ABSTRACT

Ophiolite metamorphism occurs before, during or after obduction onto the continental margin. Models of submarine hydrothermal alteration for modern ridge-crests have been developed from heat-flow measurements, seismic data and direct sampling of oceanic crust and mantle from DSDP drilled and dredged samples. The heat-flow data require convective heat-transfer to explain the low measured values, thus requiring the development of a fracture system from thermal contraction of the cooling crust and the tensional environment. The hydrothermal activity will persist if the fracture system remains open to accommodate fluid access as the crust ages. We do not know the distribution and depth of oceanic fracture systems away from the ridge crest and transform faults. Extent of metamorphism of the oceanic crust is strongly dependent on: (1) age, (2) crustal thickness, (3) permeability of sediments and crust, (4) fluid-flow parameters, (5) conductive vs. convective heat-flow regime, (6) development of major fault and fracture systems, and (7) spreading rate. Metamorphism of the upper part of a continuously emplaced ophiolite could have occurred within the oceanic environment. The difficulty with determining an oceanic provenance from the whole-rock basalt geochemistry is that (1) significant geochemical variations occur for oceanic basalts from different tectonic environments, and (2) major-element chemistry can be significantly affected by metamorphism. Trace-element data have been more successful in demonstrating the oceanic character of ophiolitic basalts. Seawater involvement in their metamorphism has been demonstrated by Li- and Na-enrichment patterns in the metabasalts and stable isotopic measurements on intrusive and extrusive mafic rocks. The presence of sulfide mineralization in dykes and mafic volcanic rocks at Troodos argues for the ability of an oceanic hydrothermal system to produce sulfide deposits. Metamorphism of the ophiolitic ultramafic layer probably occurs after obduction; heat-flow models do not predict penetration of water to oceanic mantle, and in most ophiolite sequences, metamorphism dies out in the gabbro layer. Therefore, unraveling the metamorphic history of an ophiolite is critical to understanding its tectonic evolution.

SOMMAIRE

Le métamorphisme d'un cortège ophiolitique peut précéder, accompagner ou suivre l'obduction sur la plateforme continentale. Pour les milieux sous-marins, les modèles d'altération hydrothermale aux crêtes océaniques sont construits à partir des mesures du flux de chaleur, des données sismiques et d'un échantillonnage direct de la croûte et du manteau (carottes DSDP et échantillons dragués). Le flux calorique observé implique un transfert de chaleur par convection, requérant un système de fractures à la crête, par contraction de la croûte au refroidissement et par tensions tectoniques. L'activité hydrothermale persistera tant que le système de fractures reste ouvert aux fluides au cours du vieillissement de la croûte. On ignore la distribution des cassures océaniques et leur profondeur loin de la crête et des failles transformantes. L'étendue du métamorphisme de la croûte océanique dépend fortement des facteurs suivants: (1) âge et (2) épaisseur de la croûte, (3) perméabilité de la croûte et des sédiments, (4) paramètres du flux des fluides, (5) importance relative de la conduction et de la convection, (6) développement de systèmes importants de failles et de fractures et (7) vitesse de séparation des plaques. Les suites ophiolitiques que l'on trouve sur les plaques continentales peuvent avoir été métamorphisées à leur partie supérieure en milieu océanique, mais il est difficile de le prouver à partir du géochimisme global des roches, vu: (1) la grande variation des basaltes en fonction du milieu tectonique et (2) la mobilité des éléments majeurs pendant le métamorphisme. A cet égard, les éléments traces sont plus utiles pour établir le caractère océanique des basaltes ophiolitiques. Le rôle de l'eau de mer dans leur métamorphisme est démontré par un enrichissement en Li et Na et par les isotopes stables des roches intrusives et extrusives. L'enrichissement en S dans les dykes et les coulées mafiques à Troodos montre que les systèmes hydrothermaux océaniques peuvent donner des gisements de sulfures. Le métamorphisme des niveaux ultramafiques suit probablement l'obduction; les modèles de flux de chaleur ne prévoient pas de pénétration de l'eau jusqu'au manteau. Dans la plupart des cas, le métamorphisme disparaît dans le niveau gabbroïque. Pour comprendre les événements tectoniques de l'histoire d'une ophiolite, il faut connaître son évolution métamorphique.

(Traduit par la Rédaction)
INTRODUCTION

Ophiolites are stratigraphically connected sequences of igneous and sedimentary rocks (Conference Participants 1972) that are usually considered to be part of the oceanic crust (Fig. 1). These rocks are partly to completely metamorphosed (Table 1) and have been tectonically emplaced onto continental crust.

The purpose of this paper is to examine the metamorphism of individual ophiolite sequences to ascertain the relative timing of metamorphism and to establish criteria for recognizing the metamorphic effects on the primary igneous stratigraphy. In the process of doing this, various models of the oceanic crust will be examined to determine whether metamorphism of the oceanic crust could occur in situ prior to tectonic emplacement, or whether metamorphism of the obducted oceanic slab occurs during or after tectonic transport. This work will build on the discussion of ophiolite metamorphism summarized by Coleman (1977).

