

## Zeolite facies metamorphism of basaltic rocks from the East Taiwan Ophiolite

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### Abstract

The basaltic rocks of the fragmented East Taiwan Ophiolite have been subjected to "ocean-floor" zeolite facies metamorphism. Depending on the bulk composition and mode of occurrence, various mineral assemblages occur: thomsonite + analcime + chabazite; pumpellyite + chlorite + laumontite in veins of the pillow cores; and pumpellyite + chlorite + K-feldspar; pumpellyite + laumontite + thomsonite; and prehnite (about 5 weight percent  $\text{Fe}_2\text{O}_3$ ) + hematite in veins of the pillow matrices. Plagioclase phenocrysts were replaced by albite + pumpellyite + Ca-zeolites, pumpellyite + K-feldspar, analcime + chabazite + thomsonite, or by K-feldspar alone, the olivine phenocrysts by brown chlorite + serpentine  $\pm$  pumpellyite, but the pyroxenes are well preserved. Except for local palagonitization along fractures and rims, the pillowed glassy rims are perfectly fresh. The variations in mineral association among pillow cores, rims, and matrices are believed to have resulted from local variations in compositions of the circulating fluid, in the extent of recrystallization, and in temperature. The occurrence of potash feldspar suggests that the basaltic rocks have been subjected to local metasomatic exchange with a hydrothermal solution under zeolite facies conditions.

Pumpellyites characteristically contain up to 25 weight percent total Fe as FeO, higher than most reported pumpellyite (except juldite). Calculation of their structural formulas indicates that they may contain iron dominantly in the ferric state. Substitution of  $\text{Fe}^{3+}$  for Al in this phase evidently enlarges the pumpellyite *P-T* stability field relative to the zeolite facies assemblages under oxidizing conditions. Pumpellyite probably crystallized directly from palagonitized basaltic glass. The basaltic rocks of the East Taiwan Ophiolite were hydrothermally metamorphosed at  $T = 150\text{--}250^\circ\text{C}$  and depths of 0.6 to 1.6 km, probably at some distance away from the ridge.

### Introduction

Recent petrologic studies of rocks dredged from ocean floors and ridges have thrown new light on the metamorphic processes taking place near divergent lithospheric plate boundaries. The major parts of the oceanic crust have been subjected to mid-ocean ridge metamorphism immediately after their creation at a spreading center. Varieties of this recrystallization include zeolite, greenschist, and/or actinolite-calcic plagioclase and amphibolite facies metamorphism (e.g., Miyashiro *et al.*, 1971; Bonatti *et al.*, 1975; Coleman, 1977). If on-land ophiolites are fragments of oceanic crust created at mid-oceanic ridges and later transported to convergent lithospheric plate boundaries (e.g., Coleman, 1977), the effect of ridge metamorphism may still remain to some extent even where later metamorphism is superposed on such rocks.

The lithologic and chemical similarities of the East Taiwan Ophiolite to present-day oceanic crust and upper mantle material suggest that all lithologies of this ophiolite originated in an oceanic basin (for details, see Liou *et al.*, 1977). However, the ophiolite is not simply a fragment of normal oceanic crust of the sort formed at ridge crests, because the petrology of the interbedded red shales suggests that they were deposited at significantly greater depth—below the carbonate compensation depth (Suppe *et al.*, 1977). Detailed field and petrographic features suggest that the observed stratigraphy of the ophiolite formed near the base of a deepwater fault scarp in preexisting ocean crust—possibly a transform fault—that has given rise to the plutonic mafic-ultramafic breccia deposits and has been the site of later tholeiitic extrusive volcanism and metamorphism. Therefore, study of mineral parageneses of basaltic rocks of this

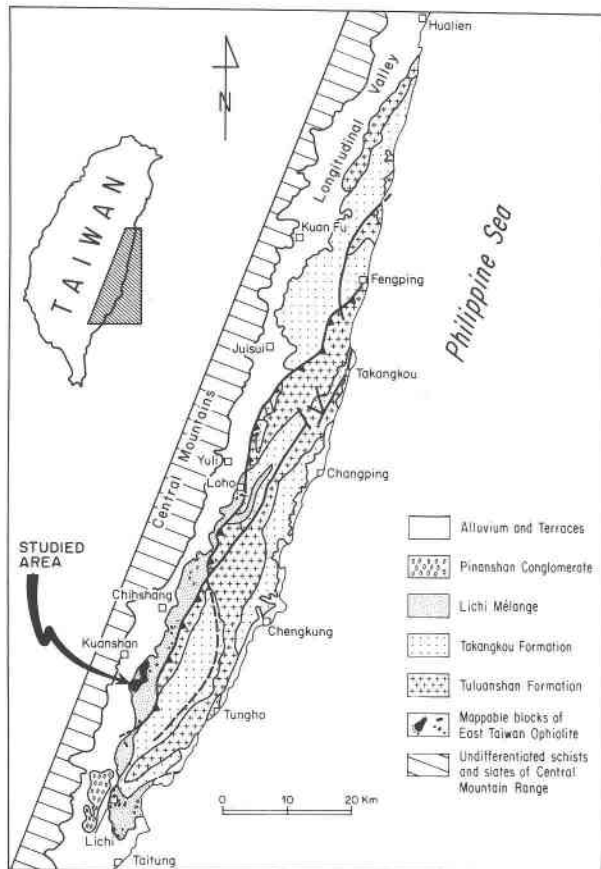


Fig. 1. Generalized geologic map of the Coastal Range of Taiwan (modified from Liou *et al.*, 1977). The investigated rocks of the East Taiwan Ophiolite are from the area shown by an arrow.

ophiolite is important to an understanding of metamorphic and tectonic processes prevailing in the vicinity of a transform fault in an oceanic basin.

The characteristic features of the zeolite facies metamorphism of extrusive basaltic rocks from the ophiolite are the absence of penetrative deformation, the occurrence of Fe-rich pumpellyite, and selective alterations. The primary igneous structures and textures are well preserved and most rocks are only partly altered, and recrystallization is incomplete. Such characteristic features are consistent with those described by Miyashiro *et al.* (1971) and by many other investigators (*e.g.*, de Wit and Stern, 1976; Coleman, 1977) for "ocean-floor" metamorphism.

#### Geologic setting and field relations

The fragmented East Taiwan Ophiolite occurs as allochthonous slabs and blocks up to one kilometer across in the Pleistocene Lichi muddy mélangé, which is exposed on the western slope of the southern Coastal Range as a long and narrow belt extending

from Locho to Taitung (see Fig. 1). The ophiolite blocks are sporadically distributed in the mélangé, but part of the primary ophiolite sequence is well preserved within individual large slabs exposed in the Kuanshan District. The field relations and petrology of the ophiolite in this district have been described in detail by Liou *et al.* (1977); a summary of field relations is given below.

The schematic stratigraphic relations are shown in Figure 2. The composite sequence passing upward is as follows: (1) angular polymict plutonic breccias of gabbro, peridotite, and related rocks, with interbedded red shale; (2) a thin (about 20 cm) layer of red shale; and (3) an extrusive tholeiitic sequence consisting largely of still-glassy pillow lavas with interbedded thin red shales. Plagiogranitic and diabasic partly rodingitized mafic dikes cut mafic-ultramafic breccias. Some of these dikes are thought to represent feeders for the extrusive basalts. The brecciated

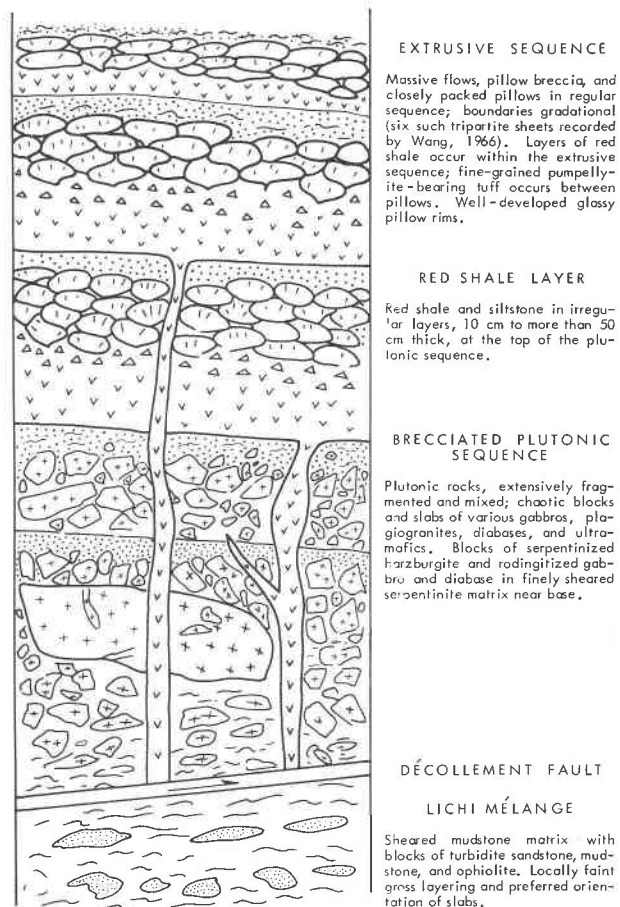


Fig. 2. Schematic interpretive geologic column of the East Taiwan Ophiolite as exposed in the Kuanshan area (after Liou *et al.*, 1977).

