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METAMORPHISM IN THE ALEUTIAN ARC: THE FINGER BAY PLUTON, ADAK, ALASKA

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ABSTRACT

The metamorphism of the small, early Tertiary Finger Bay pluton on Adak Island in the central Aleutians occurred under low pressure (<3 kbar) and subgreenschist-to-greenschist facies conditions (325–400°C); it is similar to that of other Aleutian early-stage arc rocks. The primary cause of metamorphism was a hydrothermal circulation of meteoric water associated with later magmatic activity, rather than a separate discrete event such as oceanic-ridge subduction. On all scales, the intensity of metamorphic recrystallization correlates with the proximity of fracture zones. Variable assemblages and compositions of metamorphic minerals, both within and between samples, and the maintenance of igneous whole-rock compositions, suggest only a very local redistribution of elements, consistent with low water-to-rock ratios and the low solute-content of the circulating water. Although very local equilibrium only has been achieved, the metagabbroic rocks appear to be characterized by the assemblage prehnite–chlorite–actinolite. Prehnite, which usually forms as a breakdown product of biotite, is common in the northern and central parts of the pluton. The more siliceous rocks in the south do not contain prehnite, probably owing to compositional effects, but perhaps because they were at temperatures above that of prehnite breakdown. Grandite (grossular–andradite) garnet associated with chlorite occurs in the least recrystallized rocks. Throughout the pluton, high-temperature ilmenite lamellae in titanomagnetite are commonly replaced by titanite and Mn-rich ilmenite. Aluminous actinolite related to the cooling (postmagmatic) history of the pluton is also recognized. The same pattern of metamorphism is expected in accreted island-arc terranes, which are common components of the continental mosaic in areas such as the northwestern part of the Pacific basin.

Keywords: Aleutian island arc, low-temperature metamorphism, hydrothermal circulation, prehnite, grandite, high-Mn ilmenite, oxygen-isotope ratio, Adak Island, Alaska.

SOMMAIRE

Le petit pluton de Finger Bay (île d'Adak, archipel Aléoutien central), d'âge Tertiaire précoce, a été métamorphosé à basse pression (<3 kbar) dans le facies schiste vert (ou à un facies inférieur), de 325 à 400°C. Ce métamorphisme ressemble ainsi à celui des roches du stade précoce du développement de l'arc Aléoutien. C'est la circulation hydrothermale de l'eau météorique, déclenchée par l'activité magmatique locale, qui est principalement à l'origine du métamorphisme, et non un événement distinct, comme la subduction d'une ride océanique. À toutes les échelles, l'intensité de la recristallisation métamorphique montre une corrélation avec la proximité des zones de fractures. La variabilité dans les assemblages et les compositions des minéraux métamorphiques, dans un seul échantillon aussi bien qu'entre échantillons, et la conservation des compositions globales de ces roches ignées, indiqueraient une redistribution très localisée des éléments, compatible avec un faible rapport de roche à eau et une faible concentration de soluté dans l'eau en circulation. Quoique l'équilibre atteint soit seulement local, les roches métagabbroïques contiennent l'assemblage caractéristique prehnite–chlorite–actinote. La prehnite, qui se présente surtout comme produit de décomposition de la biotite, est répandue dans la partie nord et le centre du pluton. Les roches plus siliceuses du Sud sont sans prehnite, probablement pour des raisons de composition, ou peut-être parce qu'elles ont atteint une température excédant celle du champ de stabilité de la prehnite. Un grenat granditique (à grossulaire + andradite) se trouve associé à la chlorite dans les roches les moins recristallisées. Partout dans le pluton, des lamelles d'ilménite formées à haute température dans la titanomagnétite sont communément remplacées par de la titanite et de l'ilménite manganifère. On trouve aussi une actinote alumineuse associée au refroidissement du pluton, et donc à l'évolution post-magmatique de celui-ci. On peut s'attendre à voir la même signature métamorphique dans les blocs d'accrétion des arcs insulaires, qui sont des composants répandus de la mosaïque continentale des régions comme le Nord-Ouest du bassin Pacifique.

(Traduit par la Rédaction)

Mots-clés: arc insulaire Aléoutien, métamorphisme de basse température, circulation hydrothermale, prehnite, grandite, ilménite manganifère, rapport d'isotopes d'oxygène, île d'Adak, Alaska.

INTRODUCTION

The metamorphic history of northern Adak Island in the central Aleutian island arc is recorded in the Tertiary Finger Bay pluton, the oldest known in the arc (Fig. 1a). Studies of Aleutian metamorphism are few, and this paper is the first detailed study to include chemical data bearing on the metamorphic minerals. Metamorphic mineral assemblages reported in the USGS Bulletin 1028 series on the Aleutian Islands (*i.e.*, Fraser & Snyder 1959, Coats 1956) and in later work by Hein & McLean (1980) and Lankford & Hill (1979), together with other field work in the central Aleutians (S.M. Kay, J.L. Rubenstone and others, unpubl. data), indicates that the metamorphic history of the Finger Bay pluton is similar to that of metavolcanic Tertiary rocks of

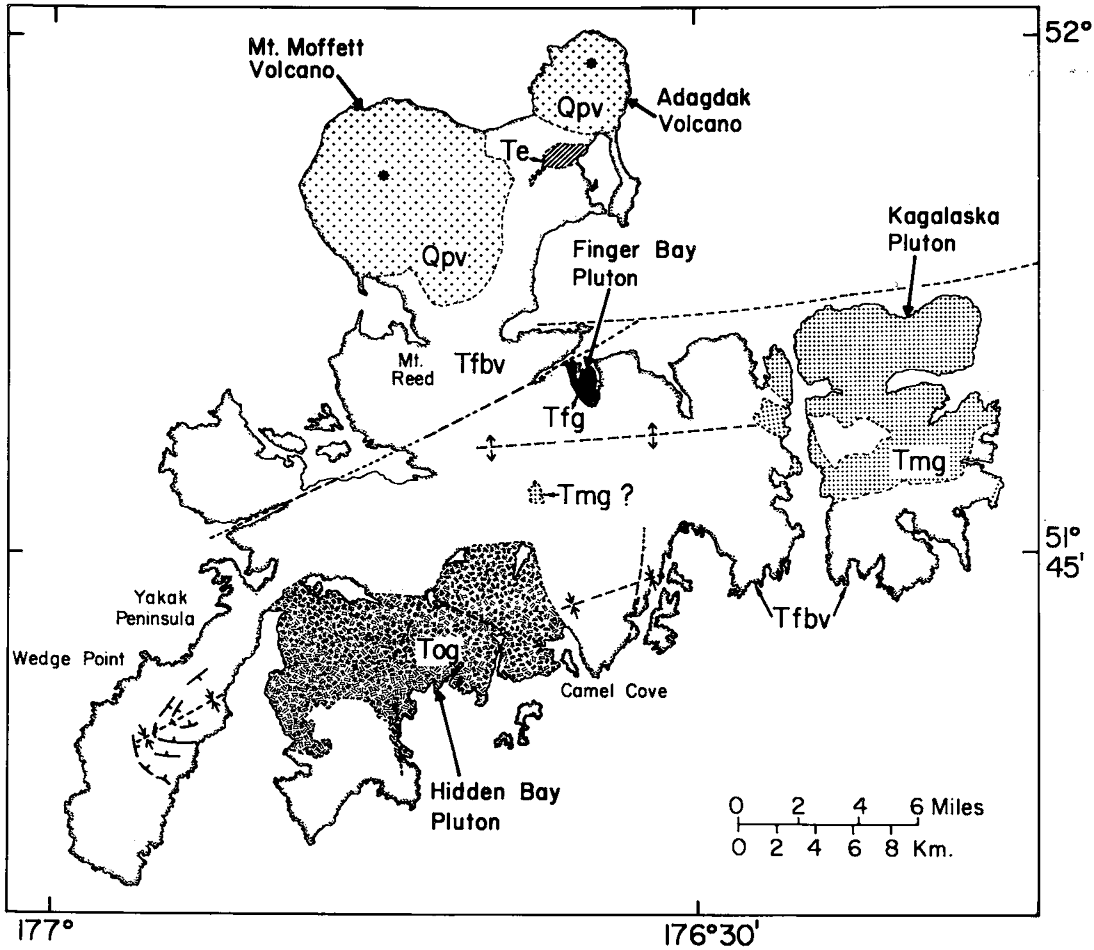

FIG. 1a. Map of Adak Island in the central, oceanic part of the Aleutian arc (176°15′to 177°E latitude, 51°30′ to 52°N longitude) showing the location of the Finger Bay pluton and other localities referred to in text. Units, in order of ascending age, are: Tfbv lower Tertiary Finger Bay volcanic suite, Tfg lower Tertiary Finger Bay pluton, Te Eocene Andrew Lake Series, Tog Oligocene Hidden Bay pluton, Tmg? Miocene(?) (Gannet Lake pluton), Tmg Miocene Kagalaska pluton, and Qpv Plio-Pleistocene volcanic suite. Fold axes (dashed lines with arrows) and faults (dashed lines, finer where inferred) are shown where mapped. Information from Coats (1956), Fraser & Snyder (1959), Hein & McLean (1980) and Citron *et al.* (1980).

the eastern and central Aleutians. Early and middle Tertiary rocks of the western Aleutians (*e.g.*, on Attu Island) have a different igneous and metamorphic history (Rubenstone 1982, Rubenstone *et al.* 1982).

The Finger Bay pluton on Adak Island (Fig. 1a, b) is intruded into the Finger Bay volcanic suite (Coats 1956), which forms part of the early to middle Tertiary Aleutian 'initial series' (Scholl *et al.* 1975). The exact age of the pluton is unknown, but other nearby plutons (Fig. 1a), including the Hidden Bay pluton (33 Ma: Citron *et al.* 1980) and the Kagalaska pluton (13.5 Ma: Citron *et al.* 1980) are unmetamorphosed and presumably younger. A K–Ar date of 31 Ma on the Finger Bay pluton, quoted as a personal communication from B.D. Marsh by DeLong *et al.* (1978), is interpreted by DeLong *et al.* to be the age of the metamorphic overprint.