OCEANIC HYDROTHERMAL MODELS


Evidence for the presence of hydrothermal circulation in the oceanic environment is derived from: (1) the presence of metalliferous sediments at oceanic ridges (Bostrom et al. 1971, Dymond et al. 1973, Piper 1973, Francheteau et al. 1979) and in the Red-Sea brines (Bischoff 1969, Shanks & Bischoff 1977), (2) heat-flow measurements at ridge crests and flanks (Williams et al. 1974, Wolery & Sleep 1976, Anderson & Langseth 1977, Herman et al. 1977, Anderson et al. 1979), (3) decay in intensity of magnetic remanence away from a ridge axis (Irving 1970), (4) microearthquake swarms at mid-ocean ridges (Sykes 1971), (5) presence of an active hydrothermal system such as at the Galapagos rift (Corliss et al. 1977, 1978, Lonsdale 1977, von Herzen et al. 1977) and TAG hydrothermal field (Lowell & Rona 1976). The sediment mounds associated with the Galapagos hydrothermal system are inferred to have formed from slow percolation of hydrothermal fluids circulating through the basaltic crust and overlying sediments (Corliss et al. 1978), though direct vertical connection between the mounds and the hydrothermal vents has not been shown to occur (Natland et al. 1979). Hydrothermally circulating seawater also has been directly observed from the subaerial geothermal fields at Reykjanes, Iceland (Björnsson et al. 1972, Mottl et al. 1975). Under the assumption that ophiolites are ancient oceanic crust, arguments have been put forth to use the occurrence of massive sulfides within the ophiolite (Parmentier & Spooner 1978, Rona 1978) and whole-rock isotope geochemistry to infer the presence of a former hydrothermal circulation system (Spoonier 1978), ostensibly while at or near an oceanic ridge.

Heat-flow models

In his heat-flow model for oceanic ridges, Sleep (1975) postulated that vertical heat-conduction from the ridge axis is a function of spreading rate. Cooling due to conduction (heat generated from magma accreting at ridge axis) is more rapid near a ridge axis with slow spreading-rates because the heat is conducted almost vertically. However, heat-flow measurements indicate that conduction is not the predominant mode of heat transfer in young oceanic crust.

Vertical heat-transfer at the ridge can be accelerated by hydrothermal circulation systems, as they add a connective term to the total heat-dissipation. Williams et al. (1974) and Wolery & Sleep (1976) postulated that hydrothermal circulation is responsible for the anomalously low heat-flow observed at ridges. Wolery & Sleep (1976) estimated the rate of hydrothermal heat advection by computing the difference between a theoretical heat-production associated with sea-floor spreading and the observed heat-flow. This calculation leads to the extremely high calculated flow-rate of seawater through the oceanic hydrothermal system of $1.3 - 9.0 \times 10^9$ g/y. Wolery & Sleep (1976) postulated that hydrothermal activity would...
TABLE 1. ROCKS OF THE CLASSIC OPHIOLITE SEQUENCE

<table>
<thead>
<tr>
<th>Oceanic Layer</th>
<th>Unmetamorphosed Rocks</th>
<th>Metamorphic Equivalent</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Marine sediments</td>
<td>Chert</td>
</tr>
<tr>
<td>2</td>
<td>Basalt</td>
<td>Spilitite (keratophyres)**</td>
</tr>
<tr>
<td>3</td>
<td>Gabbro</td>
<td>Metagabbro or amphibolite</td>
</tr>
<tr>
<td>Meltline</td>
<td>Ultramafic rock</td>
<td>Serpentinite (rodinglites)**</td>
</tr>
</tbody>
</table>

**Sheeted dikes may or may not be present between the intrusive gabbroic layer and the extrusive basalt layer (Fig. 1).

**Commonly associated rocks produced during metamorphism.

FIG. 1. Stratigraphy of two well-described ophiolite sequences compared to oceanic layer 1, 2, 3 and 4 (mantle).

for Galapagos spreading-centre, 10–15 m.y. for the East-Pacific rise, 50–70 m.y. on the Mid-Atlantic ridge). This transition is dominated by the development of a certain sediment-thickness (≥ 300 m) and a change in composition from carbonate to siliceous deposition, which effectively decreases the bulk permeability of the sediment cover. Also, hydrothermal flow decreases with time owing to the clogging of fracture systems within the crust by alteration minerals (Sleep & Wolery 1978). Therefore, other physical variables, such as permeability and flow parameters, are also strongly a function of age and spreading rate. By means of very closely spaced heat-flow surveys, Anderson et al. (1979) have been able to calculate both the conductive and convective components of heat flow through oceanic crust and sediments (Indian ocean), further substantiating the importance of convective processes in the heat-flow regime of young oceanic crust.
Depth of fluid penetration

Heat-flow models (Lister 1972, 1974, 1977, Ribando et al. 1976) have been postulated in which the depth of penetration of water is approximately 5 km, based on the wavelength of the heat-flow variation observed at the Galapagos spreading-centre (Williams et al. 1974). Permeability alone limits the access of fluids to about the upper 200–300 m of the oceanic basement (Wolery & Sleep 1976). Therefore, the permeability of the oceanic crust and sediments is not sufficient to explain the heat-flow measurements; most heat-flow models also call upon a fracture system, developed through cooling and static fatigue (Lister 1974), to explain the anomalously low heat-flow values. Widespread porous flow has been shown to be an unlikely mechanism for fluid circulation (Sleep & Wolery 1978).

DSDP drilling results in the Atlantic oceanic crust show a general absence of hydrothermal alteration regardless of crustal age, which led Hall & Robinson (1979) to conclude that the hydrothermal circulation system must occur at depths greater than 600 m and that it is probably irregularly distributed. However, in the Pacific, large fault-scarps about 1 km apart (Williams et al. 1974, Rosendahl 1976) dominate the topography and provide zones of high permeability. Also, in DSDP leg 54, Hekinian et al. (1977) reported intensely fractured and brecciated oceanic crust on the southern flank of the Galapagos spreading-centre. The development of a fracture system in the oceanic crust is related to the thickness of the basaltic crust, its brittleness (a function of composition) and the spacing of major fault patterns (Rosendahl, pers. comm. 1979).

