

## PHENGITE-3T IN HIGH-PRESSURE METAMORPHOSED GRANITIC ORTHOGNEISSES, SEWARD PENINSULA, ALASKA

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### ABSTRACT

Orthogneisses of granite, granodiorite and tonalite composition, in the low-grade schist terrane of the Seward Peninsula, Alaska, underwent structurally coherent, progressive regional blueschist-facies metamorphism in the mid-to-late Jurassic, followed by partial re-equilibration during decompression under high-greenschist-facies conditions. The mineral paragenesis of the blueschist facies in granitic orthogneiss is quartz + albite + microcline + phengite + biotite; almandine-rich garnet occurs in tonalitic compositions in the absence of microcline. No blueschist-facies minerals, such as jadeite, ferro-omphacite, or ferro-glaucophane, or obvious pseudomorphs thereof, occur in the orthogneisses. Diagnostic blueschist-facies parageneses are found, in adjacent metapelites and metabasites, in parts of the peninsula. The phengite in orthogneiss is uniaxial (3T) and Si-rich; maximum formula content of Si is 7.2 atoms for phengite in central and southern Seward Peninsula. The Si content of phengite varies considerably within samples, in some cases continuously, in others bimodally. The low-Si frequency maximum at 6.4 atoms is interpreted as being due to the greenschist-facies decompression and is accompanied by growth of biotite. Positive correlations between tschermak, paragonite, and (possibly) trioctahedral mica components in the phengite are ascribed to crystal-chemical interactions. The granitic orthogneiss paragenesis is ideal for phengite geobarometry, but existing calibrations differ considerably from one another. Microcline - low albite pairs give equilibration temperatures in the range 420 - 500°C. One garnet-biotite pair gives 390°C. Accessory minerals are: epidote, allanite, titanite, zircon, apatite, and calcite. The titanite contains between 7 and 27% of  $\text{CaAlSiO}_4\text{F}$ .

**Keywords:** phengite-3T, celadonite-rich muscovite, granitic orthogneiss, blueschist facies, microprobe analysis, phengite geobarometer, two-feldspar geothermometer, blueschist biotite, Seward Peninsula, Alaska.

### SOMMAIRE

Les orthogneiss (en composition: granite, granodiorite et tonalite) des terrains schisteux, à faible degré de métamorphisme, de la péninsule Seward (Alaska), subirent un métamorphisme de facies schistes bleus, régional, progressif et structurellement cohérent, au Jurassique moyen-à-tardif. Il fut suivi d'une ré-équilibration partielle au cours de la décompression dans des conditions de facies schistes verts de haut degré. La paragenèse des minéraux de ce facies dans l'orthogneiss granitique est: quartz + albite + microcline + phengite + biotite. En l'absence de microcline, on observe un grenat riche en almandine dans les compositions

tonalitiques. Aucun des minéraux du facies schistes bleus, tels que jadéite, ferro-omphacite et ferro-glaucophane, ni aucun de leurs pseudomorphes habituels, n'ont été relevés dans les orthogneiss. Dans certaines parties de la péninsule, on trouve des paragenèses diagnostiques du facies schistes bleus dans des métapelites et metabasites contiguës. La phengite de l'orthogneiss est uniaxe (3T) et riche en silicium; le nombre maximum d'atomes Si dans la formule est 7.2, noté pour la phengite du sud et du centre de la péninsule. Cette teneur en Si de la phengite varie considérablement à l'intérieur d'un même échantillon, soit d'une façon continue, soit de deux façons. Le nombre le plus bas d'atomes de Si est 6.4; on l'attribue à la décompression du facies schistes verts et la croissance de biotite l'accompagne. Des corrélations positives entre les composantes tschermak, paragonite et (peut-être) mica trioctaédrique dans la phengite s'expliquent par interactions cristalochimiques. La paragenèse de l'orthogneiss granitique semble idéale pour la géobarométrie de la phengite, mais les calibrations qui ont été faites diffèrent considérablement l'une de l'autre. Les paires microcline - albite ordonnée donnent une température d'équilibration qui varie de 420 à 500°C. Une paire biotite-grenat a donné 390°C. On note, comme minéraux accessoires: épidote, allanite, titanite, zircon, apatite et calcite. La titanite contient de 7 à 27% de  $\text{CaAlSiO}_4\text{F}$ .

(Traduit par la Rédaction)

**Mots-clés:** phengite-3T, muscovite celadonitique, orthogneiss granitique, facies schistes bleus, analyse à la microsonde, géobarométrie (phengite), géothermométrie (deux feldspaths), biotite des schistes bleus, Péninsule Seward, Alaska.

### INTRODUCTION

Most blueschist terranes in western North America are composed of a characteristic assemblage of rock types of ocean-crust provenance: metagreywacke, metashale, metachert, metacarbonate, metadiabase, metabasalt, and serpentinite. The protoliths of the more inboard, Seward Peninsula blueschist terrane are, on the other hand, a continental platform assemblage of sediments, into which granitic rocks and, more abundantly, MgAl- and FeTi-rich gabbros and diabases were emplaced (Forbes *et al.* 1984, Thurston 1985). The presence of granitic orthogneiss in the high-pressure metamorphic rock assemblage invites comparison with the blueschist belts of the Alps, in which pre-Alpine continental basement rocks, along with Mesozoic ophiolites and sediments, were involved in high-pressure metamorphism associated

with continental collision (*e.g.*, Dal Piaz *et al.* 1972, 1983, Compagnoni & Maffeo 1973, Frey *et al.* 1976, Compagnoni 1977, Saliot 1979, Gibbons & Horak 1984). The Seward Peninsula blueschists experienced an erosional uplift, curving clockwise P-T-path (Forbes *et al.* 1984), manifested in a partial greenschist-facies overprint, an additional contrast with subduction-complex (Franciscan-style) blueschists.

The presence of granitic protolith in the Seward Peninsula terrane provides bulk compositions and mineral parageneses that are not commonly encountered in high-pressure metamorphic belts, particularly in North America. The typical orthogneiss paragenesis is: quartz + albite + microcline + phengite + biotite; almandine garnet, epidote, allanite, chlorite and calcite are also present in some cases. Microcline is absent from the more granodioritic and tonalitic bulk-compositions. No jadeite occurs anywhere in the Seward Peninsula, and the granitic orthogneisses themselves do not contain typical high-pressure minerals such as ferro-omphacite, lawsonite, or ferro-glaucophane (but see p. 152). The orthogneiss parageneses provide opportunities for potentially informative attempts at geothermometry and geobarometry, including two-feldspar and garnet-biotite geothermometry, and phengite geobarometry. A review of eclogite-facies meta-acidites has recently been given by Black *et al.* (1986).

Over much of the central and southern parts of the Seward Peninsula (Fig. 1), the metamorphic grade is indicated by the assemblage glaucophane + almandine + epidote in metabasites, and chloritoid + glaucophane + albite + epidote  $\pm$  garnet in pelitic schists of the Nome Group (Forbes *et al.* 1984, Thurston 1985). To the north and southwest, the rocks are provisionally assigned to the greenschist and albite-epidote amphibolite facies, although more work needs to be done to be sure that the absence of sodic amphibole in these places is not caused by the lack of appropriate whole-rock compositions, or a more intense decompression-related recrystallization. The lack of discontinuous reactions, capable of serving as isograds reasonably independent of whole-rock composition, means that regional P and T variations in the Seward Peninsula must be determined by the few reliable mineral geothermometers and geobarometers available in this P-T range. This study is a first attempt in that direction; others, using for example calcite-dolomite geothermometry, are in progress (Patrick *et al.* 1985).

The granitic orthogneisses are also of interest in that they provide a bulk composition favorable for the presence of biotite and the most celadonite-rich phengites under the prevailing metamorphic conditions. It is well known that high-pressure metamorphism restricts the stability of biotite-bearing parageneses, and favors instead those containing

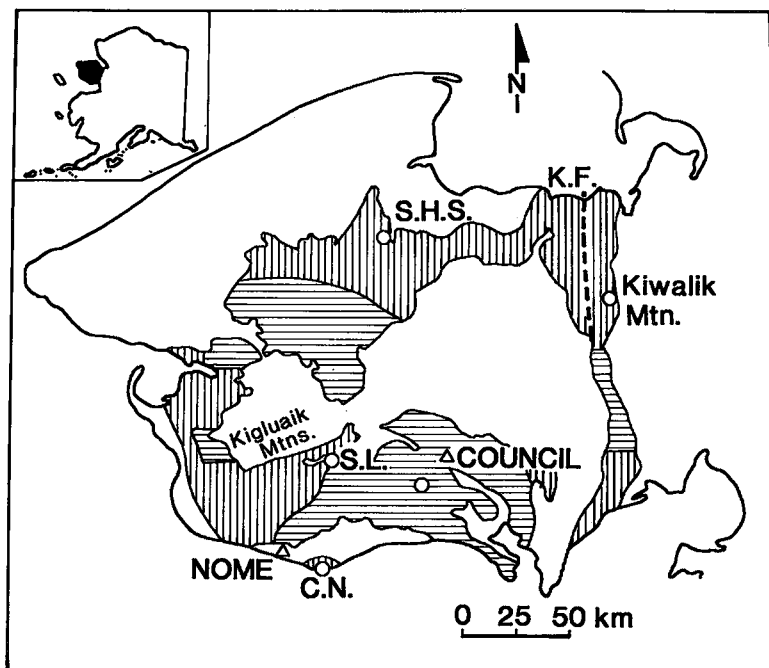


FIG. 1. Distribution of low-grade metamorphic rocks on the Seward Peninsula (shaded areas). Horizontal lines indicate areas of known blueschist-facies parageneses. Circles show localities discussed in the text; S.L. Salmon Lake, C.N. Cape Nome, S.H.S. Serpentine Hot Springs, K.F. Kugruk fault.

