

COMPOSITION AND STRUCTURAL STATE OF ALKALI FELDSPARS FROM CHARNOCKITIC ROCKS ON WHITEFACE MOUNTAIN, NEW YORK

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ABSTRACT

Eleven potassic feldspar concentrates from mesoperthites in charnockitic gneisses near the anorthosite contact on Whiteface Mountain in the Adirondack Highlands, after computer cell refinement, show a range between Δbc 0.78 and 0.96. The more ordered feldspars are closer to the contact although a regional gradient is possible. At the 0.99 confidence level greater variation in Δbc exists between samples than within samples.

Microprobe analyses of coexisting phases in mesoperthite are typically $Or_{90.0}Ab_{9.9}An_{0.1}$ and $Or_{0.05}Ab_{94.6}An_{15.4}$ with the former in good agreement with the cell volume compositions. Using the T/T_c values for mole fractions of Or and Ab calculated by Waldbaum and Thompson (1969) from the G_{ex} function, a set of binodal curves for various alkali/alumina ratios, pressures, and intermediate structural states was compared with the observed Or contents of the potassic phase. For peralkaline compositions at 7 kb in the Δbc range 0.82–0.94, the mean maximum temperature of equilibration for all samples was $588^\circ \pm 20^\circ C$.

Magmatic temperatures in Adirondack charnockites have been estimated at $750\text{--}800^\circ C$ and were followed by pyroxene granulite facies metamorphism in the eastern Adirondacks of $650 \pm 50^\circ C$. The temperatures and structural states determined for mesoperthite-microperthite from Whiteface Mountain suggest that minor internal changes within a feldspar crystal could occur at temperatures somewhat below those controlling external phase equilibria. Minor amounts of microcline in the recrystallized matrix of some gneisses indicate final equilibration occurred at $375 \pm 50^\circ C$ or below.

INTRODUCTION

Recent additions to the feldspar literature, especially in crystallography, crystal chemistry, and determinative techniques, have made this group an especially attractive object of study to petrologists and students of mineral paragenesis. The anorthosite and charnockite series of the Adirondack Highlands provide unusual opportunities to describe and interpret feldspar relationships of the utmost complexity with a magmatic and subsequent metamorphic history. Enough is known about critical mineralogical and isotopic equilibria in the Adirondacks to establish the approximate upper P - T limits of prograde metamorphism; a significant residual problem is the nature of re-equilibration between and within mineral phases. The remarks by Buddington (1969b) on the physical conditions prevailing during the magmatic and metamorphic evolution of Adirondack rock were concerned (Buddington, personal communication, 1971) with "cases where unmixing has gone so far as to yield readily mechanically separable grains, each of which could be inferred to have equilibrated approximately with each other." The present study, on the other hand, is primarily concerned with re-equi-

libration on the intragranular scale, which apparently persisted to lower temperatures and therefore presumably longer than reactions controlling the mineral assemblage as a whole. Thompson (1969, p. 341–342) has called attention to the petrologic value of investigations on homogeneous equilibria which are in part responsible for the structure and chemistry of the feldspars described here.

The geologic setting and sampling procedure, the petrography, and optical character of the feldspars are outlined below.

Samples were collected from the upper 1000 feet of Whiteface Mountain at chained intervals using a diamond drill or sledgehammer. Rock exposure is nearly continuous along a slide trending about S.70°W. from the summit. Two diamond drill cores spaced one meter apart up to 10 cm in length were taken every 100 meters beginning at 3850 feet elevation. As the anorthosite was approached, the sampling interval was reduced to 50, 12, and 1.2 meters from the contact which lies at 4510 feet elevation and 462 meters from the starting point. A total of 13 charnockite cores was obtained. Supplementary hand specimens were collected on the southeast side of the summit along the line trending about S.80°E. cut for the upper ski lift from 4100 feet elevation upwards to the actual contact.

The reader may wish to consult a geologic sketch map and cross-section of Whiteface Mountain and vicinity included in Crosby (1968, p. 34–35). It is sufficient to visualize, for the present purposes, a plate of gabbroic anorthosite gneiss of low to moderate dip, of which a 125 meter thickness is exposed on the summit of Whiteface Mountain, resting upon perhaps as much as 800 meters of charnockitic gneisses to the south and west, of which conceivably the upper third have been sampled in this study.

Megascopically, rocks of the charnockite series sampled in the above profiles are typically medium-grained, gray-green or olive green to rose or pink gneisses. By charnockite series (de Waard, 1969, p. 83–90) is meant the hypersthene-bearing equivalents of granite (charnockite), farsundite (adamellite), mangerite (monzonite), and others. Many facies are equigranular in the range from 0.5 to 2 mm but some varieties have small megacrysts of white mesoperthite up to one cm in diameter. Foliation ranges from inconspicuous to obvious (“pinstripe”) laminations formed of dark minerals spaced 5–10 mm apart. The color of the rock is largely dependent upon that of the feldspars and is commonly variable within a few meters. Some of the pink varieties are alaskite and somewhat coarser-grained than the surrounding rocks. They appear to vein or irregularly merge into the dominant graygreen gneisses; color boundaries may transect the foliation and suggest metasomatic controls, possibly oxygen fugacity gradients accompanying the widespread hydration of pyroxenes to hornblende, biotite, and chlorite.

It should be remarked here that the ortho- and clinopyroxenes can be, and have been interpreted, as of either magmatic or granulite-facies metamorphic origin. No evidence for two generations of pyroxenes has been detected in this section.

More attention should be paid to the paragenesis of hornblende in

many Adirondack charnockites. Buddington (1963, p. 1170) is convinced that the variations in ferromagnesian minerals correspond to those formed during magmatic consolidation in the highest intensity metamorphic zone (Zone E) of the Adirondacks which includes Whiteface Mountain. Subsequent recrystallization has produced new hornblende in the pyroxene granulite facies especially in response to the growth of garnet as well as new from old biotite. Buddington (personal communication, 1971) also approvingly quotes Sen to the effect that hornblende and pyroxene in the hornblende granulite facies in the Northwest Adirondacks and the Madras charnockites are coeval metamorphic products. Hornblende rims to calcic pyroxene grains and aggregates in the charnockites of Whiteface Mountain lead me to the conclusion that at least some of the hornblende is later than pyroxene. Furthermore, this rimming hornblende does not appear to be distinguishable superficially from other hornblende in the rock. It seems probable that two generations of hornblende, either an igneous and a metamorphic or two metamorphic, are present, but proof must await more subtle tests.

A much longer span of time than originally thought is now perceived to have elapsed between the magmatic and metamorphic history of the Adirondack orthogneisses. Spooner and Fairbairn (1970) have determined Rb/Sr ages of 1465 ± 85 and 1336 ± 71 m.y. for charnockites at two localities compared with earlier Rb/Sr dates of 1055 ± 31 m.y. (Heath and Fairbairn, 1969) and 1092 ± 20 m.y. (Hills and Gast, 1964). An interpreted uranium/lead zircon age of 1130 ± 10 m.y. for Ticonderoga charnockitic gneisses is given by Silver (1969) who found pyroxene granulite facies metamorphism still active as late as 1020 ± 10 m.y. ago. It appears likely that Silver determined metamorphic ages only, and that the Spooner and Fairbairn Rb/Sr dates record an igneous pre-history. Buddington (personal communication, 1971) notes that Zartmann found an age difference of 250 m.y. between a pink hornblende granite containing possible relics of mesoperthite having a temperature of formation perhaps over 700°C and a metamorphic assemblage that still yields 650°C .

