References

Ames, L. L. (1959) The genesis of carbonate apatite. *Econ. Geol.*, **54,** 829–841.

Le Geros, Z. R. (1965) Effect of carbonate on the lattice parameters of apatite. *Nature* **206,** 403–404.

McConnell, D. (1952) The problem of the carbonate apatite. IV Structural substitutions involving CO_2 and OH *Bull. Soc. Franc. Minéral. Crystallogr.* **75,** 428–445.

——— (1965) Deficiency of phosphate ions in apatite. *Naturwissenschaften* **52,** 183.

Simpson, D. R. (1964) The nature of alkali carbonate apatite. *Amer. Mineral.* **49,** 363–376.

——— (1965) Carbonate in hydroxylapatite *Science* **147,** 501–502.

THE AMERICAN MINERALOGIST, VOL. 52, MAY-JUNE, 1967

Fe/(Fe+Mg) RATIO IN BIOTITE FROM GEORGIA GRANITIC ROCKS

L. D. Ramspott[1] and E. L. Plunkett, Jr., *Department of Geology, University of Georgia, Athens, Georgia.*

Introduction

Preliminary observation of samples from quarries in the Elberton granite district, Georgia, revealed that biotite in gneiss and schist inclusions had a different refractive index than biotite in adamellite taken from the same quarry; whereas biotites in adamellite taken from scattered quarries had similar refractive indices.

Uncertainties in the determination of the refractive indices of biotite have been reported; Engel and Engel (1960) comment extensively on this, but Peikert (1963) implies that the determination is not particularly difficult. Because of absorption in the biotite, a match with the oil is difficult to tell unless the grain size is very fine. Due to these uncertainties $d(060)$ was determined in addition to $\beta(\approx\gamma)$ refractive index on biotite from 45 specimens of Elberton adamellite and one specimen each of Gray granodiorite, Stone Mountain adamellite, Danburg porphyritic adamellite, Columbia County adamellite, and a gneiss and a schist inclusion in Elberton adamellite.

Geology and Petrology of Rocks

The Elberton Granite District (Ramspott, 1964a), from which most of the samples came, extends over 200 square miles in Elbert, Oglethorpe, and Madison Counties, Georgia. Fine- to medium-grained adamellite bodies cross-cut the surrounding sillimanite-grade gneiss and schist, but are themselves faintly metamorphosed (retrograde and dynamic), as shown by microscopic metamorphic textures, chlorite-sericite alteration,

[1] Present address, Lawrence Radiation Laboratory, Livermore, California.

and the occurrence of zeolites and related minerals along fracture surfaces. Some quarried bodies are dike-shaped, being as narrow as 200 feet, and many are considerable elongated, trending generally north. Migmatite veins and dikes are common throughout the area.

In the fine- to medium-grained Elberton adamellites, oligoclase (median, An_{22}; range, An_{11} to An_{27}), microcline, and quartz are present in approximately equal amounts. Biotite constitutes about 5 percent of the rock; the color index is about 6. Common accessories are magnetite, allanite, muscovite, epidote, apatite, hematite, chlorite, and zircon. Sphene, monazite, and calcite are present locally.

The Danburg coarse-grained porphyritic adamellite (Crawford and others, 1966) is a 5- by 10-mile elliptical pluton trending N25E through northeastern Wilkes County and northern Lincoln County, Georgia. The rock weathers to a distinctive saprolite and appears to be uniform in composition and texture. An average of two modes gives a mineral composition of oligoclase (An_{24}), 30; microcline, 37; quartz, 25; others, 8 percent (including biotite, magnetite, sphene, with minor zircon, apatite, muscovite, chlorite, and hematite).

The fine-grained Gray granodiorite (Jones County, Georgia) is noted as having a high magnetite content in saprolite by Hurst (1953). A modal analysis is given in Vistelius and Hurst (1964) showing quartz, 38; K feldspar, 18; plagioclase, 40; and others, 6 percent.

The Columbia County coarse-grained porphyritic adamellite has not been studied quantitatively.