plutonic sequence grades upward from (a) predominantly serpentinized harzburgite slabs in the serpentinite matrix, to (b) angular polymict breccia of various types of gabbro, harzburgite, diabase, and plagiogranite, locally with minor red shale cement. Pelagic red shales occur as irregular layers of 10 cm or less to more than 50 cm thick at the top of the brecciated plutonic sequence and within the volcanic rocks. The contact between mafic-ultramafic breccia and overlying red shale layers is depositional, with the shale filling in around the blocks at the top of the breccia.

The extrusive sequence progresses upward from minor massive flows to brecciated and dominant close-packed pillows with several layers of interpillow red shale. Pillow structures are well developed with minor interpillow pyroclastic matrix. Individual pillows consist of three main parts: (1) a glassy outer rim 1 to 20 cm thick containing scattered microphenocrysts of spinel, olivine, and plagioclase; (2) a spherulitic intermediate zone characterized by skeletal growth of spherulites in a glassy matrix; and (3) a compact crystalline core of diabase with intergranular texture. The outer glassy rim is dark in color and contains many black vitreous rounded nodules and dull interstitial palagonitized matrices, which contrast with the light grey massive interiors of the pillow. The chilled margins of black glass are well developed around most pillows, and in some cases these glassy selvages are as much as 50 cm thick. The conspicuous differences in crystallinity, color, and structure accentuate the classical pillow structure. Some pillows are closely packed and contact each other directly, whereas others may have interpillow matrices. Pumpellyite-bearing green matrices and hematitic red shale are locally abundant.

Concentric or radiating fractures in pillows are commonly filled by late-stage calcite, zeolites, and prehnite (Juan *et al.*, 1964). At the outer margin the glass, which exhibits a round conchoidal fracture, is traversed by a polygonal network of shrinkage cracks. No vesicular structure has been found in any glassy rims of these pillowed basalts. However, small vesicles in the pillowed cores are usually partly or wholly infilled to form amygdules. In order of abundance, amygdaloidal minerals include chlorite, pumpellyite, natrolite, stilbite, thomsonite, and analcime. Many of these secondary minerals are also found as veinlets and irregular masses within the body of the pillow. The lack of visible vesicular fragments and the minor content of fine pyroclastic tuffs in the extrusive sequence of the East Taiwan Ophiolite

Table 1. Bulk rock compositions and mineral contents of some basaltic rocks of the East Taiwan Ophiolite

Sample No.	T-99C (pillow core)	T-231	T-112B	T-112C <sup>*</sup> (massive basalt)	T-112Z
SiO <sub>2</sub>	47.36	49.05	42.19	40.42	44.86
Al <sub>2</sub> O <sub>3</sub>	13.96	12.58	13.00	8.93	15.25
TiO <sub>2</sub>	0.89	0.93	0.93	0.67	0.99
Fe <sub>2</sub> O <sub>3</sub> <sup>*</sup>	10.24	10.11	14.39	14.76	12.63
MnO	0.20	0.20	0.16	0.21	0.14
MgO	9.23	10.10	10.94	22.19	7.88
CaO	12.89	10.11	9.80	7.11	11.06
Na <sub>2</sub> O	1.66	2.89	1.43	1.02	1.83
K <sub>2</sub> O	0.02	0.38	0.10	0.07	0.12
Ignition loss	3.39	3.27	6.43	4.14	4.47
Total	99.84	99.62	99.37	99.52	99.23
olivine				10	
plagioclase	55	30	60	55	45
clinopyroxene	20		15	15	20
opaques	4	7	5	3	4
spinel	1				
chlorite	5	5		5	10
celadonite	2	3	tr	2	
carbonate				1	
pumpellyite		5			
zeolites		tr**			
serpentine				10	
devitrified glass	13	47	20		20

\*Total Fe as Fe<sub>2</sub>O<sub>3</sub> \*\*thomsonite + analcime + chabazite.

ite suggest a non-explosive deep-water (abyssal) extrusion (Moore, 1965; Jones, 1969).

Representative chemical compositions of pillow core, glassy rim, and massive flow are shown in Table 1. Characteristically, the basaltic rocks are extremely low in K<sub>2</sub>O and TiO<sub>2</sub> content and are high in CaO and MgO/FeO\* ratios. They are mineralogically and chemically similar to oceanic basalts and appear to be typical abyssal tholeiite (for detail, see Liou, 1974; Liou *et al.*, 1977).

#### Petrographic descriptions of metamorphosed basaltic rocks

In contrast to the virtually unaltered glassy basalt in the pillow rims (Liou, 1974), both pillow cores and pillow matrices of the East Taiwan Ophiolite contain abundant secondary minerals including albite, pumpellyite, K-feldspar, chlorite, smectite, celadonite, analcime, and various calcium zeolites. The presence of such secondary minerals strongly suggests low-grade metamorphism.

With the exception of augite, all primary phases in the pillow cores and massive basalts are partly replaced by metamorphic minerals. Many plagioclase phenocrysts have been albitized and zeolitized. Along with albite, laumontite, thomsonite, and chlorite are the most abundant secondary minerals. Analcime, pumpellyite, potash feldspar, and other Ca-zeolites were identified. Olivine phenocrysts were relatively rare; where they occurred, they are totally replaced

Table 2. Estimated modes for pillow matrices from the East Taiwan Ophiolite

Mineral	T-97C <sup>1</sup>	T-99D	T-231D	T-232A	T-232B
Clinopyroxene clasts			1		
Plagioclase clasts			15		
Hornblende clasts			10		
White mica		5			
Chlorite	15		10		5
Pumpellyite	65	5		20	
Prehnite					28
Adularia	20	20		5	tr
Albite		tr		tr	tr
Laumontite		tr		5	
Thomsonite				5	
Analcime				5	
Carbonate		20		4	5
Opaque		3	2	1	2
Sphene			2	54	
Matrix <sup>*</sup> and devitrified glass		45	60	55	60

<sup>\*</sup>Matrix consists mainly of smectite with minor chlorite, celadonite and opaques.

by serpentine or talc, brown chlorite, and a trace of pumpellyite.

Among the different rock types in the East Taiwan Ophiolite, the most characteristic low-grade mineral assemblages are best developed in the pillow matrices. The initial pyroclastic fine-grained aggregates, presumably soaked with abundant H<sub>2</sub>O, have been pervasively recrystallized to fine-grained mixtures of Fe-pumpellyite (or Fe-rich prehnite) + smectite + chlorite ± laumontite and/or other zeolites. The identified minerals in the pillow matrices are listed in Table 2.

Most pillow matrices are fine-grained, homogeneous, and yellowish-green with disseminated fine-grained iron oxide dust. Palagonitized glassy shards were found only in texturally inhomogeneous samples. Palagonite forms as long irregular belts with color bands reflecting different degrees of palagonitization and recrystallization. The fine-grained palagonitized mesostasis consists of a mixture of smectite + pumpellyite as identified by X-ray diffraction and broad-beam microprobe analyses. Some basaltic fragments in matrices exhibit thin layers of dark to light green, fine-grained Fe-pumpellyite aggregates around fragmental rims. Apparently, pure pumpellyite was recrystallized from the palagonitized glassy rims. Some euhedral plagioclase phenocrysts are entirely replaced by adularia, whereas aggregates of phenocrystic olivine are replaced by pumpellyite and chlorite + white mica. The groundmass is recrystallized and contains minute brown sheaves of pumpellyite set in a field of finely crystalline smectite + chlorite dusted with sphene granules. Along fine

cracks, well-crystallized fine-grained and distinctly pleochroic Fe-pumpellyite occurs together with euhedral adularia. Textural evidence suggests that these two phases grew simultaneously.

