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EMPLACEMENT AND METAMORPHISM OF OPHIOLITES

Introduction

Present day workers now consider that ophiolite represents oceanic crust generated at mid-ocean ridges, from whence it slowly migrates by ocean floor spreading toward continental margins — there to be subducted into the mantle. Under some circumstances at plate houndaries, slabs of oceanic lithosphere have become detached and overide (obduction) continental margins (Coleman, 1971 a). The actual mechanism of ophiolite emplacement along continental margins is strongly debated, but most geologists agree that slabs of ophiolite are allocththonous and that they originate in an environment distinctly different from where they occur today (Davies, 1971; Dewey, 1974; Coleman and Irwin, 1974; Zimmerman, 1972; Gass and Smewing, 1973; Mesorian, 1973). This modern view contrasts with older debates regarding the ultramafic parts of ophiolites, which centered on the concept of their intrusion as magmas (Hess, 1938). These past concepts considered the ophiolites to represent the earliest magmatic phase of an ensialic geosyncline, which required that the ophiolites be autochthonous and interlayered with the geosynclinal sediments (Brunn, 1961; Aubouin, 1965; Kay, 1951) (Fig. 1).

Nearly all on-land ophiolites have undergone some kind of metamorphism as a result of their situation in the oceans or as a direct result of tectonic transport (subduction or obduction). These modifications of the primary igneous nature of ophiolites is an important element in understanding the complete history of ophiolites in ancient sutures. The metamorphic processes can be divided into two broad categories: 1) Internal metamorphism includes those metamorphic events modifying only the ophiolite mineral assemblages such as serpentinization and oceanic hydrothermal metamorphism. 2) External metamorphism includes metamorphic events affecting associated country rocks as well as the ophiolites. The tectonic history of

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emplacement as well as later orogenesis of ophiolites often superimpose externally derived metamorphic events that affect a whole region where ophiolites may occur.

The main theme of this paper will then be the emplacement of ophiolites and their metamorphism. The origin or igneous petrology of these rocks will not be discussed and it is assumed that most ophiolites represent fragments of oceanic crust formed at oceanic spreading ridges or by spreading in marginal oceanic basins.

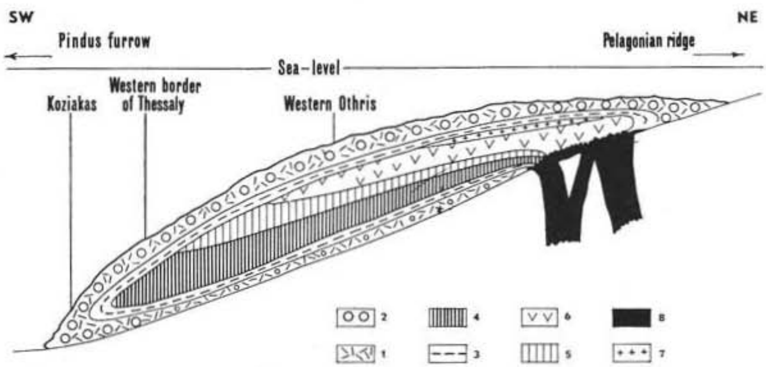

Fig. 1. — Diagram of the distribution of petrographic types in a submarine ophiolite flow: 1 - basalts, 2 - pillow lavas, 3 - dolerites, 4 - peridotites, 5 - pyroxenites, 6 - gabbros and diorites, 7 - the more acid members (quartz, diorites, etc.), 8 - eruptive fissures (after AUBOUIN, 1959, Fig. 17).

Emplacement Tectonics

Processes leading to the emplacement of ophiolite are most easily tied to plate motions rather than in-situ igneous intrusions. Interactions between plates are considered to give rise to orogeny and many of the younger orogenic belts can be related to zones of plate convergence (Fig. 2). Within certain zones of convergence such as the Alpine-Tethyan orogenic belt, allochthonous ophiolite masses are imbricated with nappes whose origin is quite different from that of the ophiolite (GANSSER, 1974). Formation of ophiolite at spreading ridges, marginal basins or rifts requires that these fragments of oceanic crust have been transported and incorporated into an orogenic zone. Only a small fraction of new crust formed is ever tectonically emplaced along a consuming or transform plate boundary. Reconstruction of spreading in relationship to consuming margins during the Phanerozoic requires that most of the oceanic crust developed at the accreting boundaries has been consumed by subduction and destroyed in the asthenosphere. The amount of oceanic crust incorporated into the orogens of continental margins is an extremely small percentage (probably less than 0.001 %) of the total oceanic crust formed and thus it is apparent that tectonic emplacement relates somehow to a major peturbation of the plate motions.

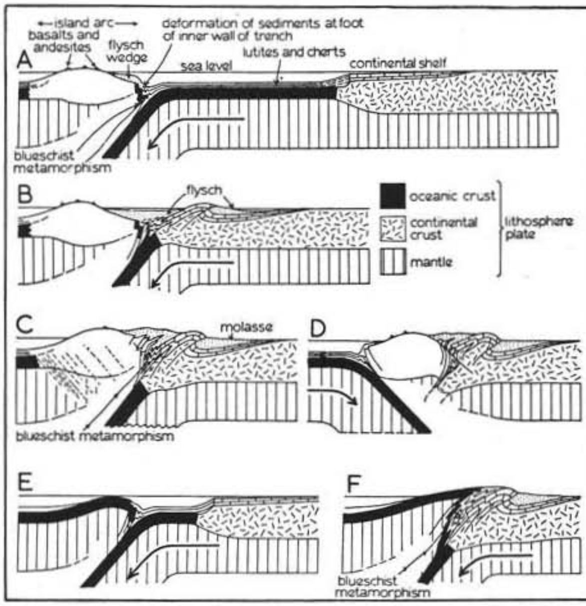

Fig. 2. — (A-D) Schematic sequence of sections illustrating the collision of a continental margin of Atlantic type with an island arc, followed by change in the direction of plate descent. (E-F) Proposed mechanism for thrusting oceanic crust and mantle onto continental crust (after DEWEY and BIRD, 1970, p. 2641, Fig. 12).

Obduction - Subduction

Steady state subduction is required to consume the large amounts of oceanic crust developed at the spreading centers. The consumption is visualized as a bending of the oceanic plate downward and its sinking into the mantle where it is assimilated and reincorporated into the mantle (DICKINSON, 1971 ab; TURCOTTE and OXBURGH, 1972; OLIVER et al., 1969). However, steady, state subduction does not allow parts of the oceanic crust to become detached and incorporated into the edge of the continent (COLEMAN, 1971 a). It is visualized that pelagic sediments resting on top of the oceanic crust are not consolidated enough to withstand deformation and may be scraped off and incorporated into the clastic trench sediments by a series of underthrust faults whose movements are synthetic to the subduction zone (ROEDER, 1973; HELWING and HALL, 1974; OXBURGH, 1974). However, none of these synthetic faults appears to penetrate into the rigid oceanic crust even though this has been suggested as an ophiolite emplacement by (ROD, 1974) for the Papuan - New Guinea ophiolite (DAVIES, 1971). ERNST (1974, 1973, 1970) has proposed that both sediments and oceanic crust can be subducted as part of the downgoing slab to be metamorphosed under high P — low T conditions and later exhumed into the orogen. This requires return to the Earth's surface within the convergence zone and could only be accomplished by a cessation of steady state subduction. The occurrence of blueschist assemblages within the Zermatt-Sass ophiolite of the western Alps provides evidence that some parts of the subducted oceanic crust can be resurrected at convergent margins (BEARTH, 1967; DAL PIAZ, 1974); however, the tectonic conditions involving transport of metamorphosed slabs up from such great depths remains obscure.