TABLE 1. APPROXIMATE MODES OF ORTHOGNEISS SAMPLES

	Sample number	Qtz	Ab	Kfs	Phe	Bt	Grt	Accessories
Cape Nome	AB81-80B	30	30	20	10(3T)	6	0	Ep Aln Ttn Zrn Ap
	AB81-80C	30	25	10	25(3T)	7	0	Aln Ttn Zrn Ap
	SL81-1-1	30	45	5	10(3T)	<5	0	Aln Ttn Zrn
	SL81-5-2	30	30	25	10(3T)	<5	0	Aln Zrn Cal
Salmon Lake	SL81-24-1	45	45	0	5	<5	0	Ep Zrn Py Rt Ap
	SL81-94-1	30	35	20	10(3T+2M <sub>1</sub> )	<5	0	Ep Ttn Zrn Cal
	SL81-127-4	25	50	0	6(3T+2M <sub>1</sub> )	8	8	Ep Ttn Zrn Ap Cal Py
S.H. Springs	SP81-324	25	40*	5	5(2M <sub>1</sub> )	10	0	Ep Ttn Zrn Ap Cal(15) Py
	SP81-113A	45	20*	15	10(3T+2M <sub>1</sub> )	8	0	Aln Ttn Zrn Cal
Kiwalik Mtn.	SP81-221	25	45*	8	10(3T+2M <sub>1</sub> )	7	0	Aln Ttn Zrn Cal

Ep epidote, Aln allanite, Ttn titanite, Zrn zircon, Ap apatite,  
 Cal calcite, Py pyrite, Rt rutile  
 \* oligoclase also present,  
 Polytype of phengite from XRD powder scans (after Bailey 1984, Table 2)

phengitic white mica (*e.g.*, Ernst 1963, Velde 1965, Guidotti & Sassi 1976). As a result, biotite in the blueschist facies occurs only in lithologies such as metagranites and certain calc-silicates and amphibolites that are low in Al relative to other components. Furthermore, the less common, uniaxial trigonal (3T) polytype of K-rich white mica is typical of the Seward Peninsula orthogneiss phengites, in contrast to the 2M<sub>1</sub> polytype in the other lithologies.

#### OCCURRENCE

Bodies of metamorphosed granitic, granodioritic, and tonalitic rock enclosed in metasediments of the Nome Group (Till 1984) range from structurally conformable layers only a few metres in thickness (Hummel 1962, Thurston 1985) to large, mappable masses several km across; the Cape Nome and Kiwalik Mountain orthogneisses (Fig. 1) fall into the latter category (Hudson 1977). The orthogneisses in the Salmon Lake area (Pollock 1982, Thurston 1985) are reliably assigned to the blueschist facies on the basis

of interlayered assemblages and related structures, whereas those of Cape Nome are only indirectly assigned to it (enclosed mafic rocks contain albite, sodic actinolite, epidote and chlorite, but glaucophane is common to the north). The orthogneiss at Serpentine Hot Springs is fault-bounded (Hudson 1979). The Kiwalik Mountain orthogneiss is associated with rocks in the high greenschist to low amphibolite facies east of the Kugruk fault zone (Dumoulin & Till 1985). Original ages of emplacement of the granitic rocks range at least from very late Precambrian (Armstrong *et al.* 1986) to Devonian (Till *et al.* 1983).

#### PETROGRAPHY

Textures are medium- to coarse-grained, gneissic to blastomylonitic in the Salmon Lake area, and crystalloblastic elsewhere. K-feldspar and albite typically range from ½ to 6 mm in cross-section, whereas quartz, phengite and biotite are finer grained. The micas are segregated into irregular, deformed mica-

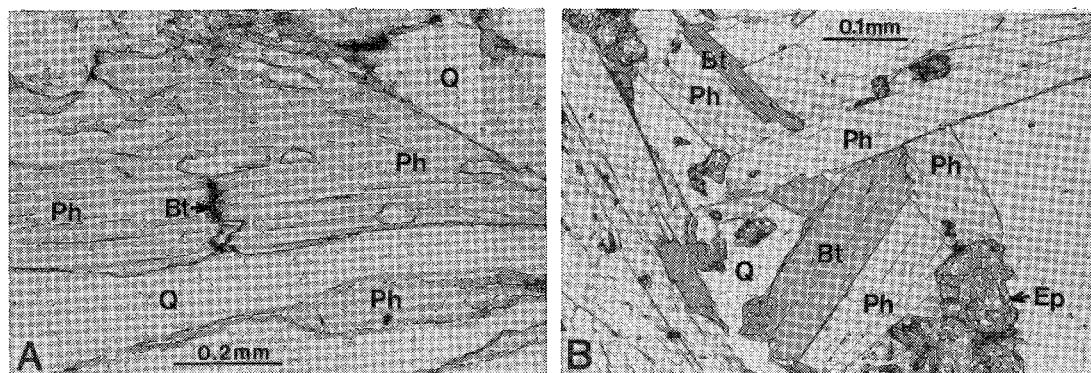


FIG. 2. Granitic orthogneiss samples SL81-94-1(A) and AB81-80B(B). Plane light, phengite (Ph), biotite (Bt), quartz (Q), epidote (Ep).

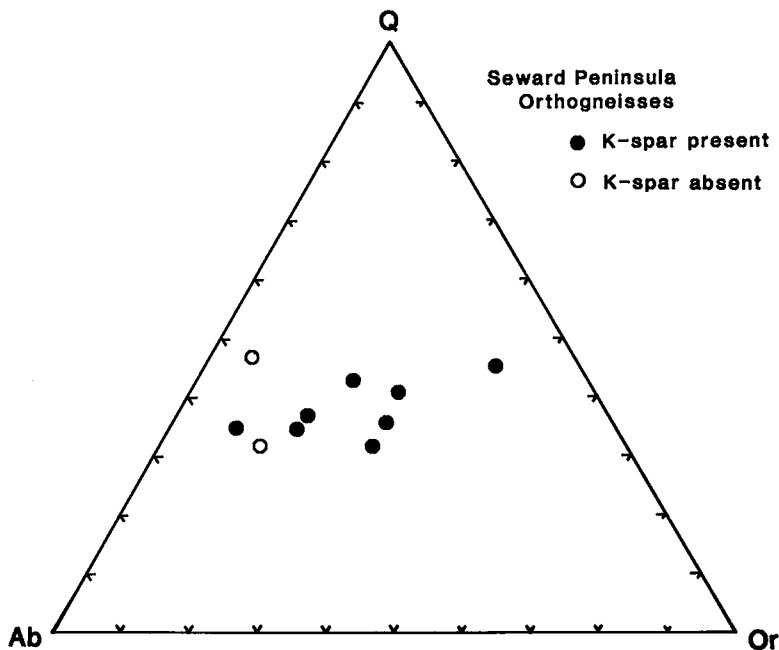


FIG. 3. Whole-rock compositions of analyzed orthogneisses in terms of normative Q, Ab and Or (Mol%).

rich layers in the Salmon Lake examples (Fig. 2A), and evenly distributed in a largely postkinematic fabric in the Cape Nome samples (Fig. 2B). The micas in the Serpentine Hot Springs and Kiwalik Mountain samples form a crystallization schistosity.

The phengite is pale green in thin section. Conoscopic observation of  $2V$  in thin sections cut

parallel to schistosity show that coarse grains of phengite in the microcline-bearing orthogneisses from Salmon Lake and Cape Nome are generally uniaxial (more than 90% of measurements). Microcline-free specimens from Salmon Lake contain biaxial phengite with a range in its optic-axial angle; in addition, about 15% of the phengite in

TABLE 2. WHOLE-ROCK COMPOSITIONS OF ORTHOGNEISSES

Weight percent	Cape Nome			Salmon Lake				S.H.S.		Kiwalik Mtn.	
	AB 81-80B	AB 81-80C	SL 81-1-1	SL 81-5-2	SL 81-24-1	SL 81-94-1	SL 81-127-4	SP 81-324	SP 81-113A	SP 81-221	
SiO <sub>2</sub>	73.78	75.20	75.55	76.30	79.70	72.19	70.53	72.29	72.48	72.93	
TiO <sub>2</sub>	.25	.26	.21	.18	.14	.29	.55	.27	.23	.24	
Al <sub>2</sub> O <sub>3</sub>	13.40	13.63	13.22	12.94	11.43	13.91	14.29	13.69	13.12	13.79	
*FeO	1.79	1.94	1.55	1.57	1.16	2.26	3.46	1.79	2.08	1.46	
MnO	.04	.03	.00	.01	.00	.03	.07	.02	.04	.04	
MgO	.35	.81	.47	.23	.59	.39	.93	.57	.39	.35	
CaO	.87	.24	.11	.28	.00	.86	1.65	1.82	1.77	1.75	
Na <sub>2</sub> O	3.50	3.59	6.11	3.16	5.23	3.89	5.16	4.41	1.19	4.75	
K <sub>2</sub> O	4.69	3.44	1.53	4.76	.96	4.67	1.98	2.68	5.91	2.71	
L.O.I.	.94	1.43	.59	1.07	.73	.73	1.57	2.41	2.39	1.97	
	99.61	100.57	99.34	100.50	99.94	99.22	100.19	99.95	99.60	99.99	
Mol. Norm											
il	.4	.4	.3	.3	.2	.4	.8	.4	.3	.3	
or	28.3	20.7	9.1	28.7	5.7	28.2	11.9	16.3	36.9	16.4	
ab	32.2	32.9	55.2	28.9	47.6	35.7	47.0	40.7	11.3	43.6	
an	4.1	.9	.2	1.1		4.0	8.0	9.0	8.9	8.5	
di										.1	
hy	3.5	5.0	3.4	2.9	3.3	4.3	7.4	4.1	4.3	3.0	
co	1.2	4.1	1.6	2.5	2.1	1.1	.9	.4	1.9		
qz	30.2	36.0	30.0	35.6	41.3	26.2	24.1	29.0	36.3	28.1	

\*total iron P<sub>2</sub>O<sub>5</sub> less than 0.05% throughout. Analyst: A. J. Irving.

SL81-127-4 is uniaxial. Phengite in microcline-bearing orthogneiss from Kiwalik Mountain is predominantly uniaxial (more than 80%), whereas phengite from the Serpentine Hot Springs sample is entirely biaxial. These findings are in agreement with polytypes determined in XRD scans of phengite separates (Table 1).

