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Anisotropy of v_P and v_S in Rock-Forming Minerals

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Abstract. The compressional and shear wave anisotropy coefficients of 33 minerals of igneous and metamorphic rocks were calculated from published elastic constants and tabulated together with the orientation of velocity extremes in single crystals. The most abundant minerals of crustal crystalline rocks – alkali feldspars, plagioclases, quartz, micas and hornblende – have higher anisotropy coefficients than the upper mantle minerals – olivine, pyroxenes and garnets. Due to the orientation of mineral grains and their velocity extremes in a stress field, however, the olivine-rich ultramafites belong to the most anisotropic rocks and, in contrast, the crack-free anisotropy of crustal crystalline rocks is generally low, with the exception of metamorphic rocks rich in micas, hornblende and calcite.

Key words: Elastic wave velocities – Anisotropy coefficients – Rock-forming minerals – Preferred orientation

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Introduction

Elastic anisotropy, usually characterized by differences in compressional wave velocities, is one of the physical parameters which reflects the inner structure of rocks very sensitively. The large-scale seismic anisotropy of the continental upper mantle (Bamford 1977) and some regions of the Earth's crust (Burdock 1980) may yield important information on directions and intensity of the stress fields and geodynamic processes acting in geological history, providing the orientation of the structural elements, especially the preferred orientation of minerals at depth, is resolved from the in situ seismic observations.

There are, generally, only vague ideas among geophysicists about the degree of elastic anisotropy of the individual minerals. The purpose of this paper is, therefore, to calculate the coefficients of elastic anisotropy of the main rock-forming minerals from the elastic constants published by different authors and to discuss the possible elastic anisotropy and the preferred orientation of minerals in a stress field.

Methods of Investigation of Elastic Parameters

The experimental methods are either static or dynamic. The static methods are performed under isothermal conditions and are based on observations of static deformations (bending, twisting, compressing) of oriented crystals which mostly have the shape of a bar or a plate. The number of measurements needed to determine the elastic constants depends on the symmetry of crystals (Nye 1960). The accuracy of the static methods is given by the accuracy of determining small deformations (optical, electrical, mechanical methods) and by the size of the specimens. At present static methods are used mainly for the investigation of metals, glasses and construction materials which provide homogeneous samples of larger size.

The dynamic methods utilize high-frequency vibrations in crystals. Due to the very high frequencies, the process itself, as well as the elastic constants obtained, can be treated as adiabatic. Ide (1935) introduced the dynamic method of resonance vibrations of cylindrical specimens. The principal modes of vibration are extensional, flexural, and torsional. Recent sophisticated techniques can be used to determine the elastic properties of spherical or parallel-sided specimens as small as 1 mm (Goto et al. 1976). Furthermore, only one piece of properly shaped specimen is sufficient for determining all the elastic parameters, because complete information is included in the resonance spectrum of a specimen. Brillouin scattering measurements yield elastic properties of crystals even smaller than 0.1 mm (Weidner et al. 1975). The technique, however, has only a limited usefulness since the orientation of crystals is not known.

By pulse-transmission methods the transit time of a high-frequency pulse through a specimen is measured in special directions. The pulses, of frequencies from 50 kHz to 10 MHz (Anderson and Liebermann 1966) are generated by crystal or ceramic transducers attached to the specimen. The pulse-transmission technique is probably also the most widely used method of determining compressional and shear wave velocities in specimens of minerals and rocks at high pressures and temperatures. Another method of measuring sound velocity in crystals is the ultrasonic interferometry method. In this method internal reflections of the same wave train are made to interfere so that a pseudo-resonance can be achieved by suitably controlling the wavelength of the imposed pulse. A more detailed account of the methods can be found in Anderson and Liebermann (1966), and Belikov et al. (1970).

The choice of the method of investigation depends mainly on the size and the quality of the specimen and the possibilities for experimental work. Large crystals very often contain various inclusions, cracks and pores, and the results are then less reliable than with relatively small but gem-quality crystals.

