

Introduction

In many regions of the upper mantle and to depths attaining at least the olivine-spinel transition, seismic anisotropy has been revealed by its various signatures: azimuthal anisotropy of P- and shear waves velocities, polarization anisotropy of shear waves (Crampin et al., 1984). As was first shown by Hess (1964), based on Pn velocities reported by Raitt (1963) and Shor and Pollard (1964) in the Mendocino and Maui areas of the northwest Pacific, velocities are generally high perpendicular to ridge crests and slow parallel to them. Similar observations have been repeated in various regions of the Pacific ocean (Raitt et al., 1969; Keen and Barrett, 1971; Shor et al., 1973; Snydsman et al., 1975; Malecek and Clowes, 1978; Shimamura et al., 1983; Shearer and Orcutt, 1985) and the Indian Ocean (Shor et al., 1973). Seismic observations in the vicinity of the Mid-Atlantic Ridge (Keen and Tramontini, 1969) suggest the presence of upper mantle seismic anisotropy; however scatter in velocities resulting from rough topography makes it difficult to clearly demonstrate its existence (Whitmarsh, 1968). Continental anisotropy has been reported in western Europe (Bamford, 1973, 1977; Fuchs, 1975, 1983; Dziewonski and Anderson, 1983), central Europe (Babuska et al., 1984 a and b), the western United States (Bamford et al., 1979; Vetter and Minster, 1981; Dziewonsky and Anderson, 1983), the USSR (Chesnokov and Newskiy, 1977) and possibly Australia (Leven et al., 1981). Areas in which analysis of seismic data have failed to detect upper mantle anisotropy include the Mojave region of southern California (Vetter and Minster, 1981), the eastern United States and northern Britain (Bamford et al., 1979). Upper mantle shear wave anisotropy has also been reported as having (1) a weak azimuthal dependence (Clowes and Au, 1982) and (2) a clear shear wave splitting (Ando et al., 1983).

Surface wave studies (Forsyth, 1975; Crampin and King, 1977; Mitchel and Guey-Kuen, 1980; Anderson and Dziewonski, 1982; Woodhouse and Dziewonski, 1984; Lévêque and Cara, 1983; Nataf et al., 1984) provide additional support for the existence of anisotropy throughout much of the mantle at depths to 450 km. Recently Tanimoto and Anderson (1984) have presented a global map showing azimuthal anisotropy of 200 sec Rayleigh waves (Figure 1). The fast directions of the Rayleigh waves show little correlation with plate motions, since they are sampling mantle beneath the lithospheric plates and thus are more likely to correlate with return flow directions.

Compressional wave anisotropy can reach 10%, but generally ranges from 3% to 6%. Observation of upper mantle azimuthal anisotropy summarized by Christensen (1984) are presented in Figure 2. They show that within ocean basins fast velocities parallel transform faults in agreement with the early findings of Hess (1964). Hess ascribed this anisotropy to plastic flow aligning in mantle peridotites olivine (010) slip planes parallel to transform faults, thus relating the anisotropy to plastic flow along transform faults. Francis (1969) proposed that upper mantle anisotropy originates by asthenospheric flow at oceanic ridges, which orients [100] olivine slip lines at high angles to the ridge trend. This model for the development of preferred orientation was based on Raleigh's (1968) experimental finding that at high temperature [100] was the dominant slip line and that the {0kl} slip planes where co-zonal with this direction. Experimental

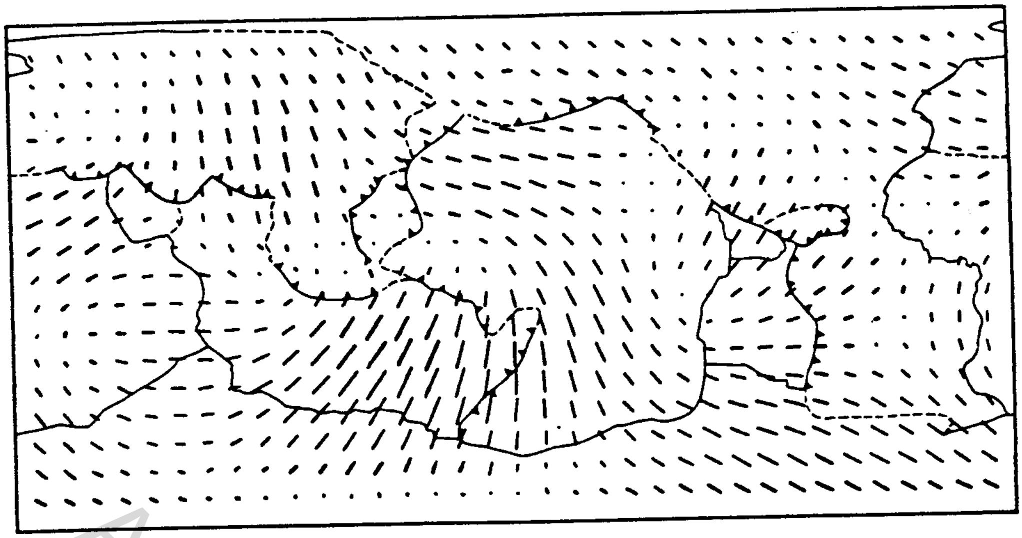


Fig. 1. Mantle anisotropy of depths of 200 to 400 kilometers. Lines are parallel to the fast propagation direction of Rayleigh waves which parallel crystallographic [100] of olivine (Tanimoto and Anderson, 1984).

studies on the development of preferred orientations in olivine aggregates (Avé Lallemant and Carter, 1970; Nicolas et al., 1973) demonstrated that distinct lattice preferred orientations were obtained for moderate strains, which were related to plastic flow and seismic anisotropy in oceanic uppermantle.