### Mineral chemistry<sup>1</sup>

#### *Albite and K-feldspar*

Extensive analyses of altered plagioclases were not undertaken. The secondary plagioclases in the pillow cores, massive basalts, and pillow breccias are characteristically albitic in composition (Ab 91–97). It should be noted that the plagioclases in pillow cores and matrix have been nearly completely albitized, whereas plagioclases in pillow glassy rims were unaltered and have compositions of An 70–72 (Liou, 1974).

Some euhedral primary plagioclase phenocrysts in pillow matrices, in some pillow cores, and in diabase have been totally replaced by adularia. Adularia also occurs together with pumpellyite as a major vein mineral. K-feldspars were analyzed in five samples. Their compositions are extremely uniform and high in Or content with very low amounts of An (0.1–0.7 mole percent) and Ab (0.7–1.8 mole percent) components and very low in minor oxides—Fe<sub>2</sub>O<sub>3</sub> (≤0.33 weight percent), TiO<sub>2</sub> (≤0.27 weight percent), and MgO (≤0.02 weight percent).

#### *Pumpellyite*

Pumpellyite occurs as a chief vein constituent in some pillow matrices and in altered pillow cores. It is typically green to deep green in color and displays characteristic anomalous blue interference colors. Strong pleochroism suggests that the pumpellyite is iron-rich. Because of this pleochroism, high relief, and distinct habit and interference colors, pumpellyite is easily differentiated from coexisting chlorite. However, because it usually occurs as clusters of very fine needles among veins, in amygdules, or along the edges of palagonitized glass shards, it is extremely difficult to obtain reliable optical data.

Although individual crystals are exceedingly small, pumpellyite usually occurs as pure monomineralic aggregates within the veins. The electron beam size for the analysis of pumpellyite is about 2 microns in diameter, which is larger than the grain size; therefore, the analyses are of crystal aggregates. For each specimen, at least 5 spots (relatively large single crys-

<sup>1</sup> The analytical methods and tabulated mineral compositions are in Liou *et al.* (1977).

tals or crystal aggregates) were selected for analysis. The results are consistently homogeneous and the compositional variations from one spot to another in each sample are small. Individual spot analyses tend to show reciprocal variation in Al and Fe; hence, most Fe is probably  $\text{Fe}^{3+}$ .

Calculation of structural formulas for pumpellyite from chemical data has been difficult, partly because of unreliable published chemical data on the fine-grained mineral separates, and partly because of the presumed presence of substantial (OH) substitution of oxygen accompanying the introduction of trivalent ions into the octahedral position. Microprobe results are further complicated by uncertainties concerning the oxidation state of iron. Different chemical formulas have been used in previous investigations (see Allmann and Donnay, 1971; Passaglia and Gottardi, 1973; Moore, 1971; Coombs *et al.*, 1976). Assuming the cation sites of pumpellyite are totally filled and without excess, Coombs *et al.* suggested the formula  $\text{W}_4^{\text{VI}}\text{X}_2^{\text{VI}}\text{Y}_4^{\text{VI}}\text{Z}_6^{\text{VI}}\text{O}_{20+x}(\text{OH})_{8-x}$ , where  $\text{W} = \text{Ca}, \text{Mn}$ ;  $\text{X} = (\text{Mg}, \text{Fe}^{2+}, \text{Mn})_{2-x}(\text{Fe}^{3+}, \text{Al})_x$ ;  $\text{Y} = \text{Fe}^{3+}, \text{Al}$ ; and  $\text{Z} = \text{Si}, \text{Al}$ . This formula was used in the present study. On the basis of 16 total cations, atomic proportions of analyzed pumpellyites were calculated. If octahedral Y sites consist only of  $\text{Fe}^{3+}$  and Al, then  $\text{Fe}^{3+}$  occupancy in Y sites should equal  $4 - \text{AlVI}$ .

For these pumpellyites, Si (5.954–6.059) is consistently close to the ideal value of 6,  $\text{Fe}^{3+}$  in the octahedral Y site is characteristically high (0.121–2.076), the occupants of the octahedral X site are greater than 2, and Ca in the W site (3.834–3.996) is less than the ideal 4. The X+W values nearly equal 6, suggesting the possible substitution of  $\text{Fe}^{2+}$  for some Ca. The total absence of Al in the octahedral X site of Taiwan pumpellyite is most characteristic, as all reported pumpellyite analyses contain very high  $\text{Al}_2\text{O}_3$  except for Russian Fe-rich pumpellyite (Zolotukhin *et al.*, 1965) and julgoldite (Coombs *et al.*, 1976). According to Passaglia and Gottardi (1973), the value of  $(\text{Mg} + \text{Fe}^{2+} + \text{Mn})$  in the X site for reliable pumpellyite analyses ranges from 0.71 to 1.55 out of 2. In other words,  $(\text{Fe}^{3+} + \text{Al})$  in the X site should range between 1.29 and 0.45 per formula unit. If this is the case, the Fe occupying both octahedral Y and X positions in the pumpellyite of the East Taiwan Ophiolite may be dominantly in the ferric state.

Compositions of pumpellyites from Taiwan and the compositional fields of pumpellyite from various known geologic occurrences compiled by Coombs *et al.* (1976) are plotted in an Al-Fe\*-Mg diagram as Figure 3. It is apparent from such comparison that

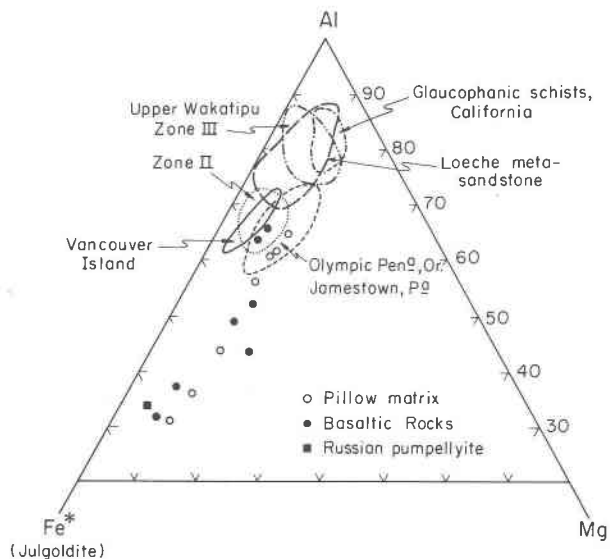


Fig. 3. Compositional variations of analyzed pumpellyites from the East Taiwan Ophiolite in Al-Fe\*-Mg diagram of Coombs (1976). Compositional fields of pumpellyites are from Coombs *et al.* (1976).

pumpellyites from the East Taiwan Ophiolite are high in total  $\text{FeO}^*$  and low in  $\text{Al}_2\text{O}_3$ . They contain up to 25 weight percent  $\text{FeO}^*$  (range 9.65–25.38 weight percent), higher than most reported pumpellyites (except julgoldite). Therefore, substantial occupancies of  $\text{Fe}^{3+}$  in the Y site must have occurred. The high iron content is also reflected by the deep blue-green pleochroic colors.

Similar Fe-rich pumpellyite associated with iron-rich prehnite and babingtonite in a metasomatized andesine diabase from Russia was described by Zolotukhin *et al.* (1965). As shown in Figure 3, this phase and pumpellyites from Taiwan are intermediate members of the pumpellyite-julgoldite solid solution series. A continuous substitution of Al for  $\text{Fe}^{3+}$  in the Y octahedral site is apparent from the analytical data. Figure 3 also shows that the Mg contents in most pumpellyites from Taiwan are nearly constant and are higher than in Fe-rich pumpellyite from Russia and julgoldite. The Taiwanese pumpellyites are Fe-pumpellyite according to the classification scheme of Coombs *et al.* (1976) or ferropumpellyite according to that of Moore (1971).