Anisotropy of Elastic Wave Velocities

Table 1 contains the elastic parameters of rock-forming minerals – the maximum and the minimum velocities of compressional and shear waves, as well as the coefficients of elastic anisotropy. The elastic constants can be found in the papers referenced, or, in the handbook by Simmons and Wang (1971), for most of the listed minerals. The coefficient $k=(v^{\max}-v^{\min})/v^{\text{mean}}$, where v^{mean} is an average value of the maximum and the minimum velocities. The corresponding directions of wave propagation and the polarization directions of the shear waves are given by Miller indices and those directions which cannot be defined by simple indices, are determined by the angles φ and ϑ (see Fig. 1). The original data are definitely of varying quality, depending mainly on the quality of the investigated crystals and the method of measurement. As mentioned in the previous section, the accurate determination of the elastic parameters requires crystals of a reasonable size. This is one of the reasons,

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Table 1. Elastic wave velocities (v_P, v_S) and coefficients of elastic, anisotropy (k) in rock-forming minerals

Mineral	Symmetry	Density ($g \cdot cm^{-3}$)	Compressional waves					
			v_P^{max} (km/s)	Propagation direction	v_P^{min} (km/s)	Propagation direction	v_P^{mean} (km/s)	k%
Ortho- and Ring Silicates								
Olivine (Fo 92. 72)	Orthorhombic	3.31	9.89	[100]	7.72	[010]	8.81	24.6
Zircon	Tetragonal	4.70	9.00	[001]	7.41	[010]	8.21	19.4
Garnets								
Pyrope (PY-1)	Cubic	3.705	8.90	[001]	8.85	$\vartheta=55^\circ$ $\varphi=45^\circ$	8.88	0.6
Almandine (AL-4)		3.930	8.76	[001]	8.73	$\vartheta=55^\circ$ $\varphi=45^\circ$	8.75	0.3
Spessartine (SP-1)		4.172	8.55	[001]	8.54	$\vartheta=55^\circ$ $\varphi=45^\circ$	8.55	0.1
Grossular (GR-1)		3.659	8.95	[001]	8.74	$\vartheta=55^\circ$ $\varphi=45^\circ$	8.85	2.4
Kyanite	Triclinic	3.62	9.88	[001]	9.00	[100]	9.44	9.3
Topaz	Orthorhombic	3.52	10.10	$\vartheta=90^\circ$ $\varphi=56.25^\circ$	9.15	[001]	9.63	9.9
Staurolite	Monoclinic	3.369	10.09	[100]	6.61	[001]	8.35	41.7
Epidote	Monoclinic	3.40	8.38	[010]	7.24	[011]	7.81	14.6
Beryl	Hexagonal	2.64	10.61	[100]	9.76	$\vartheta=36^\circ$ $\varphi=0^\circ$	10.19	8.3
Tourmaline	Trigonal	3.10	9.38	$\vartheta=94^\circ$ $\varphi=33.75^\circ$	7.29	[001]	8.34	25.1
Chain Silicates								
Pyroxenes								
Bronzite	Orthorhombic	3.34	8.30	[100]	7.04	[010]	7.67	16.4
Diopside	Monoclinic	3.31	8.60	[001]	6.94	[101]	7.77	21.4
Hedenbergite		3.41	6.55	[001]	6.26	[010]	6.41	4.5
Augite		3.32	8.36	[101]	6.81	[010]	7.59	20.4
Aegirine		3.50	8.21	[001]	6.75	[101]	7.48	19.5
Hornblende	Monoclinic	3.15	8.13	[001]	6.18	[101]	7.16	27.2
Sheet Silicates								
Muscovite	Monoclinic	2.79	8.06	[110]	4.44	[001]	6.25	57.9
Biotite		2.89	7.80	[010]	4.21	[001]	6.01	59.7
Framework Silicates								
Feldspars								
Microcline	Triclinic	2.56	8.15	[010]	5.10	[100]	6.63	46.0
Orthoclase	Monoclinic	2.54	7.64	[010]	4.76	[101]	6.20	46.4
Albite (An 9)	Triclinic	2.61	7.26	[010]	5.31	[101]	6.29	31.0
Oligoclase (An 29)		2.64	7.55	[010]	5.50	[101]	6.53	31.4
Labradorite (An 53]		2.68	7.80	[010]	6.06	[100]	6.93	25.1
Anorthite		2.76	8.61	[010]	6.01	[101]	7.31	35.6
Quartz	Trigonal	2.66	7.00	$\vartheta=130^\circ$ $\varphi=90^\circ$	5.36	$\vartheta=72^\circ$ $\varphi=90^\circ$	6.18	26.5
Nepheline	Hexagonal	2.62	7.12	[001]	5.61	[100]	6.37	23.7
Non-Silicates								
Rutil	Tetragonal	4.264	10.65	[001]	8.00	[100]	9.33	28.4
Spinel ($MgO \cdot 3.5\ Al_2O_3$)	Cubic	3.63	10.31	[101]	9.10	[001]	9.71	12.5
Magnetite		5.18	7.46	$\vartheta=55^\circ$ $\varphi=45^\circ$	7.29	[001]	7.38	2.3
Pyrite (2)		5.013	8.40	[001]	7.78	[110]	8.09	7.7
Calcite	Trigonal	2.717	7.55	$\vartheta=70^\circ$ $\varphi=90^\circ$	5.43	[001]	6.49	32.7