The petrography of the charnockite section on the west side of the mountain is given in Table 1. Modes were determined by 1000-point counts at a spacing of 1 mm. As the anorthosite is approached, comparison of the sample nearest the contact (WA-7) with the mean of 12 samples more remote shows an increase in ferromagnesian minerals and ores, a decrease in quartz and an increase in plagioclase relative to mesoperthite; this is also true of samples from the east side of the mountain and is characteristic of the contact everywhere in the Adirondacks. Although the average rock of Table 1 is a charnockite, there are representatives of

TABLE 1

Modes of Charnockite 12 to 462 Meters from Anorthosite Contact, West Side of Whiteface Mountain Compared with Sample near Contact

Mineral [†]	Mean (N=12)	Fs. prop. whole rock*	Standard deviation	Sample 1.2 m from contact (WA-7)
Mesoperthite	48.93	---	5.45	40.3
K sp. comp.	---	32.78	---	---
Plagioclase	23.28	39.43	7.40	37.9
Quartz	18.63		7.86	7.1
Clinopyroxene	0.53			4.0
Orthopyroxene	0.26			---
Hornblende	5.26			3.7
Garnet	1.56			3.6
Ores	0.74			1.9
Total	99.19			98.5
Color index	8.95		3.24	

[†] Minor accessories omitted from tabulation. Ore minerals are chiefly titaniferous magnetite.

* Proportion of feldspars in whole rock based upon point count of mesoperthite grains (see text for details).

farsundite and mangerite as well, if allowance is made for the alteration of most of the hypersthene to secondary minerals.

Four mesoperthite grains from WA-1A, which is representative of all samples, were point-counted at a 0.1 mm interval to determine the proportion of exsolved plagioclase blebs to the host K feldspar phase. The mean plagioclase content is 33.1 ± 6.03 percent. Another large grain from WA-4A gave 32.0 percent plagioclase. The term mesoperthite implies a 1:1 ratio (Michot, 1948) but in the Adirondacks is generally used rather liberally in a generic sense. Under the microscope the host potassic feldspar phase contains ovoid plagioclase blebs, seldom showing twinning, averaging 50 μ in diameter and spaced 100–200 μ apart. Less commonly, plagioclase occurs as strings or rods covering an equivalent area. Some of the host crystals have discontinuous, apparently exsolved, plagioclase rims. In others, neither blebs nor rims of exsolved plagioclase are found in the outermost zone. According to Tuttle's (1952) classification based upon size of the exsolved phase almost all such host crystals could be described as micropertthite (size range 5–100 μ).¹

¹ Buddington (personal communication, 1971) considers the feldspar described here to be a micropertthite and cites as *bona fide* mesoperthites those illustrated in the following photomicrographs: (Buddington, 1939; Pl. 9, Fig. 2; Pl. 16, Fig. 1; Pl. 17, Fig. 1; and Pl. 19, Fig. 2). He acknowledges, however, the broad usage the term has received in the Adirondacks, and accepts my expression "mesoperthite-micropertthite" for the Whiteface feldspars.

Plagioclase grains, none of them antiperthitic, nearly as large as most of the "mesoperthite" crystals form nearly a quarter of the typical charnockite (see Table 1). If the feldspar ratios in the mesoperthite-microperthite are used to recalculate the percentage of potassic feldspar and plagioclase in the whole rock, the values in column 2 of Table 1 result. These are based on the modal averages, but calculations on the modes of individual samples (WA-1A, -4A) also show more plagioclase than potassic feldspar in the whole rock indicating that enough plagioclase is present to form a 1:1 mesoperthite with a slight excess of plagioclase. It seems reasonable to conclude that most of the plagioclase has exsolved from a single homogeneous feldspar. The composition of the discrete plagioclase grains, which presumably are produced by granule exsolution during deformation at an appropriate temperature, is consistent with this hypothesis. A check of several such grains from WA-7 gave An contents of 12 and 15 percent using the Nieuwenkamp (1966) method with Berek compensator. These compositions lie within the range of microprobe determined plagioclase grains in mesoperthite although, unfortunately, their Or content is not known.

The mesoperthite-microperthite grains show variable optical character under crossed nicols. The larger grains show the homogeneous extinction characteristic of monoclinic feldspar but patches of shadowy and irregular to vaguely rectilinear extinction exist and indicate that triclinic domains with crossing albite and pericline twins are large enough to be optically resolved. In the more highly ordered samples (Fig. 1, Table 4) and those nearest the anorthosite contact, finer-grained feldspars showing the full microcline quadrille twinning are found on the margins of the mesoperthite grains; these are interpreted as recrystallization products developed at a late stage.

ANALYTICAL PROCEDURES

In preparation for electron microprobe analysis, sections were cut and polished from the cores used for making thin sections. Mesoperthite grains of interest were marked under a stereomicroscope in such a way that they could be identified in the electron microprobe, a 3 channel ARL instrument. Simultaneous analyses for K, Na, and Ca were made using an accelerating voltage of 15 kv and a beam current of 0.15 or 0.30 microamps. Ten or more spots were analyzed from one or more grains in each sample using a beam 10 to 20 μ in diameter and a fixed counting time of ten seconds. To minimize the effects of sodium volatilization on counting rates for this element, the beam size was adjusted until 5 consecutive 10 second counting periods on the same spot showed a stable number of counts. Since it was desired to obtain the most potassic and sodic coexisting feldspars in each perthite grain and remove, insofar as possible, the bias resulting from the electron beam overlapping two phases, the samples were moved under the electron beam using cathodoluminescence and reflectivity as a guide to the phases. Comparison of counts for K, Na, and Ca on the typewriter printout usually insured an appropriate distribution of analyses. Standards selected from those listed in Smith and Ribbe (1966, Table 4) were used to pre-

pare calibration curves. The average of five 10 second counts on 2 to 4 standards repeated after 2 or 3 unknowns were corrected by the factors for pure feldspars and for trace amounts of Fe, Ba, and Sr given in Table 3 of Smith and Ribbe (1966). The raw counting statistics on unknowns were adjusted by factors interpolated from estimates of K, Na, and Ca. On the assumption that the unknown feldspars fall approximately within a ternary system, only those analyses which total nearly 100 mole percent were accepted (Smith, 1965, p. 860-861). It is likely that iron, which is responsible for the ubiquitous green or pink coloration of the feldspars in the charnockite series, represents the chief neglected component. Of the 17 analyses given in Table 3, only 5 feldspar molecule totals differ more than ± 2 percent from 100 percent and none more than 8 percent.

The separation of potassium feldspar for X-ray analysis from the whole rock was accomplished by the procedures described by Scotford (1969, p. 1082), *viz.*, crushing to pass 200 mesh screen, settling in beaker of water, drying, and two-stage centrifugation of four half gram lots in a bromoform and N-Dimethyl-formamide solution having a specific gravity of $2.60 \pm .005$. This process resulted in a potassium feldspar concentrate having a size range of 30 to 75 μ which was sufficiently fine to remove most of the impurities. X-ray patterns show variable but negligible quantities (in terms of peak intensities) of quartz and plagioclase. Quartz with about the same dimensions as exsolved plagioclase appears under the microscope in some mesoperthite grains.