Information on shear anisotropy is provided by Love and Rayleigh wave data, since surface waves are sensitive primarily to shear wave velocity. Surface wave studies of both polarization (e.g., Forsyth, 1975; Schlue and Knopoff, 1977; Yu and Mitchell, 1979; Anderson and Regan, 1983) and azimuthal anisotropy (e.g., Forsyth, 1975; Mitchell and Yu, 1980) show that anisotropy extends to considerable mantle depth. The recognition of polarization anisotropy, based on inconsistency of Love and Rayleigh wave dispersion data, requires a faster horizontal velocity than vertical velocity in the asthenosphere. In general, azimuthal surface wave anisotropy of Rayleigh and Love wave data for the Pacific lithosphere show fast

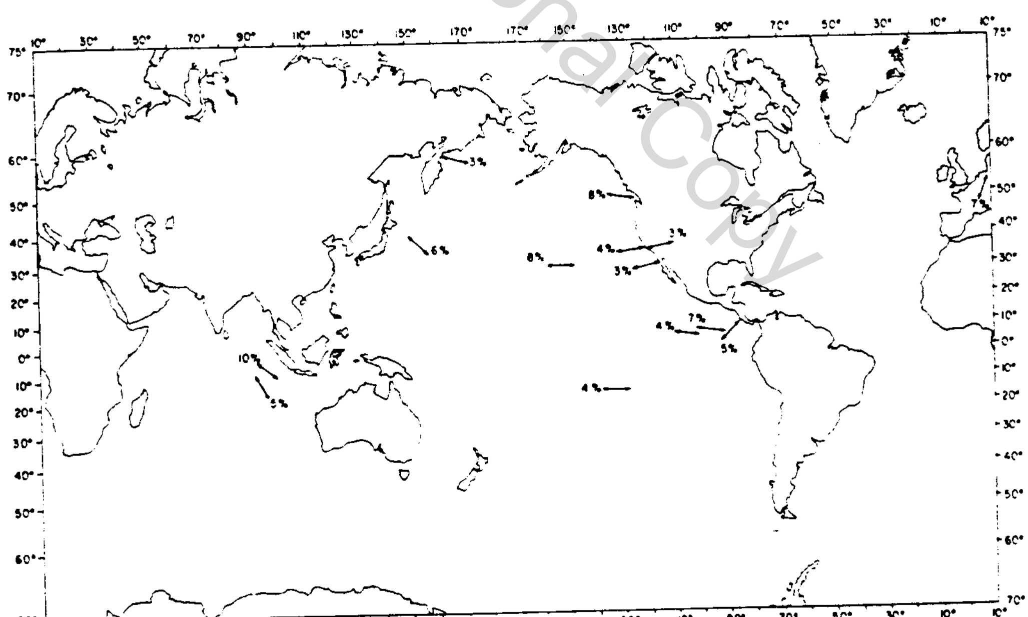


Fig. 2. Observed upper mantle anisotropy. Arrows are parallel to fast directions (Christensen, 1984).

velocities approximately perpendicular to magnetic lineations (Forsyth, 1975; Yu and Mitchell, 1979; Mitchell and Yu, 1980). These studies are, however, complicated by their poor resolution and the non uniqueness of modeling to satisfy both Rayleigh and Love wave data (Knopoff, 1983).

In this paper, we review experimental and theoretical studies which have related seismic anisotropy to preferred mineral orientation and upper mantle flow. The review relies primarily on recent investigations of natural deformation of peridotites and laboratory studies of peridotite anisotropy, which relate to the interpretation of seismic anisotropy in terms of uppermantle structure and dynamics.

Crystallography and Velocities: Theory and Experiments

Hess' (1964) discovery was stimulated by laboratory measurements in Dunbar laboratory at Harvard University of olivine single crystal elastic constants (Verma, 1960) and peridotite compressional wave velocities by Birch (1960). The single crystal data were obtained at ambient conditions from velocities of acoustic wave propagation and measurements of Young's moduli along principal crystallographic axes. Verma's (1960) study demonstrated that single crystal olivine was highly anisotropic with a maximum compressional wave velocity of 9.87 km/sec for propagation parallel to crystallographic [100] or a and a minimum compressional wave velocity of 7.73 km/sec parallel to crystallographic [010] or b. Birch (1960) measured compressional wave velocities in three orthogonal directions for several dunites and noted the presence of significant anisotropy. Although fabric data were not presented, the anisotropy was attributed to preferred olivine orientation (Birch, 1961). Following these pioneering studies, much has been learned about wave propagation in upper mantle rocks and minerals and the nature of upper mantle seismic anisotropy.

Anisotropy of Olivine and Orthopyroxene

Geophysical constraints, together with petrological and chemical limitations (e.g., Ringwood, 1975), restrict the dominant mineralogical composition of the upper mantle to an assemblage of olivine and pyroxene. Thus velocities and orientation of olivine and pyroxene are to a first approximation responsible for upper mantle seismic anisotropy. Several detailed experimental studies have reported elastic properties of these minerals at ambient conditions and at elevated temperatures and pressures.

Following the measurements of Verma (1960), elastic constants of single crystal forsterite were determined to hydrostatic pressures of 1000 MPa and temperatures in the range of 300°K to 700°K (Graham and Barsch, 1969). In addition, Kumazawa and Anderson (1969) measured the elastic constants of forsterite and olivine (Fo93) at pressures to 200 MPa and temperatures between 24.5°C and 33°C. Single crystal data are available for orthopyroxene of bronzite composition (Ryzhova et al., 1966 and Kumazawa, 1969). Measurements of single crystal elastic constants of bronzite as functions of pressure and temperature have been reported by Frisillo and Barsch (1972).

