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Structure and Petrology of an Alpine Peridotite on Cypress Island, Washington, U.S.A

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With 23 Figures in the Text

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Abstract. A study of the structural petrology of a peridotite exposed on Cypress Island in Skagit Co., Washington, U.S.A. has been carried out. The Cypress peridotite is, by virtue of its composition, structure and associations, typical of ultramafics of the alpine type. It contains relict layering and accumulative textures which show it to have originated by crystal settling from a magma of unknown initial composition. Parallelism of lineations and *b*-axes of folds in the layering with well-developed [100]-maxima in the fabrics of olivine crystals is considered to have arisen through a penetrative deformation of the mass accompanied by plastic flow or recrystallization of the olivine. An indication of the minimum temperature of the deformation is provided by cross-cutting veins of pyroxenite which have not participated in the folding. The most satisfactory interpretation of the overall fabric of the peridotite is that it was deformed, possibly during intrusion, as a crystal mush, and that filter pressing due to compaction of the solid particles by plastic flow or recrystallization removed all but a small percentage of the magmatic fraction which then crystallized following cessation of the movements.

A. Introduction

Several features of peridotites of the alpine type set them aside from the generality of igneous rocks and have provoked considerable speculation as to their origin. Paradoxically, although they are composed of the most refractory suite of minerals in igneous rocks, the country rocks around their contacts in most cases show no strong heating effects. The contacts are, in fact, generally faulted. The peridotites are commonly foliated and the chromitite layers lineated. Microscopically, undulatory extinction in the olivines provides additional evidence of penetrative deformation.

Compositionally, the peridotites of the alpine type have high Mg/Fe ratios and are impoverished in lime and alumina compared to ultramafic bodies of the stratiform type. The peridotites, associated with spilitic pillow basalts, cherts and greywackes are located exclusively in orogenic belts, both in island arcs and in mountain chains of a variety of ages. These characteristics as well as numerous others have been summarized by THAYER (1960) and have been noted by HESS (1938, 1955) and BOWEN and TUTTLE (1949) among others in attempting to account for the origin of the peridotites. There is no very general agreement between different investigators, but it seems clear from BOWEN and TUTTLE's (1949) experimental work that the peridotites cannot have been intruded into their present positions either as dry or hydrous magmas. Both HESS (1955) and BOWEN and TUTTLE conclude, therefore, that emplacement by solid or crystal mush flow followed by flow in a marginally serpentinized zone is probable for at least some alpine peridotites. DE ROEVER (1957) proposes faulting of a shallow peridotitic upper mantle into eugeosynclinal sediments. At the other extreme, BAILEY and MCCALLIEN (1953) and MAXWELL and AZZAROLI (1962) find evidence that some alpine ultramafics originate by submarine extrusion of ultrabasic lava. CLARK and FYFE (1961) have shown that at H_2O pressures consistent with emplacement in ocean deeps a partially liquid peridotite can exist at temperatures of 1300° C or greater. Later tectonism may be assumed to give rise to the discordant, faulted nature of the serpentinite blocks observed *in situ.*

The fact that so many workers have found evidence which brings them to such varied conclusions may well be due to variation in the modes of origin of the peridotites. One field observation which is reported fairly consistently, however, is that alpine peridotites are deformed. This aspect of the problem has not been treated in very great detail before and therefore is given special attention in this study of an alpine peridotite. The evidence from examination of its microscopic textures, preferred orientation of minerals and macroscopic structures points strongly toward a mechanism of intrusion which may be of wider applicability.

Acknowledgements. The research on which this paper is based was carried out in partial fulfillment of a Ph. D. program at the University of California at Los Angeles. Financial support was granted by the Shell Foundation and the National Science Foundation and the field work was supported by grants from the Penrose Bequest of the Geological Society of America and the Society of Sigma Xi. The author has benefitted from discussions with Professor D. T. GRIGGS, Drs. J. M. CHRISTIE, E. D. JACKSON and D. H. GREEN.

B. General Geology

1. Location and Previous Work

The Cypress peridotite crops out over the southern two-thirds of Cypress Island, a member of the San Juan Islands group in the northwestern Washington (Fig. 1) Serpentinite makes up several small islands including Burrows, Allan, the westernmost peninsula of Fidalgo Island and a few other smaller islands at a somewhat greater distance from the main mass (McCLELLAN, 1927). The work reported here was confined to outcrops on the islands named above except for a cursory examination of the serpentinized diopside pyroxenite occurring on Blakely Island (Fig. 1). McCLELLAN mapped the San Juan Islands and named the formations in 1927. A more recent map of the area based on McCLELLAN's work and revisions from several unpublished sources appears on the Washington State Geological Map (HUNTTING 1960). Cypress Island was remapped by the author.

The ultramafic rocks in the field area were named the Fidalgo Formation by McCLELLAN (1927) from the occurrence of serpentinite on Fidalgo Island (Fig. 1). Fresh peridotite is exposed in the area only on Cypress Island where the largest continuous mass of ultramafic rocks occurs. Therefore, the name, Cypress peridotite will be applied to the ultramafic rocks exposed on Cypress Island and serpentinites occurring on islands to the south will be referred to individually by the name of the island. There is no direct evidence for the various ultramafic occurrences having once been part of the same mass, although their proximity and the lack of other rock types in these occurrences suggests continuity.

McCLELLAN (1927) used the name, Leech River Formation, for the greywackes and black shales occurring on Cypress Island and elsewhere in the San Juan Islands (Fig. 1). Mapping by the author indicates that the sediments and the volcanic rocks, the Eagle Cliff porphyrite of McCLELLAN, were deposited contemporaneously and deserve the same formation name. Therefore, until a new name is proposed from a better exposed section, the name Leech River Formation will be retained for both the sedimentary and volcanic members.

2. *Geologic Setting*

The Cypress peridotite lies on a broad belt of ultramafic intrusions extending along the west coast of North America from Alaska to Southern California (Noble and Taylor, 1960). The Cypress mass has close affinities to the other major alpine peridotites in western Washington — the Twin Sisters dunite (Ragan, 1963) and several smaller bodies to the southeast. The vertically zoned ultramafic intrusions of southeastern Alaska in the western part of the belt are different both in structure and in overall composition from the Cypress peridotite and others in the eastern part of the belt.

The rocks of the San Juan Island region consist of a pre-Devonian basement overlain by sediments ranging from Devonian to Eocene in age. Amphibolite

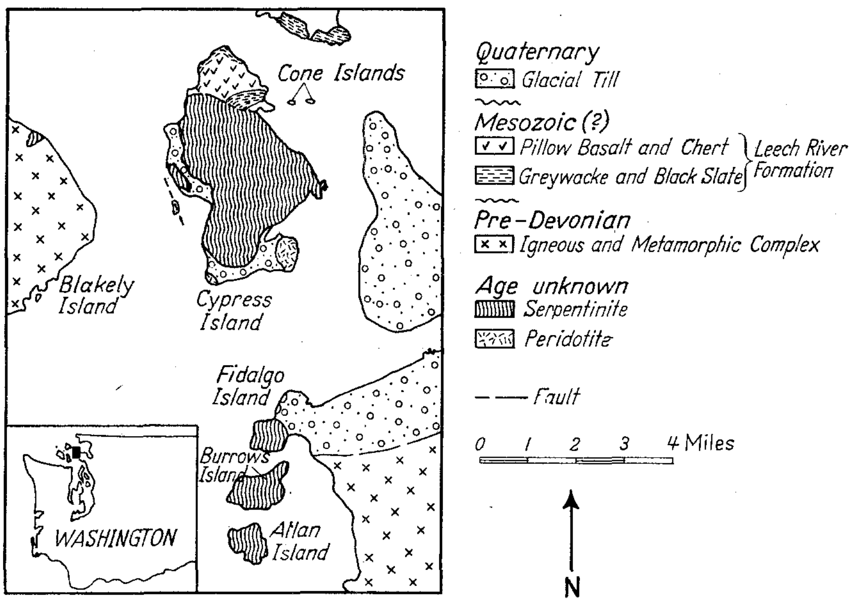

Fig. 1. Geologic map of the eastern San Juan Islands area of northwestern Washington. Geology taken in part from McClellan (1927) and from mapping by the author

gneiss and plutonic igneous rocks together with cross-cutting diabase and dacite dikes compose the basement (Vance and Ellis, in ms.) which was named the Turtleback complex by McClellan (1927). The complex has been subjected to a low-grade regional metamorphism which Vance and Ellis consider contemporary with the metamorphism of the Easton Formation on the mainland.