Biotite (including secondary chlorite) is minor in amount in the Salmon Lake and Cape Nome samples, and more abundant in those from Kiwalik Mountain and Serpentine Hot Springs. K-feldspar is patchily grid-twinned. The albite is moderately albite-twinned. Chlorite appears in almost all cases to be retrograde after biotite. Common accessory minerals are: epidote, allanite, calcite, apatite, zircon (idiomorphic; surrounded by pleochroic haloes in both biotite and phengite), titanite and pyrite (Table 1). With the exception of oligoclase in the Kiwalik Mountain and Serpentine Hot Springs samples, all the principal minerals are judged on the basis of their textures and compositions to be metamorphic in origin.

#### WHOLE-ROCK COMPOSITION

Ten of the samples whose mineralogy has been investigated in detail were analyzed by inductively coupled argon-plasma optical emission spectrometry (ICP). Major-element contents and norms (Table 2) indicate that protoliths are compositionally equivalent to tonalite, granodiorite and granite (Fig. 3). Calcium, iron, titanium and manganese correlate inversely with %  $\text{SiO}_2$ , although it is clear that our samples do not represent a formerly interrelated group of igneous rocks. The ratio  $\text{K}_2\text{O}/\text{Na}_2\text{O}$  varies widely (in comparison to  $\text{K}_2\text{O} + \text{Na}_2\text{O}$ ) and is perhaps for some samples indicative of exchange of alkalis prior to or during metamorphism. In the AKF diagram (Fig. 4), microcline-free samples plot outside the triangle biotite-phengite-microcline.

#### COEXISTING FELDSPARS

Microprobe analyses (Table 3) show that the feldspars in the granitic orthogneisses coexist across a

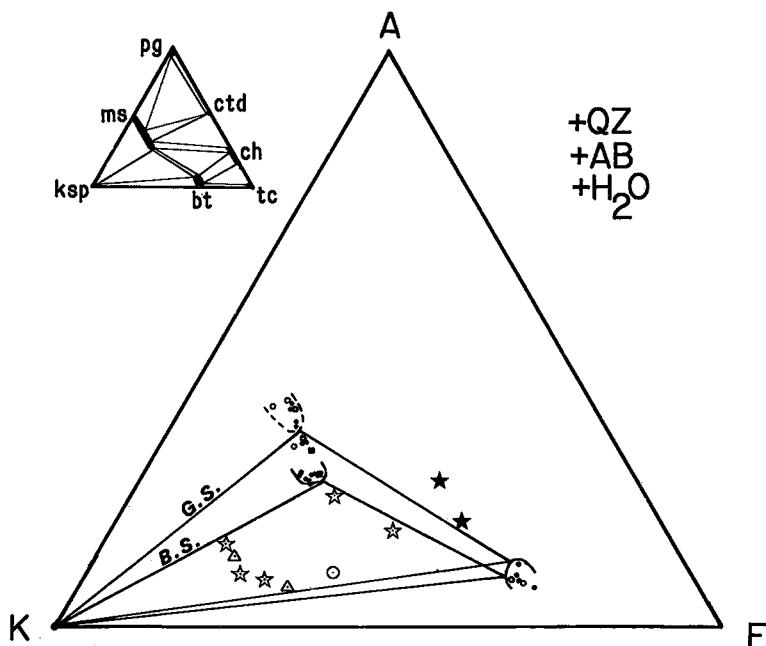


FIG. 4. AKF diagram for Seward Peninsula orthogneisses.  $A = \text{Al}_2\text{O}_3 + \text{Fe}_2\text{O}_3 - \text{K}_2\text{O} - \text{Na}_2\text{O}$ ,  $K = \text{K}_2\text{O}$ ,  $F = \text{FeO} + \text{MgO} + \text{MnO}$ . Whole-rock compositions plotted as large symbols: stars are specimens from Salmon Lake and Cape Nome, filled stars are orthogneisses lacking K-feldspar; open circle from Serpentine Hot Springs; open triangles from Kiwalik Mountain. Phengite and biotite compositions are plotted as small symbols, using data from Tables 4 or 5; dots are from Salmon Lake and Cape Nome; circles from Serpentine Hot Springs and Kiwalik Mountain; filled square is amphibolite from near Council. Dashed line encloses greenschist-facies (G.S.) phengite compositions; continuous line encloses blueschist-facies (B.S.) phengite compositions. Inset: schematic diagram for epidote-zone blueschist facies; garnet and glaucophane are possible substitutes for chlorite, depending on Fe/Mg ratio, T and P.

TABLE 3. AVERAGE COMPOSITION OF FELDSPAR (ELECTRON-MICROPROBE DATA)

	Cape Nome		Salmon Lake				Serp. H.S.	Kiwalik Mt.		Council		
	AB81-80B	AB81-80C	SL81-1-1	SL81-5-2	SL81-24-1	SL81-94-1	SL81-127-4	SP81-324-113A	SP81-221	AB80-163-3		
<b>Plagioclase</b>												
Mol. % An	0.3	0.2	0.1	0.1	0.1	0.2	0.2	0.4	0.1	0.5	0.9	29.1
Or	0.6	0.7	0.4	0.6	0.4	0.5	0.4	0.6	0.7	0.5	0.9	1.6
Ab	99.1	99.1	99.5	99.3	99.5	99.3	99.4	99.0	99.2	99.0	98.2	69.4
<b>K-feldspar</b>												
Mol. % An	0.0	0.1	0.0	0.0		0.0		0.1	0.1	0.2	0.4	
Or	94.4	94.1	94.6	95.5		94.7		94.2	91.9	86.5	90.9	
Ab	5.6	5.6	5.2	4.1	absent	5.0	absent	4.9	7.6	7.1	8.7	
Ce	0.0	0.2	0.2	0.4		0.3		0.8	0.6	6.2		
T(°C)*	454	454	445	421	-	442	-	440	489	488	500	

\*Temperature based on Whitney & Stormer (1977) for P = 10 kbars

AB80-163-3 is an amphibolite; all others are acidic orthogneisses. Relict plagioclase in orthogneisses excluded. Microprobe standards: Tiburon albite, Miyake anorthite, orthoclase OR-1, synthetic celestine.

The feldspar mineralogy of samples AB81-80B and SL81-5-2 has been determined from the cell parameters (Cinlar-Hagg camera, CuK $\alpha$  radiation). The K-feldspar in both cases is well-ordered microcline ( $\gamma$ : 0.98, 1.00, respectively); composition, based on unit-cell volume, is 94.6 and 97.7 mol.% Or. The albite also is well ordered ( $\gamma$ : 0.100, 0.99, respectively); composition, based on the  $\beta^* - \gamma^*$  plot, is 100 and 99 mol.% Ab, 0 and 1 mol.% An, respectively).

highly asymmetric binary solvus; in mol.%, albite defines the range  $An_{0.1-0.5}Ab_{99.0-99.5}Or_{0.4-0.7}$ , and microcline the range  $An_{0-0.2}Ab_{4.1-7.6}Or_{86.3-95.5}Ce_{6.2-0}$ . Assuming that samples AB81-80B and SL81-5-2 are typical, the structural states of both feldspar species are within 2% of being fully ordered (Table 3). Apparent equilibration temperatures, assuming a pressure of 10 kilobars and ideal low albite, have been calculated (Table 3) after the formulation of Whitney & Stormer (1977, equation 8). Temperatures for the Salmon Lake area, Cape Nome, and Serpentine Hot Springs overlap within probable uncertainty, in the range 421–454°C. Those for Kiwalik

Mountain are higher, consistent with petrographic findings on the rocks east of the Kugruk fault zone (Dumoulin & Till 1985), although the difference narrows when lower pressures are adopted for Kiwalik Mountain (see below). The geothermometer for low albite - microcline pairs contains no solution for albite activities in the vicinity of the peristerite miscibility gap, and is for use at temperatures below the range of experimental calibration (Bachinski & Muller 1971). As is the case for a good many geothermometers in current use, relative temperatures can exploit their high precision, whereas absolute temperature estimates are subject to unknown, perhaps

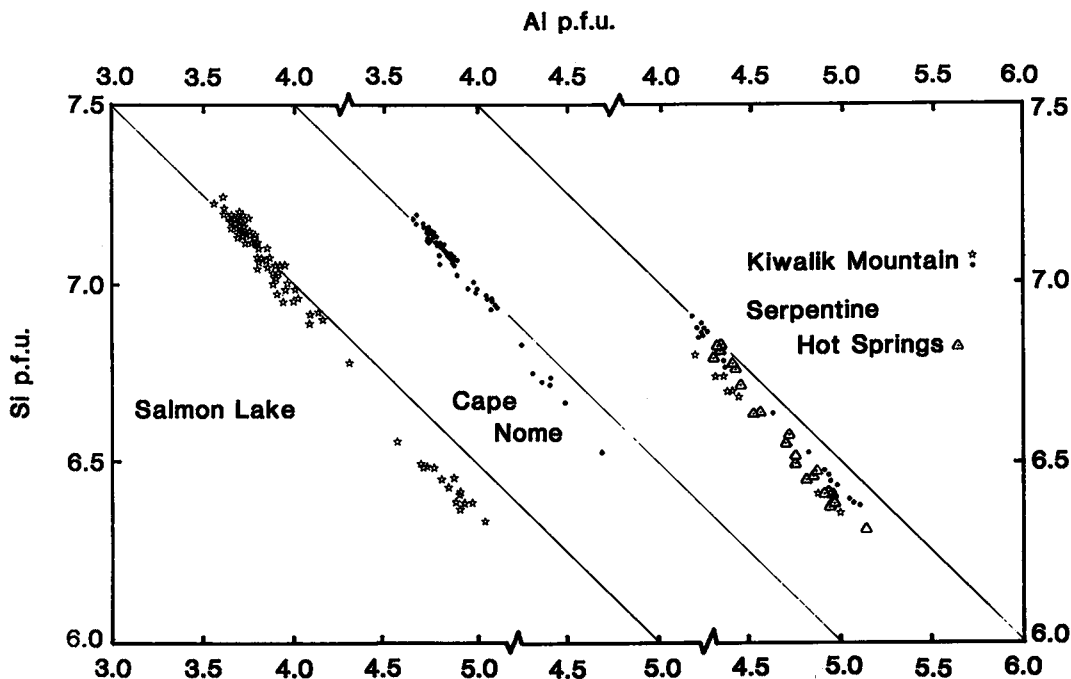


FIG. 5. Al versus Si (per formula unit) in microprobe spot-analyses of orthogneiss phengite; 45° line corresponds to the ideal muscovite-Al-celadonite join, reflecting only tschermak substitution.

large, uncertainties. In the present case, agreement with calcite-dolomite temperatures (Patrick *et al.* 1985) is good (*cf.* Nesbitt & Essene 1982). We believe the two-feldspar temperatures represent an estimate of the highest temperature reached during the high-pressure metamorphic cycle.