The Stone Mountain adamellite is well studied, the most recent publications being Herrmann (1954) and Wright (1966). It differs from the other rocks, as it has an 11 to 1 ratio of muscovite to biotite. Neither of the above papers mention magnetite, although Hurst (1953) reports it in the saprolite, and the writers separated small amounts while obtaining the biotite. The Stone Mountain adamellite has only about 1 percent biotite, whereas the other rocks studied contain about 5 percent biotite.

The two inclusions are from the Acme Quarry (Ramspott, 1964b). The schist inclusion is a biotite-oligoclase-quartz schist with muscovite, apatite, allanite, sphene, magnetite, and carbonate. The gneiss inclusion has the same mineralogy, plus the addition of microcline metacrysts. The field occurrence indicates origin of the gneiss by metasomatism of the schist.

All of the rocks studied contain magnetite and either microcline or muscovite. Hematite was observed in nearly all as a very minor constituent. These rocks thus contain the critical assemblage biotite, magnetite, and sanidine or muscovite, as well as containing both components of the "buffer" Fe_3O_4-Fe_2O_3.

The most conspicuous occurrence of hematite is as thin, translucent flakes within biotite. It occurs scattered in very minor quantity in all of the major minerals. There is no proof that this hematite coexisted with magnetite to form a Fe_2O_3-Fe_3O_4 buffer during crystallization of the biotite. However, the values obtained in this study plot near the curve defined by plotting β index versus $d(060)$ valves for corresponding Fe/(Fe+Mg) ratios from Fe_2O_3-Fe_3O_4 buffered synthetic runs (Fig. 1, data from Wones, 1963).

Procedure

Rock samples were crushed with a Plattner mortar and sieved to obtain 80–100 and 100–120 mesh fractions. Biotite was separated by means of a Franz Isodynamic Magnetic Separator. Because of the intent to analyze by X-ray diffraction or refractive index, checks for sample purity were only qualitative.

The biotite was ground with a silicon internal standard in an agate mortar to pass 325 mesh. Cell mounts were made by packing material lightly to avoid orientation and slicing off the excess with a razor blade leaving a smooth surface. Material coarser than 325 mesh did not make good mounts by this method. Smear mounts and tightly packed cell mounts were tried, but the preferred orientation made the biotite (060) peak unmeasureable.

Diffraction peak 2θ (060) was determined by step-counting across both the silicon (311) and biotite (060) peaks, which occur respectively at about 72.4° and 77.5°. The instrumental settings were FeKα radiation; Mn filter; divergence and scatter slits: 1°; receiving slit: 0.006 inch; and scale factor of 2. Counts were accumulated at points spaced 0.1° or narrower for 132 seconds, so that 800 to 1000 counts were accumulated at points on the slopes of the peaks. The peak was bisected at 2/3 height to determine position. The silicon (311) peak will resolve into $K\alpha_1$ and $K\alpha_2$, but the biotite peak is broad and unresolved [Wones (1963) suggests that (331) may be superposed on (060)]. 2θ (060) was converted to $d(060)$ by tables (Switzer and others, 1948) using the value 1.9373 Å for FeKα.

Refractive index β was determined using Cargille index oils spaced at intervals of 0.002 n. Both white and monochromatic (589 mμ) light were used as necessary. The β index is so close to the γ that no differentiation was attempted, nor was it possible because of the absorption of the thin cleavage flakes.

One sample was determined three times by the X-ray method by running remounts of the same material on different days. The standard deviation of the repeated measurement is 0.00031 Å. Assuming a normal distribution, the measurements by this method should be within ± 0.0006

Å at the 95 percent confidence level. This is equivalent to ± 0.03 Fe/(Fe+Mg). Wones' (1963) data have a mean standard deviation of ± 0.0006 (± 0.06 Fe/Fe+Mg), indicating that this is about the range expected for this method.

Duplicate determinations of the refractive index were made for ten samples. The mean standard deviation was 0.00145. Assuming a normal distribution, the measurements by this method should be good to ± 0.003 at the 95 percent confidence level, equivalent to ± 0.023 Fe/(Fe+Mg). This is the value obtained by Peikert (1963) for 104 duplicate measurements and Wones (1963) commonly reports the γ index to ± 0.002 or ± 0.003.

