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HYDROTHERMAL ALTERATION AND DYNAMIC RECRYSTALLIZATION OF FELDSPAR IN AN ARCHEAN IRON-FORMATION

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ABSTRACT

Associated with iron-formation at the Adams mine, Kirkland Lake, Ontario, are massive, thickly or thinly layered quartz-feldspar rocks, here called the feldspathic quartzite. Both layered and massive feldspathic portions contain grains of composite alkali feldspar that exhibit two types of intergrowth textures: type 1 is an intimate "zone perthite", and type 2 is a twinned end-member albite core sharply rimmed by end-member K-feldspar. Some K-feldspar (Ksp I) is untwinned and contains substantial Ba; some other K-feldspar (Ksp II) is twinned and contains little or no Ba. Fine, discrete grains of K-feldspar and albite occur in the matrix, and albite also occurs as recrystallized grains on the margins of the K-feldspar rim of coarse feldspar grains. Rock textures and chemistry suggest a precursor of rhyolitic ash for the feldspathic portions. Values of $\delta^{18}O$ quartz show that (a) quartz in feldspathic rocks is magmatic, (b) thick quartzose layers were precipitated as chert in isotopic equilibrium with warm marine water (~70°C), and (c) thin quartzose layers were precipitated as chert but have variably re-equilibrated with neighboring feldspar, probably due to very fine grain-size. Values of $\Delta^{18}O_{qtz-feld}$ for feldspathic rocks yield temperatures in the range of 150–170°C (± 10°C). This range, coupled with a type-2 texture, end-member compositions of the two feldspars and the high Ba content of Ksp I, suggest that Ksp I formed by hydrothermal replacement of albite at $T < 150°C$. Some Ksp I was later converted to Ksp II, expelling barium. Finally, a limited back-reaction of K-feldspar to new albite occurred, probably related to mild deformation in the area.

Keywords: alkali feldspar, Archean, barium, chert, hydrothermal alteration, iron formation, oxygen isotopes, recrystallization, Kirkland Lake, Ontario.

SOMMAIRE

On trouve, associées à la formation de fer dans la mine Adams, à Kirkland Lake (Ontario), des roches quartzofeldspathiques massives, en strates minces ou épaisses, que l'on traite ici de quartzite feldspathique. Les portions feldspathiques stratifiées aussi bien que massives contiennent des grains composés de feldspath alcalin qui montrent une de deux textures d'intercroissance: type 1, perthite en zones fines; type 2, coeur d'albite maclée recouvert d'un liseré de feldspath potassique. Dans certains cas, le feldspath potassique (I) paraît sans macle et possède une teneur en Ba appréciable. Dans les autres (II), le feldspath potassique est maclé et contient très peu ou pas de baryum. De petits grains séparés d'albite et de feldspath potassique sont présents dans la pâte, et on voit aussi des grains recristallisés d'albite en bordure du liseré de feldspath potassique sur les grains les plus gros. Les textures et le chimisme des roches indiqueraient. comme précurseur de la fraction feldspathique, une cendre rhyolitique. Les valeurs de $\delta^{18}O$ pour le quartz montrent que a) ce minéral serait magmatique dans les roches feldspathiques, b) les strates quartzeuses épaisses étaient à l'origine du chert en équilibre isotopique avec de l'eau de mer réchauffée à environ 70°C, et c) les couches quartzeuses plus minces ont aussi été précipitées comme chert mais ont eu l'occasion de se rééquilibrer, plus ou moins, avec les grains de feldspath adjacents, probablement à cause de leur granulométrie très fine. Les valeurs de $\Delta^{18}O_{qtz-feld}$ des roches feldspathiques indiquent des températures entre 150 et 170°C (± 10°C). Cet intervalle, le développement de la texture du type 2, la grande pureté des deux feldspaths et la teneur élevée en Ba du feldspath potassique I seraient des conséquences de sa formation par remplacement hydrothermal de l'albite à une température inférieure à 150°C. Une partie du feldspath potassique I aurait ensuite été transformée en feldspath potassique II, avec lessivage du baryum. Finalement, la réaction inverse du feldspath potassique, pour produire une albite nouvelle en quantités limitées, serait le résultat d'un épisode de faible déformation.

(Traduit par la Rédaction)

Mots-clés: feldspath alcalin, archéen, baryum, chert, altération hydrothermale, formation de fer, isotopes d'oxygène, recristallisation, Kirkland Lake, Ontario.

INTRODUCTION

The occurrence of extensive rock-water chemical interaction during volcanic activity has been thoroughly documented during the last several years; such interaction has been shown to be a significant contributor to the present-day composition of many igneous rocks (*e.g.*, Spooner & Fyfe 1973, Bischoff & Dickson 1975, Andrews 1977, Humphris & Thompson 1978a, b, Fyfe & Lonsdale 1981). Massive and layered quartzofeldspathic rocks are present in the immediate stratigraphic hanging-wall of the Adams mine iron-formation at Kirkland Lake, Ontario. Examination of these unusual rocks has revealed evidence for a high degree of modification of the feldspars, probably by Archean marine water.

This was followed by an additonal change in composition induced by deformation. The aims of this paper are (a) to describe the inferred chemical changes in the feldspars, (b) to determine the conditions under which these changes may have taken place, and (c) to comment on the timing of the alteration and recrystallization with respect to the volcanism and chemical sedimentation, especially in light of the exhalative theory for the origin of banded iron-formations (Ridler 1973).