Obviously, tectonic processes related to steady state subduction cannot provide conditions that will allow detachment of ophiolite slabs up to 12 km thick from the downgoing oceanic lithospheric plate. The presence of large, unmetamorphosed

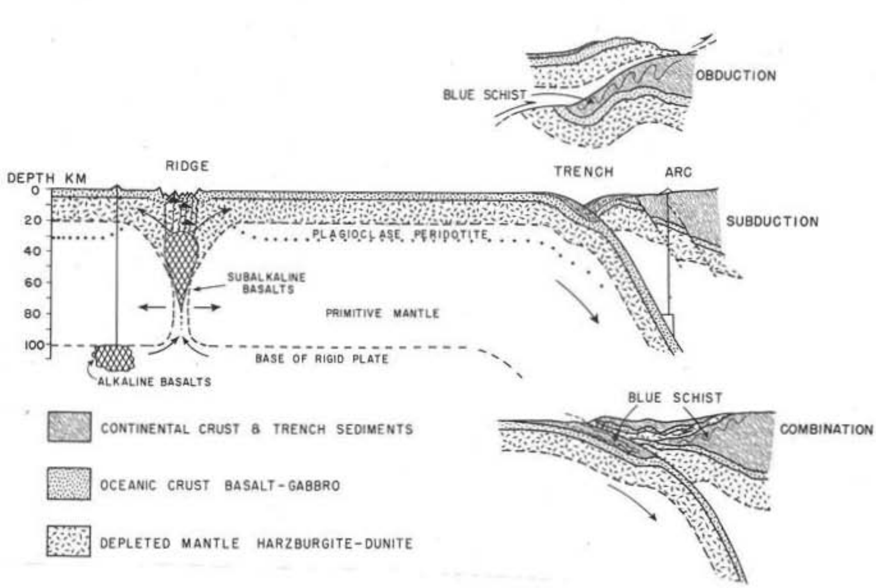

Fig. 3. — Conceptual model illustrating the development of oceanic crust at active ridges and its subduction or obduction or both at consuming plate margins. (COLEMAN, 1971, p. 1219, Fig. 6).

ophiolite slabs overthrust onto the continental margins, however, provides direct geologic evidence that at least some of the oceanic crust had escaped subduction. To provide a tectonic term that would adequately describe this process and also mark it as distinct from the more common term « Subduction », I introduced the term « obduction » (COLEMAN, 1971 a) (Fig. 3). Obduction implies overthrusting of consuming plate margins but there may be numerous tectonic situations that would allow the detachment of oceanic crust prior to overthrusting (Fig. 4). Generally, it is considered that the thickness of the oceanic plates are 60-100 km thick (Oxburgh, 1974) and the thickest known obducted ophiolite slab (12 km) is in Papua - New Guinea (DAVIES, 1971). The emplacement of such thin ophiolite slabs therefore requires some sort of detachment surface to develop within the upper portion of the oceanic crust. Areas of steep thermal gradient could produce faulting in zones of softening and eventually detachment (ARMSTRONG and DICK, 1974). Situations that could provide high heat flow are more likely to occur in rear-arc marginal basins or in small restricted ocean basins rather than where cold, thick slabs of oceanic crust are being subducted (DEWEY, 1974).

There are numerous possibilities where young hot oceanic crust could be detached and obducted. In situations where a limb of a ridge crest may be subducted, a unique

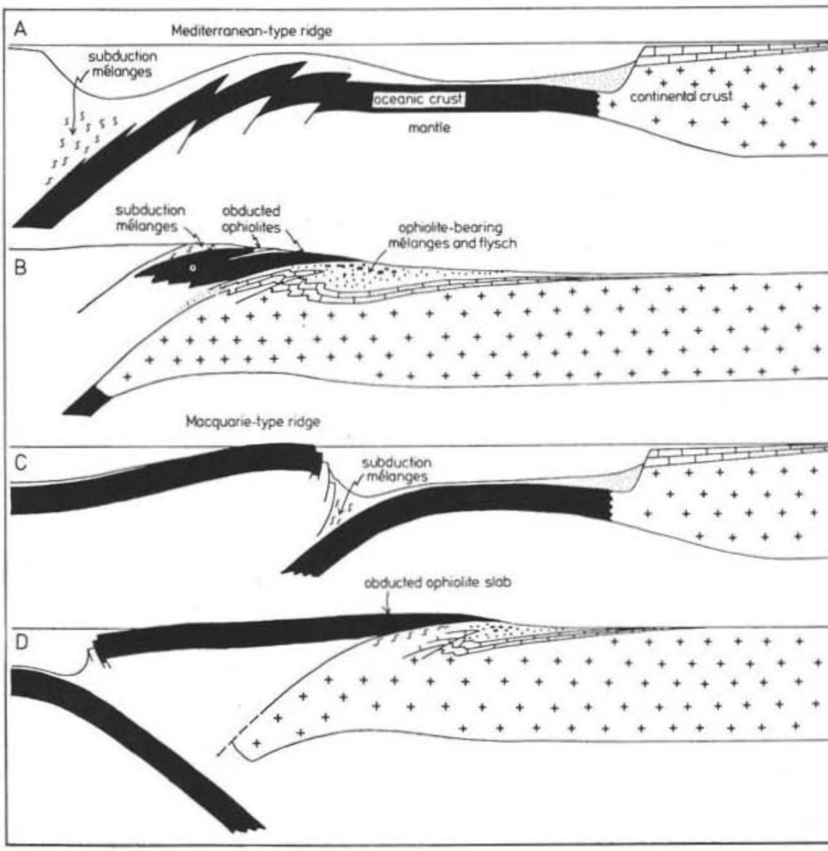

Fig. 4. — Possible mechanisms for the obduction of ophiolite sheets onto continental margins (after DEWEY and BIRD, 1971, p. 3193, Fig. 6).

point is reached where a thin, hot, mechanically weak segment of oceanic crust approaches a consuming margin. It is very possible that here faulting will produce dismembering of major segments and obduction of the outboard plates could take place (CHRISTENSEN and SALISBURY, 1975) (Fig. 5).

KARIG (1972) has suggested that the polarity of an active arc system, i.e. direction of subduction, may change and that this would lead to arc-arc or continental margin remnant-arc collisions. In this situation, high heat flow could be expected and allow shallow detachment of the oceanic crust. Destruction of marginal basins by a change in the polarity of the associated arc systems combined with continental margin collision provides a hypothetical mechanism of emplacement for many ophiolite occurrences described from the Circum-Pacific area and also explains the close association of ophiolite with island arc assemblages. Considering present day plate tectonics, obduction is not observed and so documentation is dependent on geologic evidence. DEWEY and BIRD (1971) have also recognized a lack of obduction across active trenches in which oceanic crust is being transported from the subducted plate across the trench onto the upper surface of the continent. The lack of observed obduction, combined with apparent irregular emplacement of ophiolites throughout Phanerozoic time, requires that there be other mechanisms for ophiolite emplacement. (OXBURGH, 1972) introduced the concept of « flake tectonics » where two continental crust-capped plates collide and as a result of asymmetric locking low angle crustal splits form flakes. Even though Oxburgh specifically excludes oceanic crust or mantle from his model, it seems unlikely that the low-angle detachment zone will always occur within the crust.

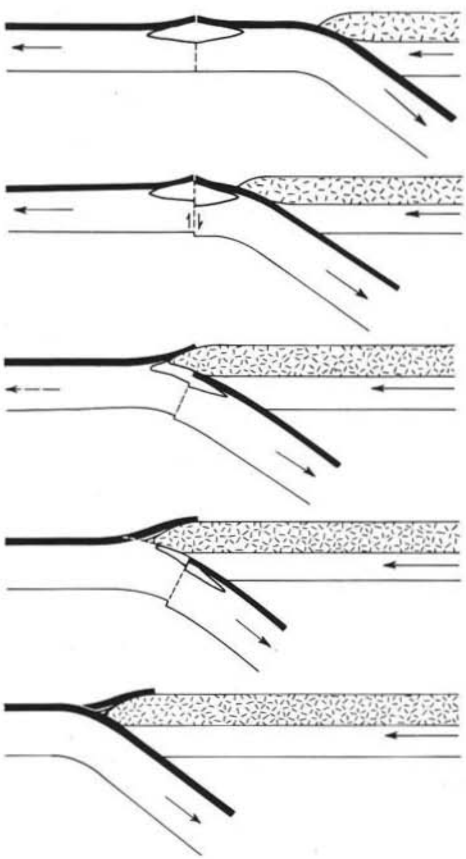

Fig. 5. — Ophiolite emplacement during subduction of a ridge crest (after CHRISTENSEN and SALISBURY, 1975, p. 79).

The work on the Lanzo lherzolite massif demonstrates that it is the bottom part of a « flake » formed along the Alpine suture (ERNST, 1973) a zone of convergence between two continental plates (BERCKHEMER, 1968, 1969; NICOLAS et al., 1972). Thus, it is possible by continental collision to emplace heavy mantle material or oceanic crust as « flakes » over continental crust. ROEDDER (1973, Fig. 1) describes « flipped » subduction zones where change in the direction of subduction develops alpine root zones, rootless rotation zones and overidden rotation zones and parts of the oceanic crust within older frozen subduction zone may be tectonically transported upwards into the orogen during changes in convergence polarity. DEWEY (1974) and DEWEY and BIRD (1971) also prefer emplacement of ophiolites by obduction, but believe that the process represents the collision of continents or arcs and continents and not an abberation of steady state subduction. DEWEY (1974, Fig. 7) illustrates the emplacement of the Appalachian-Caledonian ophiolites as a complicated series of openings and closing of marginal basins landward of a west dipping subduction zone. He visualizes the ophiolite occurrences as developing in separate rear-arc and inter-arc oceanic basins during the Ordovician and being obducted shortly thereafter.