The rocks of the Finger Bay pluton are predominantly gabbro, with lesser amounts of diorite, quartz monzodiorite and quartz monzonite, and have retained their igneous composition and much of their igneous texture through the metamorphism. The pluton follows a tholeiitic, Fe-enrichment trend of igneous differentiation, resembling that of the Aleutian Plio-Pleistocene to Recent tholeiitic volcanic centres, and is interpreted to represent the root zone of a tholeiitic volcano (Kay *et al.* 1982). The tholeiitic nature of the pluton contrasts with the later plutons on Adak, which show a calc-alkaline trend of differentiation (Citron 1980, Citron *et al.* 1980). A detailed description of the igneous mineralogy, petrology and geochemistry of the Finger Bay pluton is given by Kay *et al.* (1983).

Investigations of the metamorphic histories of relatively young oceanic island-arcs like the Aleutians are important in providing constraints on heat transfer in arc environments and on plate interaction in the past. Few such studies have been done. In the Aleutian arc, DeLong & Fox (1977) and DeLong *et al.* (1978) related the thermal history of the arc directly to the history of plate convergence in the northern Pacific. They suggested that the Aleutian 'initial series', which includes the Finger Bay pluton and the Finger Bay volcanic suite, are in the greenschist facies and that the metamorphism of these rocks accompanied uplift during an amagmatic period of oceanic-ridge subduction at about 30 Ma. However, other investigators contend that ridge subduction occurred closer to 50 Ma (Atwater & Molnar 1973) or that the ridge stopped spreading before reaching the arc (Byrne 1979). Furthermore, subsequent work has shown that the Hidden Bay pluton on Adak was intruded around 33 Ma (K–Ar ages span from 31 to 35 Ma: Citron *et al.* 1980) and that subgreenschist-facies assemblages are present elsewhere in the Finger Bay volcanic suite on Adak Island (Rubenstone & Kay 1980). In this paper, the distinctive low-grade mineral assemblages that develop in the Finger Bay pluton are used to evaluate the cause of this metamorphism.

An understanding of the style and causes of metamorphism in oceanic arc regions is also important when attempting to unravel the complex metamorphic history of arc terranes that have been accreted onto continental margins. Although this topic is not pursued in detail in this paper, it is clearly

TABLE 1. MINERALOGY OF THE FINGER BAY PLUTON

Sample	Type (SiO_2,FeO/MgO)[1]	Plag[2]	Cpx[2]	Igneous Minerals[3]	Metamorphic Minerals[3]
FB81L	gabbro (49, 2.29)	1	1	MAG,ol(?)	CHL,epi,mag,titan Mn-ilm,hem(?),rutile ?)
FB98	gabbro (50, 2.43)	1	1	MAG,ol(?),opx(?) kspar,qtz	ACT[4],CHL,grandite, qtz,mag,hem
FB8-17	gabbro (52, 2.24)	1	1	BIO,MAG,qtz,kspar	none
FB95	gabbro (52, 2.65)	1		OPX,MAG,KSPAR, qtz,bio	ACT,CHL,mag,qtz
FB96	monzodiorite (60, 4.31)	1-2	1-2	OPX,MAG,KSPAR,QTZ bio(?)	ACT,CHL,epi,titan,qtz
FB8-20	gabbro (47, 2.17)	2	2	BIO,MAG	ACT,CHL,titan,preh(?)
FB6-45, 12Q, K	gabbros (for 12Q, 51, 2.15)	2-3	1-2	MAG,BIO,ol(?), opx(?),qtz,kspar	ACT,CHL,PREH,EPI,kspar titan,albite,qtz
FB12N	gabbro	2	2	MAG,OPX,ol(?), qtz, kspar	ACT,CHL,epi,mag,hem, titan,qtz
FB8-13	monzodiorite (55, 2.62)	2-3	1	MAG,QTZ,KSPAR	ACT,CHL,mag,qtz
FB90f	quartz monzodiorite (63, 5.19)	1-2	3	MAG,KSPAR,QTZ	ACT,chl,mag,qtz
FB53	gabbro (48, 1.74)	2-3	2-3	MAG,ol(?),	ACT,CHL,epi,mag,Mn-ilm, titan,qtz
FB12M	gabbro (49, 2.46)	4	3-4	MAG,OPX,ol(?)	ACT,CHL,epi,mag,qtz
FB61	quartz monzodiorite (62, 5.34)	3	4	KSPAR,MAG,QTZ	ACT,CHL,titan,Mn-ilm,qtz
ADK27	quartz monzonite (67, 6.21)	3	-	KSPAR,QTZ,AMP, MAG, allanite	ACT,CHL,epi,mag,qtz
FB6-2	gabbro	5	5	MAG,opx(?),bio(?), qtz,kspar	ACT,EPI,preh,chl,mag, qtz,hem

[1] Numbers in parenthesis for SiO_2 and FeO/MgO ratios are from whole-rock analyses from Kay et al. (1983). [2] Feldspar and clinopyroxene alteration is ranked on a scale of 1-5 with 1 indicating almost a complete lack of alteration to 5 indicating almost complete replacement. Lines separate sample groups that generally correspond to mapping units on Figure 2b. [3] Minerals in capital letters are major components, minerals in small letters are minor components. [4]Actinolite is post-magmatic (see text). Accessory minerals are not listed. Mag magnetite, ol olivine, chl chlorite, titan titanite, Mn-ilm Mn-rich ilmenite, hem hematite, opx orthopyroxene, act actinolite, qtz quartz, bio biotite, epi epidote, preh prehnite, kspar K-feldspar.

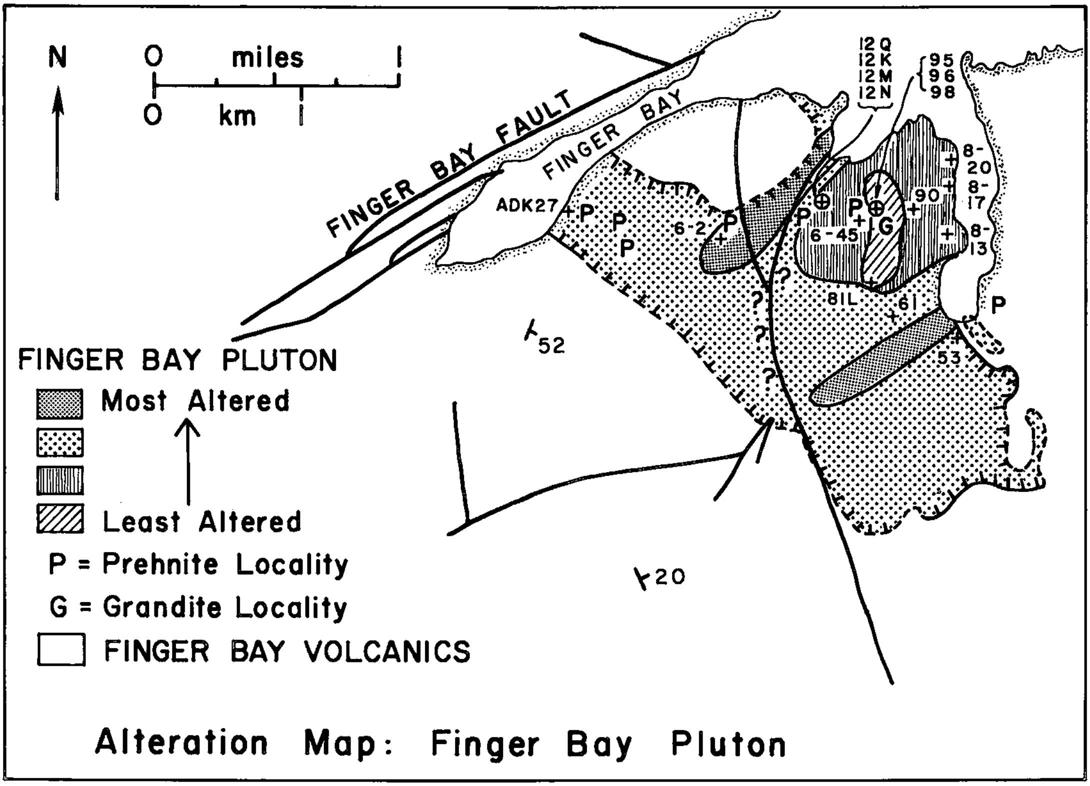

FIG. 1b. Generalized map showing alteration of the Finger Bay pluton, based on the relative alteration of plagioclase and pyroxene (see Table 1). Broad solid lines are inferred faults. Question marks along the fault through the centre of the pluton indicate regions of no outcrop, *i.e.,* where the degree of alteration is unknown. P indicates the presence of prehnite, and G, that of grandite garnet in the assemblage. Hachured boundaries are contacts between the Finger Bay pluton and the Finger Bay volcanic suite and are dashed where inferred. Strike and dip symbols indicate bedding in the Finger Bay volcanic suite. Numbers by crosses are sample localities. Circled crosses are locations of multiple samples, too closely located to be individually indicated. Dotted pattern indicates shoreline. Map base from Coats (1956).

important in any attempt to relate metamorphic and tectonic events in these terranes.

METHODS OF STUDY

Samples were collected during field mapping to determine the igneous character of the Finger Bay pluton (Kay *et al.* 1983). Sixty thin sections have been examined for their metamorphic mineralogy and twenty were selected for analytical work. Minerals were analyzed using a JEOL 733 electron microprobe with Tracor Northern automation at Cornell University. Results were refined using the Bence–Albee correction scheme with the alpha factors of Albee and Ray. Standards were obtained from the Smithsonian Institution (Jarosewich *et al.* 1979). Minerals of known composition were used as internal checks. All analyses were done at 15 kV with a sample current of 0.015 μA. Many analyses represent single points, as inhomogeneities made averaging difficult. Sodium was counted for 40 seconds; all other elements were counted for 60 seconds.