#### Zeolites

Numerous vein zeolites in the glassy basalts have been identified optically and by X-ray (*e.g.*, Juan *et al.*, 1964; Juan and Lo, 1971). These zeolites include analcime, natrolite, thomsonite, stilbite, heulandite, and laumontite. Systematic identification and analy-

ses of the zeolites were not undertaken. Calcium zeolites including natrolite, heulandite, stilbite, laumontite, chabazite, and thomsonite were optically identified as vein and amygdaloidal minerals and as fine-grained discrete patches replacing primary plagioclases. Vein zeolites are relatively coarse-grained, and their optical identification is relatively simple. However, most calcium zeolites enclosed in altered plagioclase are fine-grained aggregates, and differentiation among the different species is extremely difficult.

Except for the analcime with 2.42 weight percent CaO replacing plagioclase in T-231, the other analyzed analcimes are very uniform in composition. The weight percent of major oxides averages about 54 percent SiO<sub>2</sub>, 22.5 percent Al<sub>2</sub>O<sub>3</sub>, and 12.5 percent Na<sub>2</sub>O, with other analyzed oxides present in only minor amounts. The analcime replacing plagioclase in T-231 contains relatively high Fe<sub>2</sub>O<sub>3</sub> ( $\leq 0.31$  weight percent), K<sub>2</sub>O ( $\leq 0.43$  weight percent), and wairakite component (CaAl<sub>2</sub>Si<sub>4</sub>O<sub>12</sub>·2H<sub>2</sub>O), but analcimes from the ophiolite usually have compositions very close to the ideal stoichiometry Na<sub>2</sub>O·Al<sub>2</sub>O<sub>3</sub>·4SiO<sub>2</sub>·2H<sub>2</sub>O. However, anhydrous totals of 89.26–89.71 weight percent suggest H<sub>2</sub>O contents may be higher than the ideal value of about 8.5 weight percent. The Si/Al ratio ranges from 2.04 to 2.14, which is comparable to the range of Group C analcimes of Coombs and Whetten (1967). Analcimes from the East Taiwan Ophiolite occur mainly as fissure-fillings and replace plagioclase in the low-grade metamorphosed basaltic rocks. No quartz and dolomite were found associated with analcime. Yet the relatively low Si/Al ratio and relatively high assumed H<sub>2</sub>O content may stem from a variety of factors such as low-temperature formation for analcime, salinity, the nature of the initial solution and possibly the mineral association and the bulk rock composition (Coombs and Whetten, 1967; Wilkinson, 1968; Liou, 1971d).

Laumontites are relatively constant in composition, closely approaching the ideal formula CaAl<sub>2</sub>Si<sub>4</sub>O<sub>12</sub>·4H<sub>2</sub>O; they contain only minor amounts of K<sub>2</sub>O (0.41–0.56 weight percent), Na<sub>2</sub>O (0.09–0.16 weight percent), and Fe<sub>2</sub>O<sub>3</sub> (1.61 weight percent). The chemical compositions of three analyzed laumontites are compatible with the analysis made by Juan and Lo (1971) on laumontite from this ophiolite. Chabazite is practically pure Ca-zeolite. Substitution of Ca by Na [Ca/(Ca+Na) ratio ranges from 0.68–0.77] in thomsonite varies with the mode of occurrence and rock type. In thomsonite, K<sub>2</sub>O, Fe<sub>2</sub>O<sub>3</sub>, and MgO are present in negligible amounts.

One K-rich phase was found during microprobe analysis. This unknown K-rich phase occurs as massive, fine-grained crystals replacing a coarse-grained plagioclase of An 71 in a porphyritic basalt (TPY 6568). It is optically clear and lacks visible cleavage; the estimated birefringence is about 0.008 and the refractive index about 1.52. Chemically, it appears to be high in TiO<sub>2</sub> (3.16 weight percent), K<sub>2</sub>O (10.41 weight percent), Al<sub>2</sub>O<sub>3</sub> (20.88 weight percent), and SiO<sub>2</sub> (50.92 weight percent) and is very low in CaO (0.05 weight percent) and Na<sub>2</sub>O (0.45 weight percent). On the basis of occurrence, optical properties, and chemical composition (particularly in the high apparent H<sub>2</sub>O content of 14 percent), it may be a K-rich zeolite.

#### *Prehnite*

Prehnite is one of the most common vein minerals in the brecciated plutonic sequence of the ophiolite. However, prehnite was found in only one sample from the pillow matrix, and it is stably associated with hematite. The prehnite contains high Fe<sub>2</sub>O<sub>3</sub> (4.8–5.8 weight percent) and may have crystallized under high *f*O<sub>2</sub> conditions. This hypothesis is consistent with the occurrence of pumpellyite–laumontite-bearing assemblages in the associated pillow matrix, and with the conclusions based on previous studies that prehnite in low-grade metabasaltic rocks commonly contains appreciable Fe<sup>3+</sup> (e.g., Surdam, 1969; Kuniyoshi and Liou, 1976).

#### *Chlorite*

Chlorite occurs as tiny flaky inclusions in albitized plagioclase, as massive fine-grained aggregates in veins and amygdules, and as replacements of devitrified glass of the porphyritic basalts. In most samples chlorite is closely intergrown with zeolites or pumpellyites, and is therefore not suitable for microprobe analysis.

Chlorites from four basalts were selected for microprobe analysis. The chlorite composition is strictly a function of the chemistry of the precursor assemblage; therefore, the composition of chlorite crystallized from primary clinopyroxene in a basaltic rock may depart significantly from that of chlorite crystallized directly from devitrified glass in the same sample. Thus, a considerable range of chlorite composition was encountered in each sample analyzed.

Most analyzed chlorites are extremely low in the minor oxides TiO<sub>2</sub>, MnO, and Cr<sub>2</sub>O<sub>3</sub>, and the alkali contents are negligible. The apparent Ca contents of chlorites vary extensively. Some samples contain as

high as 1.4 weight percent CaO, which is probably due to the minute inclusion of pumpellyite or zeolites. The analyzed chlorites are higher in  $\text{SiO}_2$ , in  $\text{SiO}_2/(\text{SiO}_2+\text{Al}_2\text{O}_3)$  ratio, and in  $\text{Al}^{\text{VI}}$  compared to chlorites from the plutonic rocks of the ophiolite (Liou *et al.*, 1977). The high  $\text{SiO}_2/(\text{SiO}_2+\text{Al}_2\text{O}_3)$  ratio may be related to the lower metamorphic grade developed in the extrusive rocks (zeolite facies) compared to that of the plutonic sequence (greenschist), as this ratio has been suggested to decrease with increasing metamorphic grade (*e.g.*, Ernst *et al.*, 1970). The ratio  $\text{Fe}^{2+}/(\text{Fe}^{2+}+\text{Mg})$  varies considerably within the extrusive chlorites, and these chlorites belong to the talc-chlorite, penninite, and diabantite series of Hey (1954).

### Mineral paragenesis

Altered basaltic rocks in both pillow cores and massive flows commonly contain the following mineral associations in veins: thomsonite + celadonite; thomsonite + analcime + chabazite; analcime + chlorite + smectite; and pumpellyite + chlorite. Plagioclase phenocrysts have been replaced by albite  $\pm$  pumpellyite, pumpellyite + adularia, analcime + chabazite, or by potash feldspar alone, and the olivine phenocrysts by brown chlorite + serpentine  $\pm$  pumpellyite. Such replacement strongly suggests substantial transport of K and Fe from solution into plagioclase phenocrysts and Al and Ca into olivine.

For the pillow matrices, the mineral associations are highly variable, depending on the clast-matrix ratio, degree of palagonitization of basaltic clasts, extent of recrystallization and, most importantly, inferred chemistry of the hydrothermal solution during the alteration. The most common associations observed in veins and occasionally in amygdules are:

pumpellyite + chlorite  $\pm$  K-feldspar  
 pumpellyite + laumontite  $\pm$  thomsonite  
 thomsonite + analcime  $\pm$  laumontite  $\pm$  K-feldspar  
 prehnite + hematite + smectite + K-feldspar  
 pumpellyite + laumontite + chlorite  
 pumpellyite + chlorite + calcite

Rare associations include smectite and sphene. The prehnite + pumpellyite association was not found, whereas prehnite + hematite occur as a vein assemblage, and smectite + opaque and potash feldspar occur sporadically as replacements for groundmass and plagioclase respectively in small basaltic clasts. Quartz was not identified as a vein mineral or as one of the replacement assemblages for primary igneous minerals. Calcite occurs as central fracture fillings

and as small veinlets cross-cutting the zeolite, prehnite, and pumpellyite veins, hence it must have formed later than these minerals.