ABBATE et al. (1973), have provided compelling evidence that all the Tethyan ophiolites were emplaced nearly simultaneously during a Late Cretaceous gravity sliding or thrusting (obduction). They also point out that spreading of the Tethyan Sea had virtually stopped by Late Cretaceous and that the closing of the Tethyan Sea was not accompanied by subduction (lack of volcanism). It is their opinion that the emplacement was due to convergence of small unconnected ocean basins. This leads to another variant of ophiolite emplacement which can be related to formation of small ocean basins such as the Red Sea. These basins would have only limited size, but would maintain a high heat flow until a later compressional event closed the basin with concomitant detachment of the upper parts of the oceanic crust. Obduction of the detached oceanic crust during closing of the small ocean basins would produce ophiolite allochthons resting on the basin edge sediments and capped by pelagic sediments of the basin axis.

Diapirs

There are a number of situations where the occurrence of ophiolite apparently requires an emplacement mechanism unrelated to plate motions. MAXWELL (1970, 1973, 1974 ab) has maintained the diapiric uprise of hot mantle material through continental and oceanic crust best explains the internal location of ophiolites within orogenic zones. Maxwell's model is an extension of the submarine extrusion hypothesis of BRUNN (1961) and does provide a means of overcoming the large lateral transport of oceanic crust across subduction zones at continental margins required by obduction. Furthermore, the production of small batches of oceanic crust by this model does not later require consumption of new oceanic crust at a spreading center.

Diapiric uprise of material within the mantle and crust requires that the rising material (have a density less than its surrounding material) and a plasticity that could allow such migration. For the peridotites from ophiolites, this would entail either a magma or transformation of the peridotites to serpentinite. The paucity of contact aureoles around the boundaries of ophiolites has been the most damaging evidence against hot diapir emplacement. On the other hand, diapiric uprise of serpentinites has been documented in numerous instances (COLEMAN, 1971 b, p. 905).

Gravity Slides - Protrusions - Deep Faults

There have been numerous mechanisms invoked to explain the emplacement of ophiolite, all of which may be valid for certain situations but do not appear to have a universal application. It is easy to conceive a complicated history for an ophiolite after it becomes part of the continental crust, particularly if the peridotite member of the ophiolite becomes serpentinized and incorporated into a melange.

The tectonic evolution of the northern Apennines has been carefully worked out by numerous geologists (ABBATE et al., 1976; ELTER and GREVISON, 1973; DECANDIA and ELTER, 1969) and here ophiolite emplacement by gravity sliding has been documented. Oceanic crust formed during Jurassic times is covered by radiolarites and overlying Calpionella limestones within the Ligurian domain. Uplift of the Ligurian oceanic crust with its carapace of Jurassic cherts and pelagic limestones produced the Bracco ridge giving rise to widespread gravity sliding where large ophiolite blocks (several cubic km) were transported on lenses of breccia (olistostrome). This event was followed by overthrusting (gravity sliding?) of the Ligurian olistostrome containing coherent blocks of ophiolite over continental and marine sediments of the (Tuscan and Umbrian) depositional basins during the Tertiary. Formation of serpentinite from the Ligurian peridotites facilitated the gravity sliding of these units.

GLENNIE et al. (1974) and STONELY (1975) visualize a similar mechanism for the Semail ophiolite. However, neither explanation provides a satisfactory tectonic mechanism for elevating the ancient oceanic crust high enough to initiate gravity sliding. There is no geodynamic reason for a deep ocean basin to rise isostatically unless there is initiated an upward diapiric movement within the mantle. On the other hand, the closing of a small ocean basin by plate movement could easily elevate parts of the oceanic crust during collision of two continental plates by upwarping and thus initiate gravity sliding of ophiolite onto a continental margin. Thus, detachment of ophiolite from the upper parts of the oceanic lithosphere and concomitant gravity sliding most certainly is a viable emplacement process. Even though actualistic models of gravity sliding of oceanic crust away from mid-ocean ridges can not be verified, it seems feasible to entertain this as a possible mechanism, particularly where heat flow is high and the oceanic lithosphere still retains some of its plasticity at shallower depths (BOTTINGA and ALLEGRE, 1973).

Numerous workers (MILOVANOVIC and KARAMATA, 1960; KNIPPER, 1965; LOCKWOOD, 1971, 1972) have been impressed by the tectonic mobility of serpentine once it forms and have referred to this phenomena as cold intrusion, solid intrusion, or tectonic intrusion. LOCKWOOD (1972) has suggested the term *protrusion* to cover this tectonic process. Examples of protruded serpentinites are extremely common and the mechanism allows tectonic movement of serpentinized peridotites. Emplacement of ophiolite slabs by this process, however, seems quite unlikely. The mobility of the serpentinized peridotite within the orogen is perhaps the most important factor in the dismembering of an ophiolite and must always be considered in its tectonic reconstruction.

This brings us to consider emplacement of ophiolites along deep fundamental faults. Some Russian geologists call upon deep fundamental faults into the mantle to provide a tectonic situation to emplace solid peridotites within the earth's crust (KHAIN and MURATOV, 1969; PIEVE, 1945; REVERDATTO et al., 1967). The concept here is that the mobile orogenic zones are bounded by deep fundamental faults that persist for long periods of geologic time. Most of the movement is considered to be vertical and that the fault extends through the earth's crust and penetrates into the mantle. Association of discontinous pods of ultramafic rock and high temperature and pressure metamorphic minerals along these deep fundamental faults has been interpreted as signaling great vertical movements whereby mantle material and high temperature and pressure rocks can be brought to the surface. Perhaps the best studied deep fault is the Alpine suture which represents a fundamental structure in the Alps. It separates the Hercynian structures and metamorphism on the south from the younger Alpine nappes to the north and was earlier connected in with the concept of « root zones » (GANSSER, 1968). Significantly, several peridotite bodies such as the Finero and Lanzo are situated on the hanging wall of the Alpine suture. Geophysical studies have shown that positive gravity anomalies follow the Alpine suture and that seismic velocities indicate that exposed peridotite bodies may have roots in the mantle (BERCKHEMER, 1968, 19669; PESELNICK et al., 1974; NICOLAS et al., 1971). The evidence indicates possible solid emplacement of mantle peridotites along a deep fundamental fault. These peridotites are, however, not typical of the ophiolite sequence and contain no associated mafic rocks (NICOLAS and JACKSON, 1972). It therefore becomes obvious that there are numerous possible ways to emplace ophiolites, but confusion develops because numerous non-ophiolitic mafic-ultramafic rock bodies have been used to demonstrate emplacement for all ophiolites.

The high temperature lherzolites such as Beni Bouchera, Ronda, Lanzo, Finero are clearly different than the ophiolites and should be treated as a special class as suggested by NICOLAS and JACKSON (1972). The deep fault and mantle diapir emplacement of the high temperature lherzolites seems required to preserve their high temperature-pressure assemblages, whereas thin skin tectonics such as gravity sliding, initiated by subduction or obduction can best explain emplacement of most ophiolite slabs. Post emplacement deformation of ophiolite slabs combined with serpentinization and metamorphism produces complications that often obscure their mode of initial emplacement.