GEOLOGICAL SETTING

The Adams mine iron-formation is a member of the Boston Township iron range, a group of volcanic and volcanogenic rocks of Archean age, near Kirkland Lake, Ontario, in the Superior Province of the Canadian Shield (Fig. 1). Iron formation at the Adams mine consists primarily of alternating layers of magnetite and fine-grained quartzite, the latter having been interpreted as recrystallized chert (Dubuc 1966). In addition, layers of silicate-facies iron-formation and tuff are also common. Near the Lebel syenitic stock, skarn mineralogy has developed from a silicate-facies iron-formation precursor. Alternating with individual horizons of iron formation are volcanic flows and pyroclastic rocks ranging from ultramafic to intermediate in composition. Descriptions of the Adams mine and surrounding area can be found in Dubuc (1966) and Jensen (1978). Citing the occurrence of large inclusions of iron formation within flows, Jensen (1978) concluded that volcanism and chemical precipitation of the iron formation were contemporaneous.

At the top of the main horizons of iron formation are volcanic–sedimentary units rich in quartz and grains of alkali feldspar, collectively designated the "cherty quartzite" by mine personnel (Jensen 1978), but referred to in this paper as feldspathic quartzite. The spatial association of the feldspathic quartzite with the iron formation suggests that the two are genetically related. This is supported by the

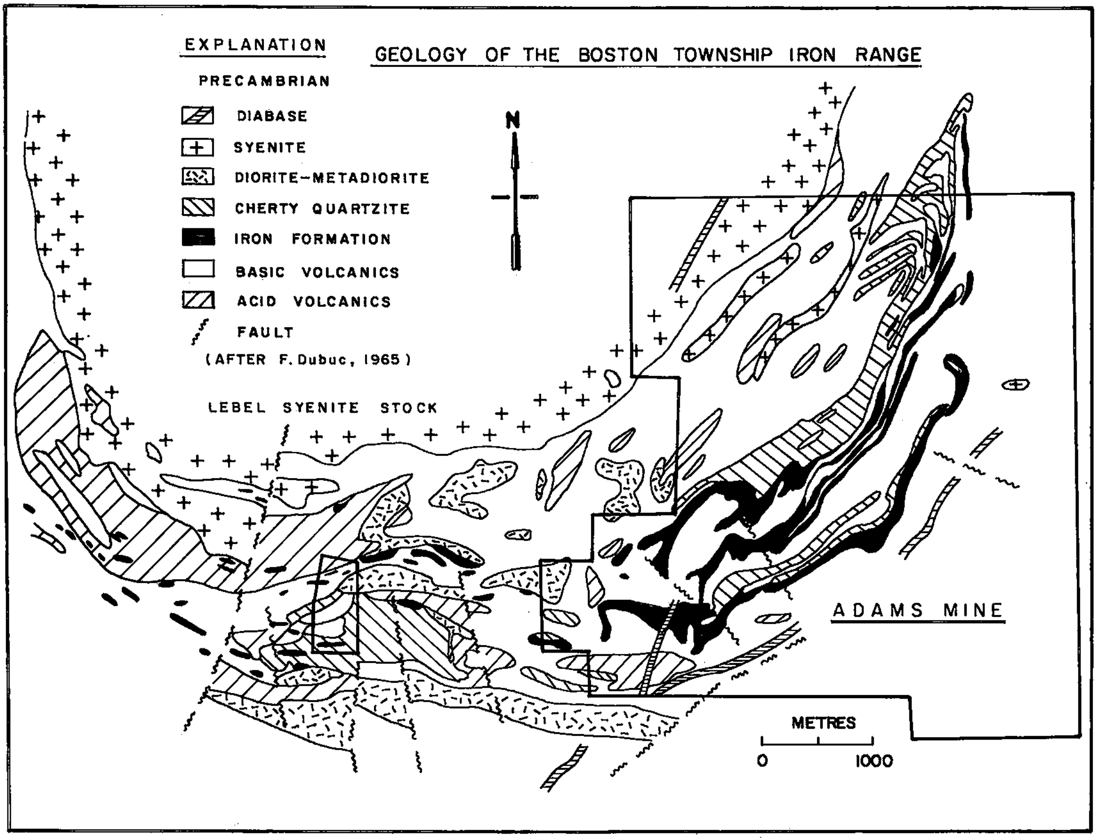

FIG. 1. Geology of the Boston Tonwship area, including the Adams mine (after Dubuc 1966). The cherty quartzite unit mapped by Dubuc is referred to in this paper as the feldspathic quartzite (see text). Note the conformity of the volcanic–sedimentary sequence to the margin of the Lebel stock.

occurrence of layers of micaceous clastic sedimentary rocks within the iron formation, implying that volcanism, clastic sedimentation and chemical precipitation proceeded at the same time.

GENERAL DESCRIPTION OF THE FELDSPATHIC QUARTZITE

The feldspathic quartzite consists almost entirely of quartz and grains of composite alkali feldspar composed of K-feldspar + albite. Accessory minerals include amphibole, chlorite, muscovite, titanite, pyrite and magnetite. Rocks range from massive to layered; where present, the layering occurs on a scale of two mm (thin) to several cm (thick), and consists of layers rich in composite alkali feldspar alternating with layers rich in quartz.