Fluid penetration obviously determines the depth to which metamorphism can occur. Heat-flow models do not predict penetration to the oceanic mantle. The seismic velocity data (Rosendahl 1976, Fig. 5) and oceanic crustal models generated from those data (Clague & Straley 1977) cannot distinguish between pristine igneous rock in layer 3 and the uppermost mantle (Fig. 1) and metagneous rocks and

---

**Fig. 2.** Ocean-ridge model for a fast-spreading centre from Sleep & Rosendahl (1979). The magma chamber (dots), which is probably filled with cumulate mush, and the conduit of upwelling mantle material (angles) are very wide. The walls of sides (half circles) or material segregated from the partial melt. The inner intrusion-zone (dense gabbroic crust (slanted lines) and the magma chamber are possible anomalous masses. The geotherms are calculated assuming conductive heat-transfer only.
serpentinite. Epp & Suyenaga (1978) do not agree with Sleep & Wolery (1978) that hydrothermal precipitation will close the fracture system and stop circulation in the crust older than the heat-flow transition zone (Williams et al. 1974). They argued that thermal contraction continues as the crust continues to cool, which allows the fracture system to remain open and to accommodate the volume occupied by the formation of hydrothermal minerals. Thus, fractures would increase in depth with age and therefore would increase the extent of metamorphism with age, perhaps allowing fluid penetration to the mantle. Serpentinites have been dredged from Atlantic oceanic ridges and fracture-zones (Aumento et al. 1971, Bonatti & Honnorez 1976). Bonatti & Honnorez (1976), Bonatti (1976), Clague & Straley (1977) and Epp & Suyenaga (1978) proposed diapiric intrusion of serpentinite into fault zones parallel to oceanic ridge-axes, which implicitly assumes that fluid access to the lower crust and mantle occurred through deep fractures produced by the tensional environment at the ridge.

**Fast versus slow spreading-rates**

The importance of fast versus slow spreading has been emphasized not only with respect to heat flow, but also to chemical and stratigraphic differences in the oceanic crust generated (Rosendahl 1976, Lister 1977, Sleep & Rosendahl 1979). Fracture patterns are distinctly different for the two rates of spreading; fast-spreading ridges are marked by axial-block topography and slow-spreading ridges by axial valleys (Rosendahl 1976). The salient structural difference between the two types of oceanic ridge is the presence or absence of deeply penetrating faults; their presence seems to characterize slow-spreading ridges. The difference in the fault patterns between the slow and fast-spreading ridges can also be inferred from the common occurrence of metagneous rocks dredged from slow-spreading ridges (Mid-Atlantic ridge) and the absence of metamorphic rocks from fast-spreading ridges (Pacific ocean-basin) (Rosendahl 1976).

Sleep & Rosendahl (1979) presented possible geothermal gradients away from the ridge axis based on evidence that a fast-spreading ridge has a relatively large, wide magma-chamber (Fig. 2), and a slow-spreading ridge has a very narrow magma-chamber (Fig. 3). The slope of the isotherms is much less steep at the fast-

---

**Fig. 3.** Ocean-ridge model for slow-spreading centre from Sleep & Rosendahl (1979). Symbols are same as Figure 2. Contrast with fast-spreading centres indicates a very small magma chamber, a thick sheeted-dyke complex, a wide zone of cooler material segregated from melt and a narrower intrusive zone. The geotherms are calculated assuming conductive heat-transfer only.
spreading ridge because of the large magma-chamber and reduced vertical heat-loss by conduction. The Nisbet & Fowler (1978) model for a slow-spreading ridge suggests the absence of a magma chamber and injection of magma by crack propagation. However, the postulated geothermal gradients for both types of spreading centre are high enough to drive a hydrothermal circulation system and to be the cause of metamorphism of the oceanic crust. The presence of a hydrothermal circulation system at the oceanic ridge would significantly alter the geotherms (Figs. 2, 3) as calculated by Sleep & Rosendahl (1979). Parmentier & Spooner (1978) have attempted to model such a convective system, and have demonstrated that the magnitude of the geotherms can decrease significantly as a function of the boundary conditions chosen.

The preceding discussion has shown that oceanic hydrothermal systems exist and that their activity is determined by such factors as: (1) age of oceanic crust, (2) crustal thickness, (3) permeability of sediments and crust, (4) fluid-flow parameters, (5) conductive versus convective heat-flow regime, (6) development of major fault- and fracture-systems, and (7) spreading rate. Unfortunately, the extent of oceanic metamorphism, exclusive of fault and fracture zones that allow fluid access, is unknown because of the technical inability to sample deep oceanic crust away from those zones. Therefore, an understanding of oceanic hydrothermal systems is critical to an understanding of the extent of oceanic metamorphism because fluid access to rock determines the extent of metamorphism. Further work is needed on the processes of formation, spacing, depth and frequency of occurrence of fracture systems in the crust, since they are a critical part of the hydrothermal system, both at and away from the oceanic ridge.

**Ophiolites**

The stratigraphic sequence of rocks making up the complete ophiolitic sequence is relatively well defined (Conference Participants 1972). The major problem in correlating ophiolites with oceanic lithosphere is that the oceanic provenance of the ophiolite is uncertain. Various models have been proposed suggesting ophiolites can be correlated with mature oceanic crust and mantle (Gass 1977), marginal-basin type of crust and mantle, or immature oceanic crust, i.e., that from a ridge crest (Rosendahl 1976).

**Geochemistry of ophiolitic rocks**

Experiments with basalt and seawater (Table 2) show major-element exchange between basalt and fluid. Table 2 also illustrates element changes in oceanic and ophiolitic basalts that have undergone greenschist-facies metamorphism. A good correlation exists between the direction and change of element mobilities between oceanic metabasalts and experimentally derived analogues. The metabasalts of ophiolite suites show the greatest variation in the major-element mobility when compared to the experimentally derived metabasalts and to the oceanic metabasalts.