Metamorphism

Introduction

The possible geological situation whereby ophiolites can be metamorphosed are numerous and overlapping. A discussion of these various situation along with a general scheme of these relationships will provide a starting point for this discussion (Fig. 6). As shown on this scheme the ophiolite is divided into two main types

METAMORPHIC SITUATION

		OCEANIC HYDROTHERMAL	SUBDUCTION	OBDUCTION		REGIONAL
OPHIOLITE PROTOLITH	BASALT DIABASE GABBRO	ZEOLITE ↓ GREENSCHIST ↓ AMPHIBOLITE (Massive Sulfides)	ZEOLITE ↓ PREHNITE - PUMPELLYITE ↓ BLUESCHIST (Type C Eclogites)	NO METAMORPHISM		ZEOLITE ↓ PREHNITE - PUMPELLYITE ↓ AMPHIBOLITE
	PERIDOTITE	LOCAL SERPENTINIZATION RODINGITES TRANSFORM FAULTS – MYLONITES	SERPENTINIZATION HT/LP: ZEOLITE ↓ LIZARDITE-CHRYSOTILE HP/LT: BLUESCHIST ↓ ANTIGORITE RODINGITES	BASAL CONTACT HOT SLAB AUREOLES GARNET AMPHIBOLITE	COLD SLAB MELANGE SERPENTINE RODINGITES	PROGRADE METAMORPHISM CHRYSOTILE ↓ ANTIGORITE ↓ OLIVINE + ANTIGORITE ↓ OLIVINE + TALC ↓ OLIVINE + ENSTATITE
CONDITIONS OF METAMORPHISM		NO DEFORMATION MINOR METASOMATISM	PLASTIC AND PENETRATIVE DEFORMATION	PLASTIC DEFORM.	CATACLASTIC DEFORM.	POLYMETAMORPHISM SUB-SOLIDUS DEFORM. OF PERIDOTITES
THERMAL GRADIENT		STEEP	LOW	STEEP	LOW	MODERATE – STEEP

Fig. 6. — Scheme of metamorphism for ophiolites.

of protolith: 1) Peridotite, 2) Basalt, Diabase, Gabbro. The peridotites are so distinctive in their bulk composition that their metamorphism should be separated from their mafic associates particularly when comparing metamorphic assemblages. Four geological situations are considered to represent the most likely to develop characteristic metamorphic assemblages: 1) Oceanic Hydrothermal resulting from hot circulating ocean waters near spreading centers. 2) Subduction of oceanic crust could give rise to high pressure-low temperature assemblages. 3) Obduction of young hot slabs of oceanic crust could produce dynamothermal contact aureoles. 4) Regional metamorphism of ophiolites after they become incorporated into orogenic belts. Oceanic hydrothermal metamorphism, serpentinization, and the formation of rodingites will be considered as internal metamorphism affecting only the ophiolites and not their associated country rocks. The section on external metamorphism will include all other types of metamorphism commonly found for ophiolites.

Internal Metamorphism

1. *Serpentinization*

The peridotites within ophiolites are usually serpentinized to some degree and it is rare to find peridotites that contain completely unaltered olivine, orthopyroxene, or clinopyroxene. The process of changing a peridotite to serpentine involves mainly a hydration reaction between water and the primary igneous minerals. Mineralogically, serpentinites consist predominately of lizardite, clinochrysotile, and antigorite with minor amounts of brucite, talc, magnetite, and carbonate. During serpentinization of dunite, harzburgite, or lherzolite assuming that only water is introduced, the possible mineral species that can be developed at these bulk compositions are limited. In the case of small ultramafic bodies being serpentinized or undergoing regional high-grade metamorphism, it is not unusual to have CO_2 or silica invade the ultramafic body at its borders to produce talc, chlorite, and carbonates within the serpentinite (CHIDESTER, 1962, 1969; JAHNS, 1967).

The assemblage lizardite + clinochrysotile + brucite + magnetite appears to be the most common one developed in the peridotites from ophiolites. Where these same serpentinized ultramafic rocks are found in terranes that have undergone external high-grade regional metamorphism, antigorite is the stable serpentine mineral. Progressive metamorphism of peridotites from Pennine region of the central Alps provides a consistent picture of the assemblage developed in the system MgO-SiO_2-H_2O (TROMMSDORFF and EVANS, 1974). If the experimental information can be used as a guide to pressure-temperature conditions, it would seem that most of the serpentinites derived from ophiolite peridotites probably formed in the temperature range 25°-300° C with only minor amounts forming at ambient temperature (WENNER and TAYLOR, 1971). The antigorite serpentinites must represent temperatures in excess of 300° C and perhaps as high as 550° and usually reflect higher grades of regional metamorphism. None of the serpentine minerals or their common associates are sensitive to pressure effects.

The average amount of iron in dunites, harzburgites, and lherzolites is about 7 weight percent FeO, with olivine and orthopyroxene containing nearly equal amounts. Small amounts of iron are contained in chromite and clinopyroxene. The Fe^{+3} and Fe^{+2} ratios of serpentines reflect the changes in the activity of oxygen during their formation. The formation of awaruite (Fe Ni) in association with or in the place of, of oxygen (NICKEL, 1959; PAGE, 1967). During serpentinization the partitioning of iron between serpentine, brucite, magnetite and awaruite depends on the availability of O_2. The alteration of ophiolite peridotites to serpentinite requires a large amount of water. Serpentine minerals contain from 12 to 13.5 weight percent water. The amount of water required to serpentinize a peridotite completely is controlled by the original proportions of olivine, pyroxene, and plagioclase and by the mobilities of Mg or Si, or both.

The development of the large serpentinized peridotites is significantly linked with crustal processes that must be part of ophiolite tectonic evolution after the peridotite emplacement into orogenic zones. The change from a peridotite to serpentinite represents a great range in physical properties. Peridotites have a density of 3.3 whereas serpentinite average 2.55 and concomintantly the compressional wave velocity Vp decreases from 8 km/s to 5 km/s (COLEMAN, 1971 b). Magnetite developed during serpentinization increases the magnetic susceptibility (K) and serpentinite masses are known to produce magnetic anomalies in excess of 500 gammas. Even though strengths of unsheared serpentinites are comparable to massive igneous rocks (RALEIGH and PATERSON, 1965), estimation of the shear strength of completely tectonized lizardite-chrysotile serpentinites in New Idria, California, by COWAN and MANSFIELD (1970) gives values of 1 bar. The weak nature of the tectonized serpentinite combined with its low density demonstrates that tectonic movement by plastic flow at low stress could easily account for the development of serpentinite melanges. Thus, low temperature hydration of massive, relatively strong peridotites to sheared, weak serpentinites will have a drastic affect on the geometry of the original ophiolite assemblage in orogenic zones. The development of serpentinite within an ophiolite can be visualized as happening in three distinct situations: 1. Within the oceanic realm particularly along transform faults. 2. During tectonic transport into the continental margins. 3. As part of regional metamorphism. The source of the water can be fingerprinted by stable isotopes and demonstrates that serpentinization is a continuous process as long as meteoric, oceanic, or metamorphic water is available.

2. *Rodingites*

Metasomatism of varied rock types associated with serpentinites is widespread in Phanerozoic orogenic belts (COLEMAN, 1966, 1967; DAL PIAZ, 1967, 1969). This calcium metasomatism is related to the process of serpentinization and tectonic history of the ultramafic emplacement (Fig. 6). Diverse rock types such as gabbro, basalt, graywake, granite, dacite and shale have been involed in the metasomatism. Through the years these metasomatic rocks have acquired the general term rodingite and for the purpose of this discussion I will use this as a general all-inclusive term for this process.

Nearly all of the rodingite occurrences have similar characteristics. The metasomatic reaction zones are always within or in contact with serpentinite and have not been observed in unserpentinized ultramafic rocks or in the high-temperature contact aureoles of such rocks. Rodingites are developed from diverse rock types associated with serpentinites. The reaction zones develop at the contact between tectonic inclusions, dikes and layered mafic rocks, and the surrounding country rock and the serpentinite. The rodingites are characteristically involved in tectonic movements that have acted on the serpentinite. Synkinematic brecciation and mylonitization of some rodingites indicates that deformation may accompany metasomatism. The chemical changes recorded in the rodingites as a result of metasomatism all trend towards a similar bulk composition. All rodingites are undersaturated with respect to silica ($\sim 45\%$ SiO_2) and are enriched in calcium (25-35 % CaO) and to a lesser extent magnesium. Where the metasomatism has been complete many rodingites have the following approximate molecular proportion $3CaO. Al_2O_3. 2SiO_2.H_2O$. Rodingites derived from silicic rocks such as sandstone and granitic rocks, often contain a secondary feldspar-rich core or zone away from the serpentinite contact, chlorite and nephrite also may form within serpentinite at its contact with the rodingite. Rodingites are small-scale localized metasomatic occurrences and are not related to a regional metasomatism or metamorphism. Metasomatic rocks having rodingite chemistry have not as yet been reported from other geologic environments. Hydrogarnet is the characteristic mineral of rodingites and is commonly associated with idocrase, diopside, prehnite, xonotlite, wollastonite, chlorite, sphene, and tremolite-actinolite (nephrite) where mafic igneous rocks have been metasomatized.

Evidence for the presence of calcium release during serpentinization is also afforded by the occurrence of calcium hydroxide waters issuing from partly serpentinized peridotites (Barnes et al., 1967; Barnes and O'Neil, 1969; Barnes et al., 1972). These small sluggish springs have travertine aprons and discharge a calcium hydroxide type ($Ca^{+2} - OH^{-1}$) water that has unusually high pH (11^{+}) values. These waters are undersaturated with respect to the minerals in those rocks that are commonly metasomatized within the serpentinized peridotites and may provide the calcium responsible for metasomatism.