Oxygen isotopes were analyzed by T.F. O'Brien using the method of Clayton & Mayeda (1963) on a Varian mass spectrometer at Northern Illinois University.

PATTERN OF METAMORPHIC ALTERATION

The degree of metamorphic recrystallization in the Finger Bay pluton is variable and, on all scales, appears to correlate with the proximity of fractures and faults. A generalized map (Fig. 1b) based on the preservation of plagioclase and pyroxene (Table 1) shows the distribution of alteration within the pluton. The most altered areas occur in zones crudely parallel to the NE–SW-trending Finger Bay fault, and may themselves be fault traces. Alteration along the central and southern part of the fault running through the middle of the pluton is difficult to assess as the outcrop is very poor in this region. The northeastern part of the pluton is the least fractured mesoscopically and the least recrystallized metamorphically.

Most fractures and faults dip steeply and have small displacements (Coats 1956). Predominant trends are from N60°E to N60°W and from N20°E to N10°W, although intermediate trends are present and some faults curve. The fracture spacing in the most fractured parts of the pluton is less than 20 m, whereas in more coherent parts, the spacing is up to 100 m. Fracturing on the scale of several cm occurs in some intensely recrystallized areas, and veins containing epidote, actinolite, chlorite and quartz are found in these zones. The faults and fractures cut the Finger Bay volcanic suite as well as the pluton, but not the Plio-Pleistocene volcanic rocks (Coats 1956). The correlation of fracturing with vein fillings and the degree of metamorphic recrystallization suggest that at least some of the fracturing is contemporaneous with, or predates, the metamorphism. Subsequent fracturing may have followed predefined zones of weakness.

Table 1 indicates the state of alteration of primary feldspar and clinopyroxene, and the igneous and metamorphic mineralogy of representative samples spanning the range from the most to the least recrystallized. Metamorphic minerals are listed as those present, rather than as assemblages, because equilibrium has in many cases not been attained. Igneous textures are still visible in most samples. In the most recrystallized samples, igneous plagioclase is heavily saussuritized, clinopyroxene ranges from fairly unaltered to completely replaced by actinolite or chlorite (or both), and most other igneous minerals are completely replaced. Orthopyroxene and olivine are principally replaced by chlorite and

TABLE 2. REPRESENTATIVE COMPOSITIONS OF AMPHIBOLE, FINGER BAY PLUTON

	1	2	3	4	5	6	7	8	9
Sample	FB6-2	FB6-2	FB6-2	FB6-2	FB6-45	FB6-45	FB6-45	FB98	FB98
SiO_2	49.62	50.46	52.72	50.05	49.51	46.12	49.55	50.37	53.15
TiO_2	0.38	0.30	0.46	0.19	1.02	0.02	0.49	0.19	0.03
Al_2O_3	5.05	4.09	2.12	3.98	4.04	4.11	3.10	3.36	2.35
FeO	13.88	16.89	10.03	21.41	13.48	28.15	18.92	21.88	14.08
MnO	0.45	0.42	0.54	0.68	0.43	0.69	0.59	0.70	0.48
MgO	15.02	13.00	17.59	9.41	16.76	6.61	12.06	11.55	17.15
CaO	12.45	12.03	11.61	11.66	10.68	10.98	11.50	9.52	10.09
Na_2O	0.50	0.29	0.55	0.30	1.28	0.82	0.65	0.65	0.16
K_2O	0.19	0.18	0.28	0.28	0.38	0.44	0.21	0.34	0.05
Total	97.54	97.66	95.90	97.96	97.58	97.94	97.07	98.56	97.54

	10	11	12	13	14	15	16	17	18
Sample	FB12K	FB12Q	FB96	FB90	FB90	FB61	ADK27	ADK27	ADK27
SiO_2	51.40	52.38	47.14	50.89	47.88	53.07	51.36	50.74	47.64
TiO_2	0.72	0.54	0.98	0.21	1.30	0.31	0.89	0.48	0.99
Al_2O_3	4.70	2.37	4.39	1.98	5.32	2.08	4.13	2.77	5.02
FeO	10.85	17.49	24.10	22.32	19.24	14.31	10.35	18.00	18.71
MnO	0.62	0.52	1.09	0.65	0.63	0.87	0.76	1.01	0.82
MgO	17.32	13.44	8.83	9.93	11.16	14.86	17.35	13.00	12.04
CaO	11.28	11.52	9.78	11.76	10.66	11.71	10.23	10.60	10.41
Na_2O	0.99	0.65	1.49	0.39	1.37	0.86	2.42	0.87	1.84
K_2O	0.36	0.25	0.47	0.16	0.58	0.20	0.81	0.40	0.70
Total	98.24	99.16	98.27	98.29	98.14	98.27	98.30	97.87	98.17

1) Small bladed actinolite needles in epidote (Table 3, #3) in gabbro. 2) Actinolite after clinopyroxene in gabbro. 3) Core of actinolite (#4) in gabbro. 4) Actinolite rim (#3) next to epidote (Table 3, #3) containing prehnite (Table 3, #7) in gabbro. 5) Actinolite in contact with biotite altering to prehnite (Table 3, #6) in gabbro. 6) Blue-green actinolite enclosed in light green actinolite in gabbro. 7) Actinolite after clinopyroxene in gabbro. 8) Epitactic actinolite on clinopyroxene in gabbro. 9) Fibrous actinolite surrounded by chlorite (Table 2, #5) containing grandite in gabbro. 10) Actinolite next to biotite altering to prehnite (Table3, #5) and chlorite in gabbro. 11) Well-crystallized actinolite in intergranular area with chlorite (Table 2, #9) in gabbro. 12) Separate actinolite grain in monzodiorite. 13,14) Small actinolite grain in quartz monzodiorite. 15) Bladed edenite in association with chlorite-titanite-quartz in quartz monzodiorite. 16) Separate actinolite grain in quartz monzodiorite. 17) Separate grain in quartz monzonite. 18) Most common actinolite composition in quartz monzonite.

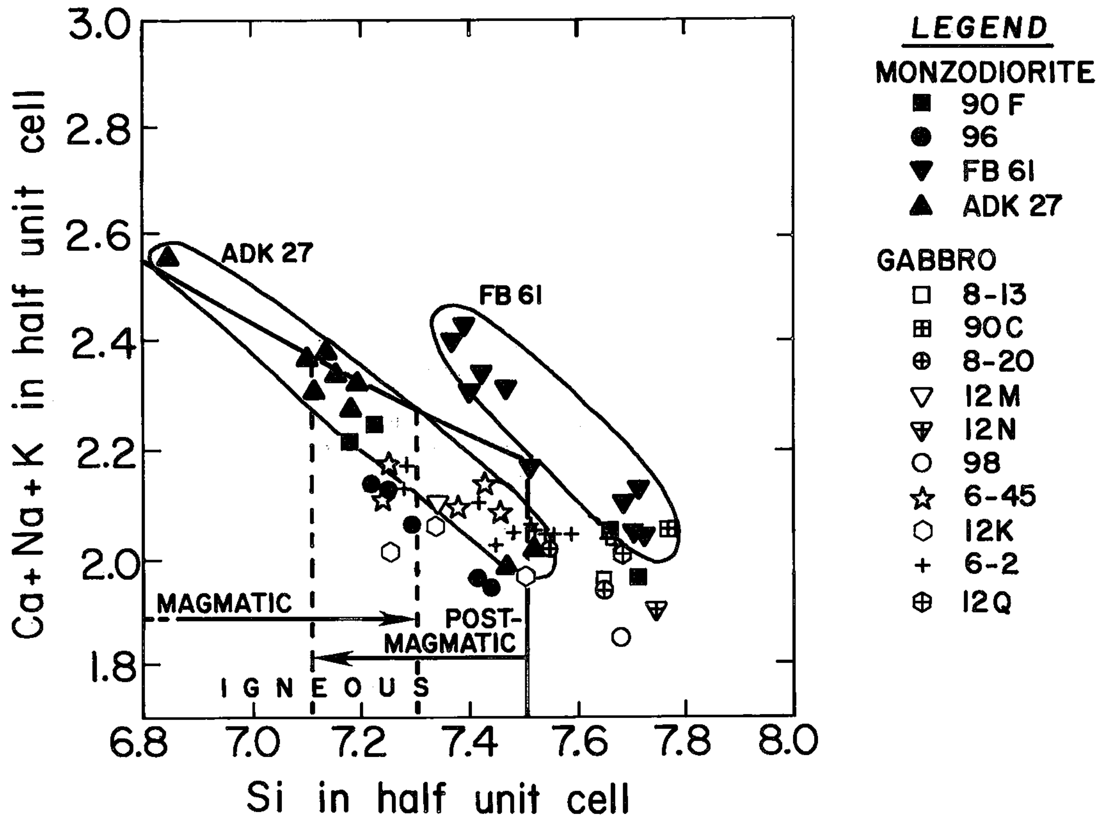

FIG. 2. (Ca + Na + K) *versus* Si for amphiboles from the Finger Bay pluton. The plot shows that most amphiboles in the pluton are not of magmatic origin. Magmatic and postmagmatic fields (dashed and solid lines) are defined on the basis of compositions published by Leake (1971). Finger Bay amphiboles from monzodiorites, quartz monzodiorites and quartz monzonites are indicated by solid symbols. Those from gabbros are shown as open symbols. The same symbols are also used in subsequent figures. Some amphiboles from quartz monzonite ADK27 (circled field) plot on the upper limit of the magmatic field and are magmatic in origin. Others in the same sample appear to be secondary. Amphiboles in quartz monzodiorite FB61 (circled field) plot clearly outside the magmatic and postmagmatic fields and have secondary textures. Amphiboles plotting in the overlap region between the magmatic and postmagmatic fields have secondary textures.

magnetite, and biotite is replaced by chlorite, actinolite, prehnite, K-feldspar and titanite. Small veins and patches occur, principally composed of chlorite with magnetite and fibrous quartz, or of some combination of chlorite, actinolite and epidote. Hematite is found sporadically in some very highly altered areas. In less recrystallized samples, plagioclase and clinopyroxene are almost completely unaltered. In some of these samples (*e.g.*, FB98, Table 1), chlorite with small inclusions of grandite garnet surrounds partly decomposed actinolite. In the freshest samples, olivine(?) is completely replaced by chlorite and magnetite, whereas igneous orthopyroxene, biotite and amphibole may be only partly replaced. In all samples, titanomagnetite commonly retains the composition it obtained during postmagmatic cooling, although ilmenite lamellae and blebs included in the titanomagnetite are often partly or completely replaced by titanite and Mn-rich ilmenite.