Microscopic examination of textural relations in several sections has identified the following alterations and crystallization sequence of secondary minerals in mafic extrusive rocks of the ophiolite:

- (a) palagonitization of basaltic glass and devitrification of glassy mesostasis;
- (b) recrystallization of palagonitized glass to pumpellyite, and alteration of plagioclase phenocrysts to albite  $\pm$  pumpellyite (or calcium zeolites), or K-feldspar  $\pm$  pumpellyite, and olivine phenocrysts to chlorite + serpentine  $\pm$  pumpellyite;
- (c) crystallization of chlorite + pumpellyite  $\pm$  laumontite along the inner linings of the fissures or amygdules, which are succeeded in turn by laumontite + analcime + thomsonite, analcime + chabazite + thomsonite, and finally by calcite.

All listed mineral associations may be depicted graphically in the system  $\text{CaAl}_2\text{Si}_2\text{O}_8\text{-Na}_2\text{Al}_2\text{Si}_2\text{O}_8\text{-H}_2\text{O}$  and in an ACF diagram, as shown in Figure 4. Three-phase assemblages such as thomsonite + laumontite + analcime, thomsonite + analcime + chabazite, pumpellyite + laumontite + chlorite were observed as shown in the diagrams. The zeolite paragenesis of Figure 4A is consistent with the zeolite facies metabasalts from the ocean floor described by Miyashiro *et al.* (1971) and metabasalts from on-land ophiolites (*e.g.*, Spooner and Fyfe, 1973; Gass and Smewing, 1974); the zeolite facies assemblages have been considered as close to chemical equilibrium. Phillipsite and clinoptilolite, commonly reported in deep-sea sediments, do not appear to occur in the zeolite facies metabasalts from the East Taiwan Ophiolite and dredged oceanic basalts. Evidently these zeolite facies rocks have been recrystallized at higher temperatures than those accompanying the crystallization of phillipsite and clinoptilolite in deep-sea sediments.

Most characteristic of the extrusive sequence is the occurrence of Fe-rich pumpellyite in veins and amygdules and as a replacement of both olivine and plagioclase. The significance of such occurrences is discussed in detail in the next section. On the ACF diagram of Figure 4B, compositional fields for Fe-rich pumpellyite are plotted according to the present chemical analyses; the compositional field of average basalt after Coombs' (1963) data, and compositions of the altered basaltic rocks listed in Table 1 are also



## MINERAL ASSEMBLAGES OF THE EXTRUSIVE ROCKS

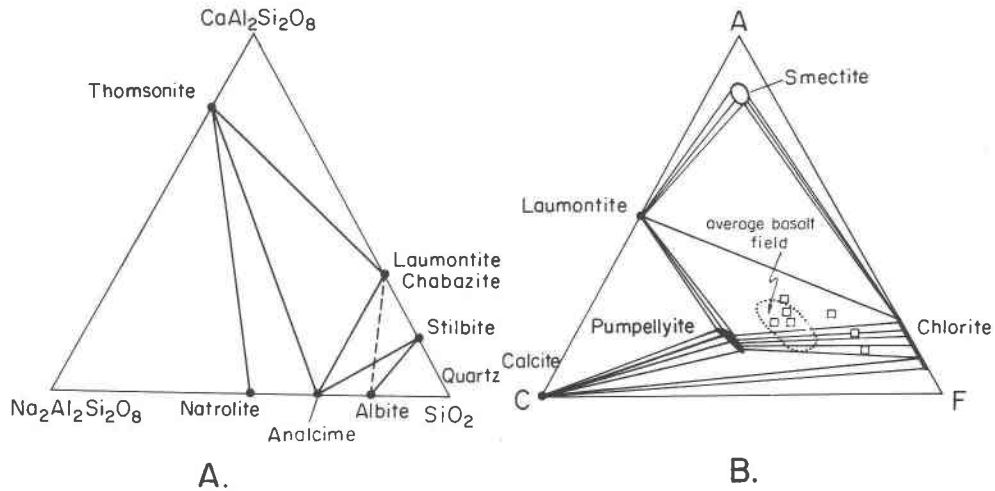


Fig. 4. Parageneses of zeolite facies assemblages of the basaltic rocks, East Taiwan Ophiolite. Diagram A illustrates zeolite assemblages on an anhydrous projection  $\text{CaAl}_2\text{Si}_2\text{O}_8$ - $\text{Na}_2\text{Al}_2\text{Si}_2\text{O}_8$ - $\text{SiO}_2$ . The most common zeolite assemblages in the metabasalts are thomsonite + chabazite + analcime and thomsonite + laumontite + analcime. The dashed line connecting laumontite-albite is a higher-temperature join than analcime-stilbite. Diagram B is an ACF diagram illustrating the zeolite facies assemblage for basaltic composition. The dotted circle refers to the basaltic compositions compiled by Smith (1968) after Coombs' data (1963). Compositional fields of pumpellyites are plotted after the data of Liou *et al.* (1977). Pumpellyite + chlorite  $\pm$  laumontite are common in the zeolite facies metabasalts from the East Taiwan Ophiolite.

shown. Apparently, under zeolite facies conditions, Fe-rich pumpellyite + chlorite and Fe-rich pumpellyite + chlorite + laumontite are the most common assemblages for basaltic compositions. The assemblage pumpellyite + chlorite was found to replace the entire basaltic clasts of T-97C', except for initial plagioclase microphenocrysts which have been transformed to potash feldspar. In addition, pumpellyite + chlorite + laumontite (or other Ca-zeolites) commonly occur in veins and as replacements after plagioclase in some rocks. The observed mineral assemblages are apparently compatible with the proposed relations shown in Figure 4B. Retrogressive changes in mineral sequence from pumpellyite + chlorite  $\rightarrow$  pumpellyite + chlorite + laumontite  $\rightarrow$  laumontite + pumpellyite  $\rightarrow$  laumontite + calcite or prehnite + calcite observed in the vein parageneses suggest the fluid composition is enriched in  $\text{CaO}/(\text{Fe,Mg})\text{O}$  ratio and in  $\text{CO}_2/\text{H}_2\text{O}$  ratio with time. Such chemical variation is consistent with metasomatic trends for the Ordovician lavas of Australia proposed by Smith (1968).

The phase relations illustrated in Figure 4 may also apply to the mineral association observed in the interflow red shales from the ophiolite (Suppe *et al.*, 1977). Excluding the detrital minerals (plagioclase, pyroxene, quartz), interpillow shale layers character-

istically contain the zeolite facies mineral assemblage laumontite (or wairakite)  $\pm$  pumpellyite  $\pm$  chlorite + smectite + hematite. No phillipsite and clinoptilolite were found in the red shale. The finding of wairakite in the red shale and one pillow matrix (H-3c) led us to conclude that wairakite may also be present in the basaltic rocks of the ophiolite.

### Discussion

#### *Occurrence of Fe-pumpellyite and its P-T- $f\text{O}_2$ relations with epidote*

Widespread occurrences of pumpellyite in low-grade metamorphic terrains are increasingly being recognized. The composition of pumpellyite may be used to characterize different metamorphic conditions. Fe-pumpellyite with total Fe as  $\text{Fe}_2\text{O}_3$  of about 10 to 23 weight percent characteristically occurs in rocks of the zeolite and prehnite-pumpellyite facies, whereas Al-pumpellyite with total Fe as  $\text{Fe}_2\text{O}_3$  less than 10 weight percent invariably is associated with blueschist and pumpellyite-actinolite facies metamorphism (Seki, 1961; Coombs *et al.*, 1976). Although the correlation of the Fe content of pumpellyite with depth of burial (*i.e.*, pressure) has been suggested from the compilation of available chemical data, the quantitative effect of pressure on pumpelly-

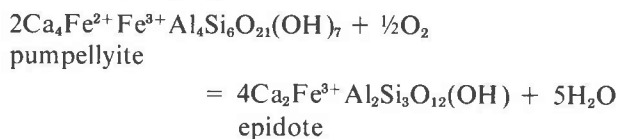


ite composition and on the stability of Fe- and Al-pumpellyite has not been investigated in relation to the effects of bulk composition, temperature, and fluid chemistry.