In the layered rocks, the feldspathic layers are chemically and mineralogically indistinguishable from the massive portions of this unit; therefore, the

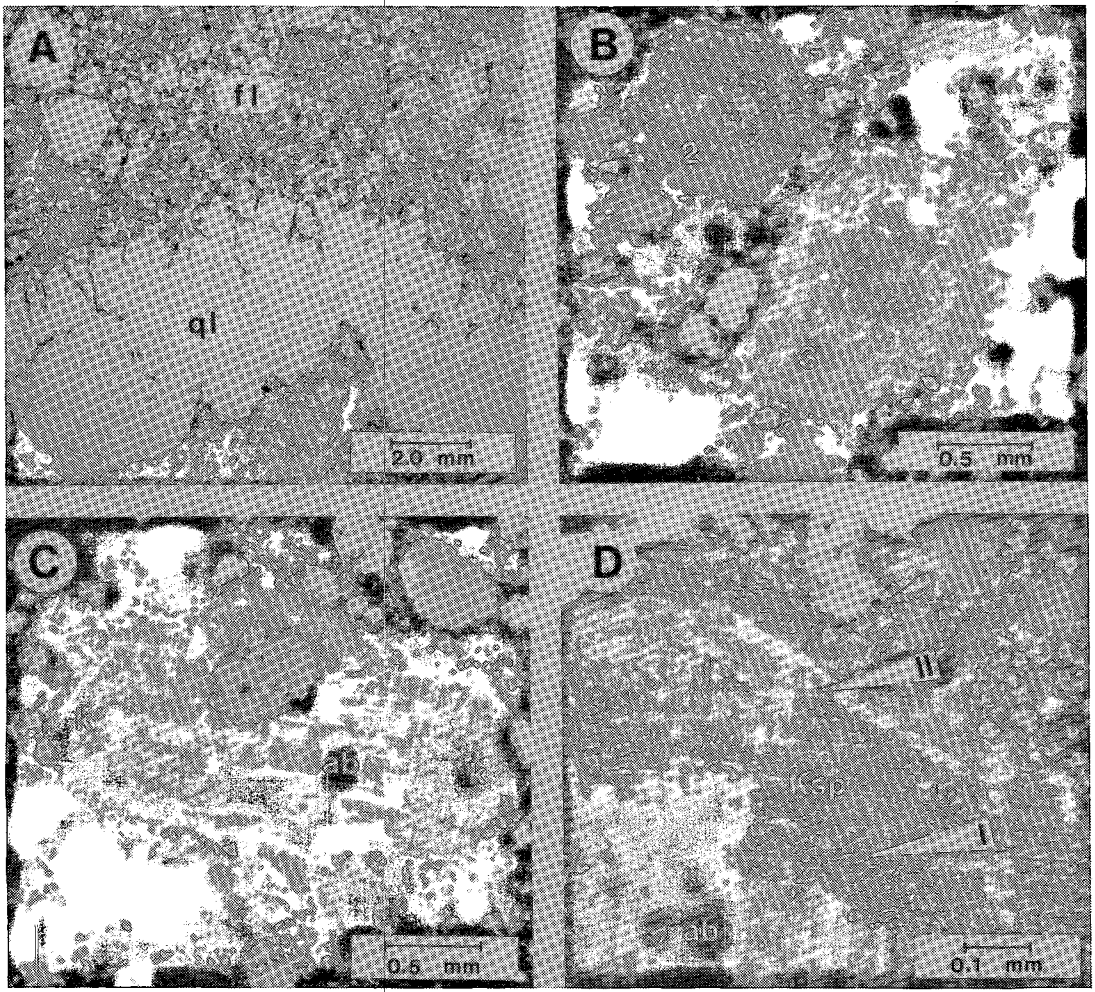

FIG. 2. Microstructural relationships between albite and K-feldspar in the feldspathic quartzite. A. A feldspathic layer (fl) is stained and shows up darker than the quartzose layer (ql). Thin "ribs" of fine-grained feldspar extend into the quartzose layer. Microphotograph taken in unpolarized light. B. Grains 2 and 3 (Tables 1, 2, and Fig. 4) exhibiting type-1 intergrowth, the presumed equivalent of "zone perthite" described by McTaggart (1979). The lighter grey albite is concentrated in the core of grains. C. Type-2 intergrowth exhibited by grain 1 (Tables 1, 2, and Fig. 4) showing a core of twinned albite (ab) sharply rimmed by K-feldspar (k). Contrast this with the type-1 intergrowth shown in B. D. Closeup of left end of Grain 1 (rotated clockwise 90°) showing untwinned (Ksp I) and twinned (Ksp II) K-feldspar (labeled I and II, respectively). Photomicrographs in B, C and D taken under partly crossed polars.

two can be considered equivalent. The feldspathic layers consist of variable porportions of quartz and composite alkali feldspar, with feldspar dominant. Composite feldspars occur as euhedral to anhedral clasts in a matrix of fine-grained alkali feldspar and quartz.

The quartzose layers consist of quartz ± minor feldspar, with quartz accounting for 95% or more of the layer. Quartz grains range in size from <0.1 mm to 3 mm. The larger grains exhibit undulatory extinction, subgrains and highly sutured boundaries, whereas the fine grains are equant and strain-free. The microstructures suggest that the fine-grained quartz was derived from the coarser quartz, involving a reduction in grain size by dynamic recrystallization (*cf.* White 1976) by at least two orders of magnitude.

Extending into the quartzose layers from, and continuous with, the feldspathic layers are long narrow "ribs" of fine-grained alkali feldspar (Fig. 2A). These ribs are subnormal to layer boundaries, each narrowing to a point within a quartzose layer. They pinch out primarily in only one direction, even in successive pairs of layers, and do not cut across boundaries of quartz grains. Texturally these ribs resemble sand fillings in mudcracks, possibly having developed as a result of successive influx and settling of feldspar-rich volcanic ash above a chert layer that had undergone subaqueous cracking. If so, the feldspathic quartzite may have arisen by accumulation under water due to a series of alternating events, each dominated by either chemical precipitation of silica or deposition of feldspathic volcanic ash. The large size and euhedral nature of many of the feldspar grains suggest that they were blown out during an explosive subaerial eruption, traveled a short distance in the air, landed in the water and settled downward in a relatively quiet basin.