The rocks most closely associated with the peridotite on Cypress Island are pillow basalts, chert and greywacke of Mesozoic (?) age (Vance, personal communication). Occurrences of such rocks, lying in northwesterly trending belts, are commonplace throughout the Pacific coastal ranges. The greywackes in the map area (Fig. 1) were folded, presumably during a late Cretaceous orogeny mentioned by McClellan (1927).

3. *Leech River Formation*

Outcrops of volcanic and sedimentary members are found on the northern part of Cypress Island and on the small Cone islands bordering Cypress on the east. On the easternmost of the Cone islands spilitic pillow basalt lies conformably above and below slate of the sedimentary member of the formation. The basalts contain albite, altered clinopyroxene, pumpellyite, chlorite and calcite. On the south of Cypress Island volcanics and sedimentary rocks are in fault contact with serpentinite. The sedimentary rocks, consisting of black slates, greywacke, pebble conglomerate and grey or white banded chert, are almost unaffected by metamorphism. Thin sections of the greywacke and pebble conglomerate show a preponderance of angular quartz grains and fragments of chert, volcanic rocks and some shale.

4. Basalt Dikes

The ultramafic rocks are intruded in several places by dikes of basalt which are not found in the country rock. The dikes are ten to twenty feet thick and commonly are faulted along their margins. In thin section the dikes from the serpentinite contain partly altered plagioclase and augite in an intersertal texture whereas in the single dike occurring in the peridotite the labradorite and augite are unaltered.

5. Cypress Peridotite

Unaltered peridotite is best exposed along low coastal cliffs around the outcrop areas shown on the map (Fig. 1). Toward the north on Cypress Island the peridotite shows a gradual increase in the content of serpentine, being wholly serpentinized at the country-rock contact. On the map the boundary between peridotite and serpentinite is drawn at a content of about 15 per cent of serpentine, an amount sufficient to render the peridotite entirely black on a fresh surface, a useful criterion for mapping purposes. Both the peridotite and serpentinite are blocky and massive in outcrop.

Harzburgite containing 10—20 per cent of enstatite is the major rock type. Dunite, minor olivine chromitite layers, along with cross-cutting veins of dunite and enstatite pyroxenite constitute the remainder of the peridotite. Olivines throughout the peridotite have compositions ranging from Fo_{89} to Fo_{93} and the enstatite near En_{90} ($2V_Z = 80°$). The olivines were determined from the X-ray curve of Hotz and Jackson (1963). Diopside ($2V_Z = 56°$, $Z \wedge \underline{c} = 38°$) is present only locally in harzburgite, and constitutes less than $^1/_2$ per cent of the rock in every case. The chromite varies in colour from a translucent reddish brown to an opaque black variety which shows reddish colours on a thin edge.

The dunite, harzburgite and olivine chromitite bands occur in alternating layers with contacts which may be sharp or gradational. Large, massive bodies of dunite or peridotite also occur. The chromite grains are disseminated throughout the ultramafics but are also distributed in planar segregations from one grain to a few cm in thickness. The segregations are generally confined within layers of dunite and lie parallel to the dunite-harzburgite contacts. In a few cases, cuspate contacts between chromitite and dunite layers may be observed (Fig. 2). The compositional layering is irregular and discontinuous. A single layer of chromitite or a contact between dunite and harzburgite can rarely be traced more than about 15 metres. The dunite and harzburgite layers vary in thickness from 1 cm to 1 to 2 metres and the thicknesses are commonly variable along the outcrop of a given layer.

The dunite veins which cut the layering are a few cm in thickness and do not carry chromite in planar segregations as do dunites in the regular layered sequence. Much more common are the veins of pyroxenite which transect the layering at various angles and commonly cut another. These veins, up to 30 metres long ranging from one cm to nearly one metre in width, contain enstatite and emerald green diopside grains up to several cm in length. The diopside is generally segregated in the middle of the thicker veins.

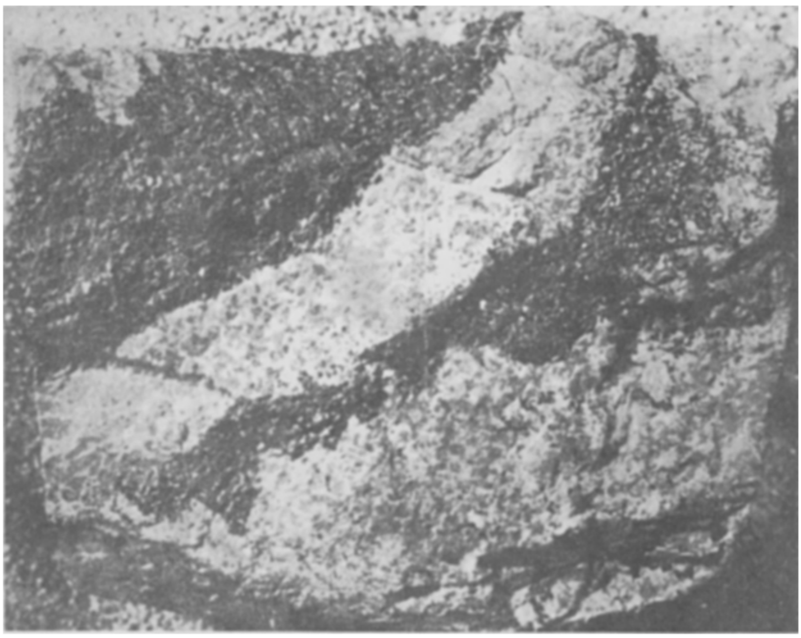

Fig. 2. Hand-specimen of interlayered chromitite and dunite. Note cuspate margin of the lower chromitite layer. Natural size

Within the serpentinite all the structures including folded layering and veins are preserved. Faults, which are common in the peridotite, are very numerous in the serpentinite. The ultramafic rocks are foliated immediately adjacent to faults but elsewhere are quite massive. The fault dividing serpentinite from the country rock in the northern part of Cypress Island is marked by a zone 10 metres wide consisting of well-foliated serpentinite.

C. Structure

1. Geometry of Layering in the Ultramafic Rocks

Determination of the structure of the layering presents problems similar to those encountered in structural studies of some metamorphic rocks. The principal difficulty is the lack of stratigraphic marker units continuous and large enough to be traced over significant distances and mapped. To overcome this lack, the techniques used in the structural analysis of metamorphic tectonites (TURNER and WEISS, 1963) have been utilized. Such analysis provides information about the spatial distribution of structural elements such as fold axes or foliations having a given orientation and also, indirectly, about the symmetry of the movements which gave rise to the structures. In complexly folded rocks geometrical analysis of such structural data may be the only means of unravelling the history of the deformations (TURNER and WEISS, 1963)

The nature of the exposures in these ultramafic rocks makes the gathering of a very complete set of structural data difficult; weathering of the inland outcrops renders the streaky chromitite layering nearly invisible. The dunite-harzburgite contacts show up well only in some coastal exposures of serpentinite and, in general, are not well-defined. In many localities around the low coastal sea cliffs the chromitite layers stand out well, however, and consequently, the statistically significant data on orientations of layering come from these areas.