The tectonic nature of some serpentinite contacts and the associated low temperature reaction zones (rodingites) is strong evidence indicating that emplacement of the ultramafic parts of the ophiolite coincide with major tectonic events (Coleman, 1967; Dal Piaz, 1967). The widespread occurrence of rodingites within sheared serpentinites and serpentinite melanges indicates that calcium metasomatism is a normal by-product of serpentinization. The pervasive presence of calcium hydroxide waters within serpentinized peridotites by analogy is very similar to the reaction that takes in the formation of cement. Potentially, then wherever these calcium hydroxide waters encounter rocks higher in silica than the peridotites ($\sim 45\%$ SiO_2) reactions can take place whereby calc-silicate minerals will replace and invade the host whether it be an exogenous inclusion or endogenous rock within the peridotite. It is therefore important to recognize the « rodingites » as by-products of the serpentinization process and not the result of high-temperature contact phenomena related to an earlier igneous history of the ophiolite.

3. *Oceanic Hydrothermal Metamorphism*

Many mafic rocks from ophiolite assemblages have undergone a « spilitic » metamorphism that appears to be widespread and uniform within the upper parts of the ophiolite (Spooner and Fyfe, 1973; Gass and Smewing, 1973). The mineral assemblages indicate a downward increasing thermal gradient from zeolite facies to greenschist facies and perhaps to low-grade amphibolite facies (Fig. 7). The hydrothermal metamorphism appears to be restricted to the pillow lavas, sheeted dikes, and upper parts of the gabbro, and the apparent metamorphic zonal boundaries are disposed subparallel to the original horizontal layers within the constructional parts of the ophiolite (GASS and SMEWING, 1973). The metamorphosed mafic rocks retain their original igneous textures and exhibit only local shearing. The disposition of the hydrothermal metamorphism as well as its apparent shallow downward

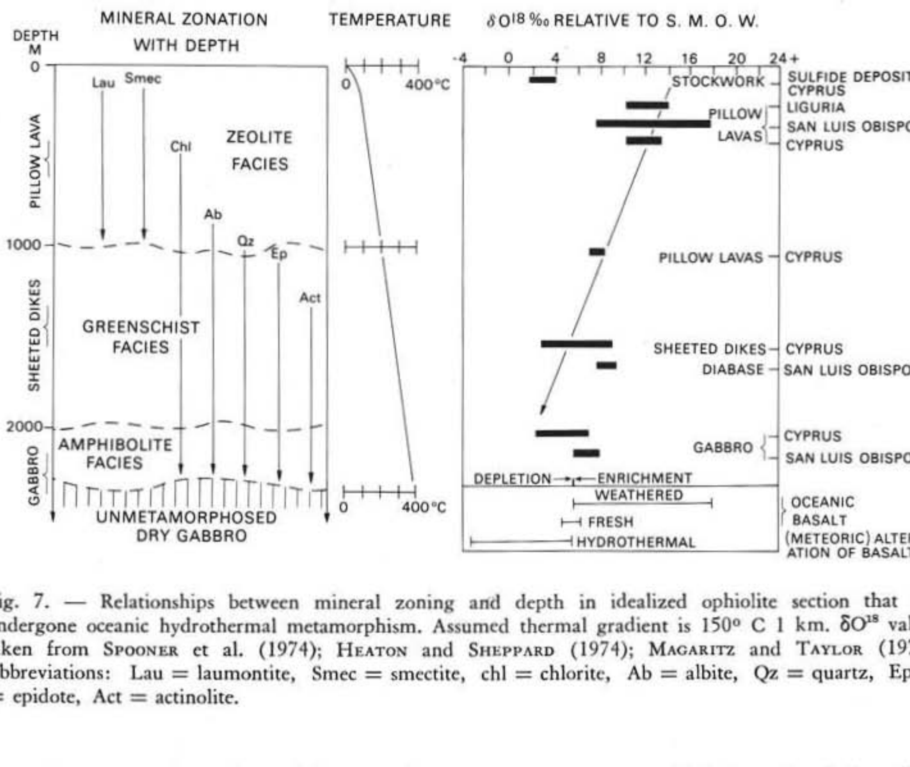

Fig. 7. — Relationships between mineral zoning and depth in idealized ophiolite section that has undergone oceanic hydrothermal metamorphism. Assumed thermal gradient is 150° C 1 km. δO^{18} values taken from SPOONER et al. (1974); HEATON and SHEPPARD (1974); MAGARITZ and TAYLOR (1974). Abbreviations: Lau = laumontite, Smec = smectite, chl = chlorite, Ab = albite, Qz = quartz, Ep = = epidote, Act = actinolite.

termination within the gabbros indicates a system controlled by circulating hot water. The lack of metamorphism in the gabbros relates to their impervious nature.

If the assumption is made that most ophiolites form at a spreading center under a cover of seawater and that high heat flow is maintained after their formation, it is possible to show that sub-sea geothermal systems operating at spreading centers are capable of producing this hydrothermal metamorphism (SPOONER and FYFE, 1973). High heat flow is symmetrically disposed on either side of present day spreading centers (LEE and UYEDA, 1965) and within the ridge axis geothermal gradients from 500° to 1400° C/km have been postulated (CANN, 1970 and SPOONER and FYFE, 1973). Even though considerable fluctuation in the heat flow is seen away from the ridge axis it has been suggested that thermal gradients of 150° C/km are maintained to at least 100 km away from the axis.

It thus sems inevitable that new oceanic crust generated at a spreading ridge will have undergone some hydrothermal metamorphism. The distribution and grade of metamorphism will be controlled by circulating seawater and the thermal gradient. Distribution of this ocean floor thermal metamorphism within on-land ophiolites from eastern Liguria, Italy (SPOONER and FYFE, 1973). Cyprus (GASS and SMEWING, 1973) and Oman (COLEMAN, unpublished data) is similar in many respects. In all of these occurrences the zeolite facies assemblages are present in the upper parts of the sequence followed downward into greenschist facies where the effects of the hydrothermal alteration disappears in the upper parts of the layered gabbros (Fig. 7). Usually the metamorphic assemblage in the upper gabbros suggests low amphibolite facies and from this we can estimate a temperature of 500° C (WINKLER, 1974). The vertical thickness affected by the hydrothermal metamorphism in Oman and Cyprus varies from 2 to 3.2 km respectively. Dividing this into a maximum temperature of 500° C gives an estimated thermal gradients from 150° C/km to 250° C/km. Such steep thermal gradients can be best explained by hot circulating sea water within the upper few kilometers of newly formed oceanic crust developed at a spreading center. Regional burial metamorphism thermal gradients are around 15° C/km and would require a depth of nearly 25 km to attain temperatures of 400° C. Therefore ophiolites that had undergone regional burial metamorphism would perhaps be too thin to have mineral isogrades developed within their limited sequence.

Thermally metamorphosed basalts and diabase from ophiolites characteristically retain their primary igneous textures. Many of the arguments for primary spilite magmas are based on the fact that retention of textures is prima facie evidence for their igneous origin (AMSTUTZ, 1974). However, careful studies of dredge hauls and drill cores from the oceans shows the presence of both « spilite » metamorphic assemblages as well as unaltered tholeiitic basalts (CANN, 1969; MIYASHIRO, 1972; MELSON and VAN ANDEL, 1966). It is the opinion of the author that « spilite » metamorphism within the ophiolite sequence results from hot circulating H_2O within the upper parts of the newly-formed oceanic crust. Tectonic transport of slices juxtapose metamorphosed basalts and diabase against sedimentary country rock that has not been metamorphosed such as reported in Cyprus and Oman (GASS and SMEWING, 1973; REINHARDT, 1969).