FELDSPAR MICROSTRUCTURES

The majority of the feldspar in these rocks consists of coarse euhedral to subhedral composite grains of K-feldspar and albite. Some grains are composed almost entirely of K-feldspar, but minor albite is invariably present. Aggregates of fine-grained feldspar are also present in which both albite and K-feldspar constitute discrete rather than composite grains, with the proportion of albite grains to K-feldspar grains approximately ten to one.

The coarse grains are all composite, exhibiting two types of intergrowths: type 1 (Fig. 2B) consists of perthitic intergrowths of the string, band or plume style; type 2 (Fig. 2C,D) is a distinct core–rim relationship in which a core of twinned albite is enclosed by a rim of K-feldspar along a sharp contact. Type 1 is more common than type 2 and is similar to "zone perthite" that has been previously described in the literature (McTaggart 1979). It is characterized by alternating perthitic albite-rich and K-feldspar-rich zones that are roughly concentric and that conform to the grain outline. In contrast, neither the K-feldspar-rich nor the albite-rich portion of a type-2 intergrowth is perthitic. In some grains the twin pattern of the albite core can be correlated with the twin pattern shown by small isolated domains of albite within the K-feldspar rim.

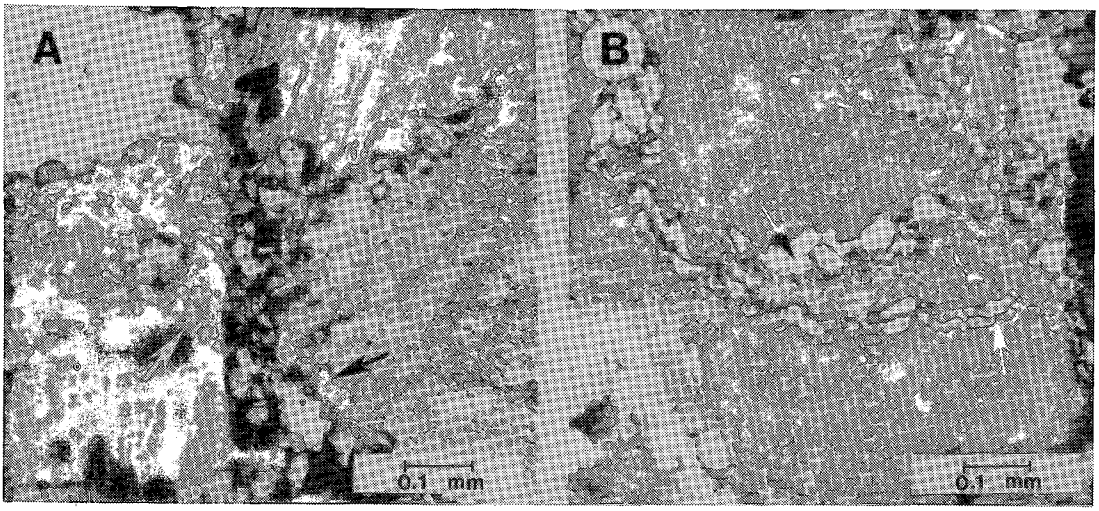

FIG. 3. Recrystallization textures of new albite forming from K-feldspar. A. Recrystallized albite (arrows) along margin of grains of K-feldspar. B. Two zones of recrystallized albite (arrows) cutting through a grain of composite alkali feldspar. Both photomicrographs taken under partly crossed polars.

The K-feldspar itself is of two textural types, an untwinned variety (Ksp I) and a twinned variety (Ksp II) (Fig. 2D). These two varieties are also compositionally distinct, as discussed below.

Fine-grained granoblastic albite occurs along the margin of the large composite grains, where it appears to have recrystallized from the K-feldspar-rich rim (Fig. 3A). These areas of fine-grained albite also extend into or completely across larger grains of feldspar (Fig. 3B).

ANALYTICAL METHODS

Major-element oxides, and Zr, Y, Nb, Cr, Ni, Sr, Rb, Ba, V, Cu, Pb and Zn were determined by X-ray-fluorescence spectrometry. Conventional procedures were used for the extraction of oxygen from minerals with bromine pentafluoride and quantitative conversion to CO_2 prior to analysis by mass spectrometry (Clayton & Mayeda 1963). Isotopic data are reported as $\delta^{18}O$ values in per mil relative to Standard Mean Ocean Water (SMOW). The overall reproducibility of $\delta^{18}O$ values averages $\pm 0.18‰$ (two standard deviations).

Fractionations, or differences in $\delta^{18}O$ among minerals, are quoted as Δ_{A-B}, defined as

$$\Delta_{A-B} = 1000 \ln \alpha_{A-B} \simeq \delta_A - \delta_B$$

where α_{A-B} is the fractionation factor for the coexisting minerals A and B.

Analyses of mineral composition were performed on a three-spectrometer MAC 400 electron microprobe incorporating a KRISEL automation system (Finger & Hadidiacos 1972), and using appropriate natural minerals for calibration standards. Operating conditions were 15 kV acceleration voltage and 0.03 μA sample current. Data were reduced using an on-line PDP-11/05 computer and the MAGIC IV data-reduction program, and program SUPREC written by J.C. Rucklidge, University of Toronto.