**Experimental work on mineral stabilities and**

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**TABLE 2. ELEMENTAL MOBILITIES DURING LOW-GRADE METAMORPHISM OF BASALTS**

<table>
<thead>
<tr>
<th>Element</th>
<th>Experimental</th>
<th>Oceanic Metabasalts</th>
<th>Ophiolitic Metabasalts</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>-</td>
<td>* or +</td>
<td>+</td>
</tr>
<tr>
<td>TiO₂</td>
<td>-</td>
<td>*</td>
<td>+</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>-</td>
<td>*</td>
<td>* or +</td>
</tr>
<tr>
<td>FeO</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>n.d.</td>
<td>+</td>
<td>+</td>
</tr>
<tr>
<td>MnO</td>
<td>+</td>
<td>+</td>
<td>-</td>
</tr>
<tr>
<td>MgO</td>
<td>-</td>
<td>+</td>
<td>-</td>
</tr>
<tr>
<td>CaO</td>
<td>+</td>
<td>+</td>
<td>+, *</td>
</tr>
<tr>
<td>Na₂O</td>
<td>-</td>
<td>+</td>
<td>+</td>
</tr>
<tr>
<td>K₂O</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>P₂O₅</td>
<td>n.d.</td>
<td>+</td>
<td>+</td>
</tr>
<tr>
<td>H₂O</td>
<td>-</td>
<td>+</td>
<td>+</td>
</tr>
</tbody>
</table>

(-) loss during metamorphism
(*) gain during metamorphism
(+) no change
n.d. not determined
Hollis (1970), (1976), Mottl & Holland (1978) (low rock/seawater ratio), comparison to unaltered starting-material basalt

---

**TABLE 3. COMPARISON OF ELEMENTAL MOBILITIES DURING LOW-GRADE METAMORPHISM OF PERIDOTITE**

<table>
<thead>
<tr>
<th>Element</th>
<th>Experimental</th>
<th>Oceanic Meta-Peridotite</th>
<th>Ophiolitic Meta-Peridotite</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>-</td>
<td>*</td>
<td>-</td>
</tr>
<tr>
<td>TiO₂</td>
<td>n.d.</td>
<td>*</td>
<td>-</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>n.d.</td>
<td>*</td>
<td>-</td>
</tr>
<tr>
<td>FeO</td>
<td>n.d.</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>n.d.</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>-</td>
<td>+</td>
<td>-</td>
</tr>
<tr>
<td>MnO</td>
<td>+</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>MgO</td>
<td>+</td>
<td>+</td>
<td>+</td>
</tr>
<tr>
<td>CaO</td>
<td>+</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Na₂O</td>
<td>-</td>
<td>+</td>
<td>-</td>
</tr>
<tr>
<td>K₂O</td>
<td>+</td>
<td>*</td>
<td>-</td>
</tr>
<tr>
<td>P₂O₅</td>
<td>-</td>
<td>+</td>
<td>-</td>
</tr>
<tr>
<td>H₂O</td>
<td>-</td>
<td>+</td>
<td>+</td>
</tr>
</tbody>
</table>

(-) loss during metamorphism
(*) gain during metamorphism
(+) no change
n.d. not determined
metamorphism of ultramafic rocks (Chernosky 1975, Moody 1976a, b, Seyfried & Dibble 1978) documents some major compositional changes during serpentinization, involving seawater under greenschist-facies conditions (Table 3). These changes chiefly involve an addition of magnesium and water and a loss in total iron, SiO₂, Al₂O₃ and FeO. Major-element data to support arguments of metamorphism of both basalts and ultramafic rocks in ophiolite sequences; thus, care must be exercised in an extrapolation from major-element chemistry to the tectonic setting of the igneous rocks. The weight of evidence for the Troodos complex is further complicated not only by metamorphism, but also by real geochemical differences reflecting tectonic environments.

Miyashiro (1975a) has also used whole-rock chemistry to classify ophiolites into three different groups based on the inferred presence of calc-alkaline or tholeiitic trends as shown by the basaltic rocks of ophiolitic sequences. However, Pearce & Gale (1977) and Sun & Nesbitt (1978) suggested, as did Miyashiro (1975a, b, c), that some real geochemical variation exists among oceanic basalts from the different tectonic environments within oceanic lithosphere (e.g., oceanic ridge-crest versus island arc versus back-arc basin). Therefore, the petrological origin of ophiolitic basalt is further complicated not only by metamorphism, but also by real geochemical differences reflecting tectonic environments.

The direction of element mobility for the
metamorphosed oceanic versus ophiolitic ultramafic rocks is the same except for SiO₂, Al₂O₃ and K₂O (Table 4). However, the preliminary results of peridotite-seawater experiments (Seyfried & Dibble 1978) conflict with the geochemistry of metamorphic rocks, especially with respect to CaO. Calcium is lost from ultramafic rocks during metamorphism in both oceanic and ophiolitic rocks, and the leached calcium is deposited in the rocks associated with serpentinites, e.g., rodingites (Barnes & O'Neil 1969, Barnes et al. 1972, 1978, Coleman 1977, Pfeifer 1977, Wenner 1979). Honnorez & Kirst (1975) described rodingites dredged from the Mid-Atlantic-ridge fracture zones; these formed by the alteration of a noritic gabbro. The element-mobility data support the formation of rodingites by fluids involved in the metamorphism of ultramafic rocks. CaO is also shown to be lost during metamorphism of basalts.