Oligoclase was found in the Kiwalik Mountain and Serpentine Hot Springs samples. It is compositionally heterogeneous, ranging from  $An_{2.4}$  to  $An_{19.1}$  at Kiwalik Mountain, and from  $An_{6.7}$  to  $An_{18.4}$  at Serpentine Hot Springs. The presence, in addition, of abundant, nearly anorthite-free albite in these samples suggests that the more calcic compositions are remnants of the primary plagioclase and are not the result of a jump across the peristerite miscibility gap during progressive metamorphism. The K-feldspar in sample SP81-221 was found to be very rich in BaO (3.3 wt. %, equivalent to 6.2 mol. % celsian component). An unusual sodic-actinolite amphibolite with high-Si phengite (see below) from southwest of Council (Fig. 1) was found to contain K-feldspar ( $An_{0.4}Ab_{8.7}Or_{90.9}$ ), oligoclase ( $An_{29.1}Ab_{69.4}Or_{1.6}$ ), and albite ( $An_{0.9}Ab_{98.2}Or_{0.9}$ ). At 10 kilobars, the alkali feldspar pair corresponds to 500°C (assuming  $a_{Ab} = X_{Ab}$ ). This somewhat higher temperature is consistent with the presence of metamorphic oligoclase, although its composition is too calcic to be in equilibrium with the albite according to Maruyama *et al.* (1982).

#### PHENGITE

The composition of phengite (Table 4) in granitic orthogneisses from the Seward Peninsula is a function of mineral paragenesis (presence or absence of K-feldspar) and geographic location. Because of the sensitivity of the extent of tschermak substitution to environmental conditions, it is the compositional parameter of greatest interest; it may be expressed as the number of Si atoms per formula unit or the percentage of the celadonite component  $K_2(Mg,Fe)_2(Al,Fe)_2Si_8O_{20}(OH)_4$  in binary solution with muscovite  $(K,Na)_2Al_4(Al_2Si_6)O_{20}(OH)_4$ . Silica content is the preferred measure of tschermak substitutions (Miyashiro & Shido 1985, p. 451-453) because Al and Mg + Fe are affected by other substitutions in white micas, *viz.*  $FeAl_{-1}$ ,  $Al_2\Box M_{-3}$ , and  $MTiAl_2$  ( $M$  is Mg and Fe). All microprobe data for phengite in this paper have been recast to an anhydrous formula containing 22 oxygen atoms (Table 4), with the assumption that all iron is in the ferrous state. Based on analyses of the coexisting biotite (see below), and comparison with a literature value of 0.37 for  $K_D$  for the partitioning of ferrous iron and magnesium (Frey *et al.* 1976, Fig. 8), we estimate the proportion of ferrous to total iron in the phengite in orthogneiss to range from 0.43 to 0.90, averaging 0.67. Titrimetric and ICP analysis of a phengite

separate from sample SL 81-5-2 yielded a ferrous to total iron ratio of 0.79. As a result, the formula content of Si, as tabulated, plotted, and discussed below, is on average about 0.04 atoms too high. Normalization of phengite analyses to 12 octahedral plus tetrahedral cations would exclude the presence of a trioctahedral mica component, which, as we shall see, becomes significant in lower-Si phengite compositions in the orthogneisses. In Table 4, spot analytical data of high-Si and low-Si phengites (see below) within individual samples have been averaged separately, in order to illustrate the differences for all the analyzed constituents. This is not meant to imply that all the orthogneisses have bimodal phengite populations.

At given P and T, phengite with maximum celadonite content can be expected in the paragenesis phengite + biotite + K-feldspar + quartz + fluid (*e.g.*, Mather 1970; also Fig. 4). It is therefore not surprising that the granitic orthogneisses from the Seward Peninsula contain unusually Si-rich phengite. The Salmon Lake orthogneisses containing microcline have phengite with 6.4 to 7.2 atoms Si (20% to 60% celadonite component), and those lacking microcline have phengite with 6.3 to 7.1 atoms Si (15% to 55% celadonite) (Figs. 5, 6). The Cape Nome microcline-bearing orthogneisses have similarly Si-rich phengite, but fewer cases of lower Si-values (Fig. 5). The microcline-bearing gneisses from Serpentine Hot Springs and Kiwalik Mountain contain less Si-rich phengite (maximum 6.9 atoms Si or 45% celadonite component) and some phengite compositions in the range 6.3 to 6.5 Si atoms, similar to those from the Salmon Lake area. Phengite containing even more celadonite component has been described from blueschist- and eclogite-facies rocks (Smith & Zientek 1979, Brown *et al.* 1978, Chopin & Maluski 1980, Chopin 1981, Saliot & Velde 1982, Sorensen 1986); in most cases these are not from K-feldspar + biotite parageneses. Black *et al.* (1986) described unusually high-Si phengite (7.1 atoms Si) in omphacite- and K-feldspar-bearing acidic metavolcanic rocks from eclogite-facies rocks in New Caledonia.

The Si-contents of phengite from Salmon Lake are bimodal (Fig. 6). The lower-Si compositions are associated with the ends of large, bent high-Si flakes, typically close to small amounts of late-stage green-brown biotite (Fig. 2A), or with rims of sericitic white mica, much as described by Saliot & Velde (1982) in metagranite from the French Alps. Our two generations of phengite are a close match to Si-values (7.2 and 6.47) recorded by Saliot (1979) from the jadeite-bearing Bonneval Gneiss. Texturally distinct generations of phengite, with celadonite contents declining with geological time, have been recognized in a range lithologies in high-pressure rocks from the Gran Paradiso by Chopin & Maluski (1980) and

from the Sesia Zone by Lardeaux *et al.* (1983). The change from high to low contents of celadonite component in the phengite from Salmon Lake is interpreted to be a record of the geological evolution of the terrane from high P/T to low P/T conditions, for which there is abundant independent petrographic evidence (Thurston 1985). The Cape Nome and Kiwalik Mountain phengites, which show less well-defined bimodal Si-contents, are interpreted in the same way. Preliminary work on phengite in pelitic schists in the Seward Peninsula has revealed

similarly bimodal compositions (*e.g.*, Si = 6.7 and 6.2).

Examples of phengite in the high P/T glaucophane-bearing pelitic schists and metabasites in the Seward Peninsula fall in the Si-range 6.2 to 7.0 (unpublished); these are primarily the  $2M_1$  variety ( $2V$  about  $40^\circ$ ). Therefore, we attribute the occurrence of the  $3T$  uniaxial polytype in the orthogneisses to the large celadonite content. Natural celadonite (Si ~ 8.0,  $Fe^{3+} > Al$ ), it should be noted, conforms to the  $1M$  type (Wise & Eugster 1964,

TABLE 4. AVERAGE COMPOSITION OF PHENGITE (ELECTRON-MICROPROBE DATA)