TABLE 1. REPRESENTATIVE COMPOSITION OF ALKALI FELDSPAR IN CORES OF GRAINS

	Grain 1		Grain 2		Grain 3	
	1	2	3	4	5	6
SiO_2	69.24	70.09	69.69	69.30	68.33	68.70
Al_2O_3	19.47	19.43	19.16	19.26	19.40	19.66
FeO^t	0.20	0.02	na	na	0.05	0.00
CaO	0.21	0.23	0.00	0.03	0.21	0.00
Na_2O	11.13	10.58	11.85	11.92	11.49	11.32
K_2O	0.26	0.14	0.04	0.02	0.07	0.24
BaO	0.05	0.00	0.00	0.00	0.00	0.01
total	100.56	100.49	100.74	100.53	99.55	99.93
number of ions based on 32 oxygen atoms						
Si	12.024	12.114	12.069	12.036	11.987	11.994
Al	3.984	3.957	3.910	3.942	4.010	4.045
Fe	0.029	0.003			0.007	0.000
Na	3.747	3.545	3.979	4.014	3.908	3.832
Ca	0.039	0.043	0.000	0.006	0.039	0.000
K	0.058	0.031	0.009	0.004	0.016	0.053
Ba	0.003	0.000	0.000	0.000	0.000	0.001
mol percent						
Or	1.50	0.85	0.22	0.11	0.40	1.38
Ab	97.49	97.97	99.78	99.75	98.61	98.62
An	1.02	1.18	0.00	0.14	1.00	0.00
Or	1.51	0.86	0.22	0.11	0.40	1.38
Ab	98.40	99.14	99.78	99.89	99.60	98.61
Ce	0.09	0.00	0.00	0.00	0.00	0.02

FeO^t total iron as FeO; na not analyzed
Or orthoclase; Ab albite; An anorthite; Ce celsian

TABLE 2. REPRESENTATIVE COMPOSITION OF ALKALI FELDSPAR NOT IN CORES OF GRAINS

	Grain 1				Grain 2				Grain 3			
	1	2	3	4	5	6	7	8	9	10	11	12
SiO_2	66.49	63.96	65.50	63.71	69.64	67.13	65.75	65.21	67.98	67.27	65.15	64.68
Al_2O_3	18.50	18.29	18.28	17.86	18.95	18.65	18.43	18.73	16.86	18.64	17176	18.73
FeO^t	0.00	0.00	0.04	0.16	na	na	na	na	1.43	0.00	0.08	0.00
CaO	0.00	0.00	0.00	0.00	0.04	0.12	0.00	0.02	0.00	0.00	0.00	0.00
Na_2O	2.90	3.05	1.77	0.25	11.60	5.57	0.44	0.13	7.43	9.42	0.56	0.17
K_2O	12.37	12.39	13.79	15.72	0.23	8.48	15.78	16.01	5.46	3.39	16.73	16.10
BaO	0.52	0.32	0.31	0.48	0.00	0.26	0.27	0.43	0.07	0.02	0.30	0.59
total	100.78	98.01	99.69	98.18	100.46	100.21	100.67	100.53	99.23	98.74	100.58	100.26
number of ions based on 32 oxygens												
Si	12.048	11.952	12.046	12.021	12.097	12.049	12.044	11.988	12.243	12.039	12.043	11.955
Al	3.950	4.027	3.962	3.971	3.879	3.945	3.978	4.057	3.578	3.931	3.869	4.077
Fe	0.000	0.000	0.006	0.025					0.215	0.000	0.012	0.000
Na	1.019	1.105	0.631	0.091	3.907	1.938	0.156	0.046	2.594	3.269	0.201	0.061
Ca	0.000	0.000	0.000	0.000	0.007	0.023	0.000	0.004	0.000	0.000	0.000	0.000
K	2.859	2.953	3.235	3.783	0.051	1.941	3.687	3.754	1.254	0.774	3.945	3.796
Ba	0.037	0.023	0.022	0.035	0.000	0.018	0.019	0.031	0.005	0.001	0.022	0.043
mol percents												
Or	73.73	72.77	83.67	97.64	1.29	49.74	95.93	98.68	32.59	19.14	95.16	98.42
Ab	26.27	27.23	16.33	2.36	98.53	49.67	4.07	1.22	67.41	80.86	4.84	1.58
An	0.00	0.00	0.00	0.00	0.19	0.59	0.00	0.10	0.00	0.00	0.00	0.00
Or	73.03	72.35	83.19	96.75	1.29	49.80	95.45	97.98	32.55	19.14	94.66	97.34
Ab	26.03	27.07	16.23	2.34	98.71	49.73	4.05	1.21	67.32	80.83	4.82	1.56
Ce	0.94	0.57	0.57	0.91	0.00	0.47	0.50	0.81	0.13	0.03	0.52	1.10

FeO^t total iron as FeO; na not analyzed
Or orthoclase; Ab albite; An anorthite; Ce celsian

GEOCHEMISTRY AND OXYGEN ISOTOPES

Composition of the feldspars

Five grains of the coarse composite feldspar were chosen for detailed study. Grain 1 exhibits a type-2 intergrowth, whereas grains 2 and 3 exhibits a type-1 intergrowth (Fig. 2). Grains 4 and 5 show evidence of recrystallization to fine-grained new albite (*e.g.*, Fig. 3). Representative compositions obtained by spot analyses of grains 1, 2 and 3 are listed in Tables 1 and 2. Collective treatment of these data (*e.g.*, by averaging) is not informative owing to the heterogeneity of feldspar compositions.