$$k=\frac{v_P^{max}-v_P^{min}}{v_P^{mean}}\cdot 100\%$$

Shear waves								Source of data
v_S^{max} (km/s)	Propagation direction	Polarization direction	v_S^{min} (km/s)	Propagation direction	Polarization direction	v_S^{mean} (km/s)	k%	
5.53	$\vartheta=45^\circ$ $\varphi=0^\circ$	$\vartheta=135^\circ$ $\varphi=0^\circ$	4.42	[010]	[100]	4.98	22.3	Kumazawa and Anderson 1969
4.87	[010]	[001]	2.94	[110]	[110]	3.91	49.4	Ryzhova et al. 1966
4.96	[101]	$[10\bar{1}]$	4.91	[001]	[110]	4.94	1	Babuška et al. 1978
4.96	[110]	$[1\bar{1}0]$	4.91	[001]	$[1\bar{1}0]$	4.94	1	
4.82	[110]	$[1\bar{1}0]$	4.80	[001]	[110]	4.81	0.4	
5.45	[101]	$[10\bar{1}]$	5.20	[001]	[110]	5.33	4.7	
6.76	[010]	[001]	4.24	[001]	[100]	5.50	45.8	Belikov et al. 1970
6.15	[001]	[100]	5.16	$[\bar{1}10]$	$\vartheta=90^\circ$ $\varphi=41.25^\circ$	5.66	17.5	Voigt 1928
5.29	$\vartheta=118^\circ$ $\varphi=45^\circ$	$\vartheta=62^\circ$ $\varphi=315^\circ$	3.23	$\vartheta=140^\circ$ $\varphi=45^\circ$	$\vartheta=50^\circ$ $\varphi=86.7^\circ$	4.26	48.4	Aleksandrov and Ryzhova 1961a
5.11	[011]	[011]	3.39	[001]	[010]	4.25	40.5	Ryzhova et al. 1966
6.25	$\vartheta=44^\circ$ $\varphi=0^\circ$	$\vartheta=136.7^\circ$ $\varphi=0^\circ$	5.34	[001]	[110]	5.80	15.7	Hearmon 1956
6.14	$\vartheta=82^\circ$ $\varphi=33.75^\circ$	$\vartheta=88.5^\circ$ $\varphi=124^\circ$	4.55	$\vartheta=172^\circ$ $\varphi=33.75^\circ$	$\vartheta 389.4^\circ$ $\varphi=119.4^\circ$	5.35	29.7	
4.99	[010]	[001]	4.27	$\vartheta=90^\circ$ $\varphi=45^\circ$	$\vartheta=90^\circ$ $\varphi=135^\circ$	4.63	15.6	Kumazawa 1969
4.83	[011]	$[0\bar{1}1]$	3.94	[110]	$[1\bar{1}0]$	4.39	20.3	Aleksandrov et al. 1963 Volarovich et al., 1975
4.72	[011]	$[0\bar{1}1]$	3.81	[100]	[001]	4.27	21.3	Aleksandrov et al. 1963
4.65	[011]	$[0\bar{1}1]$	3.48	[010]	[100]	4.07	28.7	
4.60	[011]	$[0\bar{1}1]$	3.37	[001]	[100]	3.99	30.8	Aleksandrov and Ryzhova 1961b
5.01	[010]	[100]	2.03	[001]	[100]	3.52	84.7	Aleksandrov and Ryzhova 1961c