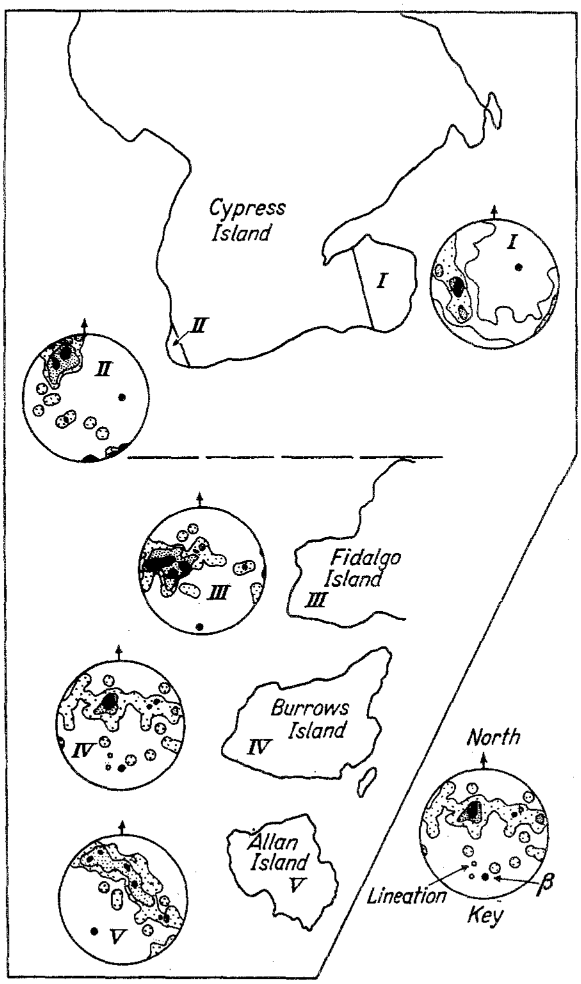

Fig. 3. Geometry of chromitite layering in the Fidalgo ultramafics. The chromitite layering in each subarea is approximately homogeneous with respect to β, the fold axis defined by the normal to the great-circle girdle of poles to the layering. Number of measurements of poles to layering and value of contours in subareas: I: 132; 1, 3, 6, 9% per 1% area. II: 37; 3,6, 12% per 1% area. III: 75; 1, 4, 8% per 1% area. IV: 66; $1^1/_2$, 6, 12% per 1% area. V: 75; 1, 4, 8% per 1% area

Small folds in the layering are locally visible, but are scarce and the lineations are also rarely apparent because the rocks do not part preferentially along chromitite layers. The structural data available then, consist mainly of orientations of the chromitite layers and are geographically restricted to coastline exposures. The lack of data from individual folds and certain parts of the area precludes a complete structural synthesis but some important results still emerge.

The measurements of orientation of layers were taken on a field notebook held in the plane of the layer. An edge of the notebook was placed along the trace of a layer on an outcrop surface; by turning the book around that edge a pencil lying along another trace of the layer could be brought into coincidence with the plane of the notebook thereby orientating the notebook parallel to the layer. The orientations thus determined are probably subject to errors of ± 5 degrees at best. The measurements were plotted on the lower hemisphere of an equal-area projection[1] as poles to the layers and were combined over areas where the layering was homogeneous with respect to a fold axis designated by β, of one orientation. β is defined as the pole to a great circle on the projection which best fits the girdle concentration of poles to the planar surface studied. β, therefore, is a line contained by all orientations of the layering in an area which the layering is homogeneous with respect to β, and may be considered as the statistical fold axis in that area.

The poles of the layering are plotted and contoured in groups according to areas in which the layering is homogeneous with respect to β or nearly so. In all the subareas except III the poles lie in rather diffuse but clearly marked great-circle girdles (Fig. 3).

In local divisions of the subareas the statistical fold axis may differ from β for the subarea as a whole by 20 degrees or so. The variations in β from one local area to another are not generally consistent with the direction of traverse. Partial diagrams of the orientation of layering in subarea I (Fig. 4) were prepared by combining data from the strips of coastline shown in Fig. 5. In order to locate the normals, β, to the girdles of poles accurately, the intersections of the great circles representing the orientation of each plane were plotted separately and contoured.

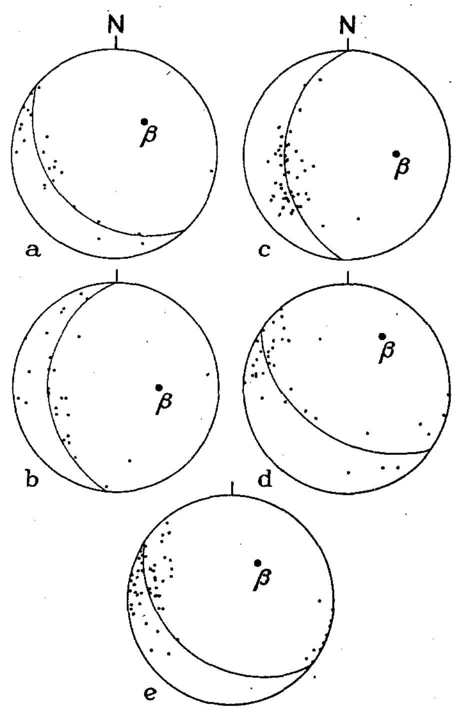

Fig. 4. Poles to chromitite layering in subarea I from local areas, a, b, c, and d shown in Fig. 5. e: poles to dunite-harzburgite layering in local area, a

The resulting maxima of these β diagrams locate the normals, β, to the girdles of poles to the layering planes (TURNER and WEISS, 1963, p. 155). Attitudes were also taken on dunite-harzburgite contacts in subarea Ia and are shown, in Fig. 4e, to have a similar orientation statistically, to the chromitite layers in that area.

2. Small Folds and Lineations in the Layering

Only a few exposures of small folds in the chromitite layers have been found in the area. The folds sketched in Fig. 6 are typical. The style is isoclinal with the thickened hinges and attentuated limbs characteristic of shear folds (KNOPF and INGERSON, 1938). A few isoclinal folds with amplitudes on the order of a few metres are visible in the dunite-harzburgite layers in subarea Ia.

Lineations having the orientations given in subarea IV of Fig. 3 were of two types. One consisted of a rod of serpentinite enclosed in a chromitite layer; the

[1] In all equal-area projections presented in this paper the data are plotted on the lower hemisphere. The primitive circle is the horizontal plane and the arrow at the top is North.

others were defined by an alignment of chromite grains alternating with chromite-free lineaments in the plane of the layering.

3. *Geometry of the Pyroxenite Veins*

Analysis of the geometry of the enstatite veins presents several difficulties. As was previously noted, the veins at some localities are gently folded, although folds are not common. The folds are not of the same style as the folds in the

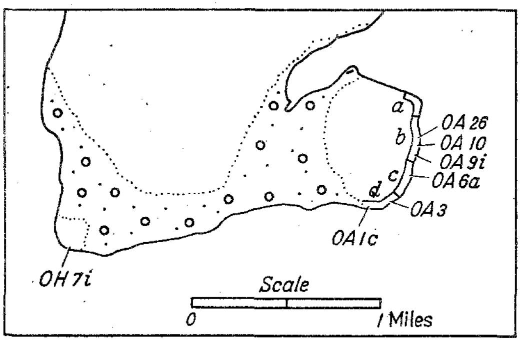

Fig. 5. Sketch map of southern part of Cypress Island showing location of subdivisions of subarea I and specimens collected for fabric analysis

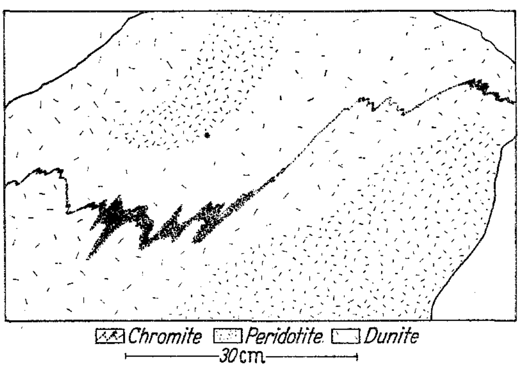

Fig. 6. Profile of small folds in chromitite layer within dunite. Viewed looking southwest

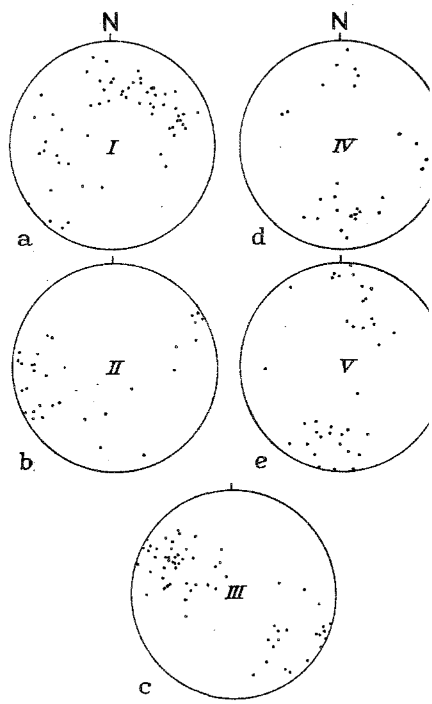

Fig. 7. Poles to pyroxenite veins in the ultramafic rocks

chromite layering; they are very open, gentle folds and it is not clear from their geometry whether they are of the flexural-slip or shear type. Intersecting veins are fairly common, indicating that the veins were initially emplaced in two or more sets. It is not known whether one set was dominant nor whether the veins in the sets were initially parallel. If the initial orientations of the veins were random and the veins were folded, plots of the poles to the veins would possibly still be random. The axes of small folds in the veins, however, should lie in a plane parallel to the slip plane if shear-folded.