Analyses of albite in the core of grains 1, 2 and 3 indicate almost pure $NaAlSi_3O_8$ (Table 1). In contrast, the alkali feldspar away from the central core contains minor to major proportions of the K-feldspar component (Table 2), suggesting either solid solution or cryptic intergrowths of albite and K-feldspar on a micrometre scale.

All grains studied exhibit the two textural varieties of K-feldspar (described above), most easily seen in grain 1 (Fig. 2D). Ksp I most commonly occurs in contact with an albite core, constituting the bulk of the K-feldspar in any one grain; it is absent from the periphery of the grains. Ksp II is commonly intimately intergrown with Ksp I, but in general tends to constitute the outer portions of a K-feldspar rim away from an albite core. In the case of grain 1, an outer rim of Ksp II has developed that is devoid of Ksp I.

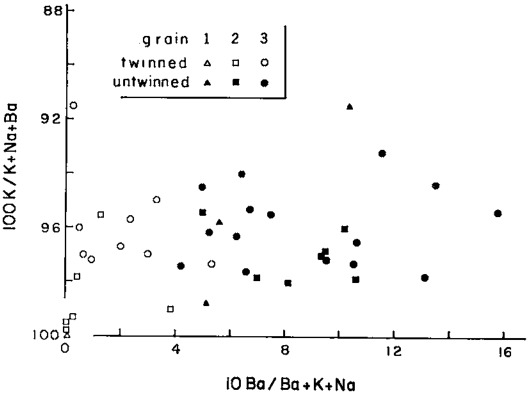

FIG. 4. Variation of Ba with K in alkali feldspar of the feldspathic quartzite. Compare the high Ba contents in untwinned portions of grains (Ksp I) with the low Ba contents in twinned portions of the same grains (Ksp II) for a given concentration of K. Each symbol represents a determination of composition at a single spot. Grain 1 is shown in Figures 2C and 2D, and grains 2 and 3 are shown in Figure 2B.

In addition to petrographic differences, Ksp I and II are chemically distinguishable by differences in barium content (Fig. 4). Ksp I contains more barium than does Ksp II, the latter commonly possessing Ba abundances below the detection limit by microprobe. On the other hand, there appears to be no systematic difference between the minor albite contents shown by Ksp I or Ksp II.

The fine-grained plagioclase that has recrystallized from K-feldspar (Fig. 3A,B) is almost pure albite, chemically indistinguishable from that in the core zone of the large grains.

Composition of the whole rocks

The composition of the feldspathic quartzite in the Adams mine is reported in Table 3. Given the petrographic relationships between albite and K-feldspar described above, it seems evident that the rocks have experienced intense chemical alteration, manifest principally by changed concentrations of Na, K and, perhaps, Ca. Determination of the original compositions of the feldspathic rocks thus requires consideration of the least mobile major and trace elements (*cf.* Floyd & Winchester 1978, Costa *et al.* 1983). For the massive and layered feldspathic rocks, the Ti/Zr ratio is in the range 6.7–6.8, and the Al_2O_3/TiO_2 ratio between 86 and 104. The rocks also exhibit low absolute abundances of TiO_2, V,

TABLE 3. COMPOSITION OF WHOLE ROCKS

	AD37	AD38A	AD38B	AD39
SiO_2	79.48	73.2	99.2	98.9
TiO_2	0.10	0.12	0.014	0.012
Al_2O_3	8.60	12.48	0.6	0.07
$Fe_2O_3^{t}$	1.40	2.50	0.07	0.02
MnO	0.01	nd	nd	nd
MgO	0.27	0.39	0.02	nd
CaO	0.22	0.09	0.05	nd
Na_2O	2.81	5.25	0.05	0.04
K_2O	2.88	4.57	0.03	0.03
P_2O_5	0.04	nd	0.05	0.01
LOI	3.19	0.80	0.07	0.05
total	99.00	99.4	100.1	99.1
trace elements in ppm				
Be	1.9	2	nd	nd
V	23	34	0.8	0.4
Cr	29	29	5	3
Ni	11	13	8	4
Cu	5	5	5	3
Zn	11	24	13	11
Pb	5	18	10	6
Rb	59	79	5	5
Sr	123	164	13	6
Y	11	15	19	8
Zr	88	107	13	10
Nb	46	39	65	52
Ba	414	583	3	nd
Ti/Zr	6.8	6.7	6.5	7.2
Al_2O_3/TiO_2	86	104	11	6

AD37:	massive quartzofeldspathic rock
AD38A:	feldspathic layer in thinly layered quartzo-feldspathic rock
AD38B:	quartzose layer in thinly layered quartzo-feldspathic rock
AD39:	thick quartzose layer (chert)

$Fe_2O_3^t$ total iron as Fe_2O_3; LOI loss on ignition
nd not detected

Cr and Ni, and high absolute abundances of Y, Zr and Nb. These geochemical characteristics are all consistent with a precursor of rhyolitic composition (Floyd & Winchester 1978).

Compared to pristine, unaltered Archean rhyolites (*cf.* Goodwin 1977, Condie 1981), these feldspathic rocks are depleted in Mg and Ca by approximately 50% and 80-90%, respectively, and enriched in K and Na by 20-100% and 0-25%, respectively. Such deduced changes in composition are commensurate with hydrothermal alteration of an inferred igneous sodic plagioclase in the rhyolite precursor to the observed albite and K-feldspar products. Bulk-rock Ba contents of 410-580 ppm (Table 3) are normal for Archean rhyolites in general, which have Ba ranges of 400 to 1000 ppm (*cf.* Goodwin 1977, Condie 1981). Differences in absolute abundances of TiO_2, Al_2O_3, Zr, *etc.*, could be accounted for solely by variation of SiO_2 contents in the feldspathic samples AD37 or AD38A (Table 3).