Tables 2 and 4 indicate that Na-enriched metabasalts (i.e., spilites) can form from the interaction of seawater with basalt. Both oceanic and ophiolitic metabasalts demonstrate an enrichment in sodium, which agrees with the experimental data on basalt-seawater interaction. A comparison of the direction of mobility for the other elements in the experimentally derived metabasalts with the direction of those in the oceanic metabasalts demonstrates very good to excellent agreement. A comparison of oceanic and ophiolitic metabasalts shows a much greater variation in patterns of element mobility.

Further evidence for seawater involvement in the formation of spilites could be inferred from the data of Shaw et al. (1977). Enriched in sodium and water, spilites are also enriched six-fold in Li compared to the unmetamorphosed basalt. Seawater, brines and volcanic waters are obvious choices for the Li- and Na-enrichment in the spilites.

The data in Tables 2 and 4 indicate that the metamorphism of basaltic rocks could occur in the oceanic crust. Also, the wider variability in the major-element-mobility data for ophiolitic metabasalts may indicate the effects of more than one metamorphic event, with involvement of a fluid other than seawater. The data for the ultramafic rocks (Tables 3, 4) are inconclusive.

Pearce & Cann (1971) used the Ti, Zr and Y contents of four different ophiolitic basalts to demonstrate that the volcanic rocks have ocean-floor provenance. Smewing et al. (1975) argued that the major- and trace-element contents of Troodos metabasalts are significantly different from those formed at present-day constructive plate-margins, and proposed an origin at a slow-spreading ridge within a small, marginal ocean-basin. Coish (1977) asserted that metamorphism had little effect on TiO₂, P₂O₅, Zr, Y, Cr or Ni contents of metabasalts from the Betts Cove ophiolite, Newfoundland, and proposed an ocean-floor origin for the basalts. Beccaluva et al. (1977) determined 12 different trace-element contents for the rocks within the ophiolitic sequence from eastern Corsica and concluded that the ophiolitic metabasalts showed an ocean-floor affinity.

Rare-earth-element (REE) data from ophiolitic rocks are sparse; rocks from the Pindos suite, Greece (Montigny et al. 1973), the Troodos complex, Cyprus (Kay & Senechal

### TABLE 5. GEOCHEMICAL CRITERIA USED BY BLAND (1978) TO DETERMINE THE ORIGINAL CHARACTER OF APPALACHIAN METABASALTS

<table>
<thead>
<tr>
<th>Chemical Discriminator</th>
<th>Rock Types Distinguished</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>12-20 wt. % CaO + MgO*</td>
<td>selects rocks of similar tectonic affinities &amp; eliminates ultramafic cumulates</td>
<td>Pearce &amp; Cann (1973)</td>
</tr>
<tr>
<td>&gt;20% CaO + MgO*</td>
<td></td>
<td>Pearce (1976)</td>
</tr>
<tr>
<td>alkali</td>
<td></td>
<td>Jakes &amp; White (1972)</td>
</tr>
<tr>
<td>K₂O &amp; MgO/K₂O ratio</td>
<td>GFB &amp; RA from (LKT, CAB, SHO, 01)</td>
<td>Rogers et al. (1974)</td>
</tr>
<tr>
<td>F₁-F₂ discriminate function</td>
<td>WFB, SHO, OFB, (LKT &amp; CAB)</td>
<td>Pearce (1976)</td>
</tr>
<tr>
<td>F₂-F₃ discriminate function</td>
<td>SHO, LKT, CAB, OFB</td>
<td>Pearce (1976)</td>
</tr>
<tr>
<td>Ti vs. Zr, ppm**</td>
<td>GFB, CAB, LKT</td>
<td>Pearce &amp; Cann (1973)</td>
</tr>
<tr>
<td>Zr vs. Ti/100 vs. (Y), ppm</td>
<td>WFB, CAB, LKT, and GFB + CAB + LKT</td>
<td>Pearce &amp; Cann (1973)</td>
</tr>
<tr>
<td>CaO wt % vs. Y, ppm</td>
<td>LKT &amp; SHO plot to the left of SCAT; O1, CON, GFB plot to the right of SCAT; CAB plot within SCAT</td>
<td>Lambert &amp; Holland (1974)</td>
</tr>
</tbody>
</table>

** coupled with TiO₂ content

SYMBOLS:

- OBF = ocean-floor basalts
- LKT = low-potassium tholeiites
- CAB = calc-alkaline basalts
- SHO = shoshonites
- O1 = oceanic-island basalts
- CD = continental-rifting basalts
- RA = rear-arc basalts
- WP = within-plate basalts
- SCAT = standard calc-alkaline trend
SERPENTINITES, SPILITES AND OPHIOLITE METAMORPHISM

1976, Smewing & Potts 1976) and the Point Sal ophiolite, California (Menzies et al. 1977) have been examined. Unfortunately, total agreement on the effect of metamorphism on the primary basalt-REE distributions is not evident. Menzies et al. (1977) summarized evidence that indicates light-REE mobility during zeolite-facies metamorphism of ophiolitic basalts. This mobility results in an increase in the total concentration of REE and a change in the profile characteristics due to Ce or La mobility. At the same time, greenschist-facies metamorphism has not changed the REE patterns of the primary basaltic lavas of ophiolite complexes (e.g., Point Sal ophiolite). Conversely, Montigny et al. (1973), Kay & Senechal (1976), Smewing & Potts (1976), Coish & Church (1978) present evidence that REE abundances of ophiolitic basalts were not modified by metamorphism for the Pindos, Troodos and Betts Cove complexes. Clearly, more REE data are needed for the total stratigraphy of an ophiolite sequence before the question of REE mobility during metamorphism can be resolved.

