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PROTEROZOIC SANIDINE AND MICROCLINE IN PEGMATITE, WAUSAU COMPLEX, WISCONSIN

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Abstract

The K-feldspar of the pegmatitic pockets in the granitic portion of the anorogenic Wausau complex, central Wisconsin, is unusually varied in its degree of Al–Si order. In spite of a Middle Proterozoic age, high sanidine, orthoclase, intermediate microcline and low microcline occur in pockets in the same pluton. The existence of disordered K-rich feldspar results from a sudden decompression and quench of the near-surface pegmatitic system. Its persistence is a reflection of the sluggishness of the ordering reaction at low temperatures, despite the presence of a mildly alkaline aqueous fluid.

Keywords: high sanidine, low microcline, granite, syenite, pegmatite, decompression, Wausau complex, Wisconsin, anorogenic magmatism.

Sommaire

Le feldspath potassique des cavités pegmatitiques de la portion granitique du complexe anorogénique de Wausau, dans le centre du Wisconsin, montre une grande variation dans son degré d'ordre Al–Si. Malgré l'âge protérozoïque moyen du complexe, on trouve sanidine, orthose, microcline intermédiaire et microcline ordonné dans les cavités d'un même pluton. L'existence de la forme désordonnée résulterait d'une décompression soudaine et d'une trempe dans ce système pegmatitique, mis en place à faible profondeur. Sa persistance résulterait de la lenteur de sa mise en ordre à faibles températures, même en présence d'une phase aqueuse légèrement alcaline.

Mots-clés: sanidine désordonnée, microcline ordonné, granite, syénite, pegmatite, décompression, complexe de Wausau, Wisconsin, magmatisme anorogénique.

Introduction

Water-saturated felsic magmas emplaced under plutonic conditions crystallize with a pegmatitic texture, generally at a temperature above 600°C. Their primary K–rich feldspar initially has a disordered distribution of Al and Si, and would be expected to contain a significant amount of Na. Because 1) the rate of cooling typically is very slow, and 2) water is available during cooling, perthitic (*i.e.*, exsolved) ordered microcline is characteristic of almost all bodies of pegmatitic granite (Martin 1982). Albite- and pericline-twinned domains attest to the primary crystallization of a monoclinic precursor. Both the perthitic texture and the twin-related domains in pegmatitic microcline are coarse compared to those in microcline from other environments; this coarseness is another consequence of very slow cooling and efficient recrystallization in the presence of an aqueous fluid medium.

The above undoubtedly applies to almost all granitic pegmatite systems, but exceptions can be expected if the normal slow cooling is interrupted, as is likely in an epizonal complex. We describe here the pegmatitic pockets in a late phase of the Wausau anorogenic syenite-granite complex, in central Wisconsin. These contain a surprising association of high sanidine, orthoclase and microcline, all in close proximity. We propose an explanation for the anomalous occurrence of disordered K-rich feldspar in a pegmatitic environment.

Setting of the Wausau Complex

The Wausau granite–syenite complex forms part of a NE–SW belt of Proterozoic anorogenic alkali granite – syenite – anorthosite – gabbro suites strung out from Labrador to the southwestern U.S. (Emslie 1978, Anderson 1983). It consists of two separate ring-structures that show opposite trends in silica saturation. The southern body is dominated by syenite and quartz syenite (Wausau and Rib Mountain plutons) in the rim, and alkali granite (the Nine Mile pluton) in the core (Fig. 1A). The ring structures, emplaced in metavolcanic, metagranitic and metatonalitic rocks of the 1.8- to 1.9-Ga Central Wisconsin complex (Van Schmus 1980), are closely related to the Wolf River granitic batholith to the east (Anderson 1980); together, they define a whole-rock Rb–Sr isochron of 1468 ± 35 Ma, $(^{87}Sr/^{86}Sr)_0$ = 0.7048(17), and a U–Pb (zircon) concordia intercept of 1485 ± 15 Ma (Van Schmus *et al.* 1975, Van Schmus & Bickford 1981). The Wausau complex is cut discordantly by a system of east–west-trending diabase dykes and concordantly by numerous pegmatite dykes, veins and lenses, generally gently dipping and of small size. Although well mapped (*e.g.*, Sood *et al.* 1980, LaBerge & Myers