Weight percent	Cape Nome						Salmon Lake					
	AB81-80B		AB81-80C		SL81-1-1		SL81-5-2		SL81-24-1		SL81-94-1	
No. of spots	21	3	23	3	7	3	20	6	11	7	22	3
SiO <sub>2</sub>	52.02	49.20	52.41	48.85	52.36	47.13	52.10	47.12	51.63	46.93	52.14	47.41
TiO <sub>2</sub>	.13	.20	.11	.17	.14	.22	.13	.25	.22	.21	.10	.14
Al <sub>2</sub> O <sub>3</sub>	23.65	26.88	23.67	27.97	23.33	29.71	23.48	29.93	25.03	30.01	23.01	28.76
FeO*	6.77	6.04	5.28	4.46	5.14	4.65	7.06	5.39	3.53	3.90	6.77	5.56
MgO	1.68	1.70	2.62	2.24	2.92	1.79	1.49	1.14	3.52	2.09	2.01	1.55
MnO	.08	.07	.06	.06	.06	.03	.08	.01	.00	.00	.06	.06
CaO	.00	.02	.00	.00	.00	.00	.00	.00	n.d.	n.d.	.01	.00
Na <sub>2</sub> O	.01	.07	.01	.13	.01	.08	.01	.13	.09	.19	.02	.13
K <sub>2</sub> O	11.21	10.73	11.13	11.05	11.33	11.13	10.94	10.81	11.17	11.20	10.92	11.11
BaO	.07	.07	.05	.05	.02	.02	.08	.08	.16	.16	.09	.09
F	.41	.34	.22	.15	.52	.25	.81	.49	.49	.30	.55	.33
Cl	.01	.01	.01	.01	.00	.00	.00	.00	.00	.00	.00	.00
H <sub>2</sub> O**	4.20	4.22	4.32	4.34	4.17	4.26	4.01	4.15	4.22	4.24	4.12	4.21
-O(=F,Cl)	.17	.14	.09	.06	.22	.10	.34	.21	.21	.13	.23	.16
	100.07	99.41	99.80	99.42	99.78	99.17	99.85	99.29	99.85	99.10	99.57	99.21
Cations/22 oxygen atoms												
Si	7.088	5.730	7.103	6.645	7.105	6.445	7.114	6.447	6.955	6.413	7.131	6.513
Ti	.013	.021	.011	.017	.014	.023	.013	.026	.022	.022	.010	.015
Al	3.797	4.334	3.780	4.484	3.730	4.788	3.778	4.826	3.974	4.833	3.710	4.658
Fe*	.772	.691	.598	.508	.584	.532	.807	.616	.397	.446	.775	.638
Mg	.341	.346	.530	.454	.592	.365	.303	.234	.706	.426	.410	.317
Mn	.009	.008	.007	.007	.007	.003	.009	.001	.000	.000	.007	.007
Na	.003	.019	.003	.034	.003	.021	.003	.035	.024	.051	.005	.035
K	1.950	1.873	1.924	1.918	1.961	1.941	1.907	1.887	1.919	1.952	1.906	1.946
Ba	.004	.004	.003	.003	.001	.001	.004	.004	.008	.009	.005	.005
F	.176	.148	.095	.065	.224	.109	.332	.213	.209	.130	.239	.144
Cl	.002	.002	.002	.002	.000	.000	.000	.000	.000	.000	.000	.000
OH**	3.822	3.850	3.903	3.933	3.776	3.891	3.648	3.787	3.791	3.870	3.761	3.856
Salmon Lake Serp. Hot Sps. Kiwalik Mountain Council												
Weight percent	SL81-127-4		SP81-324		SP81-113A		SP81-221		AB80-163-3			
No. of spots	6	2	8	13	8	8	10	4	4	6		
SiO <sub>2</sub>	52.23	46.61	50.35	47.10	50.90	47.48	48.75	46.34	53.02	50.86		
TiO <sub>2</sub>	.21	.18	.26	.50	.28	.47	.86	.76	.23	.26		
Al <sub>2</sub> O <sub>3</sub>	24.45	30.85	27.69	30.52	26.59	31.10	26.64	30.82	24.20	26.35		
FeO*	4.63	4.63	3.38	4.11	5.15	4.00	5.30	4.44	2.97	2.96		
MgO	3.12	1.50	2.79	1.72	1.83	1.06	2.04	1.29	3.98	3.45		
MnO	.00	.00	.00	.00	.00	.00	.03	.03	.00	.00		
CaO	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.		
Na <sub>2</sub> O	.03	.14	.16	.24	.14	.23	.10	.10	.06	.14		
K <sub>2</sub> O	11.07	10.88	10.85	10.84	11.15	11.35	10.56	10.72	10.68	10.72		
BaO	.17	.17	.22	.22	.09	.09	1.20	1.20	n.d.	n.d.		
F	.24	.12	.07	.05	.37	.27	.19	.12	.06	.04		
Cl	.00	.00	.01	.01	.00	.00	.00	.00	.00	.00		
H <sub>2</sub> O**	4.35	4.35	4.44	4.39	4.28	4.32	4.29	4.34	4.45	4.43		
-O(=F,Cl)	.10	.05	.03	.02	.16	.11	.08	.05	.03	.02		
	100.40	99.38	100.17	99.68	100.48	100.26	99.88	100.11	99.62	99.19		
Cations/22 oxygen atoms												
Si	7.019	6.358	6.746	6.391	6.847	6.405	6.674	6.317	7.093	6.852		
Ti	.021	.019	.026	.051	.028	.048	.089	.078	.023	.026		
Al	3.873	4.960	4.372	4.880	4.216	4.945	4.299	4.952	3.815	4.184		
Fe*	5.221	.528	.378	.466	.580	.451	.606	.506	.333	.334		
Mg	.626	.304	.558	.348	.367	.214	.416	.261	.794	.692		
Mn	.000	.000	.000	.000	.000	.000	.004	.003	.000	.000		
Na	.008	.037	.042	.063	.037	.060	.027	.027	.016	.037		
K	1.899	1.894	1.855	1.877	1.912	1.953	1.845	1.865	1.823	1.843		
Ba	.009	.009	.012	.012	.005	.005	.064	.064	.000	.000		
F	.103	.052	.030	.022	.158	.116	.083	.052	.026	.017		
Cl	.000	.000	.002	.002	.000	.000	.000	.000	.000	.000		
OH**	3.897	3.948	3.968	3.976	3.842	3.884	3.917	3.948	3.974	3.983		

\* total iron, \*\* calculated H<sub>2</sub>O, n.d. not determined. The two columns for each sample are averages of the high- and low-Si populations. Microprobe standards: muscovite M, biotite 3, Nuevo garnet, synthetic celadonite, Wilberforce apatite, synthetic F-phlogopite.



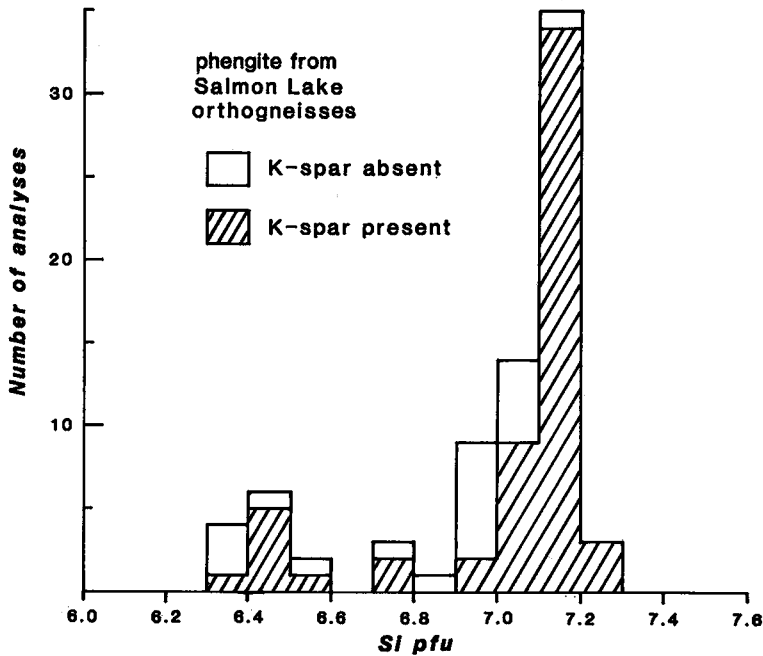


FIG. 6. Histogram of Si-contents of orthogneiss phengite from Salmon Lake.

Buckley *et al.* 1978). The compositional variability of phengite within individual samples provides a ready explanation for the mixture of polytypes observed in rocks that have evolved from blueschist- to greenschist-facies conditions (*e.g.*, Altherr *et al.* 1979, Frey *et al.* 1983, Stockhert 1985; Table 1). Only two of our orthogneiss samples from the Salmon

Lake and Cape Nome areas show  $2M_1$  reflections in addition to  $3T$  reflections, whereas  $2M_1$  reflections were found in all three samples from Kiwalik Mountain and Serpentine Hot Springs (Table 1).

The variation in Si-content along the tschermak vector  $Al_2(MSi)_{-1}$  is accompanied by small but consistent systematic changes in trioctahedral mica

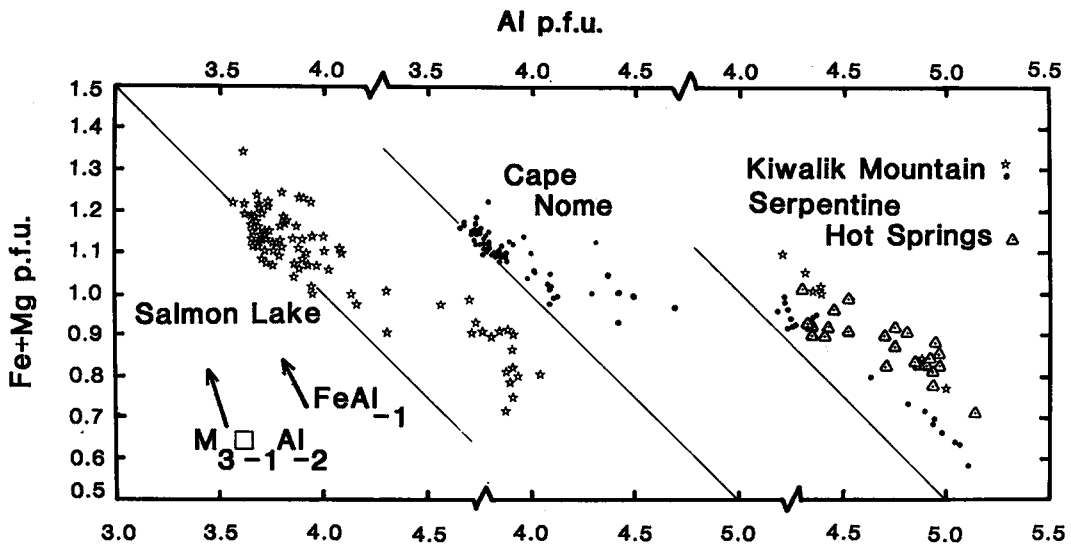


FIG. 7. Fe+Mg versus Al (per formula unit) in microprobe spot-analyses of orthogneiss phengite; 45° line corresponds to ideal muscovite-Al-celadonite join. Vectors for trioctahedral mica and ferrimuscovite solid-solution are shown by arrows.

## PHENGITE Salmon Lake Orthogneisses

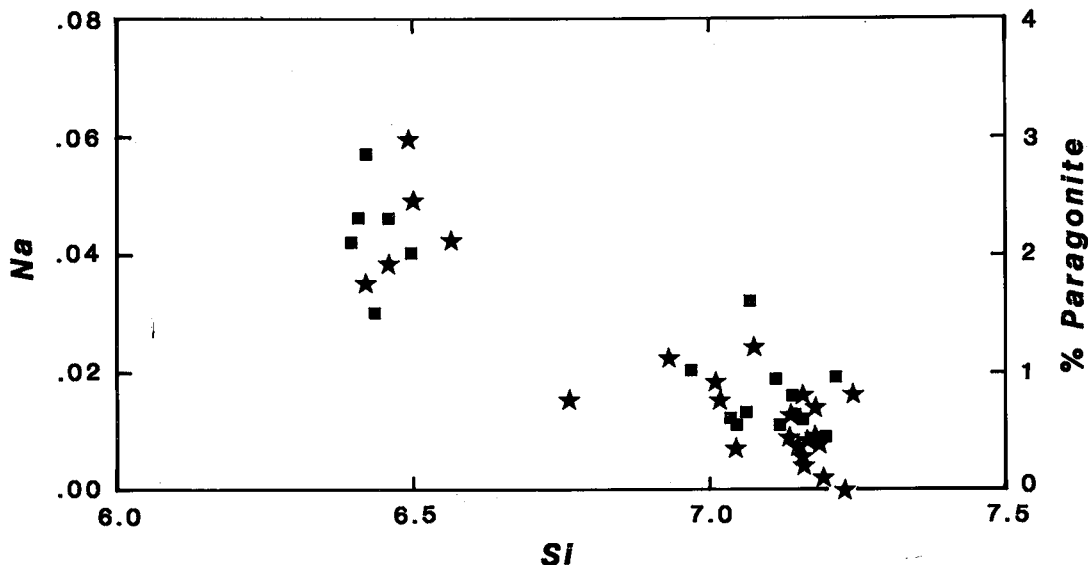
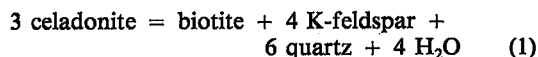


FIG. 8. Na versus Si (per formula unit) in microprobe spot-analyses of orthogneiss phengite from Salmon Lake. Squares, SL81-5-2; stars, SL81-94-1.