The ubiquitous high Fe<sup>3+</sup> content in the pumpellyites from the East Taiwan Ophiolite seems to be significant. Evidently zeolite facies metamorphism of the extrusive sequence of the ophiolite took place under oxidizing conditions. The high Fe<sub>2</sub>O<sub>3</sub> contents of the palagonitized basaltic glass may have been responsible for the Fe<sup>3+</sup>-rich nature of the pumpellyites. At least three questions related to the formation of pumpellyite need to be discussed: If zeolite facies metamorphism took place under relatively oxidizing conditions, why did pumpellyite crystallize instead of epidote? What were the precursor minerals for the formation of pumpellyite? What is the minimum temperature required for the crystallization of pumpellyite in the ophiolite?

As discussed in the petrographic sections, pumpellyite occurs along the edges of some glass shards, as a total replacement of palagonitized glass fragments in fine-grained pillow breccia, and as vein minerals in the pillow cores. Neither epidote nor prehnite were found in pumpellyite-bearing rocks. Instead, laumontite and smectite + calcite, adularia + chlorite, and laumontite + thomsonite + analcime are the major coexisting phases. Based on experimentally determined stabilities of laumontite, stilbite, and analcime (Liou, 1971a,c,d; Thompson, 1971), the pumpellyite-laumontite-bearing rocks were apparently recrystallized under zeolite facies conditions, presumably temperatures less than 250°C and *P*<sub>fluid</sub> less than 2 kbar.

For simplification, if we assume that Fe<sup>2+</sup> and Fe<sup>3+</sup> are the only ions in the octahedral X site of the pumpellyite structure and Al in the Y site, then pumpellyite possesses a simple composition such as Ca<sub>4</sub>Fe<sup>3+</sup>Fe<sup>2+</sup>Al<sub>4</sub>Si<sub>6</sub>O<sub>21</sub>(OH)<sub>7</sub>. This cation ratio is identical to the epidote Ps 33 composition Ca<sub>2</sub>Fe<sup>3+</sup>Al<sub>2</sub>Si<sub>3</sub>O<sub>12</sub>(OH) except that epidote possesses all Fe<sup>3+</sup>, whereas idealized pumpellyite contains both Fe<sup>2+</sup> and Fe<sup>3+</sup>. The following simple reaction can be written to delineate pumpellyite-epidote relations:



This reaction possibly will have a positive *P*<sub>fluid</sub>-*T* slope and negative isobaric *f*O<sub>2</sub>-*T* slope as shown in Figure 5.

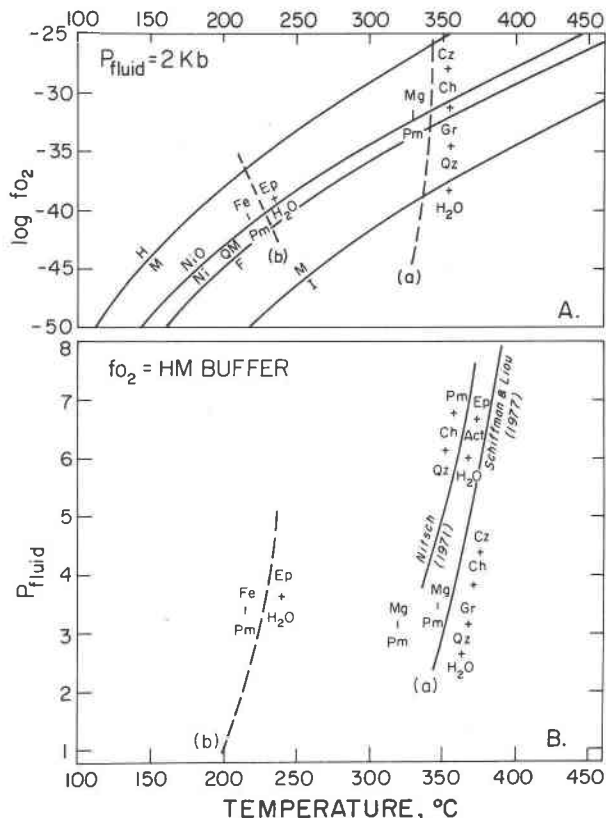


Fig. 5. Isobaric log *f*O<sub>2</sub>-*T* (diagram A) and *P*<sub>fluid</sub>-*T* at *f*O<sub>2</sub> values defined by the hematite-magnetite buffer (diagram B) showing the maximum stability relations of Mg- and Fe-pumpellyites. Solid lines are experimentally determined stability limits for Mg-pumpellyite from Schiffman and Liou (1977), whereas the broken line represents approximated *f*O<sub>2</sub>-*T*-*P*<sub>fluid</sub> limits for the reaction Fe-pumpellyite = epidote + H<sub>2</sub>O (see text for explanation).

The *f*O<sub>2</sub>-*T*-*P*<sub>fluid</sub> stabilities of Fe-pumpellyite have not been experimentally investigated. However, the low-temperature stability limit of epidote has been estimated at about 220° ± 50°C for *P*<sub>total</sub> 1–5 kbar and at 300°C, 7 kbar (Seki, 1972; Tomasson and Kristmannsdottir, 1972). The appearance of epidote in the higher-grade part of the prehnite-pumpellyite facies was suggested to be a result of the reactions (1) laumontite + hematite = epidote + quartz + Al<sub>2</sub>O<sub>3</sub> + H<sub>2</sub>O, and (2) prehnite + hematite = epidote + H<sub>2</sub>O (Seki, 1972). The decomposition of Fe-pumpellyite according to the reaction listed above should also be considered as another limiting reaction for the appearance of epidote. Evidently, as shown in Figure 5, increasing *f*O<sub>2</sub> may considerably decrease the maximum temperature of Fe-pumpellyite stability and may also increase Fe<sup>3+</sup>/(Fe<sup>3+</sup> + Fe<sup>2+</sup>) in the X site and Fe<sup>3+</sup>/(Fe<sup>3+</sup> + Al<sup>3+</sup>) in the Y site for

pumpellyite, as in the case for epidote (Liou, 1973). Therefore, at a temperature of about 250°C and pressure of 0.5 to 2 kbar, Fe-pumpellyite with high Fe<sup>3+</sup> may crystallize directly, instead of epidote, from highly oxidized and palagonitized basaltic glass.

The partition of Fe<sup>3+</sup> between pumpellyite and epidote under *P-T* conditions between the maximum stability of Mg-pumpellyite and the minimum stability of Fe-epidote (curves *a* and *b* of Fig. 5) is highly dependent on temperature. With increasing temperatures, Fe<sup>3+</sup> may preferentially concentrate in the epidote. Therefore, epidote of the prehnite-pumpellyite and blueschist facies rocks contains a higher pistacite component, whereas the coexisting pumpellyite may be enriched in Al (e.g., Karmutsen metabasalt of Kuniyoshi and Liou, 1976; Franciscan metagraywacke and metabasalts, Ernst *et al.*, 1970). However, at temperatures lower than the low-temperature stability curve of epidote (curve *b* of Fig. 5), Fe<sup>3+</sup> will concentrate in the pumpellyite structure; hence zeolite facies pumpellyites may be high in Fe<sup>3+</sup>. This suggestion is consistent with my chemical analyses of pumpellyites and with earlier published results (e.g., Coombs *et al.*, 1976; Kuniyoshi and Liou, 1976; Nakajima *et al.*, 1977). For example, pumpellyite in the zeolite facies metasandstones of Hokonui, New Zealand (Boles and Coombs, 1977) contains about 13.12 weight percent total Fe as FeO\* and was suggested to have recrystallized at temperatures about 190°C. From the analyzed pumpellyite compositions, the mineral assemblages, and the presence of fresh unaltered basaltic glass in the pillow rims of the ophiolite, we conclude that the formation temperature of pumpellyite from palagonite is 150° to 250°C.

#### Formation of pumpellyite

The suggestion "crystallization of pumpellyite directly from palagonitized glass" needs to be docu-

Table 3. Compositional comparison among fresh basaltic glass, palagonitized glass shard, and pumpellyite

Rock	Fresh glass* (Light yellow) (K-11A-1)	---Palagonitized glass---			Pumpellyite Deep green (T-99D')
		Dark brown (T-99D')	Light brown (T-99D')	Brownish green (T-99D')	
SiO <sub>2</sub>	49.86	34.98	34.05	34.22	34.46
TiO <sub>2</sub>	0.87	1.21	1.16	0.37	0.26
Al <sub>2</sub> O <sub>3</sub>	15.64	13.00	12.58	10.32	10.79
FeO <sup>†</sup>	9.45	13.83	13.57	19.72	22.51
MnO	0.18				
MgO	9.13	3.85	3.65	2.44	2.54
CaO	12.46	20.28	19.49	21.65	21.27
K <sub>2</sub> O	0.05	0.06	0.04	0.06	0.07
Na <sub>2</sub> O	2.14	0.05	0.02	0.03	0.03
Anhydrous total	99.78	87.25	84.55	88.51	91.93

\*Composition of fresh glass is from Liou (1974).