5.06	[010]	[100]	1.34	[010]	[001]	3.20	116.3	Belikov et al. 1970
4.96	[011]	$[01\bar{1}]$	2.14	[010]	[001]	3.55	79.4	Aleksandrov and Ryzhova 1962
4.45	[011]	$[01\bar{1}]$	2.33	[001]	[010]	3.39	62.5	Ryzhova and Aleksandrov 1965
4.63	[011]	$[01\bar{1}]$	2.59	[001]	[010]	3.61	56.5	Ryzhova 1964
4.70	[011]	$[01\bar{1}]$	2.65	[001]	[010]	3.68	55.7	
4.76	[011]	$[01\bar{1}]$	2.70	[001]	[010]	3.73	55.2	
4.96	[011]	$[01\bar{1}]$	2.91	[010]	[100]	3.94	52.0	Aleksandrov et al. 1974
5.06	[100]	$\vartheta=147^\circ$ $\varphi=284.5^\circ$	3.35	[100]	$\vartheta=122^\circ$ $\varphi=90^\circ$	4.21	40.1	Huntington 1958
3.92	[011]	$[01\bar{1}]$	2.83	[100]	[010]	3.38	32.2	Ryzhova and Aleksandrov 1962
6.74	[100]	[010]	3.32	[110]	[110]	5.03	68.0	Birch 1960
6.61	[001]	$[1\bar{1}0]$	4.50	[101]	$[10\bar{1}]$	5.56	37.9	Verma 1960
4.29	[001]	$[1\bar{1}0]$	4.06	[101]	$[10\bar{1}]$	4.18	5.5	Hearmon 1956
5.66	[110]	$[\bar{1}\bar{1}0]$	4.70	[001]	[110]	5.18	18.5	Aleksandrov and Ryzhova 1961d
4.77	$\vartheta=126^\circ$ $\varphi=90^\circ$	[100]	2.66	$\vartheta=36^\circ$ $\varphi=90^\circ$	[100]	3.72	56.7	Hearmon 1956

why the elastic constants of single crystals are lacking for important minerals such as chlorite and serpentine.

The minerals in Table 1 are arranged from ortho- and ring silicates, via chain silicates, sheet silicates, and framework silicates to non-silicates, as is usual in systematic mineralogy. Of the group of *ortho- and ring silicates*, olivine and garnets are among the most abundant rock-forming minerals in the lithosphere; they are important constituents of the upper mantle rocks. In contrast to olivine, garnets show the lowest anisotropy of all the minerals listed. Due to a frequent and well-developed preferred orientation of olivine, the olivine-rich ultramafites usually have a relatively high degree of velocity anisotropy, of about 10% for the compressional waves (Babuška 1972a). An admixture of garnets in these ultramafites tends to decrease their elastic anisotropy.