( $M_3\Box_{-1}Al_2$ ), paragonite ( $NaK_{-1}$ ) and fluormica  $F(OH)_{-1}$  components. For example, the Salmon Lake phengite compositions show an increase in cation content (per 22 oxygen atoms) from 13.9 to 14.15 as Si declines from 7.2 to 6.4. This shows that the departure from the tschermak substitution illustrated in Figures 5 and 7 is due to trioctahedral mica substitution and not to ferrimuscovite ( $FeAl_{-1}$ ). A compilation of analytical results with separate determinations of ferrous and ferric iron (Frey *et al.* 1983, Fig. 4; Guidotti 1984, Fig. 7) shows that low-Si muscovite is in general poorer in ferrimuscovite component than high-Si muscovite (phengite), in keeping with the typical  $Fe^{3+}$ -rich composition of natural celadonite (Wise & Eugster 1964). The high-Si phengite in the Seward Peninsula has an Na-content that is barely detectable by routine microprobe analysis (Table 4). The inverse correlation of Na with the tschermak component (Fig. 8) has been widely observed (*e.g.*, Cimmino & Messiga 1979, Fig. 2; Chopin & Maluski 1980, Fig. 2; Schliestedt 1980, Fig. 46; Enami 1983, Fig. 5; Grambling 1984, Fig. 5; Guidotti 1984, Fig. 5a), and is, of course, consistent with the reluctance of natural paragonite to accommodate more Si than the ideal 6 atoms per formula unit. The increase in paragonite content accompanying the change from high- to low-Si phengite during decompression of the Seward Peninsula blueschist terrane is therefore a crystal-chemical effect (see Guidotti 1984, p. 382) and is not to be ascribed to an increase in metamorphic temperature as greenschist replaced blueschist conditions. The decrease

in F-content of the phengite with decreasing Si (Table 4) is probably related to the accompanying growth of biotite (equation 1 below), which contains greater amounts of fluorine.

In the KMASH paragenesis phengite + K-feldspar + phlogopite + quartz +  $H_2O$ , the composition of phengite is invariant at fixed P and T. Isopleths of the Si content of phengite, governed by the equilibrium:



are a function of P and T (Velde 1967, Massonne 1981, Massonne & Schreyer 1983, Powell & Evans 1983). The Si-contents of natural muscovite range from 6.0 in high-grade Buchan-style metamorphism (Evans & Guidotti 1966) to 7.6 in blueschist- and eclogite-facies rocks (Sorensen 1986). Unfortunately, the isopleths drawn by Velde and Massonne are based on experiments that fall short of fulfilling all the desired conditions of reaction reversal and phase characterization. Their results differ considerably, particularly when considered at high pressure (Fig. 9). Velde's product micas were not characterized for composition or polytype, and Massonne's experiments were synthesis runs that produced mainly *1M* and *Md* polytypes. Massonne's charges were iron-free; since the ratio  $Fe^{2+}/Mg$  is higher in biotite than coexisting muscovite ( $K_D \sim 0.37$ ), the introduction of Fe will stabilize biotite + K-feldspar and thus reduce the celadonite content of the white

mica, displacing the Si-isopleths to higher pressure. Our data from the Seward Peninsula suggest that the K-feldspar + biotite field boundary, for the maximum P/T conditions reached, is not far removed from horizontal (Fig. 10), so that the influence of the Fe/Mg ratio on the Si-isopleths may not be large. The influence of F-substitution on the isopleths, given the amounts of F present, should also be very small; although F/OH is higher in biotite than muscovite, the effect is counterbalanced by the 3:1 stoichiometry of the micas in equation 1.

The two experimental calibrations, taken at face value, at a temperature of 450°C and with no correction for the Fe or F content of the system, indicate maximum pressures for the Salmon Lake and Cape Nome samples (maximum Si = 7.2) of 9 kbar (Velde 1967) and 14 kbar (Massonne & Schreyer 1983). The fit to Velde's data by Powell & Evans (1983) gives 1 to 2 kilobars lower readings in the higher pressure range, as compared to Velde's original isopleth diagram. The 14 kbar estimate is regarded as excessive, since, if true, it would have resulted in the widespread occurrence of jadeite in quartzofeldspathic rocks and omphacitic pyroxene in mafic rocks. So far, only three rocks containing omphacitic pyroxene have been discovered on the Seward Peninsula (between Salmon Lake and Nome; Forbes *et al.* 1984, Thurston 1985) and no signs of jadeite relics or pseudomorphs have been found. The generation of low-Si (Si = 6.4) phengite in the Salmon Lake samples corresponds to pressures of about 3 and 5 kilobars according to the two calibrations (also taken at 450°C).

The Serpentine Hot Springs sample, again taking the two experimental calibrations at face value, and assuming  $T = 450^\circ\text{C}$ , corresponds to maximum pressures of 7 and 10 kilobars. The low-Si phengites apparently formed under the same conditions as recorded in the Salmon Lake and Cape Nome samples. The slightly higher two-feldspar temperature for the Kiwalik Mountain samples translates into maximum pressure estimates of 8 and 11 kilobars, according to the two calibrations.

These results permit the preliminary conclusion that the Salmon Lake and Cape Nome areas underwent high P/T metamorphism under similar conditions, and that maximum pressures of metamorphism were 3 to 4 kilobars lower at Serpentine Hot Springs and Kiwalik Mountain. The phengite compositions at these two last localities suggest that these areas probably never experienced blueschist-facies conditions.

The variation in celadonite content of phengite as a function of metamorphic conditions and mineral paragenesis can be shown to advantage in a projection from  $\text{H}_2\text{O}$ ,  $\text{K}_2\text{O}$  and  $\text{SiO}_2$  onto the plane  $\text{Al}_2\text{O}_3\text{-MgO-FeO}$  (Thompson 1979, Miyashiro & Shido 1985), in which Si-content (= 12-Al) increases

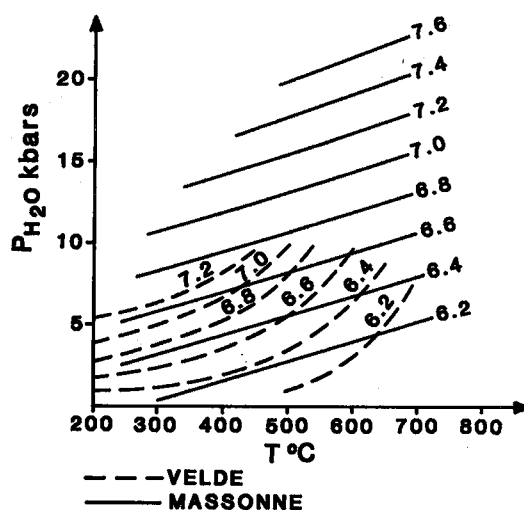


FIG. 9. Si-isopleths of phengite in the paragenesis phengite + biotite + K-feldspar + quartz +  $\text{H}_2\text{O}$  according to Velde (1967) and Massonne (1981).

from the top of the triangle downward (Fig. 10). The biotite + K-feldspar field boundary for the (maximum P/T) blueschist-facies metamorphism in the central and southern Seward Peninsula runs approximately horizontally across the diagram (though this clearly is not well constrained) at Si between 7.1 and 7.2. The fields indicated for glaucophane-, chloritoid-, and paragonite-bearing parageneses (Fig. 10) are based on preliminary investigation of phengite in other lithologies on the Seward Peninsula. With the progressive change to greenschist-facies conditions, the biotite + K-feldspar field boundary moves upward to about Si = 6.3, considerably condensing the areas of the phase fields for more aluminous assemblages. The upward trends in phengite composition for each sample in Figure 10 represent the migration of the biotite + K-feldspar field boundary (and related phase fields) during the change from blueschist- to greenschist-facies conditions. Figure 10 shows that with the exception of the Cape Nome samples, there is a tendency for the Fe/Mg ratio of phengite to increase slightly as the Si-content decreases. This observation can presumably be related to the increasingly trioctahedral character of the phengite, and is a reflection of the higher Fe/Mg ratio of the coexisting biotite. As Table 4 shows, there are some cases where total iron oxide is the same, or even higher, in the low-Si as in the high-Si variety in a given sample, despite the decrease in content of celadonite. The suggestion of an oblique trend in the high-Si phengite compositions of sample SL81-24-1 (Fig. 10) may be a reflection of the biotite + glaucophane field boundary, K-feldspar being absent. No glaucophane was found

## BIOTITE

in this sample (or indeed any orthogneiss), although a small number of idiomorphic pseudomorphs consisting of oxy-chlorite, resembling the pseudomorphs after glaucophane in Seward Peninsula metapelites and metabasites (Thurston 1985, Fig. 7D), were found.