Potassic feldspar concentrates were combined with enough BaF_2 so that the strongest peaks of the internal standard (which has four between 24° and 49°) registered as near full scale on the diffractometer charts as possible. Scans from 57° to $12^\circ 2\theta$ at a rate of $1/2^\circ/\text{min}$. were run on a Phillips instrument in the laboratories of D. B. Stewart of the U. S. Geological Survey using their standard operating procedure. A high intensity Cu tube operated at 45 kv, 35 ma is employed and the detector output is passed through a pulse height analyzing circuit to minimize background. On two samples (68-15 and WA-4B) both a strip chart and magnetic tape record were obtained, the latter accumulating the number of counts in each $0.005^\circ 2\theta$ sector of the goniometer scan which is then processed by computer. The computer output represents the integration of the diffraction peaks in terms of height (intensity), width, and area after subtracting "noise." The 9 remaining samples were X-rayed using the multiple (3 or 4) chart procedure described by Wright and Stewart (1968). After 2θ values were corrected graphically relative to the BaF_2 standard, peaks were inspected according to the visual and statistical standards outlined by Orville (1967). Only peaks of A or B quality by both criteria were averaged and used in the initial run of the least squares unit cell refinement computer program (Evans *et al.*, 1963) on an IBM 360/50.

The computer processed diffractometer data is to be preferred for its savings in time and effort and for its superior resolution, but under ideal conditions nearly as many usable peaks can be measured with the manual multiple chart method. The best results achieved in this study with the latter technique were 28 out of 29 measured peaks accepted by the cell refinement program whereas the better of the two magnetic tape records yielded 33 acceptances out of 39 measured peaks. In each case, the unit weight standard error of measured 2θ values is less than 0.02° .

ANALYTICAL RESULTS

Table 2 presents data on the direct cell parameters of eleven alkali feldspars, 8 from the sampling profile on the west side of Whiteface Mountain, and 3 from the east side. Much of the relevant structural information is displayed in Figure 1 which is an adaptation of the *b-c* plot employed by Wright and Stewart (1968). The corners of the quadri-

TABLE 2
 Direct Cell Parameters for Potassic feldspars

Specimen	Cell edges (Å) †			Interaxial angle(s)*	Cell Volume (Å ³)	Standard error 2θ‡	Reflection Measured/ Accepted
	a	b	c				
WA-1A	8.5819 ±.0024	12.9849 ±.0027	7.2008 ±.0016	115° 02.363' ± 01.230'	720.961 ±.220	0.01908	29/28 D.F. 24
WA-2B	8.5823 ±.0028	12.9752 ±.0025	7.1955 ±.0015	115° 59.844' ± 01.075'	720.186 ±.223	0.01604	25/22 D.F. 18
WA-3A	8.5887 ±.0026	12.9804 ±.0023	7.2004 ±.0013	116° 00.953' ± 01.022	721.400 ±.209	0.01447	29/23 D.F. 19
WA-4A	8.5864 ±.0023	12.9830 ±.0032	7.2054 ±.0017	116° 00.719' ± 01.525'	721.872 ±.250	0.01854	26/21 D.F. 17
WA-4B	8.5963 ±.0021	12.9772 ±.0017	7.2026 ±.0012	116° 04.375' ± 00.980'	721.729 ±.165	0.01733	39/33 D.F. 29
WA-5B	8.5799 ±.0017	12.9760 ±.0024	7.2004 ±.0012	116° 02.281' ± 00.874'	720.279 ±.183	0.01291	26/20 D.F. 16
WA-6B	8.5857 ±.0037	12.9681 ±.0039	7.2077 ±.0016	115° 59.277' ± 01.762'	721.361 ±.316	0.01688	21/14 D.F. 10
WA-7	8.5784 ±.0025	12.9711 ±.0045	7.1984 ±.0017	116° 02.133' ± 01.746'	719.691 ±.315	0.01872	25/16 D.F. 12
68-15	8.5793 ±.0021	12.9617 ±.0035	7.2138 ±.0016	116° 08.719' ± 01.213'	720.108 ±.249	0.01901	32/23 D.F. 19
68-46B	8.5791 ±.0018	12.9646 ±.0034	7.2089 ±.0015	116° 00.391' ± 01.042'	720.628 ±.234	0.01841	29/26 D.F. 22
68-18	8.5724 ±.0029	12.9540 ±.0042	7.2128 ±.0017	α89° 58.391' ±.03.092' β116° 01.254' ±.02.829' γ85° 51.090' ± 04.918'	719.764 ±.360	0.01548	22/17 D.F. 11

† CuKα (1.541781 Å) used for 2θ < 30°;
CuKα₁ (1.540562 Å) " " 2θ > 30°

* β angle is given for all monoclinic samples. α, β, and γ angles are given for 68-18 which is triclinic for lowest S.E. solution

‡ Unit weight used for all lines

lateral are formed by sanidine, high albite, low albite and maximum microcline whose *b* and *c* cell parameters are essentially those given by Stewart and Ribbe (1969, p. 450-451). The position of the samples analysed in this study are indicated by the rectangles whose dimensions are controlled by the standard errors associated with the calculated lengths of *b* and *c*.

As may be seen from Table 2, all samples except 68-18, which is nearest the maximum microcline-low albite join, give a monoclinic solution. Both monoclinic and triclinic solutions were tried for 68-15 which plots adjacent to 68-18, but the monoclinic cell gave lower unit weight standard errors between the measured and calculated 2θ values than did the triclinic. A monoclinic solution could probably be found for 68-18 as well.