The attitudes of the pyroxenite veins were measured throughout the area and plotted separately for the same subdivisions of the area as the chromite layering. The poles of the veins plotted on equal-area nets are shown in Fig. 7. The locations of the subareas correspond to those in Fig. 3. In subareas II, III, IV and V the poles of the veins tend to be concentrated in maxima although some tendency towards spreading into a great circle is apparent.

Where individual pyroxenite veins were folded the axes of the folds were measured or constructed from measurements of the attitudes of the vein around the fold. The axes in subareas I, III and V, all that are available, are plotted on equal-area projections in Fig. 8.

4. Folding in Sedimentary Rocks

Black slates and siltstones of the Leech River Formations cropping out on the Cone islands are folded and lineated. The folds are open, with amplitudes of a few metres. Intersections of cleavage parallel to the axial plane of the folds with the bedding give rise to the lineations. Cleavage having a similar orientation to that on the Cone islands has nearly obliterated the bedding in slates and greywackes on Cypress Island. In Fig. 9, a projection of the structural data from the sedimentary rocks, the average of six

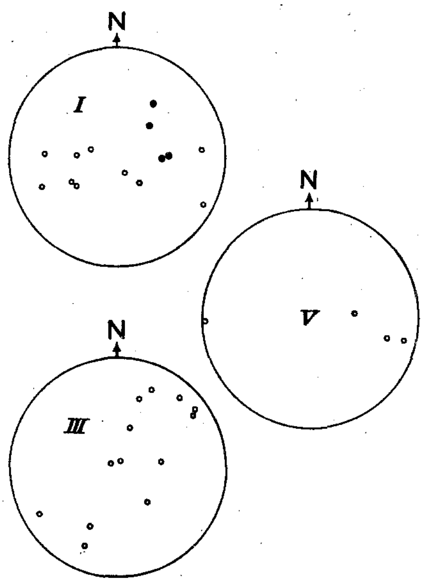

Fig. 8

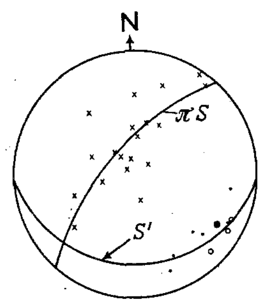

Fig. 9

Fig. 8. Axes of small folds in pyroxenite veins (open circles) in subareas I, III and V. Solid circles are β-axes of chromitite layering in local areas of subarea I

Fig. 9. Synoptic diagram of structural data from sedimentary rocks of Leech River Formation on Cypress Island and Cone Islands. Great circle, S', represents average orientation of cleavage. π-great circle contains poles to bedding (×) and is approximately normal to lineations defined by intersections of cleavage and bedding (●), fold axes (○) and long axes of stretched pebbles (⊕)

measurements of the orientation of the cleavage on Cypress Island and the Cone islands is plotted as a single great circle, S′. The fold axes measured and shown in Fig. 9 have approximately the same orientation as the large folds in Mesozoic sediments in the surrounding San Juan Islands region.

5. Discussion

The parallelism of lineations and small-fold axes with the fold axes, β, derived geometrically from the orientations of layering indicate that the peridotite underwent a penetrative deformation with folding of the layering around an axis, β. The variation in the orientation of β suggests that either 1. the layering was not parallel throughout the peridotite prior to the deformation, 2. that movement directions varied from place to place during the deformation, or 3. that later deformation has disoriented the geometry acquired during the first deformation. There is evidence from the orientation pattern of veins that later deformation has taken place. Whether later or earlier, however, the secondary folding of the layering was on a broad scale as indicated by the homogeneity of the layering in the subareas with respect to β. Had the secondary folding been intense and local, a markedly triclinic symmetry of the fabric of the layering would have been produced.

There is considerable evidence which indicates that the pyroxenite veins did not participate in the penetrative deformation resulting in isoclinal folding of the layering. Folds in the veins are broad, gentle, open structures in contrast with the appressed style of the folds in the layering. More important, the orientations of the axes of the folds in the veins bear no relationship to the fold-axes in the layering which can be accounted for by a single deformation. The spread of the fold-axes in the veins in subarea I along a great-circle girdle might be accounted for by shear folding in a plane parallel to the girdle or by compression normal to the girdle, in each case acting upon randomly oriented planar veins (Weiss and McIntyre, 1957). The fold axes in the layering in the sub-divisions of subarea I, however, tend to spread along a great cricle nearly normal to that defined by the axes in the veins (Fig. 8a). It is considered, therefore, that the veins were emplaced after the layering was folded and that a deformation of a weak, local nature produced the minor folds in the veins.

The geometry of the folding in the sedimentary rocks of the nearby Leech River Formation appears unrelated to the folding of the layering or the pyroxenite veins in the peridotite. No cleavage other than that adjacent to the faults is present in the peridotite and therefore the ultramafics are considered to have been folded in a deformation different from that suffered by the sedimentary rocks.

D. Microscopic Textures in the Ultramafic Rocks

1. Dunites and Harzburgites

Anhedral olivine and enstatite in the harzburgites average 3 to 4 mm in diameter although, locally, some olivine grains may reach several cm. The chromite grains are uniformly small, about 0.5 to 1 mm, where disseminated as anhedral grains throughout the ultramafics or as euhedral inclusions in olivine grains. The interstitial chromite grains may be polygonal or have irregular shapes controlled by the shape of the interstitial space. Within the chromite layers rounded, polygonal chromites attain diameters of 5 mm. The chromite grains tend to be equant whereas the olivines commonly are highly elongated with irregular grain boundaries (Fig. 10). Enstatite grains are equant to slightly elongate and have smooth and regular but sinuous grain boundaries (Figs. 11, 13), except where in contact with fine recrystallized olivine or where alteration to a monoclinic twinned amphibole occurs along the boundary. Inclusions of olivine occur in over half the enstatite grains whereas enstatite in olivine occurs rarely enough to be attributable to thin section effects. Embayments of one into the other are fairly common (Fig. 11). These textural relations between olivine and enstatite are similar to the reaction-replacement textures in the granular harzburgites of the ultramafic zone of the Stillwater complex (Jackson, 1961).