The presence of small amounts of low-Si phengite in some samples suggests that some biotite in the orthogneisses could be ascribed to progress of reaction (1) in the direction of dehydration (caused by

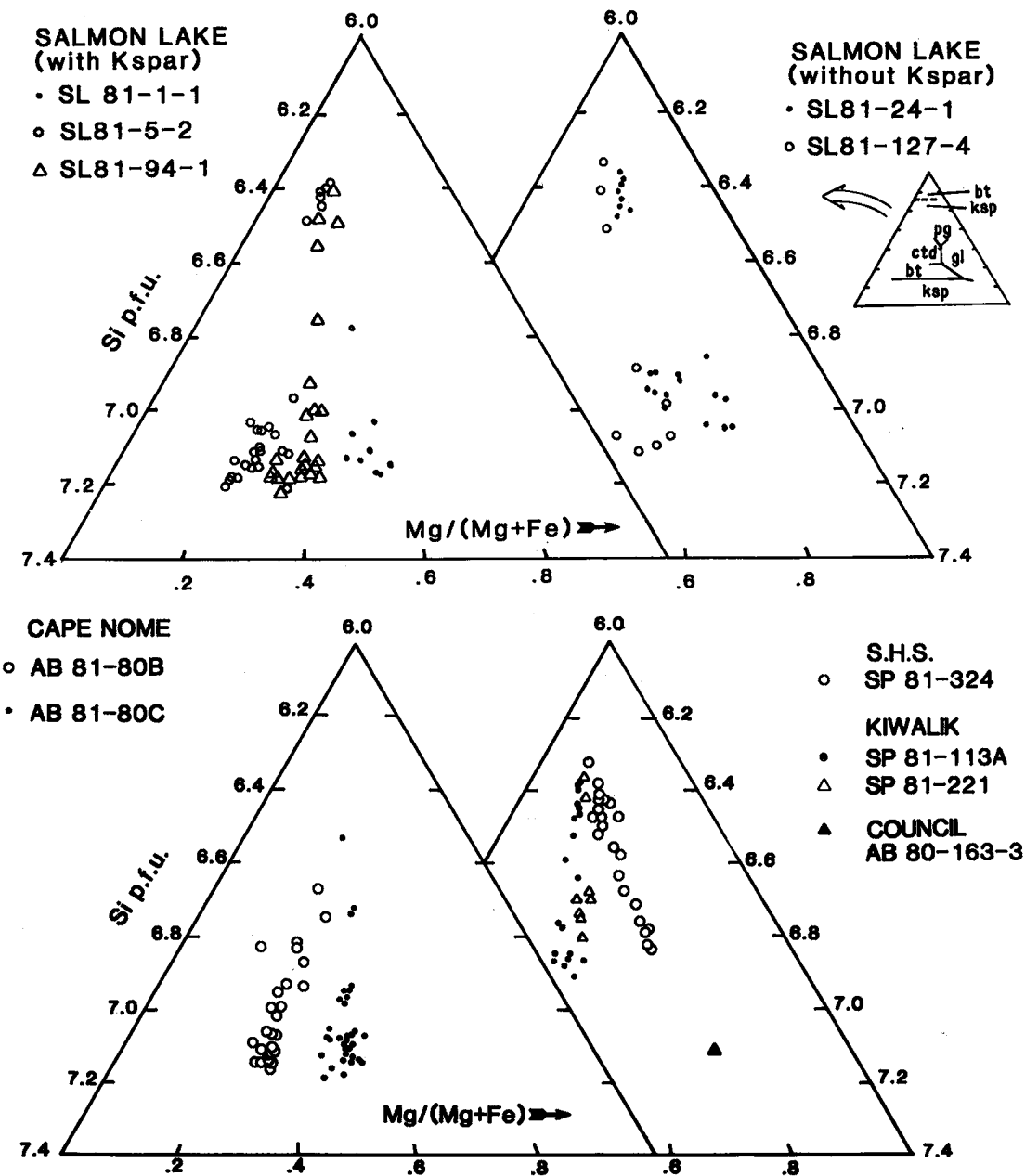


FIG. 10. Paragenesis diagram (after Thompson 1979) in terms of the projected compositions of phengite. Inset: continuous lines are field boundaries for the blueschist-facies metamorphism on the Seward Peninsula. Dashed line is the displaced biotite + K-feldspar field boundary for the greenschist-facies overprint. Chlorite could be substituted for glaucophane, depending on exact P-T conditions.

decompression). Some phengite flakes are indeed fringed by small amounts of biotite, and in many cases the biotite is in optical continuity with the phengite. A good deal of the biotite, however, appears to be part of the high-Si phengite paragenesis, since it forms independent flakes and in some cases layers, and exhibits the same textural features as the high-Si phengite. Thus, biotite is mostly prekinematic in the Salmon Lake orthogneisses, pre- to synkinematic at Kiwalik Mountain and Serpentine Hot Springs, and intergrown with phengite in a decussate texture in the Cape Nome samples (Fig. 2B).

The biotite compositions (Table 5) range from very slightly to considerably enriched in the Fe end-member [Fe/(Fe + Mg) of biotite ranges from 0.51 to 0.79], in parallel with the whole-rock and phengite iron-magnesium ratios. The higher TiO<sub>2</sub>-content of biotite in Kiwalik Mountain orthogneiss is either a reflection of the higher equilibration temperature noted previously or the absence of epidote (since the activity of TiO<sub>2</sub> in Seward Peninsula orthogneisses is controlled by the equilibrium: 4Ttn + 3Ms + 2Qtz = 2Ep + 3Or + 4TiO<sub>2</sub> + 2H<sub>2</sub>O). The low analytical totals are believed to reflect incipient chloritization and Fe<sub>2</sub>O<sub>3</sub> reported as FeO.

We believe that the composition of biotite in sample SL81-24-1 (Table 5) most nearly represents equilibration under the high P/T conditions that were responsible for the production of high-Si phengite. It is lowest in tschermak substitution (Si = 5.76) and, of all biotite-muscovite pairs analyzed here, the partitioning of F/OH most closely approaches equilibrium (a function of mole fraction phlogopite) according to experimental data (Munoz & Ludington 1977). In all other mica pairs (Tables 4, 5), measured values of *K<sub>D</sub>* for F/OH exchange suggest that the biotite is in exchange equilibrium with the later-formed (lower-Si) phengite. The compositions of analyzed biotite from Serpentine Hot Springs and Kiwalik Mountain have Si = 5.56 and 5.52 (Table 5), slightly more tschermakitic than the biotite from Salmon Lake and Cape Nome, consistent with differences in the tschermak component in the corresponding phengite. Our biotite compositions are less tschermakitic than those from pelitic rocks in the greenschist facies but not from calc-silicates and metabasites (Guidotti 1984, Fig. 35). Similarly, they are less tschermakitic than biotite from muscovite-bearing granitic rocks, but not from hornblende-bearing granitic rocks (Speer 1984, Fig. 7). Our tentative conclusion is that blueschist-facies biotite is only marginally more Si-rich than biotite from lower-pressure parageneses, that is, it does not exhibit the extreme enhancement in SiMAl<sub>2</sub> that characterizes high-pressure phengite (see also Miyashiro & Shido 1985).

TABLE 5. COMPOSITION OF BIOTITE (ELECTRON-MICROPROBE DATA)

Weight Percent	Cape Nome		Salmon Lake		Serp. H.S.	Kiwalik Mtn.
	AB81-80B	SL81-24-1	SL81-94-1	SL81-127-4	SP81-324	SP81-113A
SiO <sub>2</sub>	36.05	37.66	34.66	35.86	36.32	34.66
TiO <sub>2</sub>	1.74	.77	1.70	1.88	2.16	3.39
Al <sub>2</sub> O <sub>3</sub>	16.89	15.78	16.35	16.37	16.59	16.38
FeO*	25.66	18.70	27.59	23.70	19.30	27.13
MgO	5.37	11.92	4.65	7.03	10.56	3.95
MnO	.29	.06	.31	.26	.36	.10
K <sub>2</sub> O	.07	.03	.05	.05	.08	.05
Na <sub>2</sub> O	9.37	9.56	9.27	9.68	9.76	9.47
BaO	.02	n.d.	n.d.	.07	.08	.07
F	.46	2.25	.86	.46	.30	.33
Cl	.02	.01	.01	.03	.08	.02
H <sub>2</sub> O <sup>***</sup>	3.60	2.86	3.32	3.61	3.76	3.51
-(F,Cl)	.20	.95	.36	.20	.14	.23
	99.34	98.65	98.41	99.00	99.21	99.03

Cations/22 oxygen atoms	5.551	5.755	5.566	5.611	5.559	5.519
Si	.205	.088	.205	.221	.249	.406
Al	3.121	2.842	3.095	3.055	2.993	3.074
Fe*	3.364	2.390	3.703	3.101	2.670	3.613
Mg	1.255	2.716	1.113	1.640	2.409	.938
Mn	.039	.008	.062	.034	.067	.013
Na	.021	.009	.016	.015	.024	.015
K	1.874	1.864	1.899	1.932	1.906	1.924
Ba	.001	.000	.000	.004	.005	.004
F	.228	1.087	.437	.228	.145	.267
Cl	.005	.003	.003	.008	.021	.005
OH**	3.787	2.910	3.560	3.764	3.834	3.728

\*total iron \*\*calculated H<sub>2</sub>O n.d.: not determined  
Microprobe standards: as for phengite.

## GARNET

Modally abundant garnet occurs only in some of the tonalitic orthogneisses and not in the granitic orthogneisses; garnet appears to be incompatible with K-feldspar. Garnet in tonalitic gneiss SL81-127-4 from Salmon Lake (Table 6) is almandine-rich (67-75%) and pyrope-poor (1.5 to 3.3%); grossular content is 21-25%, and spessartine 10% (crystal centres) to 1% (margin). The high iron-magnesium ratio can be attributed to a combination of whole-rock composition and the low temperature of equilibration.

A temperature of 390°C is obtained from the Mg-Fe partitioning with biotite, using a pressure of 10 kilobars, the calibration of Ferry & Spear (1978), and the garnet-solution properties of Newton & Haselton (1981). This estimate has an uncertainty of ±30°C owing to variation in the measured values of Mg/Fe in the garnet, and of course additional uncertainties arising from the experimental calibration and the solution models adopted. The temperature is believed less reliable than those estimated from the petrogenetic grid (Forbes *et al.* 1984, Thurston 1985) and the coexisting feldspars, as discussed earlier. The possibility exists that this temperature, unlike the two-feldspar temperatures, records equilibration during the greenschist-facies overprint.