In the important group of *chain silicates* the published data enable the elastic anisotropies to be calculated for only five pyroxenes and one amphibole (hornblende). As regards *sheet silicates,* the elastic constants of single crystals can be determined reliably only for the main members of the mica group. The extremely high values of the coefficients of anisotropy in micas are explained by their structure in which a layer of octahedrally coordinated cations is sandwiched between two identical layers of linked (Si, Al) O_4 tetrahedra (Deer et al. 1966). A remarkable anisotropy of the physical properties, including a perfect basal cleavage, is a consequence of the great difference in the strength of interatomic bonds in the directions parallel and perpendicular to the layered atomic structure. The perfect cleavage, however, may decrease the accuracy of determining the elastic parameters and may cause a considerable scatter of data.

Framework silicates include the most abundant minerals of the Earth's crust – feldspars and quartz. Among feldspars the potassium members belong, along with the mica group, to the most anisotropic minerals. In the plagioclase feldspar series the coefficients of *S*-wave anisotropy are substantially higher than the coefficients for the *P* waves. The elastic anisotropy of quartz, though lower in comparison with feldspars, still classes it with the rock-forming minerals with high coefficients of anisotropy. This is in contrast to the generally very low coefficients of the crack-free elastic anisotropy of the granitic rocks (Birch 1961; Babuška et al. 1977).

Among *non-silicates* probably only calcite, in some regions of the uppermost crust, and spinel in the spinel peridotites may sporadically form substantial volumetric parts of the lithosphere. The high elastic anisotropy of calcite, together with its preferred orientation, results in a high velocity anisotropy for marbles (Babuška 1968). Due to the low *P*-wave anisotropy of spinel, any admixture of the mineral in peridotite or lherzolite would decrease the anisotropy of the rocks.

With the exception of two pyroxenes (bronzite, diopside) and olivine, all the other minerals listed in Table 1 have coefficients of anisotropy of the shear wave velocities higher than the coefficients of the compressional velocities. The largest differences in the *P*- and *S*-wave anisotropy coefficients are observed for micas, feldspars and several other minerals (kyanite, spinel, epidote, rutile, zircon).

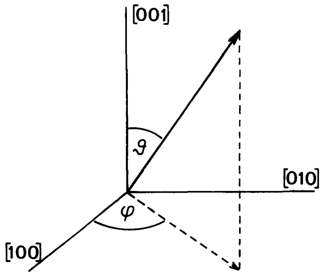

Fig. 1. System of determination of the propagation and the polarization directions given in Table 1

Discussion

Let us take from Table 1 the most abundant minerals or mineral groups of crystalline rocks and arrange them in a series according to the values of the coefficient of compressional wave anisotropy. This results in the following sequence: micas ($k = 59\%$), alkali feldspars (46%), calcite (33%), plagioclase (31%), hornblende (27%), quartz (26%), olivine (25%), pyroxenes (16%), garnets (1%). There is no significant change in this sequence if we consider the coefficients of the shear-wave anisotropy. It is remarkable that the last three members of the sequence are the main components of the upper mantle rocks. The olivine-rich rocks like dunite and peridotite, however, belong to the most anisotropic rocks. In contrast, granite or granodiorite, being composed of highly anisotropic components and often carrying signs of tectonites, including mineral-preferred orientation, generally have low coefficients of crack-free anisotropy (Babuška et al. 1977).

What then are the main factors which determine the degree of the elastic anisotropy of crystalline rocks? There are three: the values of the anisotropy coefficients of the constituent minerals and their volume in the aggregate; the character and degree of the preferred orientation of minerals and other fabric elements (e.g. oriented micro-discontinuities, alternating layers); the orientation of extremes of the elastic wave velocities in single crystals with respect to the directions which are significant for their orientation in a stress field.

Besides garnets, the anisotropy of which is negligible, all other very abundant minerals named above can contribute, according to their single crystal anisotropy, to the elastic anisotropy of crystalline rocks.