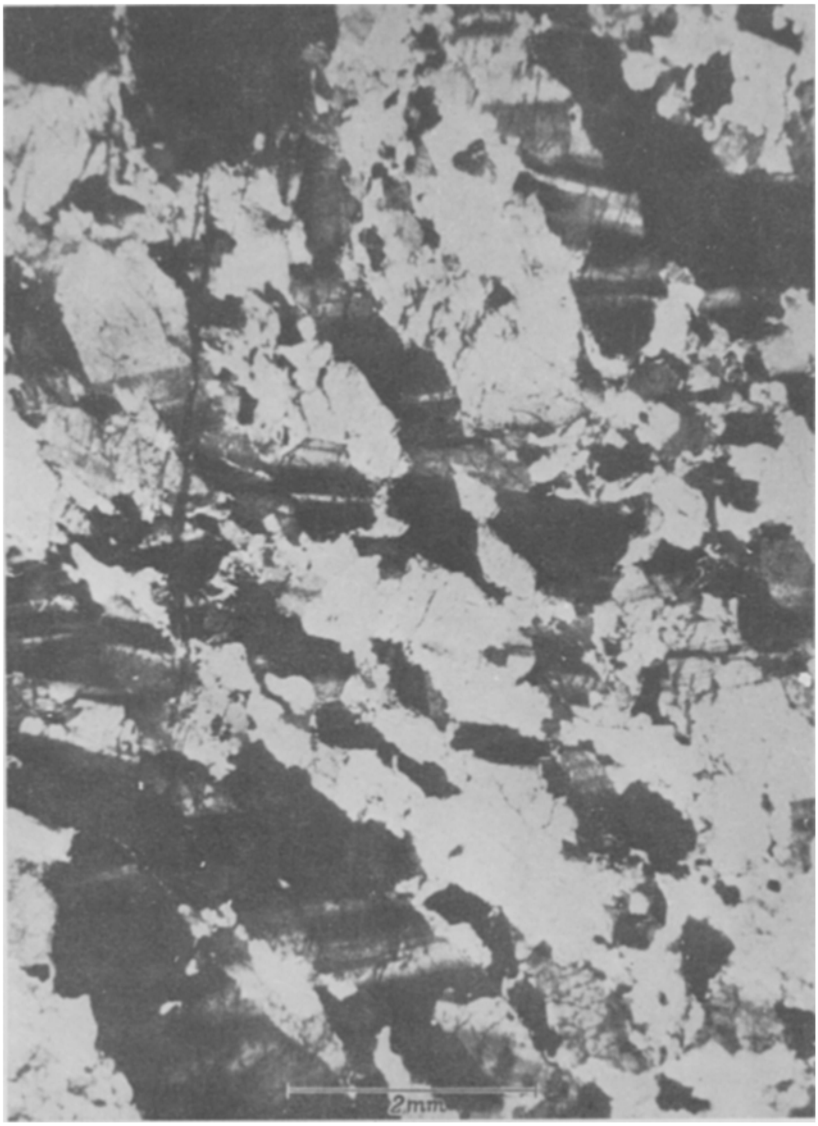

Fig. 10. Elongated anhedral olivine grains in thin section of harzburgite specimen 0A10. Extinction bands trending N-S are kink band approximately parallel to (100). Crossed polarizers

In several thin sections the olivines can be divided into two groups on the basis of grain size, one at 1 to 5 mm, the other at 0.01 to 0.2 mm. The smaller grains lie along the boundaries of the larger and embay some larger grains along kink bands parallel to (100). The texture described is suggestive of partial recrystallization with nucleation of new grains at grain boundaries and in regions of severe plastic strain.

2. *Olivine*

Subparallel bands of differing extinction position approximately parallel to (100) have been recorded in the individual olivine grains of alpine peridotites by

Fig. 11. Enstatite (central, dark grain) texture against olivine. Crossed polarizers

many workers and are profusely developed in the Cypress peridotite. CHUDOBA and FRECHEN (1950) correctly indentified them as kink bands resulting from slip in the [100] direction. Narrow lamellae lying parallel to the kink bands are of the same optical character as the kink-band boundaries. As the lamellae also separate lattices of slightly differing orientation they are considered to originate in the same manner as the kink band boundaries, i.e., by slip normal to the lamellae rather than parallel to them as was suggested earlier by TURNER (1942). A secondary set of kink bands approximately parallel to (001) appears in grains in which the Y optic direction is close to the plane of the section (Fig. 12). Kink

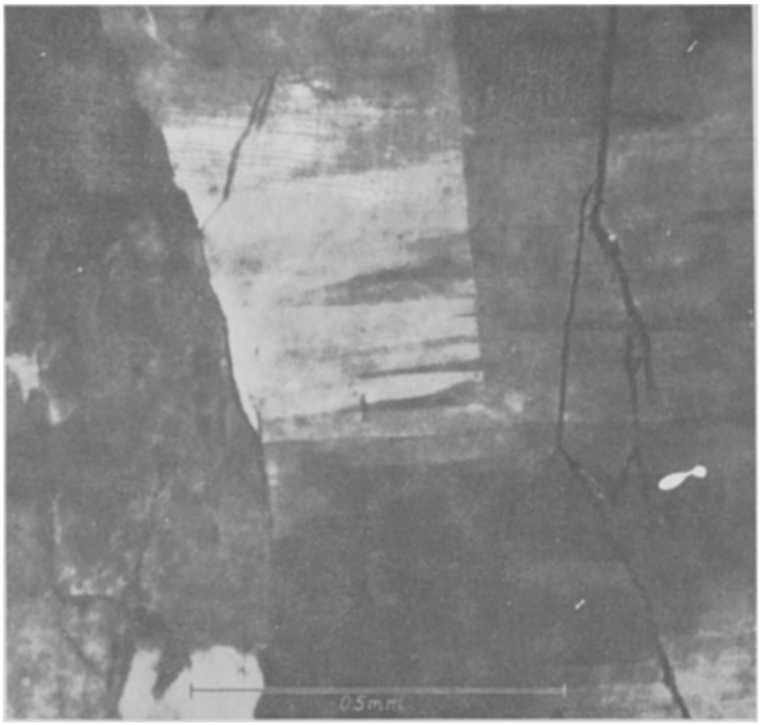

Fig. 12. (001) kink band boundary (N-S in photo) at right angles to (100) kink bands in olivine. Crossed polarizers

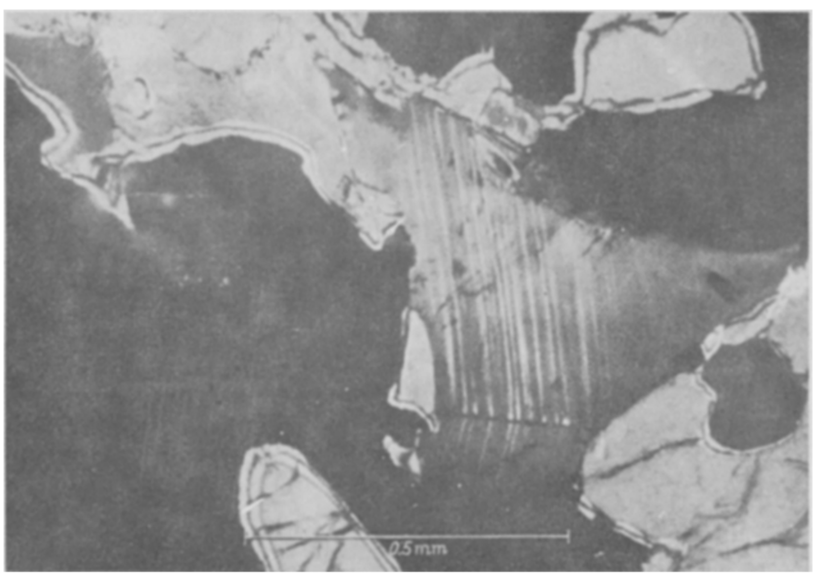

Fig. 13. Kinked enstatite showing exsolution lamellae parallel to (100) on either side of kink band boundary, Crossed polarizers

bands having this orientation have been produced experimentally (Raleigh, 1965) and result from inhomogeneous slip on {110} planes in the [001] direction.

3. Enstatite

Enstatite grains in the peridotite and in pyroxenite veins commonly contain kink bands (Fig. 13) bounded by irrational crystallographic planes containing [010]. The boundaries are symmetrical to (100) and lie at large angles to [001] in the lattice on either side. Fine lamellae parallel to (100), presumably exsolved diopsidic pyroxene, maintain their crystallographic orientation in the kink bands. Slip on the plane, (100), in the direction, [001] has given rise to the kinking as in the experimentally deformed enstatites of TURNER, HEARD and GRIGGS (1960) and RALEIGH (1965). Evidence for the inversion to clinoenstatite in the kink bands and in lamellae parallel to (100) observed experimentally is lacking in these specimens.