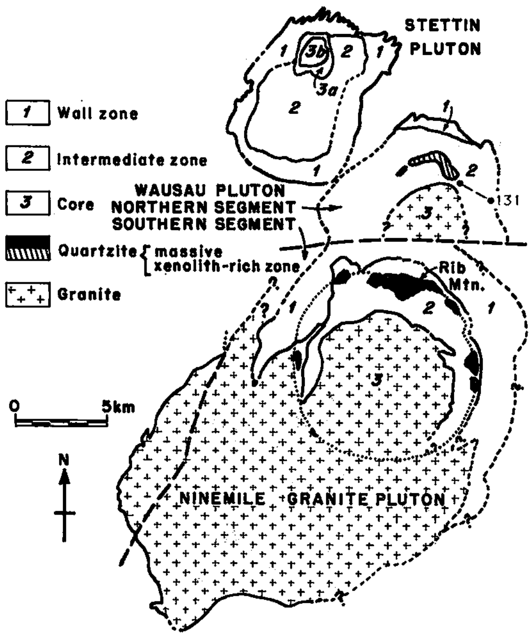

FIG. 1A. Generalized geological map of the Wausau complex in central Wisconsin (Sood *et al.* 1980), showing the silica-undersaturated Stettin pluton to the north and, south of it, the silica-oversaturated Wausau pluton. Most of the samples come from the Nine Mile granite, which forms the core zone of the southern segment of the Wausau pluton. Note location of sample 131, in the northern segment of the Wausau pluton. The map was prepared by Paul E. Myers.

1983, Myers *et al.* 1984), little systematic work has been carried out on the rock-forming minerals.

Falster (1981) briefly described the pockets in the pegmatite veins and lenses. The mineral assemblages may vary in detail from pocket to pocket, and consist of several mineral species that contain Fe, Ti and incompatible elements, *i.e.*, those that have systematically been partitioned into the felsic magma and, ultimately, into the aqueous fluid medium. Hundreds of such pockets have been mapped and sampled over an area of 5 square kilometres. This paper summarizes what is known currently concerning the feldspar mineralogy of these pockets.

Types of Pegmatite Dykes, Veins and Lenses

The study of pegmatite bodies is hampered by poor exposure owing to swamps, vegetation and low relief. The second author has nevertheless documented over 800 pegmatite bodies in the last decade. On average, pegmatite veins measure 15 × 15 × 0.4 metres, whereas the average miarolitic pocket is 0.15 m across. On the basis of careful observations and maps, the pegmatitic bodies can be classified into four groups. Some bodies show a transitional character.

Simple, schlieren-like masses (group 1)

These masses are common in the vicinity of larger pegmatite dykes. Internal zoning is generally not well developed; instead, the grain size of the feldspar and accessory quartz increases gradually from an aplitic contact-zone toward the centre, with the coarsest material in the hanging wall above the centre. The dominant minerals in the pocket zone are microcline, albite, quartz, with accessory siderite (commonly replaced), hematite and hisingerite. More rarely, phenakite, anatase, sulfides and sulfosalts appear. The samples from the following pockets are representative of group 1: 9a, 18, 35, 216 (see Fig. 1B for pocket location).

Complexly zoned larger dykes (group 2)

Dykes of group 2 are generally larger than 10 × 10 × 0.2 m and exhibit well-defined internal zones. The wall zone is more or less uniform in grain size and mineralogy. The intermediate zone commonly varies more drastically, from almost monomineralic K-feldspar to graphic (K-feldspar + quartz) to aplitic units. Some aplitic bodies are cross-cutting. The core zone consists of large masses of quartz with centimetre-size crystals of biotite at the margin of the zone. The pocket zone typically contains coarse-grained euhedral crystals and may show signs of periods of crystal growth alternating with episodes of crystal breakage. The resulting breccia commonly is cemented with a later generation of pocket minerals.