The elastic constants of single crystals are usually calculated from acoustic wave velocity measurements made by the pulse superposition technique (McSkimin, 1961). The orthorhombic symmetries of olivine and orthopyroxene require the determination of nine independent elastic constants to completely define their elastic properties. Compressional and shear wave velocity measurements parallel to the crystallographic axes provide six moduli. The remaining three moduli are usually determined from velocity measurements for three different propagation directions perpendicular to one crystallographic axis and oblique to the remaining two. Once the nine elastic constants have been determined, it is possible from the Cristoffel equation (e.g., Musgrave, 1970) to calculate velocities of three waves with orthogonal polarizations for any propagation direction within the crystal. These velocities correspond, in general, to one quasi-compressional and two quasi-shear waves.

The fastest compressional wave velocity direction in both olivine and orthopyroxene is crystallographic [100] and the slowest direction is crystallographic [010] (Figure 3). Velocity surfaces in the olivine (010) plane and orthopyroxene (100) plane are shown in Figure 4. The velocities of compressional and shear waves in these symmetry planes are particularly significant to Pn and Sn velocity anisotropy observations, because these planes tend to orient subhorizontal in the upper mantle (e.g., Nicolas and Poirier, 1976; Christensen and Lundquist, 1982).

Peridotite Anisotropy

Laboratory measurements of velocities in rocks usually employ the pulse transmission method (Birch, 1960) in which travel times of elastic waves are determined in cylindrical rock cores of known lengths. Disk shaped piezoelectric transducers, usually of natural resonance frequencies of 1 MHz, are placed on each end of the core to generate and receive the elastic waves. To investigate peridotite anisotropy, cores are taken in several directions from a single rock specimen.

The relationship of olivine orientation to the magnitude and symmetry of peridotite anisotropy was first investigated by Christensen (1966a). Olivine petrofabric diagrams obtained from thin sections cut from core ends of samples in which compressional wave velocities were measured to 1000 MPa showed concentrations of olivine crystallographic [100] axes and [010] axes parallel to fast directions and slow directions respectively, in good agreement with single crystal velo-

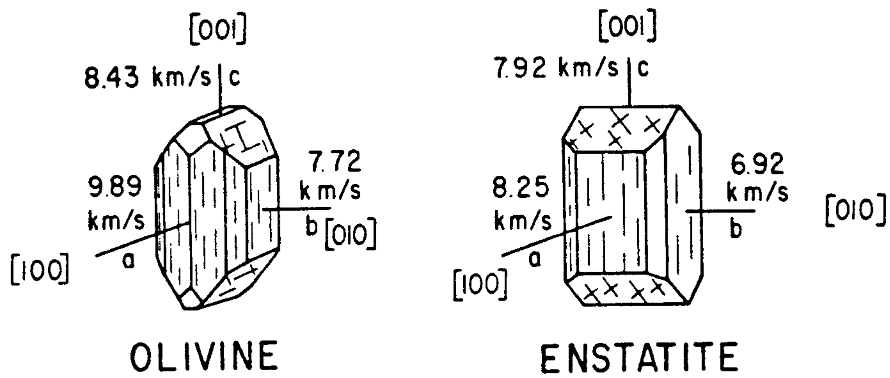


Fig. 3. Compressional wave velocities in km/sec along the crystallographic axes of olivine and orthopyroxene.

city studies. Since 1966, measurements of peridotite anisotropy have been obtained by investigations at a relatively small number of laboratories, which have both confirmed earlier results that seismic anisotropy is related to preferred mineral orientation in ultramafic rocks and provided more detailed information on shear as well as compressional wave propagation in ultramafic rocks (Christensen, 1971; Kumazawa et al., 1971; Christensen and Ramanantoandro, 1971; Babuska, 1972; Peselnick et al., 1974; Peselnick and Nicolas, 1978; Ramanantoandro and Manghnani, 1978).

Laboratory studies of shear wave anisotropy in peridotites have provided additional insight on the nature of shear wave anisotropy in the upper mantle. The behavior of shear waves in anisotropic rock is similar to that in single crystals. In general, two orthogonally polarized shear waves propagate in anisotropic rocks (Christensen, 1966b). In peridotites, the difference in velocities of the two waves is often as great as 0.2 km/sec and like compressional wave anisotropy shear wave anisotropy correlates well with preferred olivine orientation (Christensen and Ramanantoandro, 1971). Shear energy generated by a transducer attached to a core of anisotropic peridotite is split into two waves with fixed polarizations through the core (Figure 5). This phenomenon is referred to as shear wave double refraction, shear wave birefringence or shear wave splitting, details of which depend on the symmetry of mineral orientation. Sometimes directions are present which the two shear waves travel with the same velocity. These directions, referred to as conical points or shear wave singularities in single crystals (Crampin, 1981), commonly occur in planes of mirror symmetry.

Recently Ando et al. (1983) and Fukao (1984) have presented convincing evidence for shear wave splitting in the upper mantle beneath Japan. Arrival time differences for the two shear waves were up to two seconds for earthquake focal depths to 535 km. Future observations of this nature combined with compressional wave anisotropy observations will place important constraints on olivine orientation in upper mantle peridotite.

Anisotropy of Ultramafic Massifs

With increasing resolution of seismic observations, it is becoming clear that upper mantle seismic anisotropy is a common phenomenon over large regions of the lithosphere. Since laboratory measurements are made on small rock specimens, usually a few centimeters in length, caution must be exercised in correlating velocity information from a single rock sample with mantle anisotropy. This scale problem has been at least partially eliminated by studies of the structure and anisotropy of large ultramafic massifs derived from the upper mantle.