The metamorphic mineralogy of the Finger Bay pluton generally does not show a well-defined spatial distribution that can be used to determine a metamorphic gradient. The distribution of grandite-chlorite-bearing, prehnite-actinolite-bearing, and prehnite-free, chlorite-actinolite-epidote-bearing assemblages may indicate that the lowest-grade samples (Fig. 1b and later section) occur in the northern area of the pluton. However, the occurrence of prehnite (Fig. 1b) in the latter two assemblages may be a function of composition, as it is usually associated with igneous biotite, which is not uniformly distributed across the pluton (Kay *et al.* 1983). The northern and central parts of the pluton, where most of the prehnite occurs, are largely composed of gabbro and diorite (<53% SiO_2), which commonly contain biotite. The south-central and southeastern parts of the pluton, where prehnite has not been found, are largely composed of quartz monzodiorite, which does not contain biotite. The grandite–chlorite assemblage in the northern region may be lower grade (see later).

METAMORPHIC MINERALOGY

Amphibole

Amphibole occurs in most samples of the Finger Bay pluton and may either be magmatic, 'postmagmatic' (subsolidus, formed during cooling of the crystallized magma), or metamorphic in origin. Compositions show a wide range both within and between samples. Fourteen samples have been analyzed, and representative compositions and petrographic descriptions are given in Table 2. Based on the classification of Leake (1978), most amphiboles are actinolite and edenite, although a few are actinolitic or edenitic hornblende. Although ferric iron was not determined, the fact that 75–80% of the Al is in the tetrahedral sites suggests that ferric iron is a significant component of the octahedral sites (Czamanske & Wones 1973).

Compositional and textural data suggest that in most cases, amphibole is not magmatic. A plot of molar (Ca + Na + K) *versus* Si in the half cell (Fig. 2), used by Leake (1971) to discriminate between magmatic and 'postmagmatic' (subsolidus) amphiboles, shows that many of the Finger Bay amphibole compositions lie outside the magmatic field. Furthermore, many of the grains that lie within or near the magmatic field show secondary textures (Table 2, #1, 5, 10, 11, 12 and 13). The only amphibole that appears to be primary occurs in the most silicic unit of the Finger Bay pluton (quartz monzonite ADK27, Kay *et al.* 1983) and plots in the alkali-rich part of Leake's magmatic field. Other amphiboles occurring in ADK27 and the quartz monzonite samples are interpreted as secondary (Table 2, #16–17). Except for the primary amphibole in ADK27, all samples of actinolite have lower ^{iv}Al than magmatic amphiboles from plutons such as those in the Sierra Nevada (Dodge *et al.* 1968).

Distinguishing amphiboles that formed during the initial cooling of the pluton ('postmagmatic') from those formed during a later hydrothermal metamorphism is difficult. Compositionally, most Finger Bay amphiboles cannot be separated from 'postmagmatic' amphiboles in the unmetamorphosed Aleutian calc-alkaline plutons (Perfit 1977, Citron 1980), particularly in the low-Al range. Textural criteria are difficult to use in establishing the origin of amphibole that occurs as a replacement or as an epitactic growth on pyroxene (Table 2, #2 and 7) and as separate grains (Table 2, #3, 6, 12, 13, 14, 17 and 18). Actinolite in the least recrystallized samples is the best candidate for 'postmagmatic' amphibole. This amphibole has a deficiency of Ca in the *M*4 site (Table 1, FB97; Table 2, #8–9) that probably results from partial decomposition of actinolite to hydrous layered silicates under the conditions of the superimposed metamorphism. In other samples, actinolite intergrown with metamorphic phases (Table 2, #1, 4, 5, 10, 11, 15 and 16) is clearly of metamorphic origin.

Regardless of whether the actinolite is postmagmatic or metamorphic in origin, compositions are variable within individual samples (Fig. 2). Compositional ranges are illustrated in Table 2 for actinolite in one of the most altered gabbros (FB6–2, #1–4) and in one of the least altered gabbros (FB6–45, #5–7). In many samples, Ti, Na and K decrease with decreasing Al (Table 2, #8–9 and 15–18), indicating coupled substitutions (*e.g.*, Czamanske & Wones 1973). Within single samples, Fe behaves somewhat systematically with ^{iv}Al, although large differences occur between samples. In general, Fe is relatively constant with ^{iv}Al in ADK27 and FB96, decreases with ^{iv}Al in FB61 and FB90, and increases with ^{iv}Al in FB6–2.

The range of actinolite compositions is most easily explained as a reflection of very local equilibrium and direct compositional control by the phase or group of phases that the amphibole is replacing. Equilibrium on the scale of a hand specimen has not occurred, and whole-rock FeO/MgO (total Fe as FeO) ratios are not important in controlling the composition of secondary amphibole. For example, a quartz monzodiorite with a whole-rock FeO/MgO ratio of 5.3 contains actinolite (Table 2, #15-16) with a Fe content as low as or lower than actinolite in gabbros with a whole-rock FeO/MgO ratio around 2.0 (Table 2, #8, 9 and 11). Other quartz monzodiorites and quartz monzonites, with a whole-rock FeO/MgO ratio over 4.5, have actinolite with a much higher Fe content (Table 2, #13, 14, 16 and 17). These compositional trends are similar to those of the igneous pyroxene that the actinolite is replacing (Kay *et al.* 1983). Similarly, actinolite with a high *A*-site occupancy replaces biotite or relatively alkali-rich pyroxene and amphibole (Kay *et al.* 1983); therefore, the high *A*-site occupancy does not reflect a medium- to high-pressure metamorphic environment (Grapes & Graham 1978). Actinolite compositions in one of the most highly altered samples (FB6–2) show a tschermakitic trend that is typical of actinolite from low-pressure metamorphic environments (*e.g.*, in the Karmutsen volcanic suite: Kuniyoshi & Liou 1976b).

The actinolite (approximately 4% Al_2O_3, Table 2, #4, 5 and 10) associated with prehnite has a higher

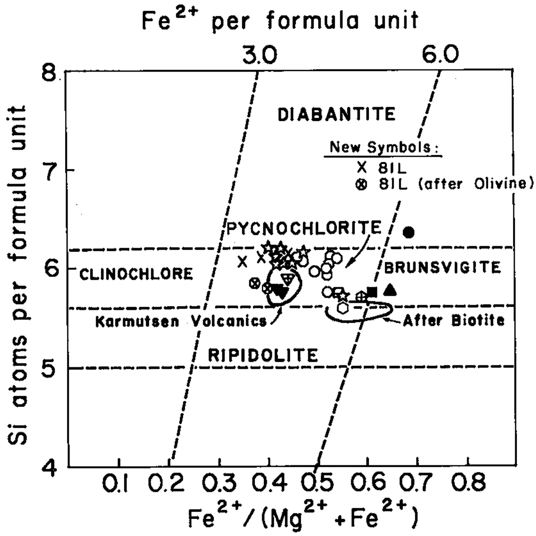

FIG. 3. Chlorite from the Finger Bay pluton, plotted on the classification diagram of Hey (1954), showing diversity of composition within single samples. Total Fe is plotted as Fe^{2+}. Symbols are as in Figure 2 with the additions noted on the figure. Field within the pycnochlorite field represents chlorite from the epidote–actinolite facies in the contact aureole in the Karmutsen volcanic suite (Kuniyoshi & Liou 1976b). Field on edge of pycnochlorite field, labeled 'after biotite', is the compositional range of chlorite associated with the breakdown of biotite, from Tulloch (1979).

TABLE 3. REPRESENTATIVE COMPOSITIONS OF CHLORITE, FINGER BAY PLUTON

	1	2	3	4	5	6
Sample	FB81L	FB81L	FB61	FB6-45	FB98	FB98
SiO_2	29.23	27.80	27.82	30.35	28.80	28.44
TiO_2	0.07	0.02	0.02	0.02	0.01	0.04
Al_2O_3	16.71	18.15	18.55	16.08	16.01	16.49
FeO	19.29	21.96	22.84	23.14	28.54	28.48
MnO	0.28	0.36	0.93	0.16	0.22	0.29
MgO	20.42	18.74	17.77	19.27	14.33	14.90
CaO	0.35	0.21	0.06	0.14	0.14	0.16
Na_2O	0.19	0.26	0.03	0.00	0.39	0.10
K_2O	0.12	0.03	0.08	0.00	0.22	0.05
Total	86.66	87.53	88.10	89.16	88.66	88.95

	7	8	9	10	11
Sample	FB12K	FB6-45	FB12Q	FB90	FB96
SiO_2	26.86	26.94	27.07	27.37	29.73
TiO_2	0.27	0.04	0.05	0.07	0.46
Al_2O_3	19.23	18.29	18.43	18.87	14.54
FeO	28.94	29.57	30.58	32.75	33.59
MnO	0.59	0.29	0.47	0.39	0.33
MgO	13.78	13.58	12.48	11.84	10.89
CaO	0.10	0.05	0.09	0.05	0.33
Na_2O	0.00	0.00	0.03	0.07	0.04
K_2O	0.06	0.00	0.04	0.07	0.48
Total	89.83	88.86	89.24	91.48	90.39

1) Green chlorite with intergranular area in gabbro. 2) Green chlorite with magnetite (Table 4, #7) after olivine(?) in gabbro. 3) Green chlorite with actinolite (Table 1, #15), titanite and quartz in quartz monzodiorite. 4,5) Brown, high-birefringence chlorite with magnetite after orthopyroxene (?) in gabbro. 6) Green chlorite enclosing grandite garnet (Table 3, #7) and surrounding actinolite after pyroxene or olivine in gabbro. 7) Green chlorite interleaved with prehnite, titanite and actinolite after biotite in gabbro. 8) Green chlorite interleaved with prehnite (Table 3, #6) after biotite in gabbro. 9) Green chlorite in intergranular area with actinolite (Table 1, #11) in gabbro, probably after biotite. 10) Green chlorite intergrown with actinolite in quartz monzodiorite. 11) Green chlorite after opx (?) in monzodiorite.