In addition to the minerals already mentioned, such pegmatite bodies contain pyrite, sphalerite and galena, sulfosalts (*e.g.*, jamesonite, boulangerite), calcite, the Be minerals, phenakite, bertrandite, bavenite, euclase and beryl, and the *REE* minerals cheralite, monazite, xenotime and synchysite-parisite. Considered representative of group 2 are pocket numbers 15, 61a, 73, 156d, 156g and 200 (Fig. 1B).

Simple dykes with an extensive vuggy region (group 3)

These generally are large dykes, exceeding 100 metres in length. Internal zonation is not as well defined as in group 2. Wall zone and intermediate zone are marked by coarse patches of quartz, as might be found in the core zone of some pegmatite bodies. Large, central pockets are rare; instead, one finds myriads of isolated tiny cavities up to 1.5 cm across in the intermediate zone. The cavities are not attributed to a late-stage leaching process, but to the

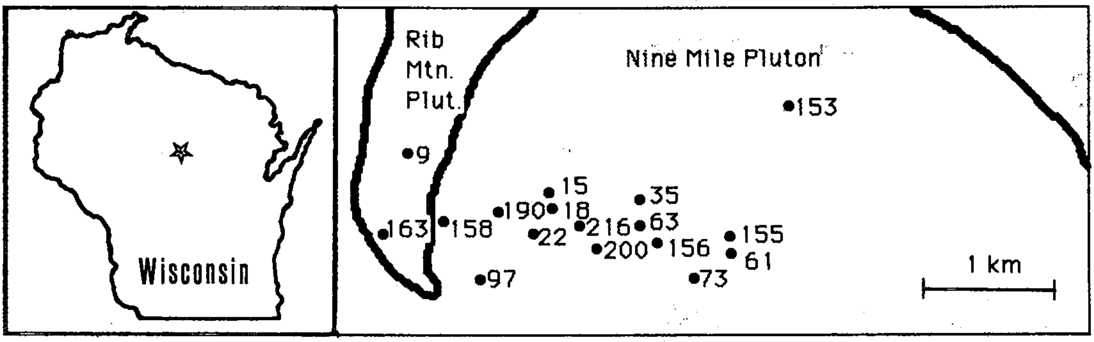

FIG. 1B. Location map of the Wausau area in Wisconsin, and location of samples in the area of the Nine Mile granite (unit 3 in Fig. 1A) inside (and to the east of) the ring of Rib Mountain quartz syenite (unit 2 in Fig. 1A).

inability of the bubbles of fluid to coalesce in the magma (owing to a lack of time?) to form one central pocket. The dominant minerals are microcline, quartz and albite. Hematite is also dominant and characterizes pegmatites of group 3. Accessory phases are sparse: fluorite, zircon, fluorapatite and cheralite. No example of this category was investigated.

Pegmatite bodies with late-stage selective etching and crystallization of accessory minerals (group 4)

These pegmatitic bodies resemble those of group 2 but differ in that they typically show an inward decrease in the proportion of quartz. Quartz has been selectively removed near (and especially below) the pocket zone, leaving vuggy masses of feldspar. Secondary growth of K-rich feldspar produced crystals having the adularia habit; this is almost always accompanied by a characteristic assemblage of rutile, anatase, brookite, zircon, fluorapatite, ilmenite, hematite, muscovite, cheralite, monazite and, in small amounts, second-generation quartz. Representative samples of such pockets are 22, 97, 153a, 163, 190a and 190e (Fig. 1B).