The hydrothermal alteration of the upper parts of the ophiolite sequence requires the presence of circulating water. Evidence from alteration is within the upper 2-3 km. Oxygen isotope geochemistry has provided important and indirect clues that the source of the water taking part in the hydrothermal alteration is sea water (SPOONER et al., 1974; HEATON and SHEPPARD, 1974) (Fig. 7). Hydrothermal alteration of a rock by a fluid will change the oxygen isotope ratios of the rock away from the original igneous values. The average δO^{18} values of unaltered basalts is usually

$+ 6‰ \pm 0.5‰$ (Taylor, 1968). Continental examples of hydrothermally altered mafic igneous rocks show negative δO^{18} shifts and represent depletions brought about by interaction of the rock with hot water of meteoric origin (Taylor and Epstein, 1963; Taylor, 1971). In contrast to these depletions, the hydrothermally altered ophiolites from Mediterranean area are enriched in O^{18} relative to unaltered basalts by as much as 7‰ (Spooner et al., 1974). Even more extreme positive shifts are recorded for weathered basalts taken from deep ocean environments, where the fractionations took place at low temperatures. The δO^{18} enrichments of the upper parts of the ophiolite pass downwards into depletions within the

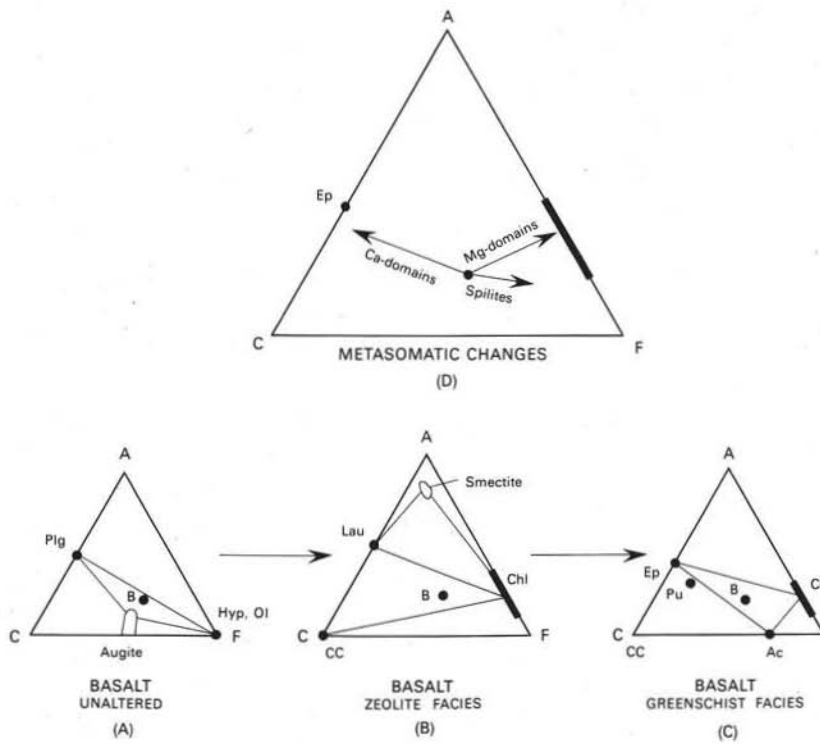

Fig. 8. — Mineral assemblages in metabasalts undergoing hydrothermal seawater alteration from fresh basalt through zeolite to greenschist facies: A → B → C. Three possible trends of metasomatic alteration are shown in D. Filled Circle represents average basaltic composition.

higher grade metamorphic zones and is related to the increasing metamorphic temperatures at depth (Spooner et al., 1974). The stable isotope data indicates that the water circulating during the hydrothermal alteration of the ophiolites was sea water rather than meteoric water. The presence of seawater within the hydrothermal system responsible for the alteration of the ophiolites provides a potential fluid to bring about large scale seawater-basalt interactions. Evidence for such interaction is found in the hydrothermally altered ophiolites where domains of contrasting composition may be present within a single exposure (Fig. 8).

There seems to be at least three important alteration domains: 1. Ca-enriched areas of monomineralic or bimineralic epidote and/or pumpellyite. 2. Spilite lithology where albite, chlorite, and calcium silicates dominate. 3. Mg-enriched areas that represent the complete alteration of glass to chlorite. These same alteration domains have been recognized in continental terrains where mafic volcanics have undergone burial metamorphism (SMITH, 1968; JOLLY and SMITH, 1972; VALLANCE, 1974). Recent experimental work by BISCHOFF and DICKSON (1974) on seawater-basalt at 200° C and 500 bars provides information on the possible species involved in this kind of hydrothermal alteration. They found that Mg was continuously abstracted from seawater and probably forming a Mg-rich chlorite or smectite from the glass. Ca precipitated as $CaSO_4$ early in the experiment but continued to increase in the seawater. Heavy metals, such as Fe, Mn, Ni and Ca are leached under the conditions of the experiment and could concentrate in the fluid in high enough concentrations to be considered potential ore forming fluids. The results of this experiment point to at least a partial explanation of the alteration domains within ophiolites. The spilite assemblage must represent the residual part of the basalt that has undergone hot seawater reaction (Fig. 8). Sodium enrichment may be related more to the loss of calcium and other elements rather than addition of sodium. BISCHOFF and DICKSON (1974) found that potassium was strongly leached from the basalt during their experiment whereas sodium remained nearly constant in the seawater. The calcium-rich domains, such as the epidosites, must represent a major movement of calcium and alumina during the hydrothermal alteration. The low temperature formation of smectites (montmorillonites) and higher temperature formation of chlorite represents an important seawater action that removes Mg from the oceans. Hydrothermal brines developed during the alteration of the upper 3 km of oceanic crust (ophiolites) must then have the potential of forming metalliferous concentrations within the ophiolites or at the interface between ocean water and points of brine discharge such as are now taking place in the Red Sea (BISCHOFF, 1969). As was mentioned in earlier chapters, this hydrothermal alteration of the upper parts of the ophiolite sequence will have a profund affect on the primary chemistry of these igneous rocks. Thus, the comparison of various igneous suites within ophiolites with unaltered volcanics cannot provide definitive answers to the igneous history of these rocks.

External Metamorphism

1. *Metamorphic aureoles*

Thin zones of high grade amphibolites are present at the base of several large obducted slabs of ophiolite (WILLIAMS and SMYTH, 1973; ALLEMANN and PETERS, 1972; GLENNIE et al., 1974) and granulites are commonly associated with non-ophiolitic lherzolite tectonic massifs (NICOLAS and JACKSON, 1972; LOOMIS, 1972 ab, 1975; KORNPROBST, 1969) (Fig. 6). The significance of these rocks has been given various interpretations over the past 50 years and up to the present time considerable controversy still surrounds their interpretation (THAYER and BROWN, 1961; THAYER, 1967, 1971). Earlier opinions on the origin of the emplacement of peridotites focused on their intrusion as magmas and the associated aureoles were considered to be the result of hot ascending ultramafic magma (SMITH, 1958; McGREGOR, 1964; MACKENZIE, 1960). Tectonic emplacement of peridotites was not, however, completely neglected (IRWIN, 1964).

It is important to first separate the ophiolite sequences and their associated amphibolites from the non-ophiolitic lherzolite tectonite massifs that have associated granulites, as was recently done by NICOLAS and JACKSON (1972). The non-ophiolitic lherzolite massifs represent undifferentiated mantle from beneath the continents and the associated graunlites are probably metamorphosed lower continental crust. Their emplacement is along deep fundamental faults initiated as mantle diapirs into the lower parts of the continental crust (KORNPROBST, 1969; LOOMIS, 1972 ab, 1975). The associated granulites are usually more extensive than can be accounted for by contact metamorphism of a rising, solid, diapir of peridotite (LOOMIS, 1972 b, p. 2492). The final tectonic emplacement of these peridotites into the crust distorts the original geometry of the peridotite and the granulites precluding a proper geologic evaluation of their formation. The non-ophiolitic lherzolite tectonites contain only peridotite in contrast to the ophiolite sequence where gabbros, diabase, and pillow lavas develop above the underlying peridotite. For the purposes of this discussion, the continental lherzolites of NICOLAS and JACKSON (1972) will be excluded.

The metamorphic aureoles associated with ophiolites are always located at the base of the peridotite and consist of narrow zones (usually less than 500 meters) of amphibolite. The metamorphic fabric of the amphibolites show polyphase deformation with the second generation schistosities and fold axes parallel to the contact between the peridotite and amphibolite. Structures within the peridotite are also subparallel to the contact and suggest a recrystallization during the second generation of amphibolite deformation. Hornfels textures that are characteristic of static igneous contacts have not been reported as primary or secondary textures within these aureoles. In Newfoundland, downward progression from high grade amphibolites into greeschist assemblages at the base of the peridotite suggest extremely high thermal gradients (MALPAS et al., 1973; WILLIAMS and SMYTH, 1973). Within the Klamath Mountains of northern California a nearly continuous narrow belt (35 km × 0.5 km) of gneissic garnetiferous amphibolite occurs at the base of the Trinity ultramafic sheet (DAVIS et al., 1965; IRWIN, 1964). The amphibolite, in the past, has been considered to be an intrusive contact aureole, but synkinematic deformation suggests that it may have developed during tectonic emplacement of the Trinity ophiolite. Metamorphic aureoles are not present at the base of all ophiolites, but often tectonic blocks of similar amphibolites are present in the melange upon which the ophiolite is resting (REINHARDT, 1969; MOORES, 1969). In

California and Oregon serpentinite melange units representing dismembered ophiolites often contain polyphase deformed amphibolites that are similar to those described as aureoles in Newfoundland (COLEMAN and LANPHERE, 1971).