Several independent studies, taking somewhat different approaches, have concluded that the large peridotite massifs possess seismic anisotropy similar to upper mantle anisotropy. An early investigation (Christensen, 1971) of petrofabrics and seismic velocity anisotropy measurements on field oriented samples from the Twin Sisters massif in Washington found that olivine orientation and compressional wave anisotropy

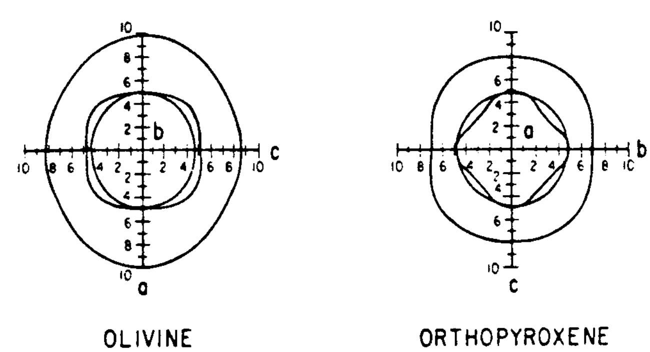


Fig. 4. Compressional and shear wave velocities within the olivine (010) plane and orthopyroxene (100) plane.

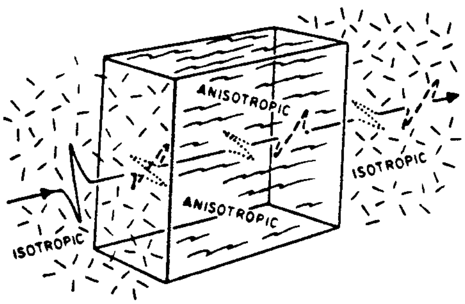


Fig. 5. Shear wave splitting in anisotropic rock (Christensen, 1984).

were uniform over approximately 100 km^2 of exposed dunite and harzburgite. This massif with its tectonite fabric has an overall anisotropy of a few percent, in agreement with measured anisotropy in the uppermost mantle.

Using the elastic constants and their pressure derivatives of olivine and orthopyroxene, compressional and shear wave anisotropy can be calculated at upper mantle depths from petrofabric data. This technique has been described by Kumazawa (1964), Crosson and Lin (1971) and Baker and Carter (1972). The calculated anisotropy was shown to be in excellent agreement with laboratory measurements.

Peselnick and Nicolas (1978) measured velocities up to 800 MPa and 275°C in 3 cores of a field oriented harzburgite from the Antalya ophiolite complex in Turkey and from olivine and orthopyroxene petrofabric data, calculated the seismic anisotropy of the sample, which was related to structural measurements of lineation and foliation. Using the calculated anisotropy, field measurements from a 15 km north-south outcrop, and a structural and kinematic model for the ophiolite (Juteau et al., 1977), anisotropy was calculated for the whole massif. The anisotropy in the restored horizontal plane was found to be similar in direction and magnitude with the refraction data of Raitt et al. (1969).

Further evidence equating seismic anisotropy of ophiolite peridotite sections with upper mantle anisotropy was provided by studies of the Bay of Islands ophiolite, Newfoundland (Christensen and Salisbury, 1979). Olivine petrofabrics of several field oriented harzburgites from widely-spaced traverses were shown to have remarkably uniform orientations, with olivine [100] crystallographic axes aligned subperpendicular to sheeted dikes and [010] and [001] axes parallel to the sheeted dikes. Compressional wave velocities computed from the petrofabrics have 5-6% anisotropy in the plane of the Mohorovicic discontinuity, with the fast direction parallel to the spreading direction inferred from dike orientations. Similar results were obtained for a traverse in the northern section of the Samail ophiolite, Oman (Christensen and Smewing, 1981).

Recently, Christensen (1984) calculated anisotropies from petrofabrics for several additional peridotite massifs, including ones from New Zealand, Newfoundland and Cyprus and found good agreement with previous studies. Considering their broad geographic distribution and varying age, it is remarkable that anisotropies of the ultramafic massifs are so similar. This suggests that seismic anisotropy is a fundamental property of the lithosphere.

Origin of Anisotropy

The abundance and the quality of studies devoted to the subject and mentioned in the introduction demonstrate that there is a strong connection in upper mantle peridotites between plastic flow, mineral preferred orientations and acoustic waves anisotropy. We wish here to consider further the relation between plastic flow and preferred orientations and that between preferred orientations and anisotropy. Conversely, is in situ seismic anisotropy in the upper mantle always and best explained by this three-terms relation?

Development of Preferred Orientations by Plastic Flow

It is now well established both by experimental and natural studies, that plastic flow induces preferred orientations in rock-forming minerals (reviews in Carter, 1976 and Nicolas and Poirier, 1976). The respective roles of deviatoric stress and plastic strain in the development of preferred orientations have been long debated and, in view of the consequences in interpreting seismic anisotropy this necessitates a summary of the fundamental physical mechanisms responsible for lattice reorientation in minerals (Nicolas, in press).

In the plastic field, crystals deform principally by dislocation slip and climb. Deformation mechanisms relying solely on diffusion (Nabarro-Herring and Coble creep favouring grain boundary sliding), which could become important at high temperature, do not contribute to the development of preferred orientations and do not seem to play a large role in natural deformation of known mantle rocks (Nicolas, 1976).

To assure continuity of a deforming crystal with its neighbours during the course of a given deformation, five independent degrees of motion are necessary (the Von Mises criterion). This can be achieved in a crystal with activation of five independent slip systems or with a combination of fewer slip systems and other modes of deformation: diffusion, heterogeneous deformation with or without lattice rotation, fracturing. In silicates, and particularly here in olivine and pyroxenes a limited number of slip systems are activated, say one or two. Consequently, the relationships between the directions of finite strain

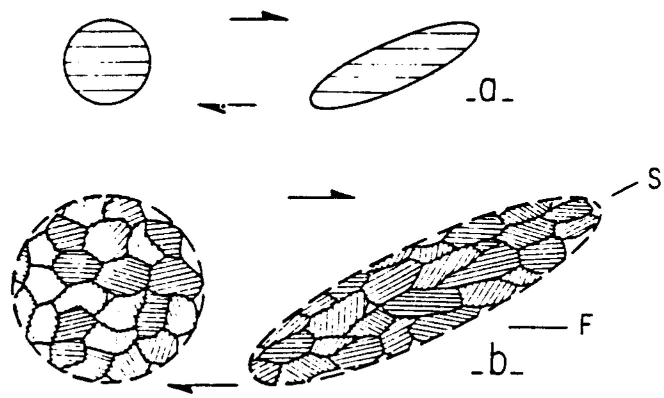


Fig. 6. Relation between the direction of finite strain (S) and that of flow (F); a) in a single crystal with a slip plane properly oriented and (b) in an aggregate of crystals with a dominant slip plane (Nicolas, 1984).