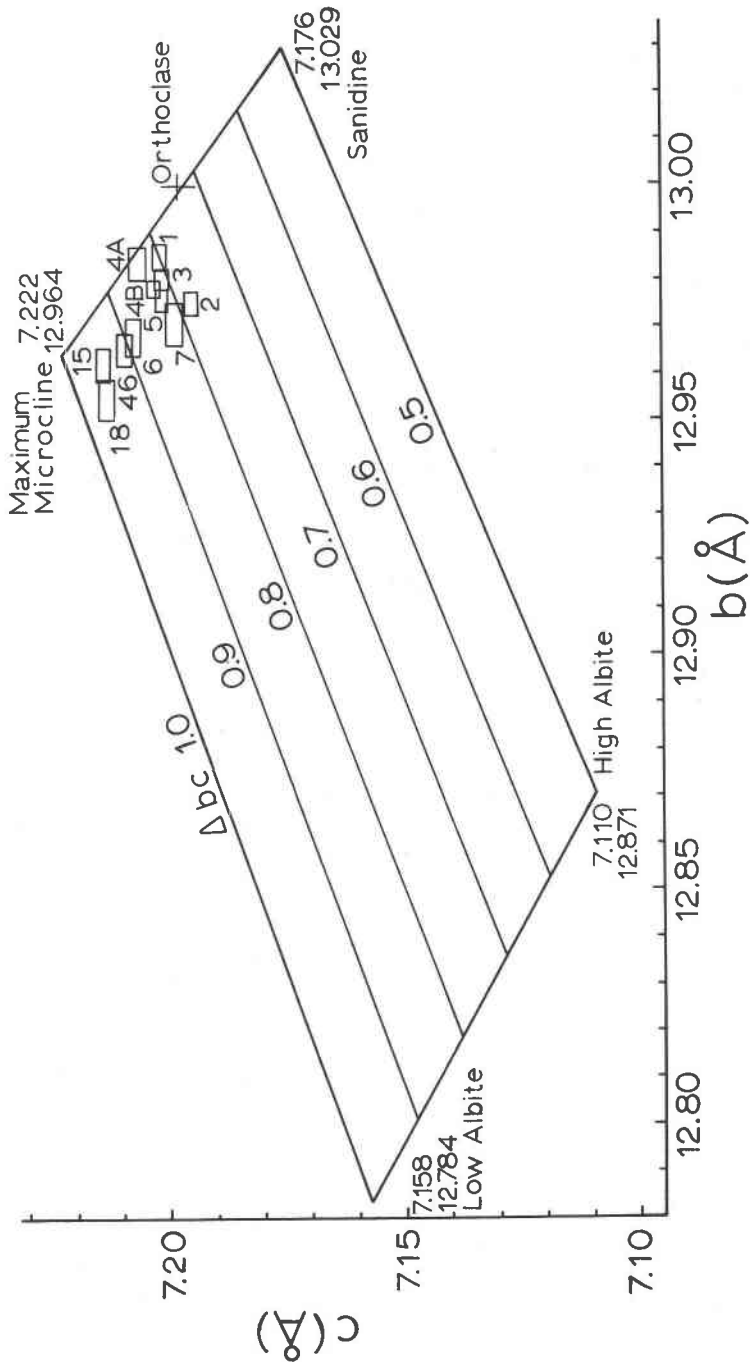


FIG. 1. Potassic feldspars from Table 2 on *b-c* plot with cell edges of end members and equal Δbc lines from Stewart and Ribbe (1969) and unpublished ms. The numbered rectangles show *b* and *c* standard errors for each sample. Prefix and suffix designations for samples have been omitted for simplicity. Orthoclase is KAlSi_3O_8 equivalent from P50-56F exchange series (Wright and Stewart, 1968, Table 7, p. 59).

It should be observed that symmetry determinations from powder X-ray diffraction data is limited by the averaging effect imposed by domains within crystals. Thus, as pointed out by Wright and Stewart (1968, p. 42), "many potassic phases of intermediate structural state may actually be triclinic, but must be treated as monoclinic by powder-diffraction methods." These workers have also found (Stewart and Ribbe, 1969, p. 452-453) that the ranges of cell parameters of triclinic and monoclinic K feldspars overlap, and that there is no single identifiable point on plots of cell angles that corresponds to the symmetry change.

Compositional data for the feldspars is provided in Table 3 as determined by both microprobe and cell refinement techniques. With respect to the cell parameters, it was found that direct cell volumes gave compositions more consistent with the microprobe data than the a cell edge did. In both cases, Figure 1 of Wright and Stewart (1968) which is a plot of a and cell volume against composition was used. The Or content indicated by the a cell edge was from 1 to 6 percent higher than that

TABLE 3. COMPOSITIONS OF COEXISTING FELDSPARS FROM MESOPERTHITE IN CHARNOCKITE

Distance from contact	Sample no.	Most sodic composition Microprobe data	Most potassic composition	Average or content of potassic phase by cell volume
West Side Profile				
462 meters	WA-1A	Or _{0.05} Ab _{83.0} An _{17.0}	Or _{84.9} Ab _{15.0} An _{0.1}	89
362 meters	WA-2B	Or _{0.7} Ab _{87.4} An _{11.9} *	Or _{87.0} Ab _{12.2} An _{0.8}	88
262 meters	WA-3A	Or _{0.1} Ab _{84.8} An _{15.1}	Or _{89.3} Ab _{10.6} An _{0.1} *	90†
162 meters	WA-4A	Or _{1.4} Ab _{82.7} An _{15.9} *	Or _{90.5} Ab _{9.1} An _{0.4}	90
162 meters	WA-4B	No data	No data	90
62 meters	WA-5A	Or _{0.05} Ab _{84.6} An _{15.4}	Or _{90.0} Ab _{9.9} An _{0.1}	88†
12 meters	WA-6B	Or _{0.3} Ab _{89.2} An _{10.5}	a	92
1.2 meters	WA-7	a	Or _{91.5} Ab _{8.3} An _{0.2}	86
East Side Profile				
290 meters	68-18	No data	No data	89
70 meters	68-14	a	Or _{88.6} Ab _{6.3} An _{0.1}	No data
70 meters	68-15	Or _{0.6} Ab _{81.9} An _{17.5}	Or _{89.0} Ab _{10.9} An _{0.1} *	90
On contact	68-46A	Or _{1.3} Ab _{76.4} An _{22.3}	Or _{89.8} Ab _{10.0} An _{0.02}	90†

* No pure exsolved phase measured.

† Normalized from feldspars totalling 100 ± 6-8%. Others normalized from feldspars totalling 100 ± 0-3%, mostly 100 ± 1%.

† Cell refinement from paired sample (A or B) one meter distant from microprobe analyzed sample.

given by cell volume indicating that the cells of most of the analysed feldspars are slightly anomalous. Excellent agreement exists between the most potassic compositions detected in mesoperthites by the microprobe and the average Or content of the potassic feldspar separate measured by cell volumes except in samples WA-1A and WA-7. It may be that these two are more heterogeneous compositionally than the rest and were inadequately sampled by the microprobe. Probe analyses reveal that variation in Or and Ab content of homogeneous phases may range up to 5 percent. A test of compositional homogeneity can be made by noting the width of the $\bar{2}01$ peak; in these samples the width of the $\bar{2}01$ peak is comparable to that of the samples where compositional agreement is better; in fact, several of the latter had somewhat broader $\bar{2}01$ peaks than WA-1A and WA-7. With but few exceptions, however, the $\bar{2}01$ peak in the charts for all samples easily met the condition for A visual quality specified by Wright (quoted by Orville, 1967, p. 61) namely that the upper 1/10 of the peak is confined to $0.1^\circ 2\theta$. Waldbaum and Robie (1971) nevertheless indicate that in their experience with microprobe analyzed alkali feldspars a sharp, narrow $\bar{2}01$ peak is not necessarily a sufficient criterion for complete homogeneity.

It is now more convenient to specify the relative structural state of alkali feldspars by their Δbc values as derived by Stewart and Ribbe (1969) than by their position relative to representative alkali exchange series (see Wright and Stewart, 1968, Table 8). To facilitate giving a numerical value to Δbc , one may, as Stewart and Ribbe (1969) describe, draw a series of equispaced lines (as has been done on Fig. 1) parallel to the sanidine-high albite and maximum microcline-low albite boundaries of the quadrilateral and assign a value of 0.5 to the former join and 1.0 to the latter. By locating the upper left and lower right corner of each sample standard error rectangle and bisecting the diagonal between them, one can measure the mean Δbc of each feldspar as well as the *within sample* standard deviation of the determination. The results appear in Table 4.