PREVIOUS WORK AND THE APPROACH USED

Some detailed information has been published on adularia from Wausau. Černý & Chapman (1984) studied four samples from pockets 61, 63 and 69 (pegmatite dykes of group 2). They refined the cell dimensions of the K-feldspar from powder X-ray-diffraction data, and found that all four consist of nonperthitic low microcline. This finding was confirmed for one sample by infrared absorption spectroscopy. An optical examination showed the characteristic grid pattern of microcline inverted from a monoclinic precursor. A chemical analysis by electron microprobe revealed a composition of 98 (§61) and 99% Or (§63).

The approach used here is X-ray diffraction (powder method). Twenty-five samples were selected to represent the various textural variants of the K-feldspar. Compositional and structural indicators (Table 1) are determined from the cell constants, refined from carefully indexed peaks corrected against a synthetic spinel standard (Guinier – Hägg focusing camera, Cu$K\alpha_1$ radiation). The results

TABLE 1. INDICATORS OF COMPOSITION AND DEGREE OF Al-Si ORDER IN K-FELDSPAR, POCKET ASSEMBLAGES, WAUSAU SUITE

brief description of sample		Or(b^*c^*)	Or(V)	t_10	
9a	quench coating (qtz)	1.07	1.05	0.29	HS
15	blocky euhedra (ab, qtz)	1.00	0.98	1.00	LM
15	small, dusted euhedra	1.00	0.96	0.98	LM
18	amazonite euhedra (ab, qtz)	1.00	0.97	0.99	LM
22	perthite euhedra (ab)	1.01	0.98	1.00	LM
35	amazonite euhedra (ab)	1.01	0.98	1.00	LM
61a	glaze	1.00	0.97	0.99	LM
61a	loose overgrowth (qtz)	0.99	1.00	0.31	HS
63	drusy crust	1.00	0.97	0.99	LM
63	spherulitic material	1.03	1.00	1.00	LM
73	orange overgrowth (ab, qtz)	0.99	0.97	1.00	LM
97	pinkish feldspar (ab)	1.02	0.98	0.97	LM
131	corroded breccia	1.01	0.98	0.99	LM
	corroded breccia	1.02	1.00	0.32	HS
153a	glaze	1.01	0.97	1.00	LM
155	glazed euhedra	1.00	0.97	1.01	LM
156d	muddy coating (ab)	0.97	0.99	0.33	HS
156d	folded crusts (ab)	0.96	0.98	0.38	OR
156g	graphic material (ab, qtz)	1.04	1.00	1.01	LM
163	curved plates (ab)	1.00	0.97	0.97	LM
190a	cracked crystals (qtz)	1.00	0.98	0.92	IM
190e	crust (qtz)	1.09	1.03	0.29	HS
190e	coarse brown crystals (ab)	0.99	0.97	0.99	LM
200	crust (qtz)	1.04	1.01	0.32	HS
200	perthitic substrate (ab, or)	0.99	0.97	0.99	LM
216	loose overgrowth (ab, or)	0.97	0.95	0.27	HS

Or(b^*c^*), the Or content (mole %) based on co-ordinates in the b^*-c^* diagram, is calculated using the end members of Kroll & Ribbe (1983) and the equation of Blasi (1977). Or(V), the Or content based on the unit-cell volume, is calculated using the equations of Kroll & Ribbe (1983) (for the LA-LM or the HA-HS series). An Or content greater than 1 may indicate an expanded cell owing to the incorporation of larger-radius cations than K and Al. The degree of Al-Si order is quantified by t_10, the proportion of Al in the T_10 position, calculated according to the equations of Blasi (1977). In low microcline, t_10 is 1, whereas in high sanidine, it is 0.25. Estimated maximum standard errors: ±0.005 in both Or and t_10. See Fig. 1 for sample locations. Nomenclature: high sanidine HS: $0.5 < 2t_1 < 0.666$; low sanidine (not encountered): $0.667 < 2t_1 < 0.74$; orthoclase OR: $0.74 < 2t_1 < 1.0$ (Ribbe 1983); intermediate microcline IM: $t_10 \leqslant 0.92$; low microcline LM: $0.92 < t_10 \leqslant 1.0$. In parentheses, coexisting phases in small sample X-rayed.

shed light on the dominant feldspar (or feldspars) in a small (approximately 1 mm^3) representative specimen. Those that are perthitic are shown to coexist with albite in Table 1. The complete set of refined cell-parameters is available at nominal cost from the Depository of Unpublished Data, CISTI, National Research Council of Canada, Ottawa, Ontario K1A 0S2.