axes and that of the activated slip systems in the deforming crystal are simple; this relationship is univocal in the case of a single slip system (Figure 6). The following rule can be proposed, based on numerous studies of naturally and experimentally deformed rocks. In the statistically homogeneous deformation of a specimen composed of minerals characterised by a dominant slip system the preferred orientations of slip planes and slip directions in these minerals tend to coincide respectively with the orientations of the flow plane and the flow line. Consequently, the flow regime can be deduced from the relative preferred orientations of the slip systems and the finite strain directions. This is illustrated in Figure 6a for deformation of a crystal by a single slip system correctly oriented with respect to the applied shear stress. The rules consist in extending to the aggregate the behaviour of this individual (Figure 6b). With increasing strain, the crystals will progressively be reoriented to mimick the single crystal behaviour. Let us examine how this reorientation proceeds, in a planar situation with single slip.

Simple shear in a crystal rotates all the lines attached to the crystal except those which are contained in the slip plane (Figure 6). Let us suppose that the deforming crystal illustrated by Figure 7 is embedded into an aggregate deforming by flattening (coaxial regime). In the aggregate, the principal finite strain axes Xa and Za remain parallel during the course of deformation and impose this constraint on the crystal strain axes Xc and Zc ($X \geq Y \geq Z$). Accordingly for each $d\beta$ strain increment producing a $d\beta$sr clockwise rotation due to shear, the Xc, Zc axes must rotate by bulk rotation of a $d\beta$br anticlockwise angle equal to $d\beta$sr; for a given finite strain we have $\beta sr = -\beta br$. The bulk rotation produces a lattice reorientation which in the present case would produce in the aggregate a statistical preferred orientation of the slip planes close to Xa. The following conclusions are derived:

1) The cause of lattice preferred reorientation in a crystal is the requirement to maintain continuity with its immediate neighbours. Inasmuch as the neighbours deform approximately in a similar fashion as the aggregate, so must the considered crystal.

2) Responding to neighbours solicitations the crystal reorientation is not a direct consequence of the stress applied to the aggregate. At the crystal scale, reorientation obeys to local dynamical considerations due to this geometrical requirement.

3) With homogeneous strain and single slip in crystals, it is impossible to achieve continuity from one deforming crystal to the next. This problem receives different solutions in low- and high-T temperature deformations: lattice deformation, cleavage, fracturing and recrystallization at low-T and diffusion-controlled processes (dislocation climb, grain boundary migration, ...) at high-T. Larger degree of freedom at high-T explains the strength of some fabrics (Nicolas and Prinzhofer, 1983; Nicolas, in press).

Effects of Peridotite Minerals and Their Preferred Orientation on Seismic Anisotropy

Mantle peridotites occurring in basalt and kimberlite xenoliths and orogenic massifs (Nicolas, 1976) are dominantly lherzolites composed of 65-70% olivine, 20-25% orthopyroxene, 5-10% clinopyroxene and ~5% aluminous phase (feldspar, spinel or garnet depending on depth). Studies of ophiolite, however, suggest that the top 20 km of oceanic mantle are harzburgitic and locally dunitic, with on average 75% olivine, 20% orthopyroxene and 5% spinel and diopside.

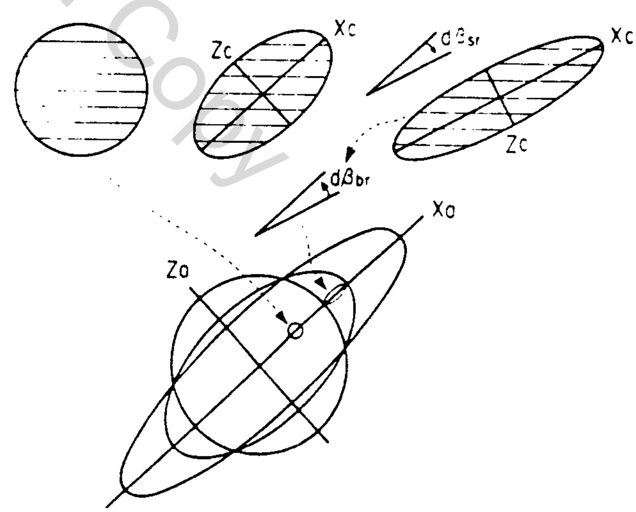


Fig. 7. Lattice reorientation in a crystal with a single slip system during progressive pure shear (Nicolas, 1984).