The possible variability of structural state within individual host potassic feldspar grains has been implied by the description of their optical appearance. A mixture of structural states in the powder sample would be reflected in the broadening of certain diffraction peaks of which 131 in the $30^\circ 2\theta$ range is the most diagnostic. It would also be apparent in a reduced number of peaks acceptable to the unit cell computer refinement program. The size of the standard error rectangles in Figure 1 would become somewhat larger as peaks having k and l indices grew more diffuse. Some evidence of these effects is given by the samples analyzed here, particularly those of lower structural state. Perhaps the best ex-

TABLE 4. ANALYSIS OF VARIANCE OF STRUCTURAL STATE IN POTASSIC FELDSPARS FROM MESOPERTHITE

Specimen No.	Mean Δbc	S.D. ($\sqrt{V_2}$) Δbc	Group mean	V_1/V_2^a	F value ^b	D.F.	Significance ^c
Group I							
WA-1A	0.800	0.018					
WA-2B	0.781	0.018					
WA-3A	0.810	0.019					
WA-4A	0.839	0.023	0.814	1.056	3.87	6, 7	NS
WA-4B	0.829	0.014					
WA-5B	0.824	0.015					
WA-7	0.818	0.025					
Group II							
WA-6B	0.888	0.022	0.898	0.500	161	1, 1	NS
68-46B	0.907	0.020					
Group III							
68-15	0.946	0.019	0.952	0.125	161	1, 1	NS
68-18	0.958	0.022					
East Side Profile							
68-46B, 68-15, 68-18			0.937	1.875	4.39	2, 3	NS
West Side Profile							
WA-1 to WA-7			0.824	2.677	3.50	7, 8	NS
Synoptic							
Groups I, II, III			0.855	9.592	4.54	10, 11	VS

^a V_1/V_2 Variance between samples/variance within samples.

^b F values for degrees of freedom (next column) from Snedecor and Cochran, 1967. Table A14, Part I, p. 560. All for 0.05 level except in synoptic case which is 0.01 level.

^c NS not significant, S significant at .05 level, VS significant at .01 level.

ample is provided by 68-18 which plots nearest the maximum microcline-low albite boundary. The width of the 131 peak at half-height in this sample is about 0.7° as compared to 0.3° 2θ in sample WA-2B which has the highest structural state. Furthermore, one of the 68-18 charts shows two minor digitations or subordinate peaks on the flanks of the still overwhelmingly dominant monoclinic 131 peak. If these are interpreted as the 131 and $\bar{1}\bar{3}\bar{1}$ peaks of a triclinic phase stabilized during the ordering process, we can measure a separation of about 0.50° 2θ which would, according to Goldsmith and Laves (1954), give an obliquity

(Δ value) of 0.612. From Table 8 in Orville (1967, p. 83) we see that the Δ of maximum microcline adjusted for a composition of Or_{90} is 0.917. The Δ value thus calculated for 68-18 is in reasonable agreement with the position shown on the b - c plot. It should, nevertheless, be recognized that the 131 and other structurally dependent peaks in most samples is merely a measure of, statistically speaking, the most abundant structural states during a continuous ordering process controlled by falling temperature or other variables. It may be that the ordering process temporarily slowed or stopped in some feldspars, resulting in broader peaks, sometimes with subordinate reflections on the shoulders. The form of the peak would then be the smoothed or integrated curve of a whole set of nested peaks.

In order to ascertain at what level most of the Δbc variance occurs (i.e., within sample *vs.* between sample variance), analysis of variance of the results plotted in Figure 1 was carried out, following the model described by Krumbein and Slack (1956, p. 755 and 759) and standard statistical texts. Table 4 summarizes these data. Samples were grouped according to their position on the $b=c$ diagram (Groups I, II, and III) as well as according to their collection site (east or west side profile on Whiteface Mountain) so that statistical tests on whether a given set of measurements are homogeneous within themselves could be made. The hypothesis to be tested could be stated thus, "There is no significant variation in Δbc measurements between samples compared with the variation within a given sample." Successive tests for each group or set of measurements showed that at the 5 percent significance level, the observed F value (which is the ratio of the variance between samples over the variance within samples) is less than the critical F value for the degrees of freedom involved. It may therefore be stated that in all these cases there is no significant variation in Δbc between samples compared with individual sample variation. It may be further noted that the Δbc variation in samples one meter apart (*cf.* WA-4A and WA-4B) is as small as or smaller than Δbc variation in samples spaced as much as 100 meters apart.

When all samples are compared as a group in the synoptic case, however, it appears that the variation in Δbc between samples compared to within sample variation is significant at the 1 percent level; in other words, there is greater than 99 percent probability that all of the samples do *not* belong to a single normally distributed population. The samples 68-18 and 68-15 deviate most in Δbc (group mean 0.952) from the mean of Group III (0.814) so it is apparent that some statistically nonrandom process is probably responsible for the difference.

INTERPRETATION OF RESULTS

We shall now attempt a geological explanation of first, the variation in structural state of potassic feldspars, and second, the composition or K-Na partitioning of coexisting feldspars in mesoperthite.

The structural state is a measure of ordering of aluminum between the tetrahedral positions of the feldspar structure as outlined by Stewart and Ribbe (1969, p. 440), "with ordering Al goes first into the identical and energetically favored T_1 sites of monoclinic feldspar from the T_2 sites and then into T_1O , mainly at the expense of T_1m ." For monoclinic feldspars (including as a first approximation, those described in this paper), Δbc may be used to estimate the Al in T_1 . Thus the T_1 sites of the feldspars under study here are from 78 to 96 percent filled with Al.

There is an extensive and disputatious literature commenting on the physical and chemical controls on Al/Si ordering in the alkali feldspars of which the tenor is aptly described by the title of Laves and Goldsmith (1961). Goldsmith and Laves (1954) established the approximate temperature (500°C) at which triclinic $KAlSi_3O_8$ inverts to the monoclinic polymorph. Uncertainty arises from the nature of the transformation; if it is displacive rather than diffusive, no definite temperature can be set (see Waldbaum and Thompson, 1968, p. 2016) because displacive symmetry changes are metastable. The microcline-orthoclase transition is affected by the sodium (and possibly calcium) content of the feldspar which may stabilize the higher symmetry form at lower temperatures (MacKenzie and Smith, 1961, p. 63-64), although Laves (1961, p. 65) disputes this point. Heier (1957, 1961) has applied these concepts to metamorphism across the amphibolite-granulite facies boundary in Norway, Evans and Guidotti (1966) to the upper amphibolite facies in Maine, and Scotford (1969) to the greenschist facies in Turkey with some inconsistencies. Heier (1961) discusses most of the problems involved in ordering of the potassic feldspars in terms of the three independent variables operating in any system: T , P , and X . Composition (X) not only includes feldspar bulk composition but also the activity of pure H_2O which may be of great importance as a catalyst¹ as well as defining, of course, the boundary between the amphibolite and granulite facies. Martin (1969) has also shown that peralkaline fluids are influential in the ordering process. Shearing or differential stresses conceivably could have as yet unevaluated crystal chemical effects.