†Total Fe as FeO.

mented in detail, as this feature has not been described before. Angular fragments of glass shards have been devitrified and recrystallized during palagonitization. Depending on the extent of recrystallization, differently-colored layers of palagonite were formed and are transected by irregular pumpellyite veinlets. However, most pumpellyites are concentrated in a layer adjacent to and subparallel with the layer of light brown palagonitized glass. The dark brown layer appears to be isotropic, whereas the light brown and brownish-green palagonites have a faint birefringence suggesting incipient crystallization. Textural relations suggest the following paragenetic sequence: (1) palagonitization of glassy basalts (incipient recrystallization of mineraloids + smectite) took place through the interaction of glass with sea water; (2) pumpellyite crystallized from palagonitized glass at the edges of fragments or along the fractures through the reaction of palagonite + solution under zeolite facies metamorphic conditions; and (3) crystallization of laumontite, followed by calcite, occurred along open fractures.

Electron microprobe analyses (Table 3) yield an astonishing similarity in composition between pumpellyite and its probable precursor, palagonite. In particular, the brownish-green devitrified glasses next to the pumpellyite layer possess compositions nearly identical to that of pumpellyite. The green palagonite is homogeneous and extremely fine-grained. However, the characteristic deep green tint and the composition shown in Table 3 suggest that the palagonite may consist essentially of aggregates of minute pumpellyite crystals (M. Vuagnat, personal communication, 1976).

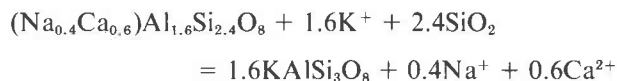
From the above discussion, we conclude that the pumpellyite did not recrystallize directly from unaltered basaltic glass under zeolite facies condition. Instead, when the palagonitization process has changed the original glass toward enrichment in Fe<sub>2</sub>O<sub>3</sub>, Al<sub>2</sub>O<sub>3</sub>, and CaO and depletion in SiO<sub>2</sub>, Na<sub>2</sub>O, and MgO, Fe-rich pumpellyite may recrystallize from such devitrified glass. These pumpellyites, therefore, may contain high Fe<sup>3+</sup> and low Al<sub>2</sub>O<sub>3</sub>. These processes may account for the total replacement of certain glass shards by pumpellyite in some samples.

#### Occurrence of potash feldspar

Occurrences of potash feldspar as an authigenic mineral in low-grade metamorphic rocks have been extensively reported in geothermal fields, in rocks affected by burial metamorphism, and in hydrothermal aureoles of ore deposits. Assemblages such

as quartz + albite + adularia, prehnite + adularia + sphene, and laumontite + adularia + sheet silicates are characteristic. Adularia commonly occurs in hot brines, as in the geothermal fields of Iceland (*e.g.*, Kristmansdottir, 1975). It has been reported only once in oceanic basalts, in metabasalt from the North Atlantic Ocean (Jehl *et al.*, 1976). This may be partly because oceanic basalts characterized by very low  $K_2O$  are not chemically favorable for the formation of potash feldspar, and partly because of the optical similarity between secondary K-feldspar and albite.

Potash feldspar was identified in five rocks from the East Taiwan Ophiolite, and was analyzed by the electron microprobe. It occurs sporadically as pseudomorphs after plagioclase microphenocrysts in pillow cores, diabase, and pillow matrices. It also crystallized together with pumpellyite along fractures. The patchy nature of replacement may be due partly to the variable permeability of the rocks. For example, potash feldspar was not found in the fresh basaltic glassy rims, whereas the basaltic pillow cores and pillow matrices contain the characteristic zeolite facies minerals potash feldspar + pumpellyite + laumontite + smectite. The replacement of plagioclase (*e.g.*, An60) by potash feldspar can be interpreted as a consequence of metasomatic exchange involving the addition of K and  $SiO_2$  according to the reaction:



Experimental studies by Hemley (1959) on the system  $K_2O-Al_2O_3-SiO_2-H_2O$  indicate that precipitation of potash feldspar at low temperatures (*e.g.*, less than  $250^\circ C$ ) requires high  $K^+/H^+$  ratios in the hydrothermal solution at low fluid pressures. Therefore, although the host oceanic basalts may be depleted with  $K_2O$ , potash feldspar may locally precipitate through interaction of basaltic rocks with K-rich solutions at relatively shallow depths. The occurrence of potash feldspar in the basaltic rocks suggests, therefore, that these rocks have been subjected to local metasomatic exchange with a hydrothermal solution under zeolite facies conditions.

#### Selective alteration

All pillow glassy rims, pillow cores and the interstitial pyroclastic matrix must have been subjected to identical  $P-T$  conditions during zeolite facies metamorphism. Yet plagioclase, olivine, and basaltic glass in the pillow rims remain virtually unaltered and unmetamorphosed (Liou, 1974), whereas the pillow

cores, massive flows, and pillow matrices are highly altered and possess in common the mineral assemblage albite + chlorite + pumpellyite + K-feldspar + sphene. What caused such contrasts in mineral assemblages and the process of alteration of primary plagioclase?

Similar features have been widely reported in numerous low-grade metamorphic sequences (see Zen and Thompson, 1974). The extent of albitization of primary plagioclase has been shown to be related roughly to one or more of these factors: (1) depth of burial resulting in increased temperature and pressure; (2) availability of fluid during the alteration; (3) the composition of the fluid (for a review, see Kuniyoshi and Liou, 1976). Differences in  $P-T$  conditions alone do not determine the extent of plagioclase alteration, inasmuch as pillow rims, cores, and matrices of the ophiolite were all subjected to the same  $P-T$  conditions. It is conceivable that a more extensive albitization may be facilitated by infiltration of a hydrous fluid phase during alteration. However, as shown by Kuniyoshi and Liou (1976), the composition of the fluid (in particular the concentration of Fe and Mg ions) is a critical factor in low-grade metamorphic reactions. This hypothesis can also be substantiated for the East Taiwan Ophiolite. For instance, there is a direct relation between albitization of plagioclase and alteration of primary olivine in the extrusive sequence. In the albitized massive flows and pillow cores, primary olivines are invariably altered to chlorite and smectite, whereas the olivines in the glassy basalts are virtually unaltered. The alteration of olivine to chlorite and smectite requires introduction of  $Al_2O_3$  and  $SiO_2$ , and release of FeO and MgO into an aqueous solution. Thus the interstitial fluid probably was highly charged with  $Fe^{2+}$ ,  $Fe^{3+}$ , and  $Mg^{2+}$  and was able to convert the anorthite component of primary plagioclase to pumpellyite + chlorite. Of course, the alteration of primary magnetite-ilmenite opaques to magnetite-sphene intergrowths is related to the albitization, as documented by Kuniyoshi and Liou (1976); this reaction may also have played a significant role during the metamorphism of the ophiolite, inasmuch as magnetite-sphene intergrowths are abundant in both plutonic and extrusive rocks.