Results of the X-Ray-Diffraction Analysis

Samples were selected for study to evaluate differences in K-feldspar mineralogy in pockets of different types of dyke and at different stages of crystallization in a given pocket. Some pockets contain evidence of three generations of K-feldspar. The outer zone typically contains coarse, blocky euhedral individuals of microcline perthite. These are commonly coated with a glaze-like nonperthitic overgrowth of microcline. In some cases (*e.g.*, 63), the overgrowth is thicker and forms small, drusy, ball-like clusters. The hollow centre may have been occupied by a mineral since removed by solution. In pocket 63, the drusy crust of microcline coats a fan-spherulite of microcline in a mosaic of narrow platy domains; the array is subtly color-banded perpendicular to the direction of the radially disposed plates, which are 5 cm long. The texture and composition may indicate growth from a feldspathic gel formed by slow decompression in that pocket. Some pockets (*e.g.*, 131) contain a brecciated aggregate of microcline perthite, locally slightly corroded before deposition of the overgrowth. Finally, and as a check of Falster's claim (1981) that sanidine occurs in some pockets, samples of indurated mud (*e.g.*, 156d), locally slumped, were studied; these deposits cover the glaze-like overgrowth of microcline and other crystals in the pockets.

In almost all samples, the K-feldspar has attained complete Al–Si order. This can be seen in a plot of

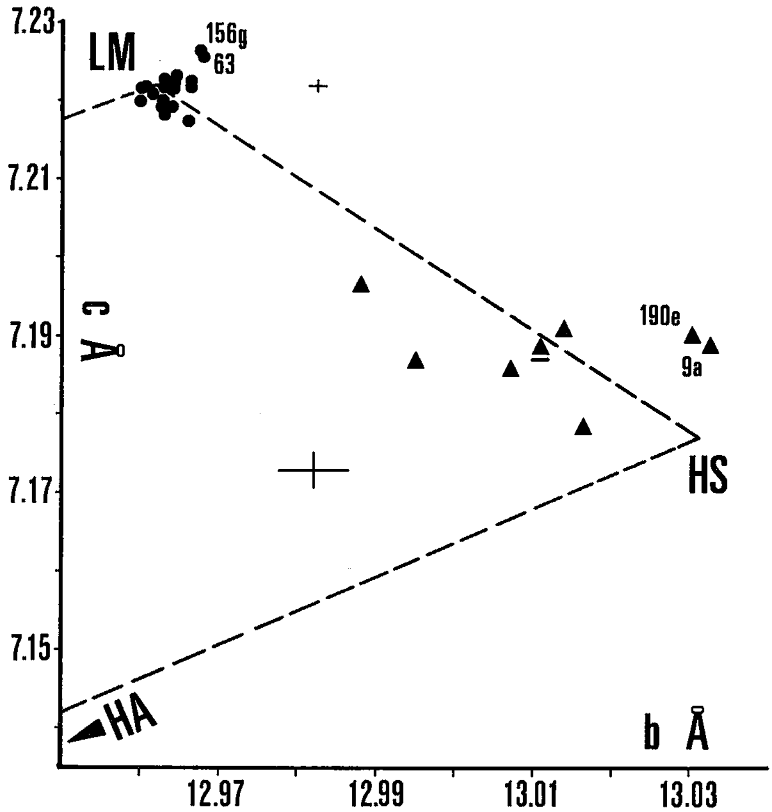

FIG. 2. The distribution of K-feldspar data-points in a plot of *b versus c* cell dimensions. An average error-bar is shown to the right of the microcline data-points and to the left of the sanidine data-points, respectively. Symbols: LM low microcline, HS high sanidine, HA high albite. The co-ordinates of the end members are those of Kroll & Ribbe (1983, Table 4).