Two main hypotheses have been advanced for the origin of Adirondack

¹ D. B. Stewart (personal communication, 1971) believes the catalysis probably occurs as protons forming OH pairs from tetrahedral oxygens, thus diminishing charge on Al and Si ions and enabling them to migrate.

charnockite: 1) it is a younger and separate intrusive series from the anorthosite series (Buddington, 1969a), or 2) it is comagmatic with anorthosite as most of those familiar with the association worldwide have believed (de Waard, 1969a). Initial $\text{Sr}^{87}/\text{Sr}^{86}$ ratios as determined by Heath and Fairbairn (1969) and Spooner and Fairbairn (1970) for anorthosite and associated charnockites from the Adirondacks fall between 0.7000–0.7056 and are indicative of a primitive magmatic source in the lower crust or upper mantle. The structural states of the potassic feldspars were probably set during the subsequent metamorphic history of the charnockitic gneisses, but the anorthosite-charnockite contact may nevertheless possess some mineralogical significance.

Ordering in the alkali feldspars generally increases as the anorthosite contact is approached (Tables 3 and 4). The samples more than 200 meters from the contact (WA-1, -2, -3) contain the least ordered feldspars of all. An exception is 68-18 from the opposite (east) side of the mountain, which, although nearly 300 meters from the contact, has the lowest structural state (highest Δbc) recorded, and does not appear to be related to any obvious geologic feature such as a fault.

A temperature gradient in the charnockite during intrusion appears improbable as an explanation for the variation in structural state of the alkali feldspars. Intrinsically, as pointed out by Wright (1967, p. 134) "potassium feldspar polymorphs yield relatively little information on the temperature of formation of plutonic rocks" chiefly because of variations in bulk composition and the length of time that the feldspar was held at the temperature of transformation. If igneous temperatures could be predicted by structural state, then the lowest temperatures (because of highest Δbc values) would occur nearest the contact on Whiteface Mountain, but this is not corroborated by temperatures obtained from the microprobe compositions fitted to the Or-Ab solvus as described below. Such temperatures (Table 5) are essentially constant, allowing for analytical and sampling error, across the profile. Furthermore, the observed temperatures of K-Na equilibration in these rocks are well below the temperatures estimated by Buddington (1969b) for unrecrystallized igneous mesoperthite-bearing granites and syenites in the Adirondacks.

The decline in structural state towards the contact could be explained by the increased likelihood of tectonic stress-produced features (granulation, fracturing) in the vicinity of a major structural discontinuity such as a contact. Walton (1953) has noted the greater ductility of anorthosite relative to charnockite; ordinarily, this would promote cataclasis in the charnockite and increased permeability to fluids which might facilitate the ordering process in the feldspars.

On the other hand, the contact may be immaterial and the variation

in structural state could reflect a regional gradient in one or more of the variables enumerated by Heier (1961). A gradient in the activity of water or in cations dissolved in water which would produce a peralkaline pore fluid seems more plausible than the alternatives.

It is apparent that the structural state on the east side of the mountain is markedly higher than on the west (Δbc for samples on the west is <0.9 ; on the east >0.9). The analysis of variance showing greater variation between all samples than within all samples at better than the 99 percent confidence level is considered to reflect primarily this trend. To give an impression of the scale of a possible gradient, the maximum and minimum horizontal distances separating the two sampling profiles are 950 and 285 meters respectively.

The degree to which the Adirondack crystalline rocks are an open or closed system to H_2O and O_2 has been disputed by Buddington (1963) and Taylor (1969), with Buddington finding evidence in the metamorphic assemblages of fixed partial pressure of H_2O from layer to layer whereas Taylor interprets the oxygen isotopic evidence to imply complete exchange. It is notable that there appear to be no abrupt, short range changes in feldspar structural state in the Whiteface Mountain section which would tend to cast doubt on the Buddington hypothesis if one assumed that hydrothermal fluids were of predominant importance in ordering.

From the $b=c$ plot, Figure 1, and from Table 3, it is apparent that the albite molecule content of the potassic feldspars studied varies within a narrow range, and it is unlikely that this factor in itself is responsible for the variation in structural state. No discernible trend is apparent in the compositions obtained from cell volumes, but the most potassic compositions determined by microprobe show decreasing amounts of Ab as the contact is approached, which, taken with the more ordered K feldspars observed there, confirm the relationships shown in Figure 3 of MacKenzie and Smith (1961). Evans and Guidotti (1966, p. 41-42) have noted, however, that the MacKenzie and Smith phase diagram is inconsistent with their results and those of Heier (1961). The slope of the symmetry change for potassic feldspars shown in a schematic $T-X$ diagram by Waldbaum and Thompson (1968, p. 2016) is the opposite of that drawn by MacKenzie and Smith, and is essentially that predicted by Laves (1961, p. 65).

Guidotti (personal communication, 1971) states that work is continuing on the problem in Northwest Maine where he finds the structural state of K feldspar depends upon bulk chemistry of the rock independent of temperature of formation. Crosby in a paper in preparation will present major element analyses of the rocks from which the feldspars studied

in this paper were separated. Preliminary results appear in Crosby (1970). Only one sample (WA-1A) shows a significantly higher Na_2O content (about one weight percent) than the mean of the west side charnockite profile. This higher Na_2O content (which appears to reflect a mode 50 percent higher in plagioclase grains distinct from microperthite than the average of other samples) might have some significance in accounting for the relatively low Δbc (0.800) of this sample, but as above is contradicted by the Waldbaum and Thompson (1968) diagram. Moreover, other samples in the same range (WA-2B, -3A) come from rocks without anomalous bulk Na_2O content.

Recent estimates of the physical conditions prevailing during the metamorphism of Adirondack rocks have been made by Buddington (1969b) and de Waard (1967, 1969b). Buddington considers the temperature range, based upon geothermometers utilizing magnetite-ilmenite equilibria, oxygen isotope equilibria in the quartz-magnetite pair and feldspar compositions to be 600–700°C. De Waard places the *minimum* temperature required for a combination of mineralogical equilibria to be 700°C (1969). He emphasizes that the actual temperature based upon single critical reactions may have been higher, such as the 800°C mentioned in his 1967 paper. The maximum pressure possible is controlled by the absence of kyanite and presence of sillimanite in Adirondack paragneisses. Buddington (1969b) estimates a pressure range of 5–7 kbar for a temperature between 600–700°C based upon the Al_2O_3 - SiO_2 system (see Richardson, Gilbert, and Bell, 1969). De Waard considers that load pressures as great as 10 kb corresponding to a depth of 35 km may have been achieved in most of the Adirondack Highlands.

As an operating assumption, the effective confining pressure for the eastern Adirondacks (where Buddington, 1965, 1969b, notes deepest burial and highest temperatures) will be taken as 7 kbar. This is a limiting case; total pressure may well have been substantially less, especially if case; total pressure may well have been substantially less, especially if equilibrium was set during the dying (retrogressive) stages of the Grenville metamorphic event or at some later time during unloading (*e.g.*, by erosion).