Experiments on the preferred orientation of *mica* carried out by Tullis (1971), showed that during crystallization or recrystallization the (001) plane mean orientation is always normal to the direction of the greatest shortening which, in most cases, coincides with the maximum compressive stress σ_1. The compressional velocity minimum in micas is perpendicular to the (001) plane and the velocity maximum is parallel to this plane (see Table 1). Thanks to these orientations and to the extremely high single crystal anisotropy, micas contribute substantially to the elastic anisotropy of aggregates. This is true of a number of metamorphic rocks such as mica schists and gneisses.

Deformation of *feldspars* and the question of the origin of their preferred orientation is not a simple problem and the data are insufficient in this respect. Borg and Heard (1970) in their experimental work on deformation of plagioclases found that, if it can be ascertained that the albite and pericline twins are mechanical in origin, it is theoretically possible to deduce the orientation of the principal stresses which produced them. It is, however, difficult to prove the mechanical origin of twin lamellae in naturally deformed, plagioclase-bearing rocks. In the case of such mechanical twins the direction of maximum stress σ_1 is at acute angles (close to 45°) to the crystallographic directions $[0\bar{1}0]$ and [001]. However, the maximum velocity in all feldspars is in the [010] direction, the velocity minimum in either [100] or [101] directions. Therefore, we cannot expect any alignment of the velocity extremes and the resulting *P*-velocity anisotropy by this mechanism of preferred orientation.

Experimental deformation of *calcite* aggregates was investigated in a number of studies which can be summarized as follows: in syntectonic recrystallization the *c* axes tend to be subparallel to the principal stress σ_1, or, the *c* axes concentrations are inclined at 20°–30° to the σ_1 direction (Turner and Weiss 1963). Because the *c* axis is the direction of the velocity minimum (also the direction of high compressibility, Birch 1966) and the *P*-wave velocities in the $x-y$ plane are very close to the maximum velocity (deWitte 1962), the directions of the velocity extremes are often preferrentially oriented, which leads to the commonly observed high elastic anisotropy of marbles.

The preferred orientation of coarse grained *hornblende* in metamorphic rocks is often observed. Schwerdtner (1964) found that the hornblende orientation in a deformed gneiss from Norway corresponds to a pattern predicted by the theory of Kamb (1959), i.e., the *c* axes tend to be oriented perpendicular to the axis of greatest compressional stress. The *c* axis is also the direction of maximum *P*-wave velocity and thus the oriented hornblende grains can contribute to an increase in the elastic anisotropy of rocks. Both in hornblende and in micas the external shape plays an important role in orientation and the resulting anisotropy of metamorphic complexes (Burdock 1980).

The investigation of the preferred orientation of *quartz* aggregates shows a complex picture of the different types of orientation developed under different experimental conditions. At lower temperatures or faster strain rates a maximum of the *c* axes of the α-quartz develops parallel to the compression direction σ_1(*c*-maximum fabric), and at higher temperatures or slower strain rates a concentration of the normal to *r* $(10\bar{1}1)$ develops parallel to σ_1(*r*-maximum fabric, Green et al. 1970). At intermediate temperatures and strain rates, the preferred orientation is a combination of both fabric types (composite fabrics). According to Green et al. (1970) at high temperatures and low strain rates, orthorhombic strain due to temperature gradients produces 'crossed girdles' of *c*-axes intersecting normal to σ_1. Taking into consideration these possibilities for the quartz orientations, which are, of course, still incomplete, and the fact that the *P*-wave velocity extremes in quartz do not coincide either with the *c*-axis or with the $[10\bar{1}1]$ direction, we cannot expect any pronounced elastic anisotropy of rocks due to the orientation of quartz grains.

Olivine in syntectonically recrystallized aggregates tends to be oriented with the [010] direction parallel to σ_1 and [100] parallel to σ_3. This orientation conforms both to Kamb's thermodynamic theory (Kamb 1959) and the experiments carried out on natural dunites by Avé Lallemant and Carter (1970). Also other mechanisms like plastic flow or shear along transverse faults, which were discussed, e.g., by Nicolas and Poirier (1976), lead to a distinct preferred orientation of olivine crystals. Olivine has a *P*-velocity anisotropy of about 25% with the maximum velocity along the [100] direction and the minimum value along [010]. These directions tend to be oriented along the minimum and the maximum stresses, respectively. Such favourable conditions result in the fact that a significant elastic anisotropy of olivine-rich ultramafites is a rule rather than an exception.