Thin quartzose layers interposed with feldspathic layers are dominated by SiO_2, but they also contain minor Al_2O_3, Fe_2O_3, K_2O and Na_2O that reflect the presence of small or trace quantities of amphibole, magnetite, feldspar and pyrite. Given the compliance of Ti/Zr (but not Al_2O_3) with the neighboring feldspathic layers, these latter groups of oxides and minerals are attributed to the addition of ~10% of a volcaniclastic component, compositionally similar to samples AD37 and AD38A, to a quartz-dominated layer. The microcrystalline sample AD39 comes from a thick quartzose layer and is essentially pure SiO_2.

With reference to a discrimination diagram proposed by Floyd & Winchester (1978), the high Zr/TiO_2 and Nb/Y values of the rocks signify that they may possess an alkaline character, raising the possibility that they are related to the neighboring Lebel syenitic stock (Fig. 1).

Oxygen isotopes

Quartz $\delta^{18}O$ values in the feldspathic rocks and layers range from 10.2 to 11.1‰ (Table 4). If our interpretation of these rocks as having a rhyolitic precursor is correct, and assuming minor or negligible shifts in $\delta^{18}O$ of the quartz, the measured δ-values are in compliance with those of quartz in felsic igneous rocks of the normal ^{18}O class (*cf.* Taylor 1968, 1978).

Primary igneous rocks are characterized by quartz–feldspar fractionations of +1.5 to +2.5‰, reflecting equilibrium at near-magmatic conditions (Taylor 1974, 1978). This is not the case for either the massive feldspathic rocks or the feldspathic layers from the Adams mine, in which $\Delta_{quartz-feldspar}$ ranges from -2.1 to -3.9‰. Quartz is resistant to oxygen-isotope exchange at temperatures <300°C, whereas feldspars readily undergo exchange reactions down to <100°C (Taylor 1974). The negative quartz–feldspar fractionations may have resulted from hydrothermal reaction of the feldspars during which the more resistant quartz retained its magmatic isotopic signature. Intimate intergrowths of albite and K-feldspar preclude a determination of their individual isotopic compositions.

TABLE 4. OXYGEN ISOTOPES OF MINERALS

Sample description and number	$\delta^{18}O$ quartz	$\delta^{18}O$ feldspar	Δ quartz-feldspar	Estimated temperature °C[†]
massive feldspathic rock AD37	10.24	14.13	-3.9	140-160
feldspathic layer AD38A	11.06	13.18	-2.1	160-180
thin quartzose layers AD38B	14.07			
AD38C	17.19			
AD38D	20.40			
thick quartzose layer AD39 (chert)	21.36			~70

[†] based on extrapolation of the K-feldspar- and albite-water fractionation equations of O'Neil & Taylor (1967), and the chert curve of Knauth & Lowe (1978) for sample AD39.

In thin quartzose layers of the layered rocks, the $\delta^{18}O$ quartz spans 14-20‰, *i.e.*, 3-9‰ enriched in ^{18}O relative to quartz in the neighboring feldspathic layers and massive feldspathic rocks. At the higher end of the range, the quartz is isotopically similar to that of quartz layers in the iron formation itself; based on this, we interpret the thin quartzose layers in the feldspathic quartzite as recrystallized chert that underwent variable degrees of partial re-equilibration with feldspar in the intervening layers. This suggests that these fine-grained quartzose layers (originally chert) were more susceptible to oxygen exchange than coarse-grained quartz of igneous origin in the massive feldspathic rock.

The thick quartzose layer of nearly pure SiO_2 (sample AD39) with $\delta^{18}O = 21.4‰$ is isotopically distinct from quartz in either massive or layered feldspathic rocks. Its isotopic composition is within the range of $\delta^{18}O$ of Archean (~2.8 Ga) marine cherts (Knauth & Lowe 1978), possibly indicative of precipitation in the presence of warm marine water at the bottom of the basin. The absence of minerals other than quartz presumably precludes isotopic exchange, despite the fine grain-size.

DISCUSSION

So-called zone perthite has previously been described from alkali-rich igneous rocks, most recently by McTaggart (1979), who concluded that the perthite intergrowths were due to two stages of exsolution by which original magmatic zoning was enhanced. This explanation is reasonable for the features described by McTaggart (1979), and may also apply to many of the grains in the feldspathic quartzite that exhibit type-1 intergrowths.

However, textures exhibited by the feldspars of the feldspathic quartzite differ from those described by McTaggart (1979) in two important respects: (a) at the scale of observation, much of the feldspar is not perthitic at all (type 2), and (b) many of the feldspar analyses indicate essentially pure end-member K or Na phases (Tables 1, 2).

Kinetic considerations coupled with equilibrium constraints on respective compositions of coexisting alkali feldspars argue against exsolution during post-magmatic cooling as a mechanism to produce the nearly end-member compositions (*e.g.*, Powell 1978, chap. 7). Instead, the mineral chemistry suggests a condition of nonequilibrium between the two phases, and the rimming of albite by K-feldspar probably represents neither magmatic zoning nor exsolution. We interpret these rims as having resulted from the secondary geochemical process of replacement. Replacement by K-feldspar (type 1 in this case) of pre-existing albite is a process that is probably more common than generally supposed (Kastner & Siever 1979), and the sharp boundary between the two phases is not texturally inconsistent with a replacement origin (Kerrich *et al.* 1981, Thomson *et al.* 1981). Additional secondary growth on an existing nucleus of partially replaced feldspar may account for some of the Ksp II, *e.g.*, the outer rim on grain 1 (Fig. 2D).