In order to arrive at a temperature of equilibration based upon the compositional data on coexisting feldspars in mesoperthite-microperthite from Table 3, we may compare the Or content of the potassic feldspars with the compositions lying on the binodal curves (solvus) calculated from Margules equations fitted to the data of Orville (1963) and Luth and Tuttle (1966) by Thompson and Waldbaum (1969) and Waldbaum and Thompson (1969). Figure 2 is a temperature-composition diagram for the Ab-Or binary system in which a number of binodal curves

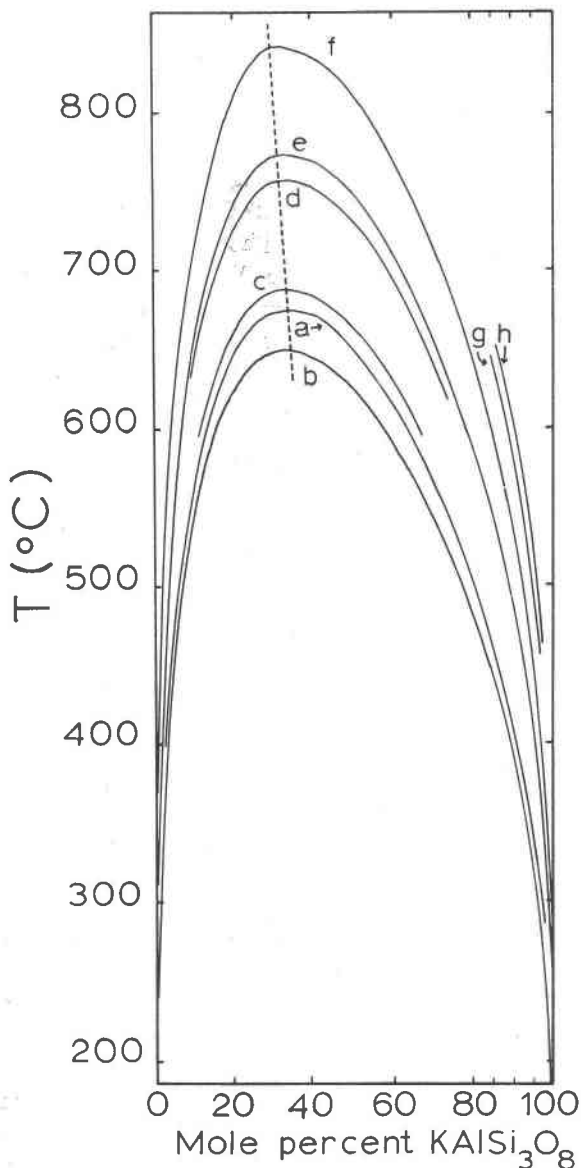


FIG. 2. Calculated binodal curves in the Or-Ab system using the $N_{2\alpha}$ and $N_{2\beta}$ values for various T/T_c ratios from Table 1 of Waldbaum and Thompson (1969) after T_c had been determined for the conditions below. Dashed line is trace of critical point with increased pressure and structural order. Curve a is for sanidine-high albite solid solution (Δbc 0.5) at 2 kbar assuming peralkaline exchange, and is identical with Fig. 13 of Thompson and Waldbaum (1969, p. 832). On Or-rich side, it is coincident with the microcline perthite data

have been drawn corresponding to various compositional, pressure, and structural state assumptions to be described below. As a starting point, the curve from Thompson and Waldbaum (1969, Fig. 13, p. 832) representing the smoothed data at 2 kbar for peralkaline exchange experiments is given.

Some uncertainty exists as to which ratio of alumina to alkalis to use in calculating other curves. If the bulk composition of the entire rock is assumed to be the reactive composition, then the customary presence of garnet (Table 1) dictates a peraluminous ratio. On the other hand, if re-equilibration has occurred in response to late stage intergranular hydrothermal solutions, then the exchange may well take place through a peralkaline medium. An intermediate 1:1 ratio may be assumed if each feldspar is re-equilibrated internally and constitutes a "closed crystal" (Thompson, 1969, p. 343-344). The peraluminous binodal curve has been drawn for 2, 5, and 10 kbar by calculating the pressure effect on the critical point from Equation 3a in Waldbaum and Thompson (1969, p. 1278). The compositional shift of the critical point toward Ab with increased pressure has been interpolated from Figure 6 in Thompson and Waldbaum (1969, p. 823). Points on binodal curves having these critical points have been plotted using the values in Table 1 of Waldbaum and Thompson (1969, p. 1281). As noted by these authors, however, (p. 1279) results can be approximately correct only for highly disordered monoclinic feldspars. Bachinski and Orville (1968) have shown that the solvus of highly Al-Si ordered feldspars has a maximum of 850°C and 30 mole percent Or at 1 bar.¹

The working curve for samples from Whiteface Mountain has accordingly been adjusted by linear interpolation between the critical point for sanidine-high albite and the Bachinski-Orville microcline-low albite value for an intermediate structural state (Δbc of 0.82, the group

¹The critical temperature for microcline solutions may be revised slightly upward by Bachinski and Muller (1971). Waldbaum and Robie (1971) state that the work of Orville (1967), Bachinski and Muller, and their own study all indicate that T_c at 1 bar for microcline solutions is less than 950°C.

←

of Bachinski and Orville (1968) at 375 and 300°C. Curve b: peraluminous exchange at 2 kbar, Δbc 0.5. Curve c: peraluminous data at 5 kbar, Δbc 0.5. Curve d: peraluminous data at 10 kbar, Δbc 0.5. Curve e: peraluminous data at 2 kbar, intermediate structural state (Δbc 0.82) corresponding to average feldspar from west side profile on Whiteface Mountain (Table 4). Curve f: peraluminous data at 7 kbar, Δbc 0.82. Curve g (partial): peralkaline data at 7 kbar for Δbc 0.82, also approximate curve for peraluminous data at 7 kbar and Δbc 0.94. Curve h (partial): peralkaline data at 7 kbar and Δbc 0.94 (the mean structural state of feldspars from East side profile on Whiteface Mountain).

mean of all west side samples from Table 4). This curve is drawn for pressures of 2 and 7 kb.

The curve for K feldspar matrices from microcline perthites in South Dakota pegmatites with compositions of Or₉₂ and Or₉₆ said by Bachinski and Orville (1968) to indicate exsolution occurred at temperatures as low as 375 and 300°C respectively is also included on Figure 2. Partial curves have been drawn in the vicinity of Or₉₀ in Figure 2 corresponding to peralkaline and peraluminous conditions for structural states of Δbc 0.82 and 0.94 (group mean of East side samples) at 7 kbar in order to specify maximum temperatures of equilibration for potassic feldspars exsolved from meso- and microperthite. Temperatures obtained from cell volumes are compared with those derived from electron microprobe analyses in Table 5 for both peraluminous and peralkaline ratios. The temperatures estimated at 7 kbar are adjusted for the variability of structure state, but those given for 2 kbar are not.