b versus c (Fig. 2) and α^* *versus* γ^* (Fig. 3). In Figure 2, note that some points fall outside the $b - c$ quadrilateral, *i.e.*, they have larger-than-expected values of the *b* and *c* cell edges. A possible explanation for this anomaly is that the *b* and *c* cell edges are expanded because of the incorporation of cations larger than K and Al in the structure. Rb, Cs and Pb become enriched at the final stages of crystallization of a granitic body (Černý *et al.* 1985a); also, Fe^{3+} may replace Al in the microcline structure (Wones & Appleman 1963). The two samples of amazonitic microcline (18 and 35) contain, respectively, 1000 and 3000 ppm Rb, 50 and 150 ppm Cs, 150 and 300 ppm Pb (E.E. Foord, priv. comm. 1983). These concentrations of Cs and Pb are too low to account for the anomalous co-ordinates in the $b - c$ plot; Černý *et al.* (1985b) have shown that rubidium substitution for potassium in low microcline causes an expansion in *a*, but none in *b* and *c*.

If a large-radius cation is responsible for the anomalously large values of *b* and *c*, *a* (and a^*) should be anomalous as well. A plot of a^* *versus* Or(b^*c^*) (Fig. 4) shows that the anomaly involves b^* and c^* much more than a^*. More work will have to be done to determine whether a structurally bound large cation, possibly Fe^{3+}, is locally significant. All samples are brick-red; some of the hematite could represent the product of a conversion of ferriferous microcline to normal microcline, but this is conjectural.

The $\alpha^* - \gamma^*$ plot (Fig. 3) shows that not all samples of microcline samples are fully ordered. Note that the symbol is roughly equivalent to one standard deviation in α^* and γ^*; samples 97 (t_1O 0.97) and 190a (t_1O 0.92) consist of microcline that is slightly disordered. These two samples come from the vicinity of the contact of the Nine Mile pluton against an arcuate septum of the older syenite (Fig. 1). The occurrence of imperfectly ordered microcline near the intrusive contact probably indicates that the rate of cooling was too rapid there to allow a complete recrystallization of the high-temperature feldspar. More detailed sampling of the K-feldspar in the pockets may show that the degree of Al–Si order could be a property that can be contoured near the contact; this would be useful to determine the sequence of emplacement of the units in such a near-surface complex.

The triangles in Figure 2 show the co-ordinates of the monoclinic K-feldspar, orthoclase ($0.74 < 2t_1 < 1.0$) or high sanidine ($0.5 < 2t_1 < 0.666$), using the definition of Ribbe (1983; see also Table 1). With one exception, this material represents the late deposit of K-feldspar that coats everything in a given pocket. In the exception (131, underlined triangle), the orthoclase coexists with microcline (and albite) in perthite and thus represents a case of arrested