Al_2O_3 content than that generally observed in low-grade metabasites. Liou *et al.* (1974) reported about 4% Al_2O_3 in amphiboles experimentally recrystallized at 450°C (uppermost greenschist facies) in a basaltic system. However, the Al_2O_3 contents of actinolite in this suite are in the same range as in amphibole associated with prehnite in the Karmutsen volcanic suite, which was metamorphosed under low-pressure conditions (Kuniyoshi & Liou 1976b). Similar low-pressure metamorphic conditions probably also apply to the Finger Bay pluton. In some prehnite-bearing samples, the Al_2O_3 content of actinolite occurring in intergranular regions with chlorite ranges between 1.5 to 2.0 (Table 2, #11). The variable Al_2O_3 content of actinolite in these samples reflects the lack of equilibrium.

Chlorite

Metamorphic chlorite is present in most samples from the Finger Bay pluton. Representative compositions and textural descriptions are given in Table 3. Using the classification of Hey (1954), shown in Figure 3, most analyzed samples are pycnochlorite. Two Fe-rich samples are brunsvigite, whereas one analyzed sample (Table 3, #10), with a somewhat poorer charge-balance, lies in an unnamed field. The presence of small amounts of CaO, Na_2O and K_2O and the low cation total (assuming all Fe as FeO and normal H_2O contents) in some samples (Table 3, particularly #1, 2, 5 and 6) are attributed to interlayered clay minerals that are suggested to occur in chlorite from very low-grade rocks (Boles & Coombs 1977, Coombs *et al.* 1977).

Chlorite within a particular sample shows a clustering of compositions (Fig. 3); however, a range of compositions does occur within a single sample (*i.e.*, FB6–45, FB81L and FB98, Fig. 3), and overlap occurs between samples regardless of the associated metamorphic minerals. Compositions of chlorite, like actinolite, appear to be largely controlled by conditions of local equilibrium and mimic the composition patterns of the igneous minerals being replaced. Chlorite after biotite is sometimes associated with prehnite and, like that reported by Tulloch (1979),

TABLE 4. REPRESENTATIVE COMPOSITIONS OF CALCIUM ALUMINUM SILICATES, FINGER BAY PLUTON

	Epidote				Prehnite			Garnet
	1	2	3	4	5	6	7	8
Sample	FB12Q	FB81L	FB6-2	FB6-25	FB12K	FB6-45	FB6-2	FB98
SiO_2	38.28	38.21	37.72	38.35	43.33	42.84	43.46	35.57
TiO_2	0.15	0.02	0.32	0.10	0.06	0.25	0.01	0.85
Al_2O_3	24.64	23.24	22.60	21.21	23.37	20.05	24.09	10.51
Fe_2O_3	12.20	12.79	14.38	16.63	3.04	6.18	1.55	18.56
FeO	-	-	-	-	-	-	-	-
MnO	0.22	0.23	0.17	0.13	0-22	0.04	0.06	0.18
MgO	0.03	0.02	0.06	0.09	-	0.15	0.12	0.43
CaO	23.42	23.36	23.30	22.00	26.28	25.78	26.27	33.73
Na_2O	-	0.04	-	0.04	-	0.00	0.00	0.07
Total	98.94	97.91	98.55	98.55	96.08	95.29	95.56	99.90
$Fe^{+3}/(Fe^{+3}+Al)$	24.0	26.0	28.9	33.4	7.7	16.4	4.0	

1) Large epidote grain of in intergranular area in gabbro. 2) Small grain of epidote in gabbro. 3) Epidote enclosing actinolite (Table 1, #1). Most common epidote composition in gabbro. 4) Epidote enclosed in chlorite in gabbro. 5,6) Well-crystallized prehnite interleaved with chlorite after biotite in gabbro. 7) Prehnite enclosed in epidote (#3) in gabbro. 8) Framboidal grandite garnet enclosed in chlorite (Table 2, #8) in gabbro.

has high Al and Fe contents (Table 3, #7–8). Mn, Ti and K are also somewhat high, reflecting incomplete replacement of the biotite. Chlorite also occurs as fibrous replacements of clinopyroxene, orthopyroxene and olivine(?) (Table 3, #2, 4, 5 and 6), as tiny platelets in altered plagioclase, and as massive aggregates in veins and interstitial areas (Table 3, #1, 3, 9 and 10).

The aluminum content of the chlorite is comparable to that in chlorite from the prehnite–pumpellyite and epidote–actinolite facies in the Karmutsen volcanic suite. In both localities, aluminum content is independent of iron content. Kuniyoshi & Liou (1976b) suggested that the aluminum content of chlorite in the Karmutsen rocks is a function of metamorphic grade. In general, in the Finger Bay pluton, chlorite in association with grandite and non-stoichiometric actinolite (*e.g.*, Table 2, #5) or occurring without actinolite (e.g., Table 2, #1) plots near the Karmutsen prehnite–pumpellyite field (Al_2O_3 14–16.5%); chlorite in association with prehnite (*e.g.*, Table 2, #7–8) plots near the Karmutsen epidote–actinolite field (Al_2O_3 16.5–19%). However, there are many exceptions, and chlorite in some samples spans both fields. For example, one prehnite-bearing sample (FB6–45) has chlorite with Al_2O_3 ranging from 15 to 18.5%. Aluminum content of chlorite seems to be controlled by conditions of local equilibrium and is not a reliable indicator of grade in the Finger Bay pluton.

Plagioclase

Plagioclase in the Finger Bay pluton ranges from fresh to altered, although textural evidence for igneous zoning and twinning is preserved in all samples. The most highly altered grains occur in samples with abundant epidote. Microprobe data in most cases (An_{10-90}) reflect original igneous compositions (Kay *et al.* 1983). Where altered, igneous plagioclase is replaced by a combination of fine-grained sericite, epidote, calcite, chlorite, and submicrometre-sized layered silicates(?) that are too fine to analyze or identify precisely. In zoned grains the amount of alteration commonly varies with the An content. Plagioclase grains with a composition ranging from An_{10} to An_{30} are usually partially altered and occur principally in the more silicic samples. This low-An plagioclase is interpreted to represent the last stage of igneous crystallization.

The occurrence of plagioclase of metamorphic origin is very minor. A few small grains of clear albite ($<An_5$) associated with chlorite, epidote and quartz have been found and are interpreted to be metamorphic in origin. No metamorphic oligoclase has been found in the pluton.

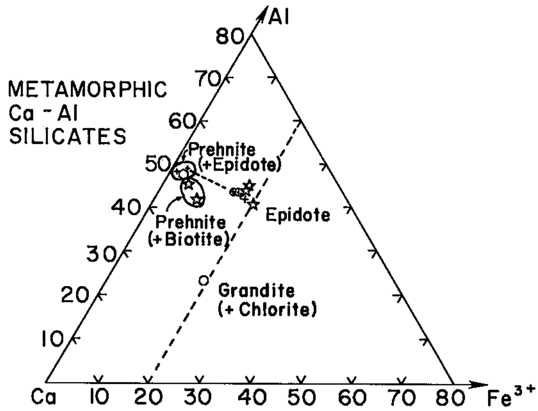

FIG. 4. Triangular plot of Ca–Fe^{3+}–Al for secondary calcium aluminum silicates from the Finger Bay pluton. Total Fe is plotted as Fe^{3+}. Symbols are as in Figure 2. Tieline connects prehnite enclosed in epidote in altered gabbro 6–2.

Secondary Ca–Al silicates

Secondary calcium aluminum silicates in the Finger Bay pluton include prehnite, epidote and grandite garnet and indicate that the metamorphism is in the subgreenschist to lower greenschist facies. Representative compositions are given in Table 4 and plotted in Figure 4. Pumpellyite was not found in any of the samples studied.

Prehnite most commonly occurs as elongate lenses parallel to biotite cleavages (Fig. 5) and is associated with titanite, chlorite, actinolite and K-feldspar. Prehnite appears to form as a breakdown product of biotite (Tulloch 1979) rather than as an intergrowth with biotite (Phillips & Rickwood 1975), and has a ratio 100 [Fe^{3+}/(Fe^{3+} + Al)] that ranges from 5 to 16 (Table 4, #5 and 6). Prehnite with a ratio 100 [Fe^{3+}/(Fe^{3+} + Al)] ranging from 4 to 6 also occurs enclosed in epidote next to actinolite (Table 4, #7) and in clumps associated with chlorite, magnetite and quartz.

Kuniyoshi & Liou (1976b) suggested that the ratio 100[Fe^{3+}/(Fe^{3+} + Al)] in prehnite increases with an increase in metamorphic grade, since a ratio from 0 to 5 characterizes prehnite from Japanese metavolcanic rocks in the prehnite–pumpellyite facies (Hashimoto 1966), whereas higher ratios occur in prehnite associated with epidote and actinolite in the Karmutsen metabasalts. However, prehnite after biotite in actinolite-free low-grade rocks may have a ratio as high as 19, reflecting the high Fe-content of the host biotite (Tulloch 1979). Thus, prehnite compositions in the Finger Bay pluton are generally consistent with the upper prehnite–pumpellyite or lower epidote–actinolite facies, whereas more Fe-rich prehnite reflects the high Fe-content of the biotite it replaces.