The exsolution lamellae on (100) could be observed in every suitably oriented grain of enstatite in thin sections of the peridotite and the pyroxenite veins. In one grain the optical orientation of blebs of exsolved material having the same extinction position as the lamellae were measured on a universal stage. The Y optic direction of the blebs was parallel with X of the host enstatite; also, Z in the blebs lay at an angle of 38° to [001] in the host. The measured orientations are similar to those of HESS (1960) and are in accord with his conclusion that exsolution lamellae having these properties are diopsidic pyroxene having their (010) plane and [001] axis in common with that of the enstatite host.

4. Serpentinite

The serpentine in partly serpentinized peridotite occurs along cracks in the olivine and enstatite grains. In specimens consisting predominantly of serpentine, olivine occurs as small islands in groups having continuous optical orientation. In the massive serpentinite, which constitutes the bulk of the serpentinites, mesh networks of chrysotile and lizardite surround cores of isotropic material. In addition to the serpentine phases, brucite is present to about 2% and magnetite to about 3% by weight of the rock (RALEIGH and PATERSON, 1965).

E. Preferred Orientations of Minerals

1. Olivine

Oriented specimens of peridotite and dunite from a number of localities on Cypress Island (Fig. 5) were sectioned for fabric analysis. In one specimen 150 grains were measured from three sections at right angles to one another to test the homogeneity of the fabric. It was found, as TURNER (1942) and BATTEY (1960) concluded, that 50 grains were sufficient to show the essential features of the fabric. The grains to be measured were selected at the intersections of a grid superimposed on photomicrographs of the thin sections. Where possible, the grid points were spaced at distances greater than the size of the largest grains to avoid repeated measurements on single grains. The grid pattern had the same ratio of length to width as the average grain shape in the plane of the section to avoid Schnitzeffekt (JONES, 1959). The data from each rock were referred to the horizontal plane and contoured with a counter having an area 1 per cent of that of the equal-area net. The maxima for the 50-grain samples are consequently stronger than they would be had larger numbers of grains been measured or a counter with larger radius been employed (KAMB, 1959).

The specimens for which fabric analyses are presented are peridotites from subareas I and II from localities where measurements of the attitudes of layering are available. Other fabrics determined are given by RALEIGH (1963). The olivine fabrics, in general, are characterized by a single strong [100]-maximum normal

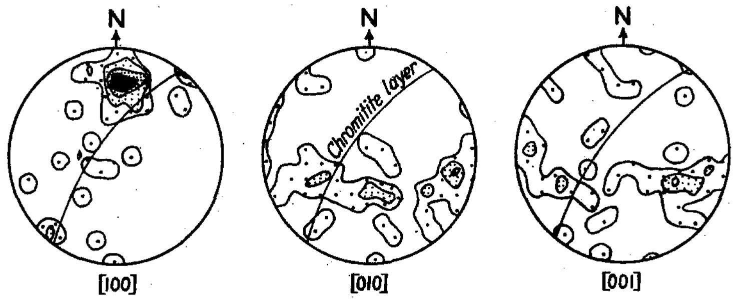

Fig. 14. Preferred orientation of 50 [100], [010] and [001] axes of olivine in specimen 0A1C. Contours 2, 6, 10, 18% per 1% area

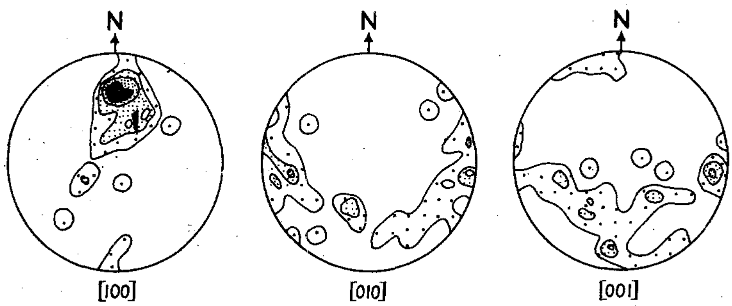

Fig. 15. Preferred orientation of 50 [100], [010] and [001] axes of olivine in specimen 0A3. Contours 2, 6, 10, 14% per 1% area

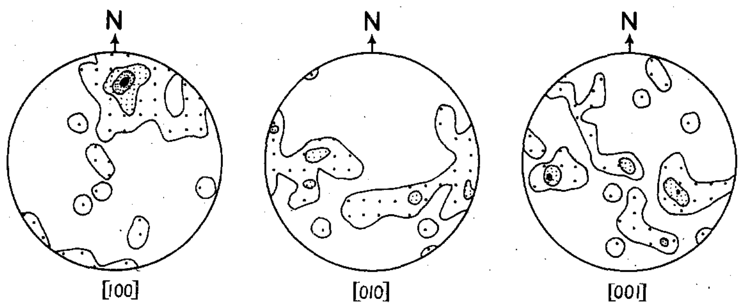

Fig. 16. Preferred orientation of 50 [100], [010] and [001] axes of olivine in specimen 0A6a. Contours 2, 6, 10, 14% per 1% area

to girdles of [010] and [001] axes (Figs. 14—19). In a few of the specimens, the [100]-maxima are not significantly stronger than the strongest [010] and [001]-maxima. The [010] and [001] girdles contain maxima and are incomplete in most cases, even in specimen 0A10, in which 150 grains were measured.

The fabric diagrams for specimen 0A10 (Fig. 18) include measurements of grains which appear to have been recrystallized. The orientation of the small recrystallized grains when separated from the larger grains and replotted give a diffuse fabric in which the strong [100]-maximum is lacking. This is a further indication of their recrystallized nature.

The [100]-maxima in rocks from subareas I and II are approximately parallel to the axis of folding of the chromite layering within the local areas from which

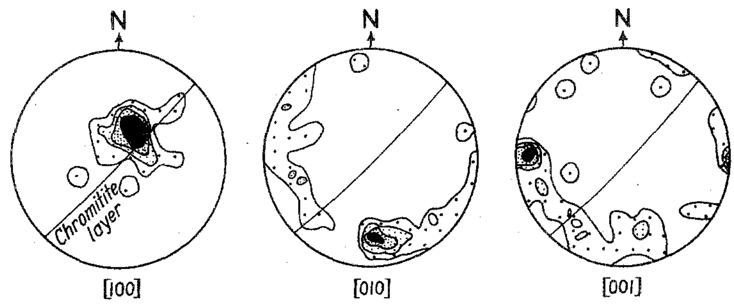

Fig. 17. Preferred orientation of 50 [100], [010] and [001] axes of olivine in specimen 0A9i. Contours 2, 6, 10, 14% per 1% area

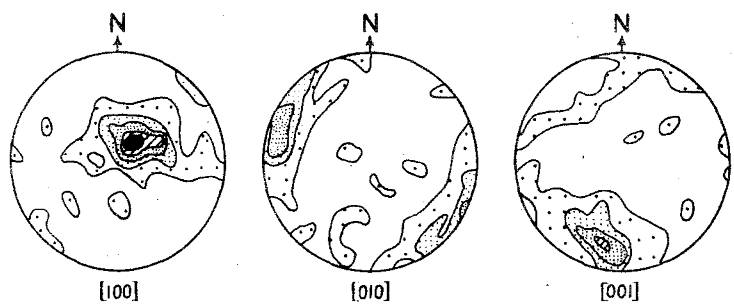

Fig. 18. Preferred orientation of 150 [100], [010] and [001] axes of olivine in specimen 0A10. Contours $1^1/_3$, 3 5, 7, 10% per 1% area

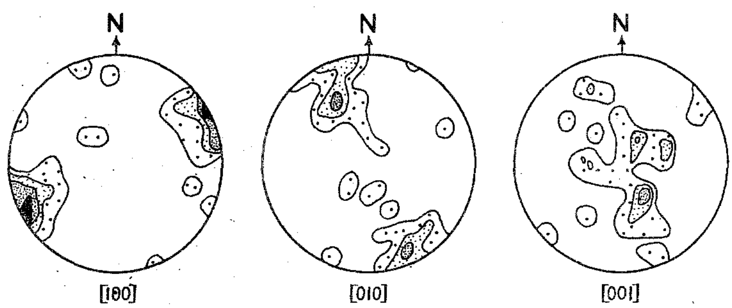

Fig. 19. Preferred orientation of 50 [100], [010] and [001] axes of olivine in specimen 0H7i. Contours 2, 6, 10, 14% per 1% area

the rocks were collected. In Fig. 20a, the [100]-maximum from specimen 0H7i (Fig. 19) lies parallel to the axis of a small fold in the layering (the attitudes of the layering are shown as great circles in the diagram). In subarea I the [100]-maxima of several specimens are approximately parallel to the statistical fold axes, β, defined by the normals to the π-great circles from the subdivisions of subarea I discussed previously (Fig. 20b).