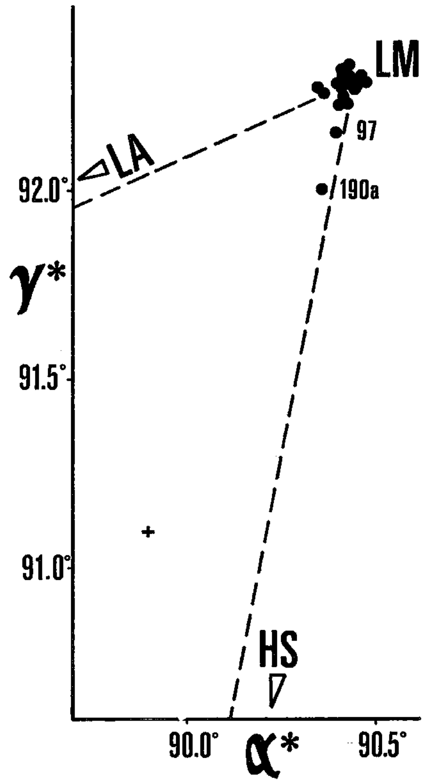

FIG. 3. The distribution of K-feldspar data-points in a plot α^* *versus* γ^*. All monoclinic samples plot at $\alpha^* = \gamma^* = 90°$. The diagram illustrates the slight disorder in some samples of microcline. The average error-bar is approximately the size of the symbol used. Symbols: LM low microcline, HS high sanidine, LA low albite. The co-ordinates of the end members are those of Kroll & Ribbe (1983, Table 4).

inversion of the primary monoclinic K-feldspar. It probably is no accident that the sample is a microbreccia, testifying to the sudden evacuation of the pocket as a result of rupture of the wall. The other triangles represent the loosely consolidated deposits that resulted from the sudden decompression. There is quite a spread in the degree of Al–Si order of these samples, high sanidine in most cases. Two of the most disordered samples, 9a and 190e, have a larger *b* and *c* than expected; hence their value of Or(b^*c^*) is anomalously large (Fig. 4), for reasons that are still not clear, as outlined above. The variability in degree of order illustrates, in our opinion, the independence of each pocket.

All the samples of K-feldspar examined are rich in K and poor in Na (Table 1, Fig. 4). The composition of most samples, based on unit-cell volume, exceeds 96% Or. Where such a composition coex-

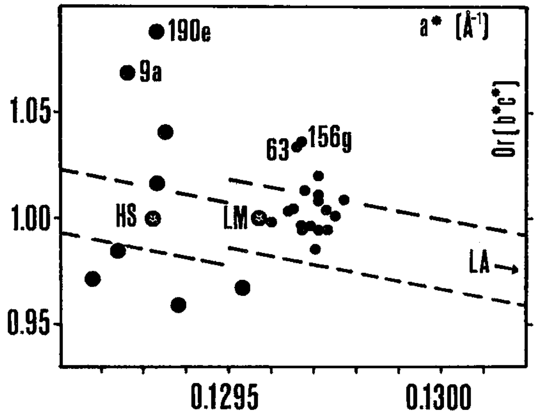

FIG. 4. The distribution of K-feldspar data-points in a plot of a^* *versus* Or(b^*c^*). Both are indicators of composition in the binary system Ab-Or. The small dots refer to microcline, which plot slightly to the right of end-member low microcline (LM), indicating the presence of a small amount of sodium in the structure. The large dots refer to samples of sanidine, shown with respect to end-member high sanidine. Samples considered normal (*i.e.*, unstrained) should plot within the dashed lines, drawn at ± 0.15 Or(b^*c^*) with respect to the lines LM–LA and HS–HA (co-ordinates of Kroll & Ribbe 1983, Table 4).

ists with albite, one can infer that the pair continued to exchange alkalis to a low temperature (at least to 250°C). The sanidine that deposited suddenly upon rupture of the pockets generally coexists with quartz and not with albite (pocket 156d is the exception). The sanidine + quartz assemblage may be close in composition to the solute in the granite system at the time of decompression, although it is conceivable that an ion-exchange reaction could have removed Na and added K at a still lower-temperature stage of pocket evolution.

DISCUSSION

The Wausau complex of plutons is genetically related to the nearby Wolf River granitic batholith (Van Schmus *et al.* 1975). Anderson (1980) showed that this massif is anorogenic and mildly alkaline. Anderson concluded that the Wolf River granitic magma crystallized in the upper crust, at a depth less than 3.8 km and in the range 790 to 640°C. Although a similar detailed study has not been carried out on the Wausau complex, there is no reason to believe that conditions were significantly different. A temperature of final crystallization of the magma near 640°C implies that the granitic melt must have approached water saturation at such a low pressure. The Wausau complex and, in particular, the Nine Mile pluton, show clear evidence of the achievement of water saturation before primary crystallization was complete. The myriads of tiny isolated cavities in dykes of group 3 suggest sudden saturation, with no time for coalescence of the vapor bubbles into a larger pocket. The differences noted between four main groups of pegmatitic dykes and also between different examples of the same group indicate that each pocket behaved as an individual system that may or may not have remained closed during the early subsolidus history of the pluton.