The narrow high grade amphibolite aureoles contain brown hornblende, clinopyroxene, garnet, calcic plagioclase in varying amounts. Usually within 10-15 meters of the contact the mineral assemblage changes to essentially green-brown hornblende and calcic plagioclase and within 500 meters the mineral assemblages are typical of the greenschist facies. Up to now there are no mineral analyses of coexisting phases and little can be said regarding the P-T conditions of formation. If we assume that the highest grade of metamorphism represented by the aureole rocks is between high-rank amphibolite and granulite then a temperature of between 600-700° C can be estimated. No realistic estimate of pressure can be made at this time.

The formation of such a narrow zone of metamorphic rock at the base of ophiolite slabs is a remarkable situation. The protolith for these narrow aureoles is difficult to establish since in most instances the tectonic transport of the ophiolite with its welded-on basal aureole is now resting on unmetamorphosed rock or a melange. WILLIAMS and SMYTH (1973) suggest that the aureole in Newfoundland was derived from fragmental mafic volcanic rocks formed along the continental margin and that a clear downward progression from amphibolite into only slightly metamorphosed mafic volcanics can be demonstrated. It must be noted here that the polyphase deformation of the high-rank aureole rocks from Newfoundland is not imprinted on the underlying mafic volcanic protolith and that there could exist low angle faults separating the high-rank and low-rank rocks under the aureole. In Oman, the high-rank amphibolite aureole of mafic composition apparently grades downward into metacherts and metashales; however, polyphase deformation of the Oman high-rank amphibolite is not present in the underlying lower rank metasediments, suggesting a cryptic discontinuity between the amphibolite and metasediments. The truly high-rank amphibolites from these aureoles appear, then, only as very narrow zones 10-20 m at the deformed base of the peridotite (ALLEMANN and PETERS, 1972).

It seems clear that these narrow aureoles are somehow related to the tectonic emplacement of ophiolites from the oceanic realm into or onto the continental margins or island arcs. WILLIAMS and SMYTH (1973) suggest a rather shallow detachment of the mantle where high heat flow allows thin slabs of oceanic crust to be emplaced. The high heat flow would provide the heat necessary to produce a thin aureole, perhaps, augmented by frictional heating produced during obduction (CHURCH and STEVENS, 1971; CHURCH, 1972). The absence of amphibolite aureoles at the base of some ophiolites could be related to emplacement of cold slabs such as that in Papua New Guinea (DAVIES, 1971). Here the basal contact is against graywackes of the Owen Stanley Range which recrystallized to high pressure blueschist assemblages.

Still another point of confusion arises when the gabbroic parts of the ophiolites have undergone metamorphism and take on the appearance of amphibolites related to the contact aureoles (THAYER and BROWN, 1961; THAYER, 1967, 1971; HUTCHINSON and DHONAU, 1969; HUTCHINSON, 1975; LANPHERE et al., 1975). Imbrication of tectonic slices of high-rank amphibolites representing lower crustal metamorphism with cold ophiolitic slabs of younger age along continental margins could also provide an alternate explanation (COLEMAN et al., 1976). Imbrication and tectonic translation of ophiolite slabs within the orogenic zones of continental margins virtually guarantees obliteration of the original geologic relationships developed during the formation of these aureoles. The following section will discuss these implications in light of metamorphism along continental margins.

2. *Continental Margin metamorphism*

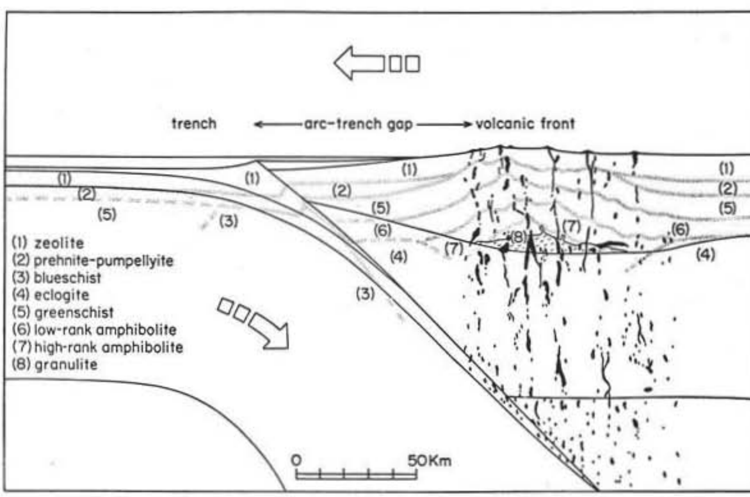

Fig. 9. — Approximate spatial distribution of metamorphic facies in crustal rocks near an active convergent plate junction. The petrographic grid of Fig. 10 has been employed with the thermal structure shown by Ernst (1974, Fig. 3). As in pressure-temperature space, the metamorphic facies boundaries are zones of finite appreciable volume within the earth (after ERNST, 1974, Fig. 5).

Nearly all oceanic lithosphere generated at spreading ridges eventually is carried to a consuming plate margin. Here it is destroyed by moving under the continental margin and finally sinking into the asthenosphere. However, within these consuming plate margins tectonic accidents transfer fragments of the oceanic lithosphere (ophiolites) into or onto the continental margins. The rheology of these consuming plate boundaries is now understood in a gross way and perhaps the most important feature is their asymmetry. This asymmetry produces a polarity in tectonic, metamorphic, and magmatic processes (ERNST, 1974; GROW, 1973; DICKINSON, 1971 b).

In this section the discussion of external metamorphism of ophiolites will be related to the possible P-T regimes within a convergent plate boundary (Fig. 6). Metamorphic belts are characterstically formed along continental margins, island arcs or where orogenic zones developed during collision of two continental blocks. Paired metamorphic belts characterize the continental and island arc metamorphic belts (MIYASHIRO, 1961; ERNST, 1974) but are lacking in the orogenic zones formed at continent-continent collision (MIYASHIRO, 1973 b). The concept has been further refined by ERNST (1974) where the interplay between active subduction, calculated heat distribution, experimental petrology, and inferred geologic processes provide

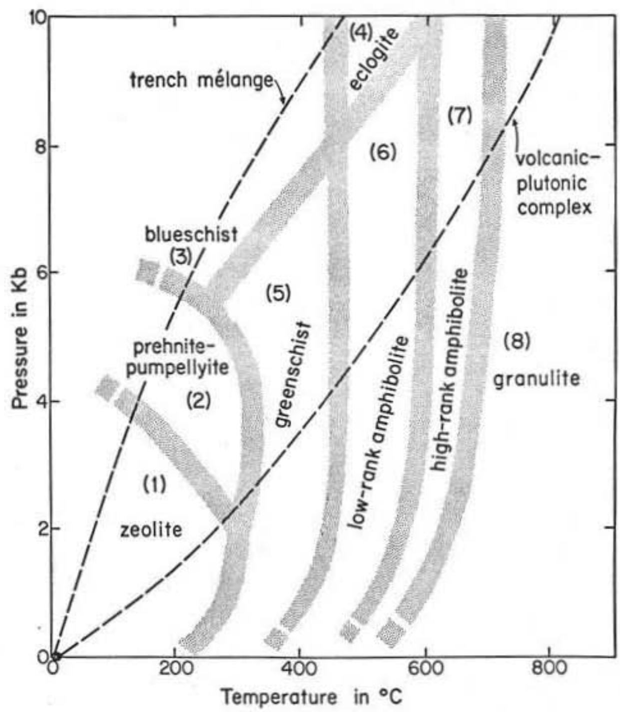

Fig. 10. — Schematic metamorphic petrogenetic grid deduced from experimental - oxygen isotope data for common rock compositions (after ERNST, 1973, Fig. 4; 1974 Fig, 4).

a three dimensional model of metamorphic facies at convergent plate boundaries (Fig. 9). From this model it is possible to consider the possible structural situation where oceanic lithosphere (ophiolites) could be metamorphosed. It also has to be assumed that decoupling or transfer of the oceanic lithosphere has to take place during or after the metamorphism in order for these rocks to be incorporated within the orogenic belts of the continental margins or island arcs.