In addition to this parallelism of the [100]-maxima with the fold axes it was found that the long axes of the olivine grains in specimen 0A10 were also parallel to the [100]-maxima. The grains in this specimen could be shown to have statistically the shape of triaxial ellipsoids with axial ratios, $Z:Y:X = 2.3:1.0:0.7$. The shape and orientation of the triaxial ellipsoid was determined by examination of aspect ratios of the grains in several thin sections, one of which was parallel to a circular section (aspect ratio of 1:1) and another which intersected the Y—Z principal plane in approximately the Z direction (Raleigh, 1963).

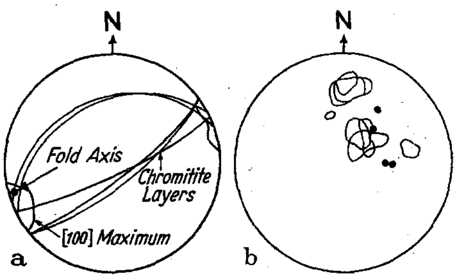

Fig. 20

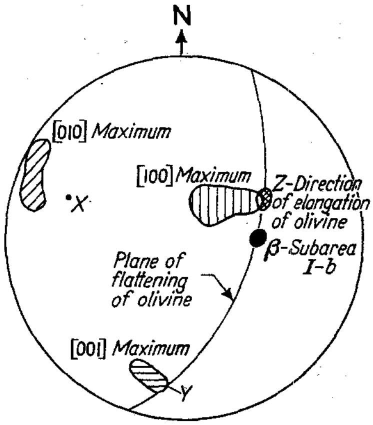

Fig. 21

Fig. 20. a [100] axis maximum in specimen 0H7i in relation to attitudes of nearby chromitite layers and the axis of a small fold. b [100] maxima from specimens in subarea I and β-axes from local areas a, b, c, and d within subarea I

Fig. 21. Maxima of [100], [010] and [001] axes in specimen 0A10 in relation to orientation of major (Z), intermediate (Y) and minor (X) axes of ellipsoid representing the average shape of the grains. The fold axis, β, from the local area in which 0A10 occurs is approximately parallel to Z and the [100] maximum

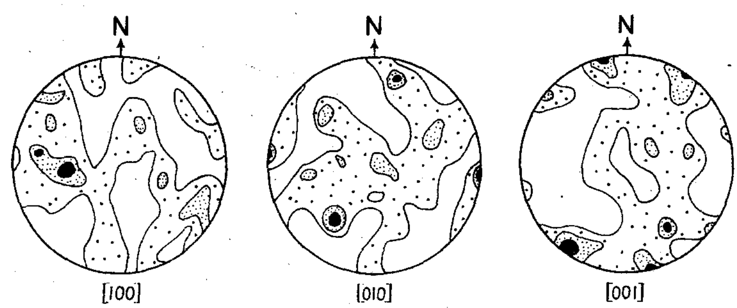

Fig. 22. Preferred orientation of 100 [100], [010] and [001] axes of enstatite from specimen 0A10. Contours 1, 3, 5% per 1% area

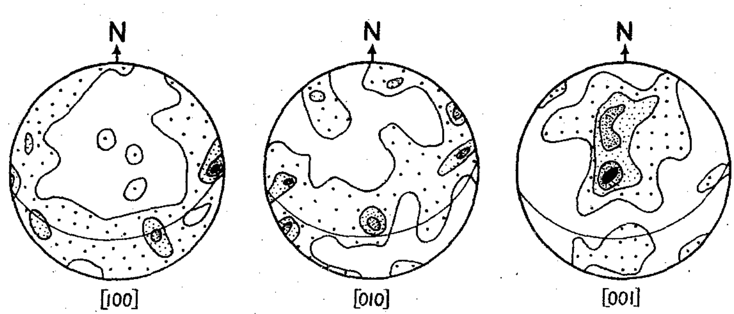

Fig. 23. Preferred orientation of 100 [100], [010] and [001] axes of enstatite from specimen (0A26) of a pyroxenite vein. Great circle gives orientation of vein wall. Contours 1,3, 5, 7% per 1% area

The least and intermediate axes, X and Y, of the triaxial ellipsoid were found to lie parallel to the principal [010]- and [001]-axis maxima of the fabric (Fig. 21). Similar relations between the grain shape and the preferred lattice orientation exist, in a qualitative way, in two other of the specimens examined. However, in other specimens no preferred shape orientation is apparent despite the strong preferred lattice orientations which they display (e.g. specimen 0A9). In none of the rocks does enstatite show any marked shape anisotropy.

2. *Enstatite*

The X, Y and Z optic directions, parallel to [010], [100] and [001] respectively were measured in 100 grains of enstatite in specimen 0A10. The fabric, shown in Fig. 22, consists of several scattered maxima of the crystal axes lying in diffuse girdles. No symmetrical relationship to the olivine fabric in the same specimen can be seen to exist.

The fabric of 100 enstatite grains from a pyroxenite vein (specimen 0A26) located nearby specimen 0A10 is shown in Fig. 23. The [100] and [010] axes lie in a plane parallel to the vein wall and the [001] axes lie at a large angle thereto. As the section was taken from within 5 cm of the vein wall it is probable that the preferred orientation developed by crystal growth controlled by the wall of the vein.

F. Discussion

1. *Origin of the Textures and Preferred Orientation and their Relation to the Tectonic History*

Observations on the nature of the layering in the Cypress Island peridotite indicate that it has a primary origin by crystal settling by analogy with textures in the Stillwater stratiform ultramafic body (JACKSON, 1961).

1. The layers of chromite, dunite and peridotite, though discontinuous, are parallel and the chromitite layers are generally confined within dunitic, rarely within peridotitic layers.

2. Cuspate textures at the contacts between chromitite and dunite layers (Fig. 2) are present as in the Stillwater, although they are uncommon.

3. Enstatite grains replace olivine, as shown by embayments of one into the other and inclusions of olivine in enstatite (Figs. 11, 12), presumably by reaction between settled olivine grains and the interprecipitate magma. The texture is similar to that of the granular harzburgites of the ultramafic zone of the Stillwater (JACKSON, 1961, p. 48).

Other minor details of the textures are comparable such as the observation that olivine contains no inclusions other than euhedral chromite, and that interstitial chromites have shapes which would result from secondary enlargement to fill interstitial volumes. In many other respects, however, the analogy is incomplete; this can probably be attributed to differences in the initial composition, rate of cooling, ambient pressure and temperature conditions, and the effects of later deformation. The lack of continuity in the layering also is attributable to the subsequent folding of the rocks, a process which in metamorphic rocks commonly gives rise to thickening and thinning or disruption of layers.

The folding of the primary layering and the parallelism of the fold axes with lineations in the chromite layers and [100] maxima in the olivine fabrics have arisen during a single penetrative deformation of the peridotite mass. The olivine grains achieved their preferred lattice orientation by plastic flow and ensuing rotation or by syntectonic recrystallization in the deforming mass.

The veins of pyroxenite were emplaced at some time following this penetrative deformation, the structural analysis having shown that the veins were not folded by the early deformation. The temperature at which the veins were emplaced thus sets a lower limit to the temperature at the time of folding of the layering if it is assumed that the rocks were not cooled and then re-heated to that temperature. The minimum temperature of formation of the veins can be taken as about 650° C, the approximate temperature below which hydrothermal alteration of olivine by addition of silica, the mechanism of pyroxenite vein formation proposed by BOWEN and TUTTLE (1949), will yield talc instead of enstatite. If the veins result from filter-pressing of interstitial liquid from a peridotitic crystal mush, the temperature would be on the order of 1300—1400° C.