TABLE 6. AVERAGE COMPOSITION OF GARNET (SL81-127-4)

Weight percent	Cations/12 oxygens		Mol.%*		
SiO <sub>2</sub>	36.8	Si	2.981	alm	70.1
Al <sub>2</sub> O <sub>3</sub>	20.9	Al	1.995	py	2.1
Fe <sub>2</sub> O <sub>3</sub> *	.26	Fe <sup>3+</sup>	.016	sp	4.8
FeO**	31.3	Fe <sup>2+</sup>	2.120	gr	22.2
MgO	.51	Mg	.062	and	0.8
MnO	2.1	Mn	.144		
CaO	8.0	Ca	.696		
	99.87		8.012		100.0

\*calculated from formula.  
Microprobe standards: Gore Mt. garnet, Nuevo garnet.

## TITANITE

Tiny grains of accessory titanite, although modally insignificant, prove to be geochemically of interest in their large content of the end-member component  $\text{CaAlSiO}_4\text{F}$ .  $\text{Al}_2\text{O}_3$  (up to 7% by wt.) and F (up to 3% by wt.) correlate inversely with  $\text{TiO}_2$  (Fig. 11), and indicate substitution of between 7 and 27 mol. % of the Al end-member along the vector  $\text{AlF}(\text{TiO})_{-1}$ . Almost the entire variation was found within a thin section of sample SL81-127-4. The small departure in Figure 11 from the 45° line may indicate that Al substitution is accompanied by minor  $\text{Fe}^{3+}$  and F by minor OH. This coupled substitution was described by Higgins & Ribbe (1976) and discussed by them in relation to cell parameters and structure.

Aluminum-rich titanite (up to 50% Al end-member) has been described in eclogite-facies rocks from Norway, Italy and Austria (Smith 1980, Smith & Lappin 1982, Rossi *et al.* 1985, Franz & Spear 1985), and the suggestion made that it is an indicator of high pressure. Al-rich titanite was found in orthogneiss from all four locations in this study, that

is, from both the blueschist- and greenschist-amphibolite-facies transition environments. As indicated earlier, the mineral paragenesis of the orthogneisses corresponds to relatively low values of the chemical potential of the tschermak exchange  $\text{Al}_2(\text{MSi})_{-1}$ . It seems preferable to us to relate the occurrence of (Al,F)-bearing titanite to an elevated fluorine potential, presumably resulting from F inherited from the granite protolith micas, and still present in some of the metamorphic micas. Titanite in glaucophane-bearing metabasites in the Seward Peninsula contains only small amounts of Al.

It is perhaps worth noting that should similar Ti-deficient titanite prove more common than hitherto realized, the assumption that titanite has standard-state composition in petrological calculations of intensive variables [such as  $T, P, X(\text{CO}_2)$ ] ought not to be routinely made.

## DISCUSSION AND CONCLUSIONS

Although they do not contain typical high-pressure mineral species, granitic and granodioritic ortho-

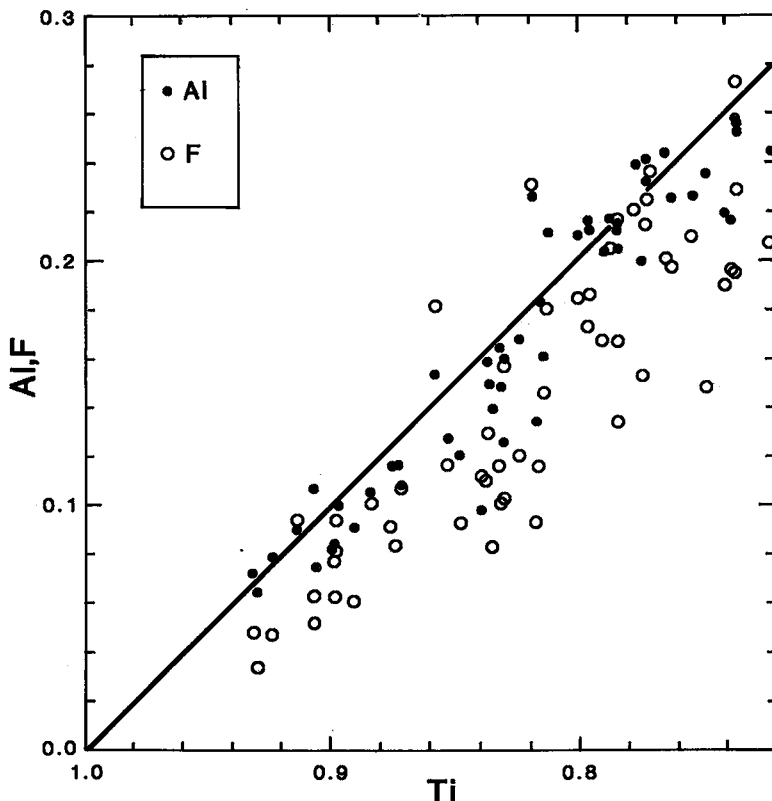


FIG. 11. Correlation between Al, F and Ti in titanite from orthogneisses; 45° line corresponds to variation along the vector  $\text{AlF}(\text{TiO})_{-1}$ .

gneisses in the central and southern Seward Peninsula possess phengite with a consistent *maximum* content of Si in the range 7.1 to 7.2 (per 22 oxygen atoms). Similar high-Si phengite compositions have been described from elsewhere in metagranites with jadeite. The quartz + albite + microcline + phengite-3T + biotite paragenesis of the Seward Peninsula orthogneisses is therefore a high-pressure one, isofacial with glaucophane-bearing parageneses in associated pelites and metabasites. The range in Si-contents of phengite in individual samples of orthogneiss is attributed to re-equilibration during decompression of the terrane, seen petrographically in most lithologies in terms of the partial replacement of blueschist- by greenschist-facies assemblages. In samples from the Salmon Lake area, the greenschist-facies "overprint" produced a second frequency-maximum of Si-contents at Si = 6.4, developed at the ends of the large primary flakes of phengite and in fringing sericite, in many cases in close proximity to small amounts of green-brown biotite. It is not known if phengite compositions with Si-contents intermediate between 7.1 and 6.4 represent equilibrium under P-T conditions between the two extremes, or indicate metastable, partly re-equilibrated phengite. In some cases, the intermediate-Si spot analyses may represent electron-beam-averaged compositions, it being necessary to use a 10-micrometre-diameter electron beam. Lower maximum Si-contents of phengite from the Serpentine Hot Springs area in the northern Seward Peninsula and across the Kugruk Fault Zone at Kiwalik Mountain in the northeastern Seward Peninsula point to lower pressures during the main stage of the metamorphism. Glaucophane has not been found in the vicinity of either of these two orthogneiss bodies.

The more pronounced bimodal character of phengite compositions in the Salmon Lake orthogneisses is very likely related to the proximity of that area to the Kigluak Complex, an east-west-trending antiformal belt of Late Cretaceous high-grade metamorphic rocks and granitic intrusive bodies. Significantly, phengite from orthogneiss sample SL81-5-2 from the Salmon Lake area has given the youngest mineral age ( $96 \pm 8$  Ma, Rb-Sr mineral/whole rock pair) of all 18 mineral ages, including K-Ar, done on the Nome Group (Armstrong *et al.* 1986). The Cretaceous greenschist-facies overprint evidently reset the isotopes but left the 3T high-Si phengite largely unchanged; this sort of effect was noted by Frey *et al.* (1983). Why so much of the high-Si phengite in all areas survived decompression through greenschist-facies conditions, despite the dehydration nature of the celadonite-breakdown reaction, is not fully understood. Fortunately, its survival proves beyond any doubt that the Si-content of natural phengite in granitic orthogneiss assemblages decreases at least by 0.8 atoms pfu during a blueschist

to greenschist PTt-path, that is, about 0.1 atom Si pfu per kilobar change in pressure, as shown by the experimental work (Fig. 9).

Not only high lithostatic pressure, but also high H<sub>2</sub>O-pressure, is required for high-Si phengite. The presence of calcite in some of the orthogneisses (Table 1) probably means that some infiltration of CO<sub>2</sub> from surrounding metasediments took place during metamorphism. However, the occurrence of titanite rather than rutile + calcite + quartz, and of microcline + epidote (in some cases accompanied by phengite + quartz + calcite, making the isobaric univariant assemblage) shows that the concentration of CO<sub>2</sub> in the fluid phase was very small, less than 1 mol. %.

Temperature estimates based on the compositions of coexisting albite and microcline, ranging between 420 and 500°C, are believed to be broadly correct, since they agree with as yet unpublished calcite-dolomite geothermometry and inferences based on the petrogenetic grid. Unfortunately, the occurrences of granitic orthogneisses are not of sufficient frequency for regional trends in metamorphic temperature, if any, to be detected from these rocks alone.

Correlations in the phengite compositions among the substitutions  $Al_2(MSi)_{-1}$ ,  $NaK_{-1}$ ,  $M_3\Box_{-1}Al_{-2}$ ,  $OHF_{-1}$  and  $MgFe_{-1}$  show great consistency in the samples studied. Some of these substitutions are driven by crystal-chemical coupling with the highly pressure-sensitive tschermak substitution. Although the muscovite-celadonite solution shows evidence of nonideality (Miyashiro & Shido 1985), the number of recent descriptions of phengite in the Si-range 7.0 to 7.6 from high-pressure rocks indicates that, at the present time, there is no compelling reason to infer a miscibility gap within the series, nor a crystal-chemical upper limit of Si in the region of 7.0. The prevalence of phengite-3T in the orthogneisses is attributed to its large content of the celadonite component, a consequence of high-pressure metamorphism of a granite bulk-composition. Phengite-2M<sub>1</sub> of lower Si content crystallized in associated pelitic and mafic rocks under the *same* high-pressure conditions. More precise conclusions on the compositional control (Si content, Fe/Mg ratio, *etc.*) of phengite polytypism will require detailed microprobe and optical study of a large number of individual mica flakes.

This and other field studies show that phengite compositions in appropriately defined multiphase assemblages are capable of providing useful information concerning pressure. Furthermore, phengite composition and polytype tend partly to survive subsequent metamorphic events, permitting a chronology of physical conditions to be determined and, in favorable situations, related to micro- and macro-structural stages.

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