Epidote occurs in variable quantities in many samples. Where abundant, it is present in vein fillings and as secondary alteration after plagioclase. Epidote also occurs as a breakdown product of biotite associated with prehnite, chlorite, actinolite and titanite. In most cases, the epidote in the altered samples has a Ps (pistacite, $100[Fe^{3+}/(Fe^{3+}+Al)]$, total Fe as Fe_2O_3) content ranging from 27 to 29 (Table 4, #3). However, like other metamorphic minerals in the Finger Bay pluton, epidote compositions are variable within the same sample. The most altered gabbro examined, FB6-2, contains abundant epidote with a Ps content ranging from 27 to 29, but also contains a large epidote next to chlorite that has a Ps_{36} rim and a Ps_{29} core. Another gabbro, FB12Q, contains well-formed intergranular epidote with $Ps_{24.5}$ (Table 4, #1) and bladed epidote near altered biotite with Ps_{27}. A partially altered gabbro, FB6-45, generally has Ps_{27-28} epidote, but has some Ps_{33} epidote (Table 4, #4) included in chlorite. In less altered samples, epidote is infrequent, and plagioclase is reasonably fresh. In gabbro lacking prehnite and actinolite (*i.e.*, FB81L), the rare epidote has a Ps content ranging from 26 to 27.4 (Table 4, #1).

The range of Ps contents encountered in epidote overlaps those in the prehnite-pumpellyite (Ps_{33}) and epidote-actinolite (Ps_{25-26}) facies of the Karmutsen metabasalts (Kuniyoshi & Liou 1976a), in low-grade New Zealand epidote-grandite-bearing metabasalts ($Ps_{24.7-33.2}$: Coombs *et al.* 1977), and in low-grade rocks in general (Ps_{33}: Miyashiro & Seki 1958). The range of Ps contents in epidote from the Finger Bay pluton thus is consistent with subgreenschist- to greenschist-facies metamorphic conditions, although variations in epidote composition within individual samples indicate that local-composition control and $f(O_2)$ (see Liou 1973) are also important.

In some Finger Bay gabbros, extremely small (up to 10 μm) grains of grossular-andradite garnet (grandite) with a framboidal texture are included in chlorite that surrounds nonstoichiometric actinolite (Table 1, FB98; Table 4, #8). Similar garnet has been described in mafic volcanogenic metasediments in the prehnite-pumpellyite facies in New Zealand

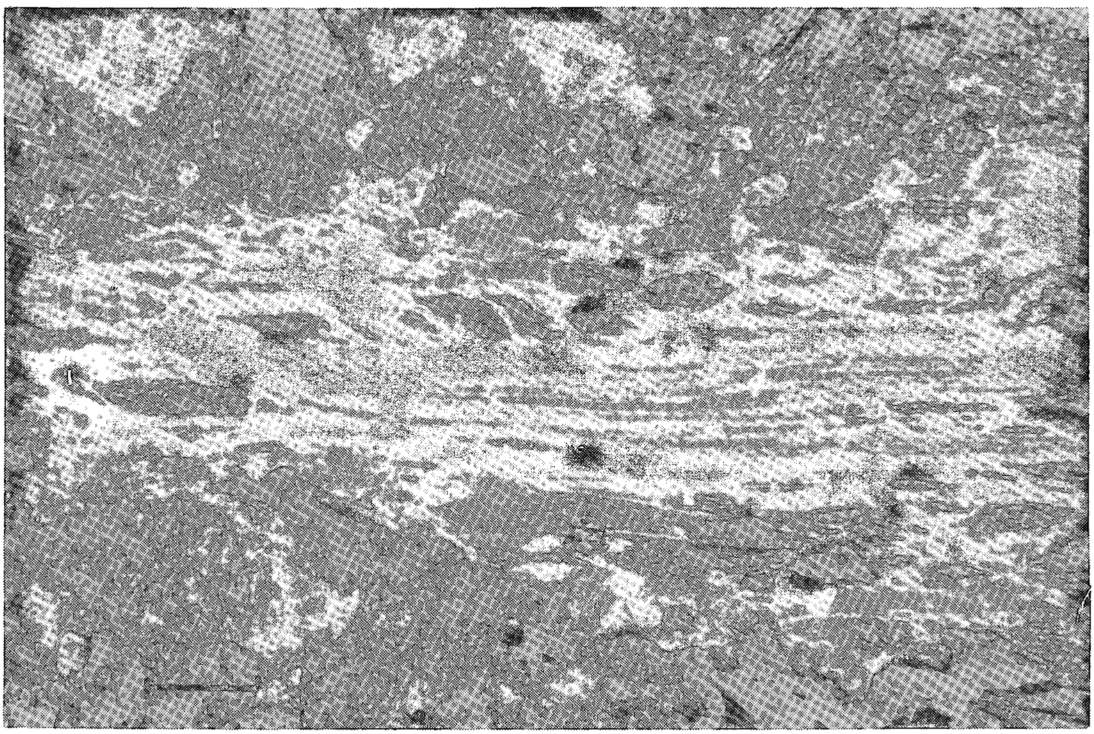

FIG. 5. Photomicrograph in plane light of former biotite in gabbro (FB8-2c), showing replacement by actinolite-prehnite-chlorite assemblage. Dark-colored layered mineral in centre is biotite largely replaced by chlorite and titanite. Lensoid grains on upper edge and included in biotite are prehnite. Actinolite is well-crystallized mineral at bottom of former biotite. Other minerals are plagioclase, K-feldspar, titanomagnetite and minor quartz. Epidote-chlorite-actinolite assemblage occurs in other areas of same sample. See text for discussion. Length of bar is 0.25 mm.

(Coombs *et al.* 1977) and as a breakdown product of biotite (Tulloch 1979). Coombs *et al.* (1977) suggested that grandite may develop in chlorite pseudomorphs after pyroxene or olivine in microdomains where the ratio of available Ca/(Fe^{3+} + Al) exceeds that of epidote and plagioclase. This may be the origin of the chlorite–grandite association in the Finger Bay pluton.

Opaque oxides and titanite

The opaque assemblages of twelve samples from the Finger Bay pluton have been examined in detail, and representative analytical data are given in Table 5. Relict titanomagnetite containing ilmenite lamellae occurs in all samples. Most host magnetite has more than 2% TiO_2 (Table 5, #5 and 8), suggesting that its present composition reflects oxidation of igneous titanomagnetite at high temperatures during cooling of the pluton (Kay *et al.* 1983). Other magnetite (*i.e.*, Table 5, #7), occurring as a breakdown product of mafic silicates, contains almost no TiO_2 and is compatible with subgreenschist- and greenschist-facies magnetite, which generally contains less than 1% TiO_2 (Abdullah & Atherton 1964). Other grains of magnetite with less than 1% TiO_2 (*e.g.*, Table 5, #6) may also be of metamorphic origin.

In general, the Mn content of ilmenite increases with differentiation index (D.I.) in igneous rocks (Neumann 1974), but only partial correspondence to this trend is found in sandwich and trellis lamellae of ilmenite and composite ilmenite grains (terminology from Haggerty 1976) enclosed in titanomagnetite in the Finger Bay pluton. As shown in Figure 6, ilmenite lamellae in most gabbros contain approximately 3% MnO, and the MnO content increases with differentiation index. However, where ilmenite is associated with titanite, the MnO content of ilmenite may reach 10% in gabbro and almost 16% in quartz monzodiorite. This Mn-rich ilmenite

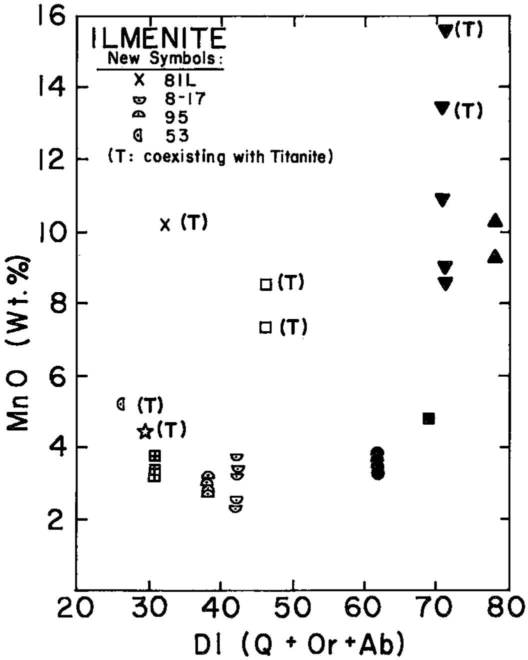

FIG. 6. Whole-rock differentiation index (D.I., normative q + or + ab + ne) *versus* MnO content of ilmenite blebs and lamellae in titanomagnetite hosts in the Finger Bay pluton. Generally, ilmenite becomes more Mn-rich with increasing D.I., but the most Mn-rich ilmenite (symbol followed by T) is associated with titanite. Symbols are as in Figure 2 with the exceptions noted on the figure.

TABLE 5. REPRESENTATIVE COMPOSITIONS OF TITANITE AND OXIDE PHASES, FINGER BAY PLUTON

	Titanite		Ilmenite		Magnetite				Composite	
	1	2	3	4	5	6	7	8	9	10
Sample	ADK27	FB61	ADK27	FB61	FB61	FB90	FB81L	FB81L	FB81L	FB81L
SiO_2	30.72	30.80	0.09	0.02	0.06	0.08	0.42	0.06	3.05	14.25
TiO_2	30.36	35.03	49.99	45.84	4.97	0.65	0.10	7.11	47.46	66.27
Al_2O_3	6.93	2.48	0.08	0.02	0.42	0.38	0.38	2.65	0.30	1.22
Cr_2O_3	0.05	0.00	0.01	0.00	0.02	0.00	0.21	0.09	-	-
Fe_2O_3	1.32	2.09	5.10	13.44	59.48	66.93	68.91	49.95	-	-
FeO	-	-	34.41	27.48	35.02	31.68	31.76	36.49	36.64	3.05
MnO	0.07	0.10	10.18	13.42	1.14	0.09	0.13	0.68	9.26	0.06
MgO	0.03	0.02	0.06	0.02	0.01	0.01	0.33	0.06	0.09	0.01
CaO	28.38	29.15	0.10	0.09	0.00	0.00	0.10	0.05	3.12	13.96
Total	97.86	99.67	100.02	100.33	101.12	99.82	102.34	97.14	99.92	98.82

1) Small grain of titanite after biotite in quartz monzodiorite. 2) Titanite associated with Mn-rich ilmenite enclosed in magnetite in quartz monzodiorite. 3) Ilmenite lamellae in titanomagnetite in quartz monzonite. 4) Ilmenite associated with titanite (#2) enclosed in titanomagnetite (#5) in quartz monzodiorite. 5) Titanomagnetite enclosing titanite (#2) and Mn-rich ilmentite (#4) in quartz monzodiorite. 6) Small magnetite in quartz monzodiorite. 7) Magnetite after olivine (?) in gabbro. 8) Titanomagnetite host related to 'postmagmatic cooling' in gabbro. 9) Bulk composition of Mn-rich ilmenite and titanite replacing ilmenite lamellae in titanomagnetite host (#8) in gabbro (Fig. 6). 10) Bulk composition of Mn-rich ilmenite, titanite, rutile(?), and hematite(?) replacing ilmenite. Same area as #9.