Some observations on the veins point, although not conclusively, to the latter mode of origin.

1. The exsolution lamellae in enstatites of the veins are composed of diopsidic pyroxene which contains lime expelled from the enstatite lattice upon cooling (cf. HESS, 1960). The presence of the lamellae suggest, therefore, a higher rather than a lower temperature of formation of the enstatite at which an excess of the diopside molecule is soluble in the lattice. Lamellae are generally absent in the orthopyroxenes of metamorphic rocks, presumably because at the temperatures at which recrystallization occurred, the solubility of diopside in the orthopyroxene is negligible. There is no experimental evidence which places limits on the relevant temperatures, however.

2. The thicker veins contain chromian diopside segregated in the middle, a relation which may be due to crystallization of Mg-rich orthopyroxene on the walls of the vein with the residuum of liquid of diopside composition segregating in the centre.

3. The mineralogy of the veins is consistent in the number and approximate proportion of phases present. The phases, moreover, are those which appear in the peridotite proper although in much different proportions. These relationships are consistent with the filter-pressing and consequent crystallization of the residual magmatic fraction of the peridotite.

Gentle flexures in the pyroxenite veins may have been produced during a later deformation of a weak and local nature. Plastic deformation in the form of kink bands in the enstatite of the veins also provide evidence of later deformation. Whether the (100) kink bands in olivine were produced primarily during the early or the later deformation is not known. The final recorded structural event is faulting of the peridotite, partly serpentinized, into its present position relative to the country rock. The complete preservation of mesh textures in the blocks of serpentinite undisturbed by faulting demonstrates that plastic flow in the serpentinized parts of the mass was unimportant (RALEIGH and PATERSON, 1965).

Adaptations of BOWEN and SCHAIRER's (1935) hypothesis of intrusion of a crystal mush which attribute the mechanism of origin of layering and lineation in alpine ultramafic bodies to flow during the intrusion of the crystal mush have recently been advanced by THAYER (1963), and LIPMAN (1964). Flow of a crystal mush saturated with interstitial magma, however the layering is produced, would be likely to give rise to folds in the layers much as those observed in rhyolite flows. Moreover, reaction-replacement textures could develop by reaction between olivine and the rest magma after cessation of intrusive flow, thereby avoiding the difficulties presented by the preservation of such textures in rocks which are assumed to have undergone extensive flow as a solid. The pyroxenite veins would then represent segregated liquid fractions filter-pressed out and crystallized toward the end of the intrusion process.

The hypothesis must be modified, however, on the grounds that no olivines of peridotites or basalts described in the literature have the crystal habit required to produce strong preferred orientations of the type observed in the Cypress Island rocks if the structures originated by flow of a crystal mush. Olivine suspended in a flowing magma has been shown by BROTHERS (1959) to take up an orientation controlled by its crystal habit. In order to produce a strong preferred orientation of the [100] axes parallel to the flow plane (plane of the layering) and the axes of small folds (normal to the flow direction) the olivine crystal habit must have been elongated parallel to [100] by analogy with TAYLOR's (1923) experiments on solid particles suspended in fluid undergoing simple shear. The elongation of olivines parallel to [100] in some of the Cypress Island specimens, having no analogue in olivine crystal habits of known magmatic origin, is therefore considered to have arisen by plastic flow or recrystallization during the folding.

Solid flow is possible in the presence of interstitial fluid if shearing stresses can be transmitted across grain contacts, that is, if the "effective" stress (TERZHAGI, 1936; HUBBERT and RUBEY, 1959) is equal to the yield stress for solid flow. If any interstitial magma at a pressure equal to that in the solid remained at the time of folding it must not have been so distributed as to prevent solid-solid contacts between olivine grains. The shear stress which could be transmitted to solid particles through a pervasive interstitial magma having a viscosity of 10^5 poises at shear strain rates as high as $10 \sec^{-1}$ is only one bar. Even at high temperatures such a shearing stress is unlikely to cause plastic yielding or drive syntectonic recrystallization in olivine. On the other hand if a liquid fraction is at a pressure *less* than that in the solids (e.g. during filter pressing of the mush when the magma is permitted to escape), differential stresses can be transmitted by the solid fraction. It is possible, then, for olivine to deform by plastic flow in two cases where interstitial magma is present: 1. where the magma occurs in isolated pockets in the impermeable solid mass, 2. where the magmatic fraction is at a pressure less than that in the solids, that is, where the fluid can escape to a lower-pressure environment during deformation. However, the effectiveness of the interstitial magma in "lubricating" the intrusion is inversely related to the proportion of the total stress carried by the grains across their contacts. Nevertheless, deformation of a crystal mush in condition (1) or (2) above provides the most complete explanation of the deformational fabric and the preservation of the igneous textures in the Cypress peridotite.

The current state of theory and experimental data concerning the development of preferred orientation in cold or hot worked crystal aggregates makes it unprofitable to discuss this aspect of the present problem. Recent theories of grain re-orientation by slip and twinning (CALNAN and CLEWS, 1950, 1952) give reasonable agreement with the observed fabrics of cold-worked metals. However, to apply CALNAN's and CLEWS' theory, knowledge is required not only of the glide mechanisms but also of the critical shear stress for glide on the different systems. Although much is now known about the glide mechanisms in olivine (RALEIGH, in preparation) the critical shear stress values for slip on the different systems are, as yet, undetermined. Consequently, attempts at relating preferred orientations of olivine to the principal strains deduced from folding in peridotites, such as those of BATTEY (1962) and BROTHERS (1962) are unlikely to prove fruitful at present.

G. Conclusions

This work has clearly demonstrated on the basis of a variety of structural and petrographic evidence that a peridotite of the alpine type has been highly deformed as a near-solid mass following its accumulation from a magma by crystal settling. The application of these results to the problem of the origin and emplacement of alpine peridotites lies in the probability that the deformation came about by intrusion of the peridotite as a crystal mush during which much of the interstitial magma was removed by filter-pressing. Given this mechanism of emplacement, a connection between the ultramafic composition of alpine peridotites and restriction of their occurrence to orogenic belts at one becomes apparent.

Accumulation by crystal settling from a basaltic or ultrabasic magma can yield a layered aggregate of olivine and minor enstatite and chromite with a pore volume of 25 to 40% filled by interstitial magma of a composition the same as or lower in calcium and aluminum and higher in magnesium than the parent magma (JACKSON, 1961, p. 89). If the parent magma is tapped and removed, the ultramafic accumulate may be dislocated and intruded as a crystal mush during which the bulk of the interprecipitate magma is expelled by filter-pressing. The final product will be ultramafic in composition. Intersection of the intrusion by an active fault in the orogenic belt could provide a path for the rapid removal of the overlying parent magma. The crystal mush, which would not be expected to rise in the crust by virtue of density differences between the mush and the country rocks, could, nevertheless, be also intruded along a reverse fault plane but more slowly. For this to occur it is required that the pressure in the leading edge of the weak, viscous tabular mush is greater than the normal stress across the fault plane, a state which may be achieved if the mush has a density less than or equal to the surrounding rock, or if an applied horizontal tectonic stress having a sufficiently large vertical gradient is present.

When a crystal mush is intruded it must be at a temperature in the neighbourhood of 1100—1400° C, so that contact metamorphism of the country rocks around the site where it comes to rest is to be expected. Serpentinization of the margins of the peridotite must occur, however, in the presence of water at a temperature of less than 400—500° C but high enough so that the reaction can go on at a reasonable rate. The serpentine will be extremely weak if the vapor pressure of the water in the serpentinizing shell is of the order of the confining pressure or, if the serpentine is re-heated to its dehydration temperature (RALEIGH and PATERSON, 1965). Tectonic stresses will then be expected to set the partially serpentinized mass in motion again by shearing of the weakened serpentinite margins and the mass will be relatively cold when the final stages of emplacement are concluded.

In conclusion, it is considered that a plausible mechanism for the origin and emplacement of alpine peridotites has been outlined which is supported by evidence from field and experimental studies. It would be desirable to test the hypothesis by examination of the structures of other, better exposed peridotites of the alpine type.

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