Falster (1981) has described the general features of the outer part of the pegmatitic pockets. The coarse euhedral crystals that line the cavities may be intact, coated with a glaze of single-phase low microcline deposited from a fluid phase, and may have been later shattered and brecciated as a result of a violent shock, possibly more than once, presumably indicating a buildup of internal pressure over confining pressure. The opportunity of a sudden decompression would be commonplace in such a shallow environment, while parts of the magmatic complex were still being emplaced. The result is a rapid deposition of the solute carried at the time of decompression by the aqueous fluid medium, mainly as high sanidine + quartz. In some cases, the attendant cooling prevented any further change, so that the degree of order remains negligible, even to this day. It is known that the midcontinent region has been a stable cratonic block since the emplacement of these anorogenic plutons (Van Schmus & Bickford 1981), and the presence of such highly disordered metastable K-rich feldspar provides convincing proof of this. An event of mild reheating would allow such a metastably preserved feldspar to order. A greater degree of Al-Si order can be obtained in the laboratory after two or three months than is found in some of the pockets at Wausau (Martin 1968).

In some pockets, the rate of cooling after decompression was slower, such that the K-feldspar was able to achieve a greater degree of order, though still remaining high sanidine (Fig. 2). Unfortunately, useful geothermometric data cannot be derived from the feldspar assemblage. The disordered sanidine did not deposit at a high temperature, contrary to what one may infer by reference to a diagram linking the degree of Al–Si order to temperature of formation (*e.g.*, Stewart & Wright 1974). As it loosely covers the microcline overgrowth, it was deposited below 400°C. The temperature of deposition could probably best be determined by fluid-inclusion thermometry on the accompanying quartz.

The sanidine that crystallizes from a felsic magma in a volcanic complex typically is efficiently transformed to orthoclase or low microcline as a result of feldspar – water interaction at the deuteric stage. Even in rhyolites of Mesozoic or Cenozoic age, high sanidine is exceedingly rare. Thus the persistence of high sanidine as a constituent of a pegmatite pocket of Proterozoic age or, even more surprising, of Archean age (Černý & Chapman 1984) is highly unusual. Foord & Martin (1979) and Martin (1982) suggested that the occurrence and persistence of high sanidine reflect 1) the sudden decompression of the pocket, with forced nucleation and disequilibrium growth from the solution on all available surfaces, leading to a poorly consolidated deposit that locally shows slumping, 2) the sealing of the pocket by deposition of material along the fissures that had allowed the fluid under pressure to escape from the pocket environment, 3) rapid cooling of the local environment, and 4) eventual re-entry of water into the system at a temperature too low for efficient ordering of the high sanidine to low microcline. This reflects the very low solubility of the K-rich feldspar at a low temperature, even though its pH may have been buffered on the alkaline side by the bulk composition of the Nine Mile granite. An alkaline fluid medium is known to hasten the ordering reaction, but even in such systems, ordering becomes very sluggish at temperatures below 250°C (Martin 1974).

If the folded and slumped aphanitic crusts are truly deposits of quenched solute, a careful study of their bulk composition would seem to offer an excellent opportunity to document the increasing level of enrichment of the highly incompatible elements in the fluid medium of a typical anorogenic granitic system after water saturation. At Wausau, the sanidine-bearing quench products occur in dykes of groups 1, 2 and 4 (samples of group 3 were not investigated). A geochemical "ranking" of the deposits should be possible, and could perhaps explain why some contain high sanidine + quartz without albite whereas others contain high sanidine + albite without quartz, and why the mineralogy of some pockets is considerably more varied than that of others.

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