A combination of major- and trace-element data is more convincing in demonstrating an ocean-floor origin for the ophiolitic metabasalts. Bland (1978) has confirmed that the tectonic origin of a metabasalt can be determined, but that more than one geochemical test must be applied in order to see through the metamorphic overprint. Table 5 summarizes the different criteria used by Bland (1978) to determine the tectonic setting of many different ages of the metabasalts in the Appalachian Blue Ridge and Piedmont. A more definitive statement about the origin of the ophiolite basalts could be made if a complete geochemical study were done on these rocks, similar to Bland’s (1978), using a statistically significant sample size. Pearce & Gale (1977) have used a combination of whole-rock chemistry, trace-element and REE data of metabasalts to classify the tectonic environment (oceanic) asso-

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Fig. 4. Summary of whole-rock oxygen-isotope data obtained from Spooner et al. (1973, 1977a, b), Wenner & Taylor (1973), Magaritz & Taylor (1974).
associated with the formation of massive-sulfide deposits related to ophiolites.

Oxygen-isotope data have been used to indicate the type of water involved in metamorphism as well as the temperature of metamorphism. Spooner et al. (1974, 1977a) argued for seawater involvement in the metamorphism of eastern Liguria, Pindos and Troodos ophiolitic basalts. However, Figure 4 shows that the ophiolitic metamorphic rocks are more enriched in $\delta^{18}O$ compared with the oceanic rocks. Some overlap is present in the data, especially for the partly serpentinized ultramafic rocks. However, these oxygen-isotope data alone do not necessarily provide conclusive evidence for seawater involvement.

Sheppard (1977) and Heaton & Sheppard (1977) presented both $\delta^{18}O$ and $\delta^D$ isotopic data for the metamorphosed Troodos pillow lavas, sheeted intrusive complex, trondhjemites and upper gabbro. The $\delta^D$ values for Troodos rocks overlap with the few available measurements of oceanic metamorphic rocks. Mineral data are combined with whole-rock values to calculate models for $\delta^D$ and $\delta^{18}O$ fractionation assuming equilibrium, with the isotopic differences reflecting different temperatures (i.e., different metamorphic grades. Heaton & Sheppard (1977) concluded that a large component of seawater was involved in the metamorphism of Troodos rocks.

Wenner & Taylor (1973) and Magaritz & Taylor (1974) documented the evidence for hydrothermal circulation. A total geochemical study of ophiolite metabasalts assemblages indicate that temperature increased with depth in the basaltic lavas. The $\delta^{18}O$ values decrease with depth but do not reach the $\delta^{18}O$ composition of primary basalt. The $\delta^{18}O$ profile was assumed to be controlled by the temperature gradient. Different oxygen-isotope values and oxidation ratios were measured for cores versus rims of metabasalt pillows, indicating that isotopic and chemical equilibrium was not achieved during metamorphism. Spooner et al. (1977a) combined the oxygen-isotope data with the oxidation ratio Fe$^{3+}$/Fe$^{2+}$ to calculate the amount of water involved in the alteration of the basalt (10$^3$–10$^4$ : 1 water:rock ratio). Their calculations assume that the fluid involved in the alteration was seawater.

Measurements of $^{87}$Sr/$^{86}$Sr ratios in the Troodos mineralized and metamorphosed dykes and mafic volcanic rocks (Chapman & Spooner 1977, Spooner et al. 1977b) yield values enriched in $^{87}$Sr relative to fresh oceanic basalts. These authors suggested that seawater was the ore-forming fluid for the sulfide ore-deposits of the Troodos complex, a hypothesis substantiated by fluid-inclusion data from vein quartz coexisting with the sulfides (Spooner & Bray 1977). They gave the occurrence of the deposits as further evidence of a major oceanic hydrothermal-circulation system within the basalts.

Spoonier et al. (1977a) for hydrothermal metamorphism of ophiolite metabasalts from eastern Liguria, Italy, is based on oxygen-isotope data, water contents and Fe$^{3+}$/(Fe$^{2+}$+Fe$^{3+}$) ratios of the rocks. The metamorphic mineral assemblages indicate that temperature increased with depth in the basaltic lavas. The $\delta^{18}O$ values decrease with depth but do not reach the $\delta^{18}O$ composition of primary basalt. The $\delta^{18}O$ profile was assumed to be controlled by the temperature gradient. Different oxygen-isotope values and oxidation ratios were measured for cores versus rims of metabasalt pillows, indicating that isotopic and chemical equilibrium was not achieved during metamorphism. Spooner et al. (1977a) combined the oxygen-isotope data with the oxidation ratio Fe$^{3+}$/Fe$^{2+}$ to calculate the amount of water involved in the alteration of the basalt (10$^3$–10$^4$ : 1 water:rock ratio). Their calculations assume that the fluid involved in the alteration was seawater.

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Spoonier et al. (1977a), Andrews & Fyfe (1976), Coleman (1977) and Rona (1978) have documented the evidence for hydrothermal systems at spreading centres with regard to seawater leaching of the basaltic oceanic crust to derive the metals for the massive sulfides of ophiolitic complexes. Spoonier (1977) and Parmentier & Spooner (1978) have developed a model for Cyprus deposits, where a strong case can be made for the location of the sulfide deposits within the metabasalt layer as a result of oceanic hydrothermal circulation. A total geochemical study of ophiolite metabasalts associated with massive sulfides in Cyprus, Oman and Betts Cove has led Pearce & Gale (1977) to postulate that the basalts formed at a spreading axis in an oceanic back-arc basin rather than at a major oceanic ridge-crest. Unfortunately, the subsequent metamorphism and deformation of the ophiolitic metabasalts of the Whalesback cupriferous iron deposit, Newfoundland, do not permit the relationship between
sulfide mineralization and sub-sea-floor metamorphism to be determined (Bachinski 1977).