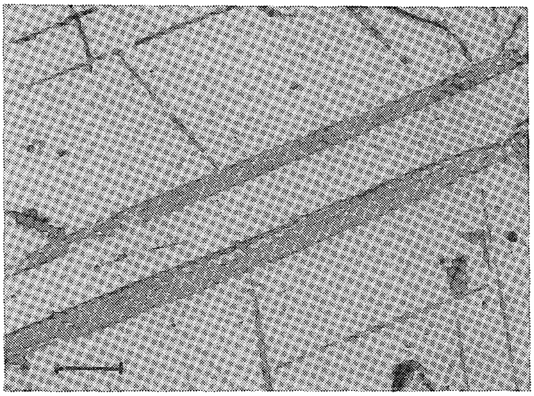

FIG. 7. Back-scattered compositional image of ilmenite lamellae in magnetite in fine-grained layered gabbro FB81L. Trellis lamellae and interiors of large sandwich lamellae are composed of titanite + rutile(?) + hematite(?) (see text). Clear areas along rims of large sandwich lamellae are Mn-rich ilmenite. Host is a titanomagnetite. Length of bar is 10 μm.

appears to result from the concentration of Mn in ilmenite produced in the titanite-forming reaction between ilmenite and Ca-bearing silicate minerals. Titanite replacing ilmenite has previously been described in the prehnite–pumpellyite facies by Zen (1974) and Kuniyoshi & Liou (1976a).

The relation between Mn-rich ilmenite and titanite is illustrated in the composition-mode backscattered-electron image in Figure 7. Former ilmenite lamellae in gabbro FB81L are replaced by Mn-rich ilmenite (Table 4, #9, ilmenite mixed with titanite), titanite, and areas of finely intergrown phases. Large lamellae have Mn-rich ilmenite and titanite on the edges, and a centre of finely intergrown phases; composite analyses (Table 5, #10) suggest that these are titanite, a Ti-rich phase (rutile?) and possibly hematite. Fractures in the centre of the well-developed large lamellae may have served as channels for fluid circulation. The host is a relatively homogeneous titanomagnetite (Table 5, #8). The composition of titanite and Mn-rich ilmenite from irregularly shaped grains included in titanomagnetite in quartz monzodiorite FB61 is given in Table 5, #2 and 4.

The metamorphic assemblage of opaque and associated minerals in the Finger Bay pluton suggests recrystallization at a higher $f(O_2)$ than that prevailing during the last stages of igneous cooling. Temperatures and $f(O_2)$ for the last stages of magmatic cooling and equilibration were obtained from coexisting lamellae of ilmenite and titanomagnetite using the curves of Spencer & Lindsley (1981). Mineral compositions vary within single samples, suggesting that reactions in various parts of the system closed before others and that equilibrium was not achieved at this time. Although values are approximate owing to extrapolations from experimental data, a level of $f(O_2)$ near the Ni–NiO buffer and temperatures ranging from 450 to 700°C are suggested to be the last recorded during magmatic cooling. Metamorphic assemblages (see below) suggest temperatures between 325 and 400°C.

The sporadic occurrence of Mn-rich ilmenite, titanite and hematite in the pluton suggests variable $f(O_2)$ and temperature during metamorphism. If rutile and hematite formed in place of Mn-ilmenite in gabbros such as FB81L (Fig. 7), a value of $f(O_2)$ near or above the Mn-poor ilmenite - rutile - hematite buffer is indicated for the metamorphic event (see Czamanske & Wones 1973) for this sample. Furthermore, the presence of magnetite formed from mafic silicates in FB81L brackets the $f(O_2)$ between the Mn-poor ilmenite - rutile - hematite buffer and the magnetite–hematite buffer. Locally, a level of $f(O_2)$ above the magnetite–hematite buffer is suggested in some highly altered areas by the occurrence of hematite in association with other metamorphic minerals. Variable amounts of oxidation by meteoric water in a hydrothermal system at temperatures of 325–400°C may explain the variable metamorphic alteration of the opaque phases, just as at a larger scale, it explains the patterns of metamorphic alteration in the Finger Bay pluton (see below).

PHYSICAL CONDITIONS OF METAMORPHISM OF THE FINGER BAY PLUTON

Metamorphic assemblages corresponding to the subgreenschist facies and to higher-temperature, postmagmatic conditions occur in the Finger Bay pluton. Grandite- and chlorite-bearing (actinolite unstable) assemblages occur in the least fractured and most unrecrystallized area, whereas more pervasively recrystallized samples from other areas commonly contain chlorite + epidote + actinolite ± prehnite. There is no clear evidence for metamorphic zonation, although the grandite-bearing rocks may suggest lower temperatures than rocks having the chlorite–epidote–actinolite assemblage. Figure 8 shows the temperature and pressure co-ordinates of some metamorphic reactions pertinent to an estimation of the metamorphic conditions in the Finger Bay pluton.

Evidence of disequilibrium makes the determination of equilibrium assemblages difficult. Compositions of metamorphic minerals, particularly chlorite and actinolite, are controlled by the phases that they replace, suggesting that only very local equilibrium has been achieved. Textural evidence suggests that several assemblages of metamorphic minerals occur in the same thin section. For example, in quartz-bearing gabbro FB76–12M, the greenschist assemblage epidote–actinolite–chlorite is observed in

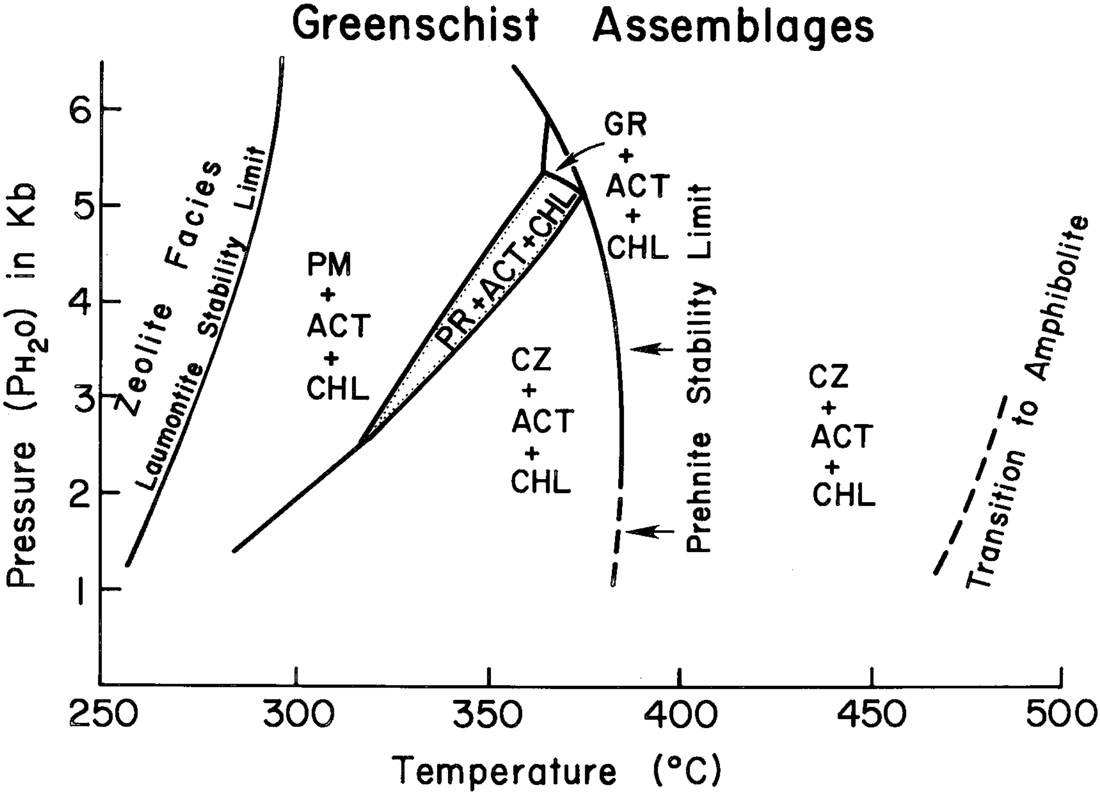

FIG. 8. Phase relations for greenschist mineral-assemblages for basaltic compositions. Prehnite and laumontite stability limits are from Liou (1971a, b). The greenschist-amphibolite transition zone is from Liou *et al.* (1974). Phase relations for act (actinolite, actually tremolite), chl (chlorite), cz (clinozoisite), pm (pumpellyite), pr (prehnite) and gr (grandite) are from the temperature-pressure (P_{fluid}) diagram for the pseudoternary system CaO-Al_2O_3-MgO of Schiffman & Liou (1980). The phase relations assume that all assemblages contain quartz + albite. Only the field of interest for the Finger Bay pluton is shown; the complete diagram is presented as Figure 10 of Schiffman & Liou (1980). The pr + act + chl field, which is critical to determining pressure-temperature conditions in the Finger Bay pluton, is patterned for emphasis. The assemblage cz + pr + chl can also be stable in this field in rocks of slightly different composition. Metamorphism of the Finger Bay pluton is estimated to have involved temperatures between 325 and 400°C and pressures of around 2-3 kbar (see text).

intergranular areas, whereas several millimetres away biotite is breaking down to prehnite, chlorite, titanite and K-feldspar. Actinolite occurs on the margin of the biotite grain and is in contact with both prehnite and chlorite in other former biotite grains in the same sample. Thus, textural evidence suggests that the assemblages epidote–chlorite–actinolite and prehnite–chlorite–actinolite are stable within a few millimetres of each other in the same sample.