The experiments of Carter et al. (1972) on syntectonic recrystallization of *orthopyroxene* (enstatite) showed that recrystallized grains have preferred orientations with [010] oriented parallel to σ_1 and the remaining principal axes arranged in girdles in the $\sigma_2-\sigma_3$ plane. The fabric is similar to that of olivine. The orientation of velocity extremes in orthopyroxene (Bronzite, see Table 1) is conformable to olivine. Therefore, provided olivine and orthopyroxene recrystallized in the same stress field, the anisotropies due to both minerals are additive for the anisotropy of rocks in bulk (Kumazawa et al. 1971). However, Helmstaedt et al. (1972) observed in a nodule of lherzolite that the crystallographic axes *a* of enstatite and those of olivine form maxima perpendicular to each other, these corresponding to the contrasting non-additive anisotropies of both minerals. As the preferred orientations of orthopyroxene are often weaker than those of olivine (Nicolas and Poirier 1976; Babuška 1972b) and the anisotropy coefficients of orthopyroxene are smaller, the resulting anisotropy of orthopyroxene-rich rocks is relatively low.

Kumazawa et al. (1971) observed that, despite a significant preferred orientation of *clinopyroxenes*, the velocity anisotropy of eclogites from diatremes of the Colorado plateau is considerably smaller than the anisotropy of peridotites. Although the coefficients of anisotropy of the omphacitic pyroxene have not been determined yet, we can suppose a value of about 20%, as for augite. On the basis of the thermodynamic theory of Kumazawa (1963), Kumazawa et al. (1971) assume that the preferred orientation of clinopyroxenes is controlled by their linear compressibility. Since in clinopyroxenes the directions of extreme linear compressibilities are different from the directions of the compressional velocity extremes, a high velocity anisotropy in eclogites cannot be expected.

Although the shear-wave anisotropy of most rock-forming minerals is larger than the compressional anisotropy, the shear anisotropy of aggregates is a complicated problem, because of the great differences in the shear velocities with different polarizations. For example, in olivine the maximum difference due to polarization for a single direction or propagation is as much as 0.46 km/s (Kumazawa and Anderson 1969) and in dunite, which is composed of 98% of olivine, the difference for a single direction reaches 0.38 km/s (Babuška 1972b), this means 10% and 8%, respectively, of the mean velocity in the corresponding directions. The shear-wave birefringence was also observed in a number of metamorphic rocks (Tilmann and Bennet 1973) and it is believed that polarization studies using three-component records will show diagnostic anomalies in the presence of in situ seismic anisotropy (Keith and Crampin 1977).

Conclusions

According to the elastic anisotropy coefficients the most abundant rock-forming minerals of igneous and metamorphic rocks form the following sequence (the values in brackets are the compressional and the shear-wave anisotropy coefficients in per cent, respectively): micas (59, 100), alkali feldspars (46, 71), calcite (33, 57), plagioclases (31, 55), hornblende (27, 31), quartz (26, 40), olivine (25, 22), pyroxenes (16, 21), garnets (1, 2).

Feldspars, quartz and clinopyroxene, the minerals which are characterized by high anisotropy coefficients, form aggregates with generally low anisotropy, although their grains are often preferentially oriented. This is probably because the velocity extremes in these minerals differ from the directions which are decisive for their orientation in a stress field. The velocity extremes are thus not aligned according to directions of the maximum and the minimum stresses, as observed, e.g., in the aggregates of olivine. Consequently, this means that we cannot expect a high velocity anisotropy in crustal igneous rocks. On the other hand, metamorphites rich in micas, hornblende and calcite have high coefficients of elastic anisotropy.

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