**Metamorphic mineralogy**

As shown previously (Table 1), metaigneous rocks similar to those in ophiolite suites have been recovered in dredge samples from oceanic ridges and fracture zones. Table 6 summarizes the essential mineralogy of serpentinites and spilites (layers 2 and 4 in the stratigraphy of the ophiolite sequences). The metamorphic mineralogy is not very helpful in contrasting the oceanic versus continental ophiolitic environment—they are the same. The interesting problem is that both the serpentinites and spilites (Na-enriched metabasalts) can undergo metamorphism in the same range of temperature-pressure conditions (Moody 1976a, b, Moody & Meyer 1978, Meyer 1978). If that setting is defined in terms of zeolite- and greenschist-facies metamorphism, then the distinction between the timing of metamorphism and occurrence of serpentinites and spilites together in the ophiolite sequence can be made.

Many models of sub-sea-floor metamorphism (Spooner & Fyfe 1973, Bonatti et al. 1975, Coleman 1977, Elthon & Stern 1978, Humphris & Thompson 1978a, b) indicate increasing metamorphic grade with increasing depth. Oceanic metamorphism is depicted in these models as dying out in the gabbroic layer at conditions of upper-greenschist or amphibolite metamorphic facies.

Ophiolitic rocks that show this increasing grade of metamorphism down through the ophiolite sequence, from sediments to extrusive and then to intrusive mafic rocks (Fig. 1) are found in (1) eastern Liguria, Italy (Spooner et al. 1974, 1977a), (2) the Sarmiento complex, Chile (Elthon & Stern 1978), (3) the Chenaillat massif, France (Mevel et al. 1978), (4) Darvel Bay, Borneo (Hutchison 1978). Table 7 summarizes the best described of the known ophiolite sequences with respect to metamorphism. The difficulty in preparing the table was that information on the metamorphism of the total stratigraphic sequence is generally unavailable. Those ophiolites that give evidence for a possible oceanic episode of metamorphism reach upper-greenschist facies, although Mevel et al. (1978) described amphibolite-facies metamorphism in an ophiolite gabbroic layer. Amphibolites formed from metamorphism of the intrusive mafic layer have been observed in the oceanic environment (Table 1). These amphibolites are not to be confused with those produced at the base of the ophiolite in a metamorphic aureole that is clearly related to its tectonic emplacement (Dewey & Bird 1971, Jamieson 1979, McCall & Kemp 1979).

Most ophiolitic serpentinites are composed of lizardite-chrysotile rather than antigorite (Wenner & Taylor 1973, Magaritz & Taylor 1974, Wicks & Whittaker 1977, Prichard 1979), indicating a lower temperature of metamorphism than the antigorite field of stability (Moody 1976a, Evans 1977). However, Liou & Ernst (1979) reported antigorite in ultramafic rocks of the East-Taiwan ophiolite, and other upper-greenschist-facies minerals from the metabasalts. Clearly, depth of penetration of circulating ocean waters is important, as are metamorphic pressure-temperature conditions. In the case of most ophiolitic ultramafic rocks, amphibolite-facies metamorphism is too high a grade of metamorphism to have taken place in oceanic environments.

Amphibolite-grade oceanic metamorphism of ultramafic rocks is uncommon; Bonatti et al. (1970) have described anthophyllite and cummingtonite in a dunite. In fact, prograde metamorphism of serpentinite with formation of “metamorphic" olivines has been postulated to

**TABLE 6. TYPICAL MINERALOGY OF OPHIOLITIC METABASALTS AND METAPERIDOTITES**

<table>
<thead>
<tr>
<th>Metamorphic mineralogy</th>
<th>Examples</th>
</tr>
</thead>
<tbody>
<tr>
<td>Chlorite, epidote, albite, actinolite</td>
<td>Vourinos, Northern Greece</td>
</tr>
<tr>
<td>± sphene, zeolites, calcite</td>
<td>Macquarie Island</td>
</tr>
<tr>
<td>Serpentinites: lizardite, chrysotile, magnetite</td>
<td>Bay of Islands, Newfoundland</td>
</tr>
<tr>
<td>± brucite, antigorite</td>
<td>Blow-Me-Down, Newfoundland</td>
</tr>
<tr>
<td>* Essential metamorphic mineralogy of oceanic metaigneous rocks and continental ophiolitic rocks is same.</td>
<td></td>
</tr>
</tbody>
</table>

**TABLE 7. METAMORPHIC HISTORY OF OPHIOLITES**

<table>
<thead>
<tr>
<th>Degree of metamorphism</th>
<th>Examples</th>
</tr>
</thead>
<tbody>
<tr>
<td>Metamorphism in oceanic crust down to gabbro layer only; serpentinitization during or after tectonic emplacement</td>
<td>Vourinos, Northern Greece</td>
</tr>
<tr>
<td></td>
<td>Macquarie Island</td>
</tr>
<tr>
<td></td>
<td>Bay of Islands, Newfoundland</td>
</tr>
<tr>
<td></td>
<td>Blow-Me-Down, Newfoundland</td>
</tr>
<tr>
<td></td>
<td>Betsu Cove, Newfoundland</td>
</tr>
<tr>
<td></td>
<td>Troodos, Cyprus</td>
</tr>
<tr>
<td></td>
<td>Sarmento complex, Chile¹</td>
</tr>
<tr>
<td></td>
<td>Point Sal, California²</td>
</tr>
<tr>
<td></td>
<td>Semail, Oman</td>
</tr>
<tr>
<td></td>
<td>Papua-New Guinea</td>
</tr>
<tr>
<td></td>
<td>Chenaillat, France³</td>
</tr>
<tr>
<td></td>
<td>E. Liguria, Italy⁴</td>
</tr>
<tr>
<td></td>
<td>Darvel Bay, Borneo⁵</td>
</tr>
<tr>
<td></td>
<td>Dunn Hm., New Zealand⁶</td>
</tr>
<tr>
<td></td>
<td>Southern Québec⁷</td>
</tr>
<tr>
<td></td>
<td>Eastern Taiwan⁸</td>
</tr>
<tr>
<td></td>
<td>Karmy, Norway⁹</td>
</tr>
</tbody>
</table>