#### Physical conditions of metamorphism

The mineral assemblages of the hydrothermally-altered basaltic sequence from the ophiolite are characteristic of the zeolite facies. Depending on the bulk composition, oxygen fugacity, and the mode of oc-

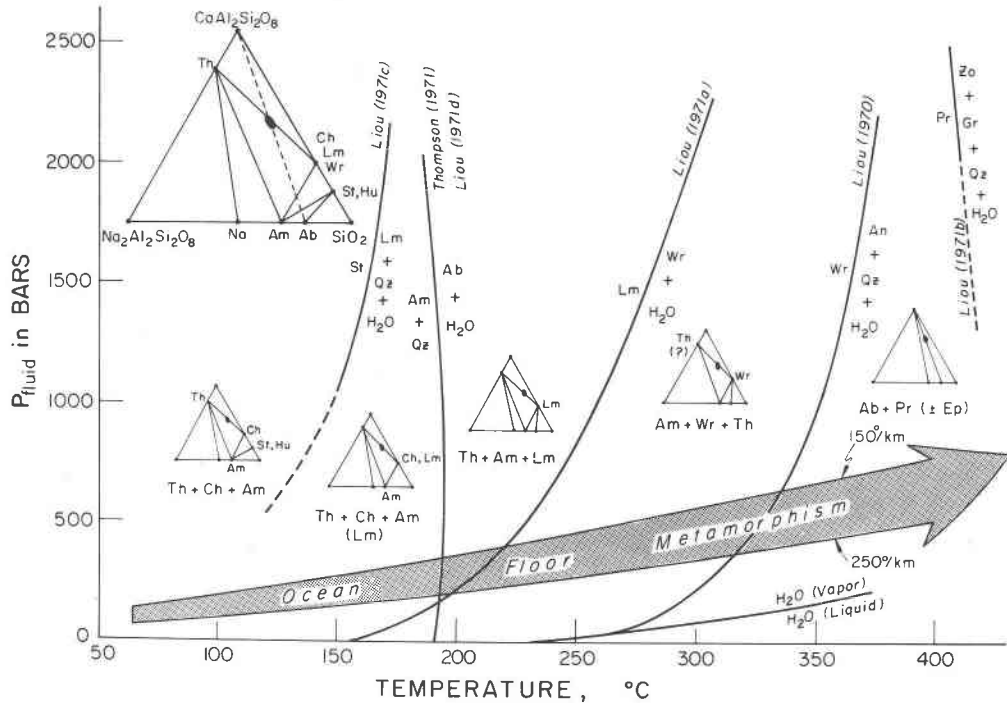


Fig. 6.  $P_{\text{fluid}}-T$  diagram showing the progressive mineral changes along the geothermal gradients suggested for "ocean-floor" metamorphism (Spooner and Fyfe, 1973; Gass and Smewing, 1973; de Wit and Stern, 1976; Coleman, 1977). The zeolite assemblages are plotted on anhydrous diagram  $\text{CaAl}_2\text{Si}_2\text{O}_8-\text{Na}_2\text{Al}_2\text{Si}_2\text{O}_8-\text{SiO}_2$ . Also shown are experimentally-determined stabilities for stilbite (Liou, 1971c), wairakite (Liou, 1970), analcime + quartz (Thompson, 1971; Liou, 1971d), laumontite (Liou, 1971a), and prehnite (Liou, 1971b). Plagioclase compositions of An 50-60 are shown as black spots. Abbreviations: Th, thomsonite; Ch, chabazite, Lm: laumontite, Wr: wairakite, St: stilbite, Hu: heulandite, Ab: albite, Am: analcime, Na: natrolite, Qz: quartz, An: anorthite, Pr: prehnite, Zo: zoisite, Gr: grossular, and Ep: epidote.

currence, calcium zeolites and analcime together with adularia, albite, sphene, smectite-chlorite, Fe-pumpellyite, and Fe-prehnite are ubiquitous. No prehnite-pumpellyite association was found in any investigated rocks.

Experimental studies suggest that zeolite facies conditions are in the range of 100-300°C and total pressures below 3 kbar, which are consistent with those deduced from reported burial metamorphic sequences (for a review see Zen and Thompson, 1974), and with directly observed  $P-T$  conditions from active geothermal areas. Based on reconstructed sequences in various metamorphic terrains, Boles and Coombs (1977) estimated the minimum temperatures of formation for laumontite, prehnite, and pumpellyite to be about 50°, 90° and 190°C, respectively. Provided these estimates are realistic, the laumontite- and pumpellyite-bearing extrusive rocks of the ophiolite have been metamorphosed under zeolite facies conditions at temperatures of about 150° to 250°C.

If ophiolite is initially created at a spreading center where geothermal gradients as high as 500°-1400°C/

km have been postulated (e.g., Cann, 1970; Spooner and Fyfe, 1973), it must have been metamorphosed at extremely low pressures. Geothermal gradients in the order of 150°C/km have been suggested as being maintained in the oceanic crust to at least 100 km from the spreading ridges (Coleman, 1977). From investigations of metamorphic assemblages developed in on-land ophiolites, Coleman and others (e.g., de Wit and Stern, 1976) concluded that geothermal gradients from 150°C/km to 250°C/km adequately explain the observed mineral parageneses from zeolite facies to low-rank amphibolite assemblages in a 2-3 km thick ophiolite sequence. If we take these geothermal gradients as having been imposed on the East Taiwan Ophiolite, at depths of about 0.6 to 1.6 km, temperatures would have reached about 150° to 250°C, enough for the recrystallization of the basaltic rocks to the observed zeolite assemblages.

The  $P-T$  conditions for the diagnostic zeolite facies minerals such as wairakite, laumontite, stilbite, and analcime have been experimentally determined (for review, see Zen and Thompson, 1974), and some of

the  $P$ - $T$  curves are summarized in Figure 6. The high geothermal gradients of  $150^{\circ}$ - $250^{\circ}\text{C}/\text{km}$  are plotted, and zeolite paragenesis in the system  $\text{CaAl}_2\text{Si}_2\text{O}_8$ - $\text{Na}_2\text{Al}_2\text{Si}_2\text{O}_8$ - $\text{SiO}_2$ - $\text{H}_2\text{O}$  and primary plagioclase composition of An 50-60 are also shown. Immediately apparent from such a plot are mineral successions encountered in the oceanic crust. They should change gradually downward from thomsonite + chabazite + analcime, thomsonite + laumontite + albite, and wairakite + analcime + thomsonite to prehnite + albite (all assemblages  $\pm$  quartz  $\pm$  chlorite  $\pm$  sphene). Except for wairakite, these zeolite-bearing assemblages have been commonly described for metamorphosed basaltic rocks from dredge hauls and from on-land ophiolites. However, to my knowledge, wairakite has only recently been reported in the hydrothermally-metamorphosed tholeiites from the north Atlantic Ocean by Jehl *et al.* (1976). The rare occurrence of wairakite in the ridge system, as opposed to its common occurrence in the active geothermal areas, may be due partly to differences in the composition of the hydrothermal solution in these two systems and partly to the difficulty of differentiating optically between wairakite and analcime.

On the basis of field occurrences of wairakite-analcime minerals in the altered volcanic rocks and active geothermal areas in Japan, Seki (1971) suggested that extensive solid solutions exist along the analcime-wairakite join (see also Surdam, 1967) at relatively higher pressure conditions, whereas in low pressure, active geothermal areas almost pure wairakite is stably associated with analcime + quartz and/or albite. As shown in Figure 6, wairakite of its own bulk composition is stable at temperatures ranging from  $210^{\circ}\text{C}$  to  $320^{\circ}\text{C} \pm 10^{\circ}\text{C}$  along the presumed geothermal gradients for the "ocean-floor" hydrothermal metamorphism. For bulk compositions possessing an excess of  $\text{SiO}_2$  (e.g., those compositions on the right side of the wairakite-albite join), analcime is not stable with respect to albite; therefore, nearly pure wairakite + albite + quartz is the dominant assemblage within the stability field of wairakite. Such assemblages commonly have been reported in active geothermal areas. However, for the bulk compositions on the left side of the wairakite-albite join, both wairakite and analcime may be stable together with albite or thomsonite (?). Very fine-grained wairakite of nearly end-member composition crystallizes ubiquitously together with smectite and albite in the interaction of seawater and basalt at temperatures of  $250^{\circ}$ - $350^{\circ}\text{C}$  and  $P_{\text{fluid}} = 500$  bars (Seyfried, 1977). Therefore the rarity of reported occurrences of

wairakite in the ridge system may be due to its fine-grained nature, which inhibits positive optical identification.

In summary, the basaltic rocks from the East Taiwan Ophiolite were hydrothermally metamorphosed at a temperature between  $150^{\circ}$ - $250^{\circ}\text{C}$  and a depth of 0.6 to 1.6 km, probably at some distance from the ridge. They characteristically contain various calcium zeolites, analcime, and smectite-chlorite together with iron-rich pumpellyite, minor prehnite and adularia. The variation in mineral association among pillow cores, rims, and matrices is believed to be the result of local variations in the composition of circulating fluids, in the extent of recrystallization, and in temperature. Mineral zonation observed in vein assemblages indicates that the hydrothermal composition changed with time.

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