The X-ray method takes about $1\frac{1}{2}$ hours per sample for ± 0.030 Fe/(Fe+Mg) precision, whereas the optical method takes less than 15 minutes per sample for ± 0.023 Fe/(Fe+Mg) precision. Thus the optical method is superior in speed to the X-ray method for similar precision.

Discussion

Figure 1 is constructed from data in Wones (1963). The curve is the locus of corresponding Fe/(Fe+Mg) values for the measurements of $d(060)$ and $\beta(\approx\gamma)$ refractive index for the buffer system Fe_3O_4-Fe_2O_3. Composition of biotite in terms of Fe/(Fe+Mg) is marked along the curve.

In the following discussion, several simplifying assumptions are made: the variation in refractive index and $d(060)$ is principally related to varying Fe/(Fe+Mg) ratio, and the observed variations are primary and not due to alteration. Numerous variables have not been evaluated, among them ferrous to ferric ratio and variation of Mn, Ti, Na, Al, and F. The work of Peikert (1963) shows a reasonably high correlation between re-refractive index and Fe/Mg ratio for similar rocks.

The scatter at the determinations shown in Figure 1 is greater than estimated analytical error, but consistent patterns of variation related to field occurrence have not yet been noted. Where it is known that two or more locations lie along the same granitic band, the variation within the band in Fe/(Fe+Mg) is as great as variation between bands. Two samples from the same quarry (C-O, Figure 1) show a spread of values outside the estimated analytical error and nearly as great as the entire range of the district. Work currently in progress indicates that there may be areal variation within adamellite bands related to the distance from the contact.

The points labeled A-1 schist and A-1 gneiss were determined on biotite from two-to three-foot diameter inclusions in the same quarry as A-1 adamellite, which contains one of the more iron-rich biotites. The adamellite sample came from within 50 feet of the inclusions. There are recent observations indicating that f_{O_2} varies across compositional boundaries in many natural high-temperature rock systems (James, 1955; Chinner, 1960; Peikert, 1963; Melson, 1966). Chinner (1960, p. 211) notes a number of examples of changes in oxidation in country rock at contacts or in inclusions. In the cases cited, there were distinct thermal

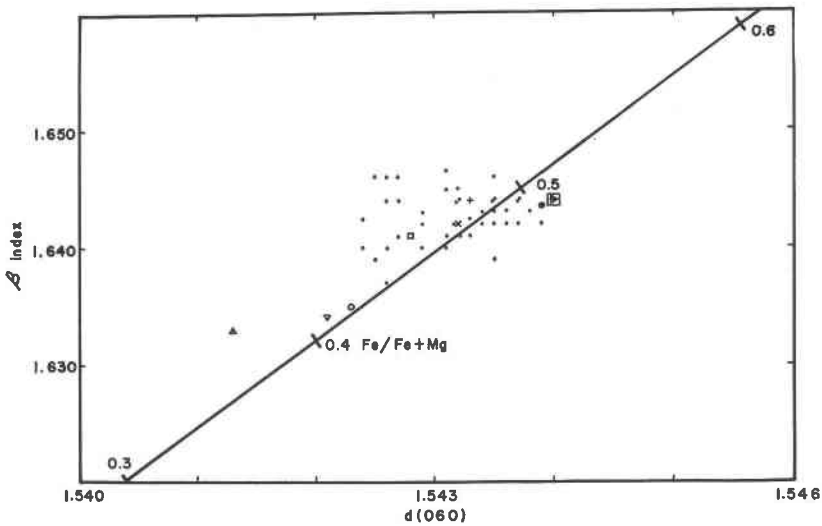

FIG. 1. Plot of β refractive index vs. d (060). The diagonal line is adapted from Wones (1963) from curves drawn for the buffer Fe_2O_3-Fe_3O_4. The ratio Fe/(Fe+Mg) is marked along the line. The symbols are as follows: D Elberton adamellite, △ A-1 quarry schist, ▽ A-1 quarry gneiss, △ A-1 quarry adamellite, □ C-O quarry, ● Stone Mountain, ○ Danburg, × Columbia, + Gray.

aureoles, whereas in the catazonal Elberton adamellites, temperature differences between the plutonic and country rocks were small or did not exist.