Estimates of the maximum temperature and pressure of equilibration for the prehnite-bearing assemblages are dependent on whether Nitsch's (1971) or Schiffman's & Liou's (1980) interpretation of the stability of the assemblage prehnite–actinolite–chlorite (+quartz+fluid) is used. The experimental data of Schiffman & Liou (1980) are preferred; Zen & Thompson (1974) and Schiffman & Liou (1980) have questioned the applicability of Nitsch's (1971) experiments to nature, since they were not buffered for oxygen, and only a few were reversed. Although the phase relations of Schiffman & Liou (1980) are developed for a specific set of assumptions, and deviation from these assumptions could significantly modify the phase relations presented, they provide some perspective on mineral parageneses in rocks of basaltic composition.

Schiffman & Liou (1980) suggested that the mineral assemblage prehnite–chlorite–actinolite (+quartz+albite) is stable between 2 and 5 kbar at temperatures of 325°– 375°C and occurs between the subgreenschist-facies assemblage pumpellyite–actinolite–chlorite and the greenschist assemblage clinozoisite–tremolite–chlorite (Fig. 8). Epidote or pumpellyite (or both) may be found with the prehnite–actinolite–chlorite assemblage in other localities (Schiffman & Liou 1980). However, pumpellyite appears to be present in Barrovian-type metamorphic terranes, but not in areas of relatively high geothermal gradients, as affected the Karmutsen volcanic rocks (Kuniyoshi & Liou 1976a). The absence of pumpellyite in the Finger Bay pluton may indicate that metamorphism also occurred under a relatively high geothermal gradient, similar to that in the Karmutsen area. The occurrence of the epidote–actinolite–chlorite assemblage in the same thin section as the prehnite–chlorite–actinolite assemblage in samples from the Finger Bay pluton may imply that temperatures were near the high-temperature boundary for the prehnite–chlorite–actinolite assemblage (Fig. 8) or, more likely, that local compositions strongly controlled reactions.

A maximum temperature for metamorphism in the Finger Bay pluton in prehnite-absent rocks is suggested by the upper boundary of the greenschist assemblage actinolite–epidote–chlorite–albite. For the Karmutsen mafic volcanic rocks, this boundary was determined experimentally by Liou *et al.* (1974) to be 475°C at 2 kbar. In the Finger Bay pluton, the lack of metamorphic oligoclase and the compositional similarity of metamorphic minerals to those occurring in the prehnite-bearing rocks suggest that the upper temperature limit lies below 475°C and that the lack of prehnite in these rocks is compositionally controlled.

The mineralogy of some of the least recrystallized areas of the Finger Bay pluton suggests subgreenschist metamorphic conditions, as grandite occurs in association with chlorite, and actinolite is either absent or partially replaced by clay minerals. Coombs *et al.* (1977) suggested that in the presence of quartz, grandite will react with chlorite to form actinolite and epidote in the greenschist facies. Although the phase relations of Schiffman & Liou (1980) suggest that pressures of 5 kbar are necessary to form grandite, field relations (Coombs *et al.* 1977) suggest that the grandite + chlorite assemblage may crystallize at pressures as low as 1–2 kbar. Schiffman & Liou (1980) pointed out that the formation of grandite is sensitive to variations of $f(O_2)$, and that this could seriously affect the topology of their suggested phase-relations. Some regions of the surrounding Finger Bay volcanic suite also contain subgreenschist assemblages (Rubenstone & Kay 1980, Rubenstone 1983).

Oxygen-Isotope Data and Metamorphism in the Finger Bay Pluton

Whole-rock oxygen-isotope data can be used to help determine the nature of the fluid associated with metamorphism. In other regions in the Aleutians, such techniques have previously defined the circulating hydrothermal fluid to be meteoric water associated with the intrusion of calc-alkaline plutons (Perfit & Lawrence 1979, Citron 1980). Whole-rock $\delta^{18}O$ values of quartz monzodiorites FB61 and FB90 are 0.54 and 1.39 per mil (relative to SMOW), respectively, suggest that circulating hot meteoric water was also associated with the metamorphism of the Finger Bay pluton. The proposed correlation between fracturing and faulting and the pervasiveness of metamorphism, as well as the common occurrence of a graphic intergrowth of quartz and feldspar, interpreted by Taylor (1971) to be the result of hydrothermal activity, also support this conclusion.

The metamorphism of the Finger Bay pluton is probably associated with hydrothermal systems set up by younger plutons on Adak and Kagalaska islands (Fig. 1). A K–Ar date of 31 Ma on the Finger Bay pluton (B.D. Marsh, pers. comm. in DeLong *et al.* 1978), interpreted to be a metamorphic age (DeLong *et al.* 1978), suggests that at least part of the metamorphism occurred in the thermal aureole of the Hidden Bay pluton (31–33 Ma: Citron *et al.* 1980). The thermal aureoles of other unexposed or partly exposed plutons, such as the undated pluton between Finger Bay and Hidden Bay plutons (Tmg? in Fig. 1a) and the Kagalaska pluton (13.5 Ma: Citron *et al.* 1980) may also have been partly responsible for the metamorphism.

Oxygen-isotope data help to constrain the temperature and pressure conditions of metamorphism in the Finger Bay pluton. However, metamorphic temperatures cannot be uniquely defined from the $\delta^{18}O$ values because alteration is incomplete, and the water–rock ratio and the $\delta^{18}O$ of the local meteoric water are unknown. Following Perfit & Lawrence (1979), metamorphic temperatures of 300–400°C inferred from the metamorphic assemblages are consistent with the oxygen-isotope data if water-to-rock ratios were about one and the local meteoric water had a $\delta^{18}O$ of −10. Since the oxygen-isotope data indicate that the hydrothermal fluid was meteoric (fresh) water, the metamorphism had to occur within the meteoric water-table during an erosional period in the arc. Although not conclusive, this suggests that pressures on the order of 3 kbar or depths of 9 km may be more realistic than pressures of 5 kbar or depths of 15 km.

Only local redistribution of elements occurred during the metamorphism of the Finger Bay pluton, owing either to low amounts of dissolved salts in the circulating meteoric water or to low water-rock ratios (or both). Evidence for only very local redistribution of elements comes from the small amount of albite produced (lack of Na metasomatism), the nonequilibrium nature of the mafic minerals, and the occurrence of nonequilibrium assemblages less than a millimetre apart. In addition, original whole-rock igneous compositions (on the hand-specimen scale) are maintained, as shown by the preservation of igneous distributions of major and minor elements (including mobile elements, as inferred from Rb, K and Ba) and the chemical similarity to the unaltered Plio-Pleistocene to Recent volcanic rocks (Kay *et al.* 1983). The evidence for limited mobility of elements in the Finger Bay pluton and also in the Finger Bay volcanic suite (Rubenstone & Kay 1980) contrasts with the evidence for substantial redistribution of elements in the Tertiary rocks on Attu Island in the western Aleutians, where the metamorphism has apparently resulted from a hydrothermal circulation system involving sea water (Rubenstone *et al.* 1982).

Conclusions

The Finger Bay pluton intruded the Finger Bay volcanic sequence and crystallized prior to the intrusion of the Hidden Bay pluton (31–33 Ma: Citron *et al.* 1980) on Adak Island and the Kagalaska pluton (13.5 Ma: Citron *et al.* 1980) on nearby Kagalaska Island. Significantly, neither of these two later plutons has been metamorphosed. During the 'subsolidus' postmagmatic cooling of the Finger Bay pluton, some actinolite developed, opaque minerals were oxidized and partial alteration of biotite may have occurred. The pluton was then subjected to a period of metamorphism apparently caused by the circulation of hot, meteoric water in hydrothermal systems set up by the intrusion of later plutons. At least part of the metamorphism appears to have occurred 31 Ma ago (DeLong *et al.* 1978), approximately the time of intrusion of the Hidden Bay pluton. The metamorphic effects varied in intensity depending on the volume of water and the density of prior fracturing in the pluton, with the most severe alteration occurring in the more highly fractured parts of the pluton.

The oxygen-isotope data indicate that the Finger Bay pluton was subareal (within the zone of circulating meteoric water) at the time of the metamorphism. Temperatures of 350 – 400°C and pressures near 3 kbar suggested by the metamorphic assemblages for much of the pluton are consistent with temperatures required to produce the near-zero $\delta^{18}O$ values. Some less-fractured and less-altered areas of the pluton and large areas of the Finger Bay volcanic sequence on Adak are of subgreenschist metamorphic grade. No systematic studies have yet been done to examine metamorphic gradients surrounding the Hidden Bay or Kagalaska plutons or on Adak as a whole.

Metamorphism of the Finger Bay pluton appears to be typical of that in other Aleutian 'initial stage' arc rocks on Adak and the surrounding islands, and suggests that the primary cause of metamorphism in the central Aleutian arc is hydrothermal circulation associated with later magmatic activity. Much of this hydrothermal circulation may be associated with meteoric water; no discrete event, such as ridge subduction, is necessary to explain the metamorphism. Work by Rubenstone (1982), Rubenstone *et al.* (1982) and Rubenstone & Kay (unpubl. data) suggests that the 'initial series' rocks in the western Aleutians (*i.e.*, on Attu Island) have different igneous affinities and have been metamorphosed under different conditions.

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