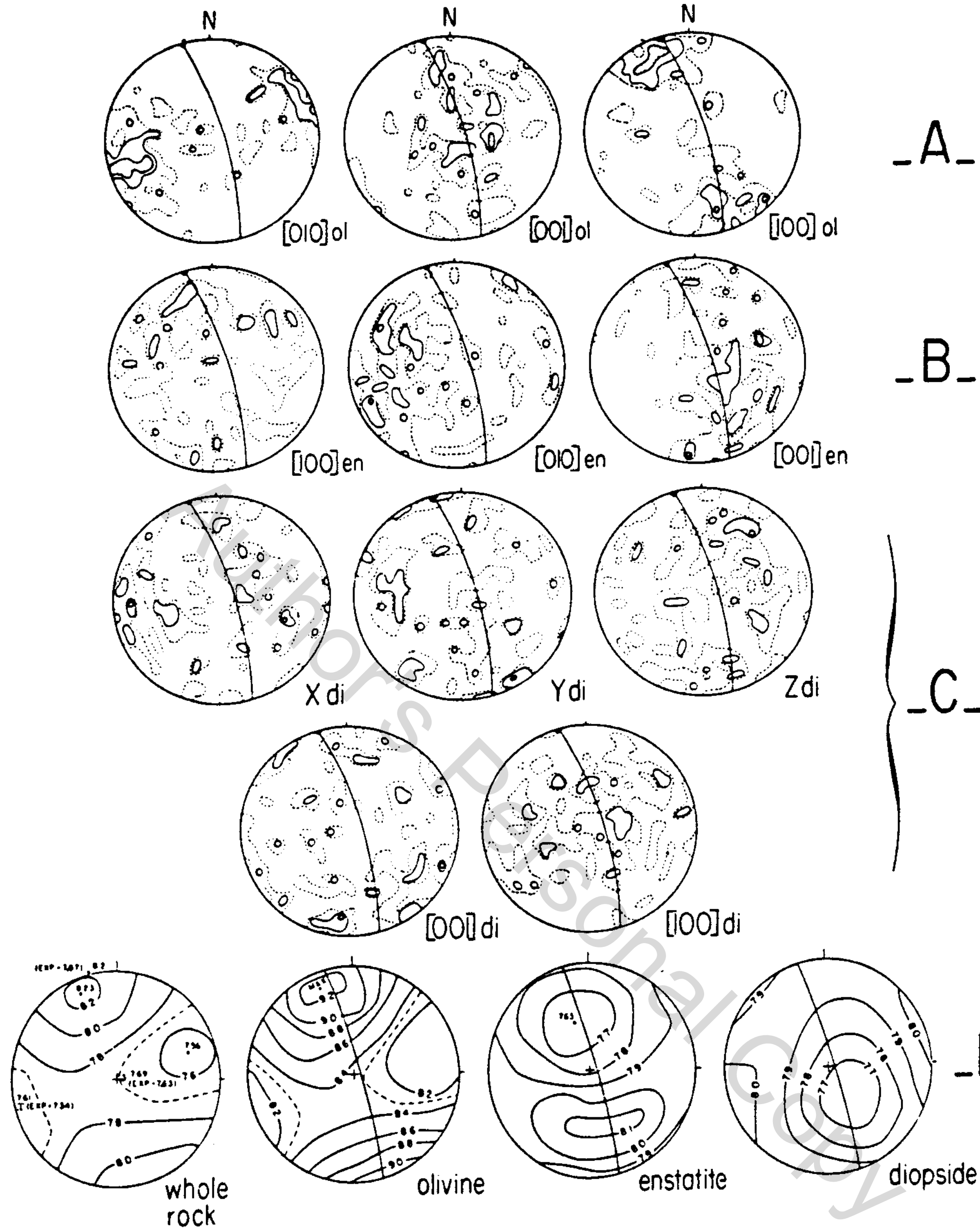


Fig. 8. Preferred orientations of olivine (A), enstatite (B) and diopside (C) in a lherzolite and corresponding calculated or measured (exp) P-waves velocities (D). Schmidt's projection in the horizontal plane (line: foliation; dot: lineation), 100 grains and contours at 1, 2, 4% area (Peselnick et al., 1974).

The volumetric predominance of olivine in peridotites, which is the same in mass since the main silicate phases have the same density, explains its influence on anisotropy. This influence is further increased by additional factors, now examined. The olivine crystal has large Vp (22%) and Vs (9-12%) anisotropies (see preceding section). Significantly, the largest Vp coincides with the [100] axis which is the dominant slip line at high-T and the smallest Vp with the [010] which is normal to a common slip plane (Figure 9). As a result, the Vp anisotropy in olivine aggregates is directly equated with the plastic flow directions. This is not true for the Vs patterns which are not simply related to the slip orientation; consequently the anisotropies between the [100] and [010] directions are only 2-3%.

In orthopyroxene, the next abundant peridotite-forming mineral, the Vp-anisotropy is still

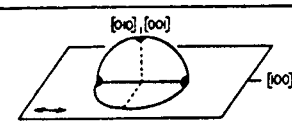
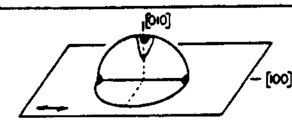
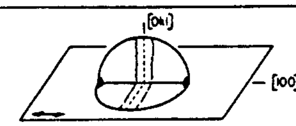
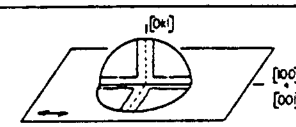
T domains	dominant slip systems	Fabrics strength, type	Relation with flow plane (horizontal) and flow line (arrow).
⩾1250°C (hypersolidus)	(010) [100] (001) [100]	- extreme - point maxima partial girdles	
>1100°C (high-T)	(010) [100]	- strong - point maxima	
∼1000°C (medium-T)	(0kl) [100]	- distinct - point maximum and girdle	
700-1000°C (low-T)	(0kl) [100] (010) [001]	- absent or weak - double diffuse girdle	

Fig. 9. Olivine slip systems and flow orientations in function of temperature for large shear strains.

large (16%), but the Vs one is small (5%). The principal Vp directions are still related to the crystallographic axes but now the largest Vp coincides with the [100] pole of the unique slip plane and the intermediate Vp with the unique [001] slip line. The Vs anisotropy shows no relation with the crystallographic axes. Clinopyroxene has significant Vp (24%) and Vs (14%) anisotropies, but in an aggregate oriented by plastic flow ([001] and (100) respectively parallel to flow line and flow plane) the anisotropy is usually small due to the development of weak fabrics. Carter et al. (1972) calculated anisotropies of only 2% in two garnet-clinopyroxenites. In a typical lherzolite, the diopside calculated anisotropy was 4% and contributed negatively to the whole rock anisotropy (Figure 8). Due to its isometric symmetry garnet is weakly anisotropic (Vp = 0.2%; Vs = 1%) and thus is unlikely to contribute significantly to upper mantle anisotropy.