Table 5 shows that peralkaline conditions set the ceiling for exsolution temperature or subsequent temperature of recrystallization. If it is as-

TABLE 5
Estimated temperatures⁽¹⁾ of equilibration in
feldspars for various chemical and physical conditions⁽²⁾

Sample	Percent Or	Peraluminous		
		2 kb	7kb	
WEST SIDE PROFILE				
WA-1A	89 (85) ⁽¹⁾	589 (634) ⁽¹⁾	492 (538) ⁽¹⁾	539 (582) ⁽¹⁾
WA-2B	88	601	506	554
WA-3B	90	573	477	525
WA-4A	90	573	477	525
WA-4B	90	573	477	525
WA-5A	88 (90)	601 (573)	506 (477)	544 (525)
WA-6B	92	552	448	516
WA-7	86 (92)	626 (542)	527 (448)	573 (491)
EAST SIDE PROFILE				
68-18	89	607	492	590
68-14	(94)	(525)	(408)	(505)
68-15	90	592	477	572
68-46B	90	585	477	557
Mean		588 (569)	487 (468)	548 (526)
Standard deviation		20.4	21.0	24.1
Two-sided t test				
95% confidence (10 D.F.)		45.5	46.8	53.7
Confidence limits as % of mean		7.7	9.6	9.8

(1) Values in parentheses are based on electron microprobe data if it differs by more than 1% Or from cell volume data

(2) Structural state compensated for in 7 kb columns

sumed that the variation present is random and does not represent any systematic trends, then the mean of all X-rayed samples for peralkaline conditions at 7 kbar is 588°C ¹ as compared with 548°C at 7 kbar and 487°C at 2 kbar for peraluminous reactions. A two-sided *t* test shows that there is 95 percent probability that these temperatures will not vary more than 46, 47, and 54°C from the mean respectively, or as a percentage of the mean value 7.7, 9.6, and 9.8 percent, respectively. The means of temperatures determined from microprobe compositions differing more than 1 percent from cell volume compositions, moreover, lie within one standard deviation of the means derived from cell volume data.

It should be further noted that we have assumed a binary system (Or-Ab) above whereas in actuality the plagioclase phase contains about 15 percent An (Table 3). The temperatures above should therefore be regarded as minimum estimates since it is impossible to predict relative solubilities in the peristerite range. A few microprobe analyses suggest the intergrowth of an An_5 phase with An_{15} .

Wright (1967) has placed the upper limit of stability of maximum microcline (Or_{95}) at $375 \pm 50^{\circ}\text{C}$. The only potential maximum microcline in the Whiteface Mountain charnockite section occurs in the granulated and recrystallized interstices between the mesoperthite-microperthite and plagioclase or other mineral grains observed in some samples as previously described. Some of the finer-grained feldspathic material here shows well-developed grid twinning whereas none of the relic larger microperthitic grains does except in patches. In this they appear to differ from the granulite facies charnockites of Madras which Howie (1955) reports contain dominantly microcline perthite. One may therefore surmise that the feldspars of the charnockitic gneisses of Whiteface Mountain acquired their final texture and mineralogy at a temperature of $375^{\circ} \pm 50^{\circ}\text{C}$ or below.

The perthitic texture of the alkali feldspars from the Whiteface Mountain charnockitic gneisses and the incipient grid twinning visible in many grains is strong evidence of their original homogeneous and monoclinic character which has been modified by prolonged annealing and deformation during subsequent metamorphism. The original feldspar was presumably "m-disordered" in the terminology of Waldbaum and Thompson (1968, p. 2010) and has probably followed a syn- and post-metamorphic path quite similar to the displacive symmetry change curve

¹ If the critical temperature for microcline-low albite solid solutions is taken to be 950°C instead of 850°C , this temperature is increased 57°C and the other temperatures in Table 5 increased in proportion (approximately $52 \pm 5^{\circ}\text{C}$). The revised value of 645°C is very close to the temperature obtained by Buddington and Lindsley (1964) for the magnetite-ilmenite pair in equivalent rocks.

" D_2 ", perhaps shifted somewhat toward " D_3 ", in Figure 6 of Waldbaun and Thompson (1968), who state (p. 2017) that orthoclases following path D_2 "would be monoclinic, triclinic, or ambiguous at room temperature, depending upon the temperature and composition at which it was formed." As the optical and X-ray evidence presented in this paper has shown, many of the Whiteface alkali feldspars have ambiguous symmetry, dependent, in part, on the extent of long range order (cf. Thompson, 1969, p. 357) in the crystal and the techniques available to resolve them (*i.e.*, a scale problem), and in part upon the variables of T , P , and X at the outcrop or subregional level.

Of those samples for which independent geothermometric data are available, the nearest geographically and petrographically to those described here is a strongly micropertthitic garnetiferous syenite gneiss from the Saranac Quadrangle (Buddington and Lindsley, 1964, Table 6, p. 337, sample no. 132). The coexistent magnetite and ferrian ilmenite in this rock give a temperature of $650 \pm 50^\circ\text{C}$ which may be interpreted as follows according to Buddington (personal communication, 1971):

"I thought of the early stage equilibration of the magnetite-ilmenite minerals in charnockite as having occurred at about $750\text{--}800^\circ\text{C}$, just below magmatic [temperatures]. The initial titanomagnetite would unmix by oxyexsolution after formation but with only very [limited] interchange between the intergrowth and isolated independent ilmenite grains. During subsequent *metamorphic* recrystallization [and] during *active* movement, the ilmenomagnetite grains would be re-equilibrated with the independent ilmenite grains at 650°C (granulite facies) . . . Since the temperature [of re-equilibration] varied so systematically regionally, I [had and] have still the impression, that when deforming action ceased so essentially did re-equilibration . . . *If* titanomagnetite is again formed at this stage it must again unmix to ilmenomagnetite by oxyexsolution at somewhat lower temperature but again with little or no interchange of composition with the independent isolated ilmenite grains. Hence the final ilmeno-magnetite grains and the ilmenite grains will give the temperature of the major part of the metamorphism, perhaps slightly below its peak."

Whereas the geothermometric data from the feldspars given here could overlap the standard error range for temperatures determined for the magnetite-ilmenite pair, it is somewhat more probable that internal re-equilibration or unmixing of K-Na between two crystallographically intergrown phases in a perthite might take place at somewhat lower temperatures than those necessary to activate the intergranular reactions which Buddington describes for Fe-Ti oxides and for feldspars. Ordering of Al and Si in the potassic feldspar component of perthite might proceed still further, perhaps to the microcline stability field, as is suggested by the bc diagram, Figure 1, and by the presence of small microcline grains on the borders of micropertthite and microcline-like patches within micropertthite. Whether continuing deformation is necessary to internal re-equilibration as Buddington believes it is to external re-equilibration remains an open question.

Another problem is the significance of the mesoperthite megacrysts in some charnockite layers. In this study, such megacrysts have been lumped together with equidimensional grains, but it is possible a distinction should be made, especially if, as Buddington (personal communication, 1971) suggests, the megacrysts are "post-dynamic" and created by the "release of strains at a temperature not far below that of the dynamic period."

Metamorphic differentiation may be responsible for pink alaskitic micropegmatites noted locally in the charnockitic gneisses of the Whiteface Mountain profiles, and is considered plausible by Buddington at temperatures above 635°C in the pyroxene granulite facies of the central Adirondacks even at very low fugacities of H₂O. He doubts the widespread anatexis in paragneisses postulated by de Waard (1967) who estimates temperatures of 800°C and a H₂O fugacity of 1.5 kb from such evidence as the minimum melting curve for granitic compositions and the breakdown curves for muscovite, biotite, and hornblende.

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