It is possible to conceive of various P.T. regimes that could be imprinted on oceanic lithosphere at consuming continental margins (Fig. 9). Within a cold sinking lithospheric slab the isotherms are deflected downward but pressure continues to increase during subduction and thus the ophiolites will undergo a progressive metamorphism from zeolite facies (Fig. 10). Thus a high-pressure, low-temperature blueschist metamorphism could be imprinted on the upper parts of an oceanic plate during subduction. There has not yet been found a completely preserved oceanic crustal slice that has undergone blueschist metamorphism with its stratigraphy intact. That is, there are numerous dismembered parts of ophiolite that have undergone blueschist metamorphism but they are presently exposed with a melange or as highly deformed nappes.

The Zermatt-Sass ophiolites within the Piemonte zone of the western Alps are perhaps the largest exposed area of oceanic lithosphere that has undergone blueschist metamorphism as a result of subduction (DAL PIAZ, 1974; ERNST, 1973). During Jurassic time the Piemonte basin formed and consisted, in part, of oceanic crust (Zermatt-Sass ophiolites). Subduction of this oceanic crust during the late Cretaceous (70-90 m.y.) converted the Zermatt-Sass ophiolites to a HP-LT assemblage consisting of blueschists and eclogites (BEARTH, 1959, 1967, 1973). The Zermatt-Sass ophiolites are now part of a nappe consisting of tectonic slabs of ultramafics, gabbros, and pillow lavas showing a complete metamorphic and structural reworking (DAL PIAZ, 1974; BEARTH, 1974). The preservation of the Zermatt-Sass ophiolites from complete destruction in the asthenosphere was brought about by oblique planes of faulting within the subduction zone that produced a backward rise along the subduction plane to finally culminate in the empacement of the Penninic and Austro-alpine nappes (DAL PIAZ et al., 1972). To further complicate the picture after emplacement of nappes, a greenschistic metamorphic event (Alpine) overprinted and irregularly destroyed the HP-LT assemblage swithin the Zermatt-Sass ophiolites. Ophiolites formed in the same basin with the Appennines were emplaced during the closing of the same oceanic basin by obduction and they only display the effects of oceanic metamorphism (DAL PIAZ, 1974; SPOONER et al., 1974). The complexities of the Alpine metamorphism of ophiolites in the Alps are further developed by DIETRICH et al. (1974).

Blueschist metamorphism is characteristically developed in rocks underlying thick slabs of ophiolite that show oceanic metamorphism in their upper constructional parts such as in New Guinea and New Caledonia (COLEMAN, 1972 b). It has been suggested that tectonic overpressures developed during the emplacement of ophiolite slabs were instrumental in the development of the HP-LT blueschist metamorphism (COLEMAN, 1972 b). OXBURGH and TURCOTTE (1974) have shown that blueschist may develop under thick overthrust sheets because of major perturbation of the regional thermal gradient — not tectonic overpressures, Where the overthrust sheets are in excess of 15 km, sufficiently high pressures and low temperatures could prevail long enough for blueschists to form and thus explain the presence of blueschists directly under thick ophiolite slabs. As shown in the previous section HT-LP, high-rank amphibolite can be more convincingly related to obduction of ophiolites than can blueschists.

The widespread occurrence of high grade blueschists and eclogite tectonic blocks of possible ophiolitic parentage within the Franciscan complex of California suggest that small parts of subducted ophiolite material escaped destruction and were transferred to the continental margins in melange units (COLEMAN and LANPHERE, 1971). Significantly there are no exposures of lithologically intact oceanic crust (ophiolite) within the Franciscan complex and those small metamorphosed fragments of oceanic crust that have undergone HP-LT metamorphism reveal tectonic transport within serpentinite melanges (COLEMAN and LANPHERE, 1971). There are, however, extensive areas of serpentinized ultramafic within the Franciscan complex that may have been involved in HP-LT subduction blueschist metamorphism. Since there are no pressure sensitive minerals within the serpentinized peridotites these dismembered parts of an ophiolite can be of no diagnostic value in reconstructing their metamorphic history. The actual volume of ophiolites that have undergone HP-LT metamorphism make up only a minor part of these metamorphic terrains giving credence to the concept that they represent tectonic accidents within steady state subduction zones where most of the oceanic crust is destined for destruction within the asthenosphere.

It is of some interest to comment here that except for the Zermatt-Sass ophiolites and other minor Alpine localities most of the ophiolites emplaced in Alpine orogenic zones as a result of the closing of the Tethyan sea during the late Mesozoic have not been affected by HP-LT metamorphism. Furthermore, Pacific-type subduction during the late Mesozoic has incorporated only minor amounts of ophiolite in orogenic zones that have been affected by blueschist metamorphism. It can be concluded that most intact ophiolite sections are obducted onto the continental margin and escape high pressure metamorphism. Whereas, oceanic crust and mantle is incorporated into the trench melange only by tectonic accidents hence such subducted, then regurgitated ophiolites are both dismembered and exhibit high pressure metamorphism.

If we now turn our attention to the arc-trench gap and volcanic front areas at the continental margins as shown in Figure 9, it is possible to consider other pressure-temperature regimes which could affect ophiolites already incorporated into the continental edge. Stepping of the subduction zone seaward or the development of a volcanic-plutonic complex on the steeply dipping subduction zone could provide the heat necessary to pass into a prograde sequence from zeolite → prehnite-pumpellyite → greenschist → amphibolites. Thus it would be possible to convert an ophiolite to a LP-HT assemblage characteristic of the high temperature side of MIYASHIRO's (1961) paired metamorphic belts. The persistent association of amphibolites with metamorphic peridotites and lack of blueschists within the cores of progressive older orogens along Phanerozoic continental margins suggests that ophiolites of these older orogens may undergo HT-LP metamorphism.

Within the Klamath Mountains of northern California and southern Oregon, there are four different belts of ophiolite becoming sequentially younger oceanward

(COLEMAN and IRWIN, 1974; IRWIN, 1973). In some instances the metamorphic peridotites from the Klamaths are closely associated with amphibolites that on close inspection reveal that they are metabasalts and metagabbros derived from the ophiolite. High angle faults juxtaposed the meta-ophiolites against island arc volcanics and sediments that have undergone zeolite, greenschist facies, or rarely blueschist metamorphism. The high-rank amphibolite facies metamorphism of these ophiolites indicates that they may have been involved in an island arc volcanic-plutonic complex where they formed the basement rocks. Continued imbrication and compression along the continental margin produced high angle reverse faults that brought slices of the basement meta-ophiolites in contact with the lower metamorphic grade island arc volcanics and sediments.

HUTCHISON (1975) describes complete ophiolite sequences from the Darvel Bay-Lubuk-Palawan belt of North Borneo and Philippines that show HT-LP metamorphism from greenschist through high rank amphibolite facies into a metamorphic peridotite. This sequence shows strong dynamothermal metamorphic fabric and even though HUTCHISON (1975) considers this as ocean floor metamorphism, it seems more likely to represent a metamorphic belt formed under an old island arc within the Sulu sea during Mesozoic types and later exposed by obduction of the metamorphosed arc basement onto North Borneo during late Cretaceous or early Tertiary.

Similar amphibolites derived from metamorphosed basalts and gabbros (ophiolites) have been reported from Yap (SHIRAKI, 1971), the Solomon Islands (COLEMAN, 1970), Puerto Rico (TOBISCH, 1968), of Yugoslavia (PAMIC et al., 1973). The recurrence of these amphibolites with metamorphic peridotites further illustrates the total metamorphism of ophiolites within the high heat flow regime of continental edges or island arcs. The important aspect of the type of metamorphism is that as long as the peridotites remain relatively dry, their deformation or recrystallization cannot be distinguished from that produced by HP-LT metamorphism in a dry environment. Confusion then arises when blueschists are associated with the same kind of peridotites as the amphibolite. Tectonic separation after upper level serpentinization of the peridotite from its metamorphosed mafic assemblage produces a confusing mixing of amphibolites with blueschists as part of serpentinite melanges (COLEMAN and LANPHERE, 1973).

To further confuse the metamorphic history of ophiolites it is not uncommon to have younger granitic plutons invade tectonically displaced metamorphosed ophiolites and overprint them with a steep contact metamorphic thermal gradient (TROMMSDORF and EVANS, 1972; DUNGAN and VANCE, 1973; EVANS and TROMMSDORF, 1970). As mentioned earlier the presence of amphibolites along the basal contacts of ophiolites has been interpreted as a dynamothermal aureole related to the emplacement of the ophiolite (WILLIAMS and SMYTH, 1973). However, it is entirely possible that tectonic movement of an ophiolite peridotite over a regionally developed terrain of amphibolite could detach slices of amphibolite. These detached tectonic amphibolites could then conceivably come to rest at the base of the peridotite, and manifest all the aspects of an aureole. Considerable data therefore are required before amphibolites associated with allochthonous ophiolites can be properly interpreted.

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