Olivine is the most ductile peridotite-forming mineral, followed by orthopyroxene. In deformed lherzolites only these two minerals develop clear and systematic patterns of preferred orientation (Boudier, 1978), that of olivine being always stronger and more coherent. For this reason and taking also into account their effective low anisotropies in plastically oriented rocks, clinopyroxene and garnet will not be further considered.

The specific contributions of olivine and orthopyroxene in peridotites have been studied by Peselnick et al. (1974), Peselnick and Nicolas (1978) and Christensen and Lundquist (1982). These results illustrated in Figure 8, clearly demonstrate the predominant role of olivine in defining the whole rock anisotropy.

The presence of serpentine in uppermost mantle peridotites is possible. It will have an important diluting effect on anisotropy as well as significantly lower mean mantle velocities. For example, Peselnick et al. (1974) measured a 6.7% Vp-anisotropy in a lherzolite with only 14% serpentine and calculated for the same specimen devoid of serpentine, a 8.1% anisotropy.

Contribution of Olivine

Strain. Studies in mantle rocks strongly suggest that simple shear is the dominant regime in the upper mantle (Nicolas, 1976). Experimental deformation in simple shear regime achieved in high-T ice shows that a clear fabric is obtained for a shear strain $\gamma \geq 0.6$ (Bouchez and Duval, 1982), a result in agreement with pure shear results in olivine aggregates where the equivalent strain corresponds to a shortening 30% (Nicolas et al., 1973). Above this moderate strain at high-T, the strength of the fabric first increases and seems to reach a quasi-steady state for strains greater than a few γ. Thus anisotropy records a shear strain $\gamma \geq 0.6$.

Temperature. Since the experimental work of Carter and Avé Lallemant (1970), it is known that different slip systems are activated in olivine depending on temperature (reviews in Carter, 1976 and Nicolas and Poirier, 1976). Studies on naturally deformed peridotites conducted by the present authors confirm this and make it possible to estimate the temperature domains using petrological geothermometers and partial melting evidence. The results are presented in Figure 9. Let us summarize the conditions favoring very strong peridotite fabrics and consequently the highest anisotropies (Figure 9). At very high temperatures fabrics can be extremely strong due to enhanced diffusion and large grain boundary mobility (Nicolas, in press). Olivine slips essentially with a single slip system, thus developing point maxima with either [010] or [001] as poles of the flow plane (Figure 9). Below 1100°C, the fabrics become weaker because of the combined effects of activating new systems (producing girdle patterns) and of syntectonic recrystallization which introduces some misorientation (Nicolas and Poirier, 1976).

Effects of Planar Discontinuities on Seismic Anisotropy

Upper mantle anisotropy may originate from effects other than preferred orientation of highly anisotropic minerals, such as warping of the Moho discontinuity and stratification in the upper mantle which deviates from horizontal. Both of these possible causes of anisotropy have been carefully analyzed by Backus (1965), who concluded that it was unlikely that either would produce the observed azimuthal dependence of mantle velocities. The anisotropy observed by Hess (1964) in the Pacific could not originate by structure of isotropic peridotite unless all the shot lines were located by chance on independent local features. For example, a model of anisotropy produced by a corrugated Moho discontinuity with north-south trending ridges would require to produce the observed Pn velocities all of the shot lines to be located above valleys and none above ridges.

In normal mantle, the only parallel planar discontinuity relevant here is the compositional layering ubiquitous in mantle rocks. Evidence from xenoliths and massifs shows that this layering is on the average very weakly represented as websteritic and dunitic layers, 1-10 cm thick and constituting less than 2% of a typical outcrop (Boudier, 1978). This layering is generally parallel to the plastic foliation, that is within less than 20° from the orientation of the flow plane (Nicolas and Poirier, 1976). If the introduced anisotropy is such that the fast velocities are in the layering plane and the slow one perpendicular to it, the net effect will be added to and difficult to separate from that due to olivine-preferred orientation.

Parallel planar discontinuities can also be formed by oriented cracks and fractures. Cracks and fractures related for instance to the extensive stress field at ridges are certainly a possible cause for anisotropy in the upper oceanic crust (Stephen, 1981; Shearer and Orcutt, 1985). Anisotropy resulting from preferred to orientation of cracks has been modeled by Crampin (1978). Cracks are, however, unlikely to produce anisotropy over extensive areas in the uppermantle since at mantle pressures in excess of a few kilobars fractures are closed (e. g., Brace, 1964). Local anisotropy related to a zone of deep fracturing is however possible in the uppermost mantle (Evans, 1984), in particular if peridotites were locally transformed into schistose serpentinites (Ramana et al., 1981).

Anisotropy and Upper Mantle Flow

Studies on uppermantle peridotites from volcanic xenoliths and orogenic massifs show that structures reflect high to very high temperature flow conditions favoring the development of the strongest preferred orientations in olivine (Nicolas , 1976). Medium to low temperature deformations are restricted to local areas like the ~1 km thick margin of mantle diapirs (Coisy and Nicolas, 1978) or the 2-10 km thick shear domains ascribed to transform faults (Prinzhofer and Nicolas, 1980; Secher, 1981).

Pyroxenes have a diluting or negative effect on the anisotropy; the other minerals (garnet, serpentine) have simply a diluting effect. The strongest anisotropies are thus to be expected in dunites (95% olivine) and in harzburgites (~75% olivine). This will be further reenforced by the fact that an olivine-rich layer is more ductile than an olivine-poor one, due to the greater ductility of olivine and the greater facility for grain boundary diffusion and migration. Thus strain will concentrate in such a layer inducing stronger fabrics.