Munhá *et al.* (1980) have recently described the formation of adularia (K-feldspar) in sea-floor rhyolites that had been subjected to low-temperature interaction with seawater under conditions of high water:rock ratio. These authors argued that K-feldspar is the thermodynamically stable feldspar to form by ion-exchange reactions of seawater K with plagioclase. At temperatures less than ~150°C, fixation of K over Na in feldspars is favored, whereas albite stability is favored at $T > 150°C$. Contemporary growth of adularia proceeds even though the Na/K in modern marine water is ~28 (Schopf 1980, p. 151). In addition to favorable temperatures, a paucity of ferromagnesian phases is also required for feldspar replacement, so that K is not consumed in forming K-rich smectite, a phenomenon known to occur during the low-temperature interaction of seawater with basalt (Andrews 1978). The latter criterion applies in the case of the feldspathic quartzite.

Adularia commonly contains significant concentrations of Ba (Smith 1974, p. 90-93). This suggests that Ksp I, which appears to have replaced plagioclase in the feldspathic quartzites, may be a disordered phase that grew at low temperatures and is thus equivalent to the adularia described by Munhá *et al.* (1980). We were not able to determine the structural state of K-feldspar in these rocks, owing to the intimate nature of the intergrowths. We suggest, however, that the process described by Munhá *et al.* (1980) may have operated in the feldspathic quartzite. This is further supported by the similarity in $\delta^{18}O$ of the feldspars at the Adams mine and those reported by Munhá *et al.* (1980), inferred in both cases to indicate equilibrium with marine water at $T < 150°C$. We cannot explain so readily the origin of Ksp II, but we suggest that much (or all) of it was derived from Ksp I. Production of Ksp II from Ksp I may have occurred by Al–Si ordering over geological time. The apparently high obliquity of the crosshatch-twinned microcline that constitutes Ksp II favors the absence of the large cation Ba from its structure (Smith 1974, p. 92). If Ba were at one time present in this feldspar, it was presumably expelled during ordering.

Recrystallization of K-feldspar to new albite was accompanied by substantial reduction in grain size. This phenomenon has been described elsewhere; to our knowledge, in every case the recrystallization has accompanied deformation (*e.g.*, Debat *et al.* 1978, Kerrich *et al.* 1980). Typically, this type of deformation-induced recrystallization initially produces a mortar texture, but zones of recrystallized albite that extend completely through original grains of perthite in a mylonitic quartz monzonite have also been described (La Tour 1979). The penetration of a grain by new albite is shown by grain 5 of the feldspathic quartzite (Fig. 3B), and is presumed to have occurred during deformation. This feature may represent a form of stress-corrosion cracking (Anderson & Grew 1977) of the grain in which the chemical reaction is catalyzed by local increase of the stress intensity, permitting crack propagation at subcritical intensities of stress (*cf.* Kerrich *et al.* 1981). The deformation most likely occurred during intrusion of the Lebel stock, which has clearly influenced the overall attitude of the iron formation (Fig. 1). This would be consistent with the relatively low states of deformation and low temperatures at which stress corrosion is thought to be active (Kerrich *et al.* 1981).

SUMMARY AND CONCLUSIONS

In the feldspathic quartzite, the zonation seen in the feldspars from albite core to K-feldspar rim to new albite outer rim may provide a thermal record for the rock. This record may be summarized as follows. Submarine alteration of the rhyolitic volcaniclastic sediment may have initially proceeded at temperatures above 150°C, accompanied by conversion of sodic plagioclase to albite. This was followed by an ion-exchange reaction of albite to K-feldspar as temperatures waned to <150°C. During burial of the rocks or intrusion of the Lebel stock, an increase in temperature enabled new albite to be formed *via* a back reaction at grain boundaries of K-feldspar. Temperature estimates are tentative, for the temperature of the K-feldspar–albite transition would depend on the fluid Na/K at each stage.

Given the abundance of chemically precipitated silica and the hydrothermal alteration of feldspar, we suggest that the feldspathic quartzite should be considered an integral part of the Adams mine sequence of chemical and detrital sediments, along with the interlayered iron-formation. This conjunction of volcanic, sedimentary and hydrothermal processes is known to give rise to iron formations *per se* (*cf.* Gross 1980, Kimberley 1979), but evidence for the same processes in the enclosing host-rocks is commonly not recognized or reported.

The intimate spatial association of quartzose and feldspathic layers is an indication of contemporaneous chemical precipitation of chert and input of rhyolitic volcanic debris. The oxygen-isotope data indicate that pure cherts (*i.e.*, negligible iron or base-metal contents) were precipitated at low temperatures (~70°C), where ambient marine-bottom waters dominated over the hydrothermal discharge. Isotopic results for the feldspars in this study, together with thermodynamic and kinetic considerations of the stability of alkali feldspar, indicate that ion-exchange reactions of igneous sodic plagioclase to albite, and subsequently of albite to K-feldspar, also occurred at this time, providing a record of temperature decrease through 150°C during convective dissipation of heat. The hydrologic regime envisaged is one of thermally driven convective cooling by marine water of hot igneous rocks constituting the oceanic crust.

Therefore, in addition to the contemporaneous primary processes of volcanism, sedimentation and chemical precipitation, secondary chemical alteration of minerals in the new sediments also occurred at this time. This complex interplay of several processes may have important consequences with respect to interpretation of the "primary" or "secondary" origin of minerals in both mineralized and barren exhalites and their contiguous volcanic rocks.

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