occur at the greenschist-amphibolite boundary (Vance & Dungan 1977, Dungan 1977) with the coexistence of antigorite and olivine. Johannes (1975) & Evans et al. (1976) determined the equilibrium breakdown of antigorite to forsterite + talc to occur at approximately 500°C at 2 kbar total pressure. Prograde metamorphism of a serpentinite to the amphibolite facies would produce the mineral assemblage forsterite + talc + tremolite (Vance & Dungan 1977, Evans 1977), not the common metamorphic assemblage found in metamorphosed ophiolitic ultramafic rocks (Table 6). Therefore, the presence of an amphibolite-facies metamorphic assemblage in the ultramafic layer of an ophiolite probably indicates a prograde regional metamorphism of the ophiolite after emplacement of the ophiolite on the continent.

**Metamorphic Models for Ophiolites**

The discussion presented in the two previous sections indicates that ophiolitic basalts have an oceanic character. Metamorphism of layer 1, layer 2 and part of layer 3 (Fig. 1) probably occurs in the oceanic crust before tectonic emplacement of the ophiolite, whereas metamorphism of the lower part of layer 3 and all of layer 4 occurs during or after emplacement (Moody 1978). Detailed models of hydrothermal circulation systems at mid-oceanic ridges are in the process of being evaluated by geophysicists, geologists and petrologists. More information is needed before an internally consistent model for both fast- and slow-spreading ridges can be developed. The extent of metamorphism at the ridge crest, and as a function of age of the oceanic crust, is still unknown.

The major consensus (Tables 7, 8) from the studies of both oceanic and ophiolitic rocks is that metamorphism of the ophiolitic sequence can occur within layers 1 and 2 but ceases at various different levels in layer 3 (gabbro, Fig. 1). The cessation of metamorphism in layer 3 is directly related to the limit of fluid penetration into the oceanic crust, which is unknown except for oceanic ridge-crests and transform faults. As further information about fluid penetration away from crests and faults is unlikely to emerge because of drilling limitations, quantitative modeling of the cooling, contracting crust will be required to shed light on the major oceanic-fracture systems away from the ridge crests.

Evidence gathered to date strongly suggests that serpentinization of the ultramafic layer occurs after and not before emplacement of the ophiolite. Certainly, the physical properties of peridotite versus serpentinite (Moody 1976a) are sufficiently different during deformation to suggest this (i.e., the tectonic emplacement of a serpentinite would take place with the production of a mélangé and complete disruption of the serpentine layer). Mélange production on a large scale has been described for subduction-zone ophiolites (Coleman 1977) but is not a widespread feature of other ophiolites. Therefore, tectonic-emplacement models point to the ophiolite metamorphism of the lower layers 3 and 4 occurring during or after emplacement, not before. The tectonic emplacement of some ophiolites (e.g., at Bay of Islands and Semail) is marked at their bases by metamorphic aureoles of medium- to high-grade rocks including mylonites.

**TABLE 8. SCHEME OF METAMORPHISM FOR OPHIOLITES**

<table>
<thead>
<tr>
<th>Ophiolite Rocks</th>
<th>Oceanic Metamorphism</th>
<th>Obduction</th>
<th>Regional**</th>
</tr>
</thead>
<tbody>
<tr>
<td>Marine Sediment</td>
<td>Zeolite</td>
<td>No metamorphism, or metamorphism in oceanic environment before tectonic transport</td>
<td>Zeolite</td>
</tr>
<tr>
<td>Basalt</td>
<td>Zeolite-greenschist (spilitic)</td>
<td></td>
<td>Zeolite-greenschist</td>
</tr>
<tr>
<td>Diabase</td>
<td>Greenschist</td>
<td>Greenschist</td>
<td></td>
</tr>
<tr>
<td>Gabbro</td>
<td>Greenschist-amphibolite</td>
<td>Greenschist-amphibolite</td>
<td></td>
</tr>
<tr>
<td>Peridotite</td>
<td>Unmetamorphosed, except local serpentinization along fracture zones + rodingites associated with serpentinites</td>
<td>a contact metamorphic aureole mélangé, mylonites, serpentinization ± rodingites</td>
<td>Serpentinitization, possible prograde metamorphism of serpentine</td>
</tr>
<tr>
<td>Conditions of Metamorphism</td>
<td>No deformation</td>
<td>Deformation at base of ophiolite slab with development of foliation there</td>
<td>Development of penetrative foliation</td>
</tr>
</tbody>
</table>

* modified from Coleman (1977)

** multiple stages of metamorphism possible, superposition of oceanic with continental regional metamorphism, including simple burial metamorphism without development of penetrative deformation.
Element mobilities during metamorphism are significant. Care must be exercised in reconstructing parent igneous rocks from their supposed metamorphic progeny. Several lines of geochemical and petrographic evidence must be assessed before definite conclusions can be reached about the parent igneous rock and its origin.

The metamorphic history of an ophiolite is an integral part of its geological history. Critical to an understanding of this metamorphism is the ability to decipher the geological events undergone by the ophiolite after emplacement on the continent, its emplacement history and then its possible oceanic provenance. It is safe to say that the unique importance of ophiolites in the understanding of global tectonic processes, coupled with the difficulty in unraveling their histories, will assure them a continuing central position in plate-tectonic studies in the foreseeable future.

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THE CANADIAN MINERALOGIST