Oceanic sampling and ophiolite studies (Nicolas et al., 1980) suggest that dunites and harzburgites predominate in the oceanic lithosphere from the Moho discontinuity to depths of 25 km. They have been plastically deformed at the high to very high temperatures prevailing close to the ridge and their strain is very large mainly within the top 2 km (Nicolas and Prinzhofer, 1983; Rabinowicz et al., 1984). A highly anisotropic level is therefore to be expected just below the oceanic Moho.

Geometry of Uppermost Mantle Flow

The anisotropy of body waves measured so far at shallow depths in oceanic lithospheres results from a frozen structure and reflects a paleo-asthenospheric flow (see preceding section and Figure 9 about the limited extent and effect on anisotropy of subsequent low-temperature mantle deformations). This is not necessarily true for P-waves anisotropies at greater depths. For instance, the anisotropy beneath Germany extending at least to 50 km (Bamford, 1973, 1977; Fuchs, 1983) could well be related with present day flow beneath the Rhine graben. The anisotropy detected by polarization of surface-waves to depths of 450 km (Nataf et al., 1984) if, as we believe it is due to preferred orientations, certainly reflects the geometry of present-day mantle flow and furthermore indicates that dislocation-creep is active at those depths.

From systematic mapping of high temperature flow structures in ophiolitic massifs and inte-

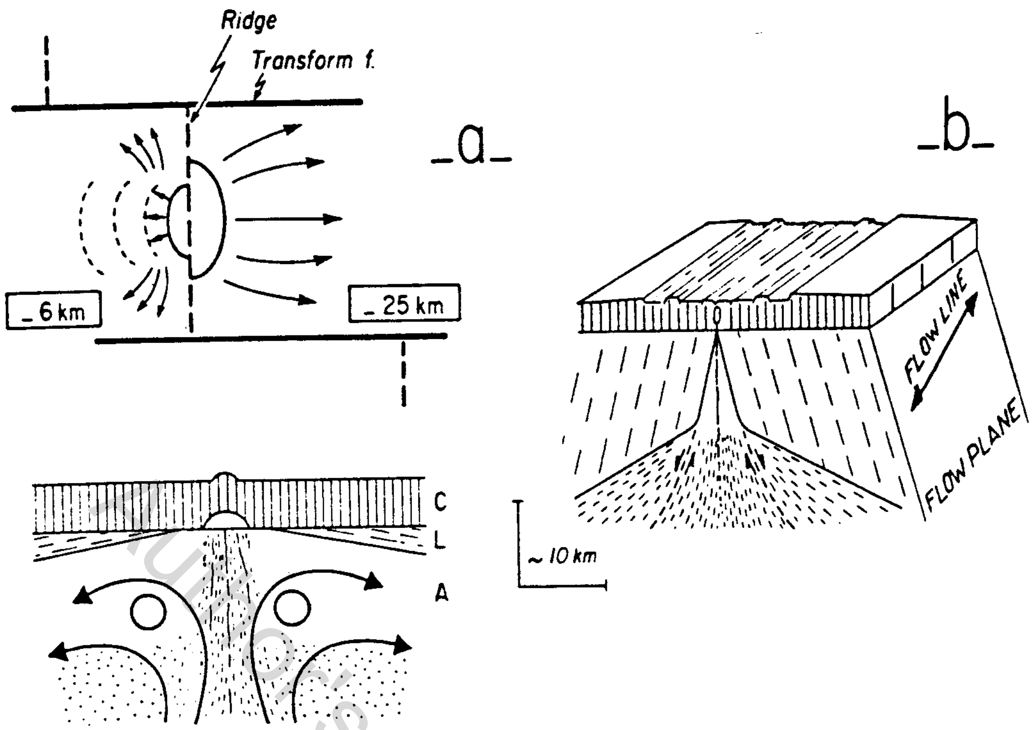


Fig. 10. Sketches of asthenosphere flow patterns at oceanic ridges for normal (a) and small (b) spreading rates. The bottom sketch in (a) is a cross section from Rabinowicz et al. (1984) showing the flat lying flow plane and the top sketch, a view in map inspired from Nicolas and Violette (1982), showing the divergence of flow lines immediately beneath the crust and at 20 km below it.

grating information over areas several thousands of km^2 in extension, it has been possible to propose models of the geometry of asthenosphere flow geometry beneath oceanic spreading centers (Nicolas and Violette, 1982; Rabinowicz et al., 1984) or along transform faults (Prinzhofer and Nicolas, 1980; Sécher, 1981). The main output of these studies for anisotropy is that the flow plane is generally subhorizontal, but that in the first kilometers below the Moho the flow line can be frozen with trends which are not always normal to that of the ridge. This is related to the mode of asthenosphere upwelling which at depths of a few tens of kilometers proceeds by the intrusion of diapirs, whose flow diverges in all directions just below the Moho. This flow is only progressively channelled parallel to the transform faults (Figure 10a). It is thus expected that for depths greater than 25 km below sea floor, the flow direction will be completely reoriented at a high angle to the ridge trend, reflecting the general asthenosphere motion. The vertically oriented flow structure of the diapir can be frozen in the first kilometers below the Moho, thus creating a band a few tens of kilometers in width and normal to the ridge with no distinct anisotropy (Figure 10a).

Below a spreading rate probably of the order of 1 cm/y (Boudier and Nicolas, 1985), this scenario is no more valid and the asthenospheric flow is frozen with a steep attitude reflecting the intrusion (Figure 10b). This situation is probably met in opening rifts. Surprisingly, in the few ophiolites with this signature and in the peridotites of the Zabargad Island in the Red Sea (Nicolas et al., 1985), the flow lineation has been found to be only moderately plunging (Boudier and Nicolas, 1986). This would generate a large anisotropy with the fast velocity parallel to the rift trend. It would be imprudent to generalize these findings but it can be observed that they are compatible with the anisotropy pattern measured in the Rhine graben area (Fuchs, 1983).

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