Figure 4 of Wones and Eugster (1965, p. 1244) suggests why the Fe/(Fe+Mg) ratios of biotite in the schist and gneiss inclusions could remain unequilibrated with those in the adamellite. A biotite with a Fe/(Fe+Mg) ratio of 0.4 (as in the schist and gneiss) is stable at any value of f_{O_2} and T on or below the 40 contour. As the biotite in the associated granite has a ratio near 0.5 (conditions appropriate for the 50 contour), the indicated temperature from the Fe_3O_4-Fe_2O_3 buffer curve is near 600°C. Therefore the biotite in the inclusions would be stable under these conditions, even if f_{O_2} in the inclusions equilibrated with that in the magma. In this example it is not possible to state that the f_{O_2} in the inclusion remained independent of that in the surrounding magma. This assumes, of course, that the temperature differences between country rock and inclusion were negligible, as the diagram shows that a rise in temperature on the order of 100°C would have a considerable effect on Fe/(Fe+Mg) ratio, even at constant f_{O_2}.

The Elberton adamellite samples all show hematite flakes along biotite cleavages. Figure 4 of Wones and Eugster (1965, p. 1244) shows that falling temperature causes the biotite to react, forming sanidine and hematite. The sanidine evidently has migrated away from the biotite as postulated by Chayes (1955) for the potash feldspar released by the biotite-chlorite transformation (which also occurs to some extent in the Elberton adamellites). This interpretation suggests that Fe_2O_3 may not have been in coexistence with Fe_3O_4 during initial crystallization of the biotite.

Peikert (1963) found β index values ranging from 1.624 to 1.663 in 430 samples of tonalite, granodiorite, and adamellite in 90 square miles of a granitic band in the Precambrian of Northeastern Alberta. In the present study, in 45 samples of adamellite composition taken from over 200 square miles, the β refractive index ranges from 1.637 to 1.647. Peikert postulates a metamorphic origin for the granitic rocks, based in part on the wide variation of the biotite indices. In the Elberton district, the restricted range of refractive index accords with other evidence indicating a magmatic origin. Assuming the magma to have been generated by selective fusion, it is not unlikely to suppose that, as the P-T conditions at a given level are likely to be fairly uniform, the anatexis was governed by water content of source rocks. This could account for the similar f_{O_2} of the adamellites implied by the Fe/(Fe+Mg) ratios.

Acknowledgments

This research was supported by an NSF Grant (GP-2143).

References

Chayes, F. (1955) Potash feldspar as a by-product of the biotite-chlorite transformation. *J. Geol.* **63**, 75–82.

Chinner, G. A. (1960) Pelitic gneisses with varying Ferrous/Ferric ratios from Glen Clova, Angus, Scotland. *J. Petrology* **1**, 178–217.

Crawford, T. J., V. J. Hurst and L. D. Ramspott (1966) Extrusive volcanics and associated dike swarms in central-east Georgia. *Geol. Soc. Amer. SE Sect.*, **2**, Athens, Ga.

Engel, A. E. J. and C. Engel (1960) Progressive metamorphism and granitization of the major paragneiss, Northwest Adirondack Mountains, New York. Part II: Mineralogy. *Geol. Soc. Amer. Bull.* **71**, 1–58.

Herrmann, L. A. (1954) Geology of the Stone Mountain-Lithonia District, Georgia. *Ga. Geol. Surv. Bull.* **61**.

Hurst, V. J. (1953) Heavy minerals in saprolite differentiation. *Ga. Geol. Surv. Bull.* **60,** 244–264.
James, H. L. (1955) Zones of regional metamorphism in the Precambrian of northern Michigan. *Geol. Soc. Amer. Bull.* **66,** 1455–1488.
Melson, W. G. (1966) Phase equilibria in calc-silicate hornfels, Lewis and Clark County, Montana. *Amer. Mineral.* **51,** 402–421.
Peikert, E. W. (1963) Biotite variation as a guide to petrogenesis of granitic rocks in the Pre-Cambrian of Northeastern Alberta. *J. Petrology* **4,** 432–459.
Ramspott, L. D. (1964a) The Elberton batholith. *SE Geol.* **5,** 223–230.
——— (1964b) Inclusions in the Acme Quarry, Elbert County, Georgia. *Ga. Acad. Sci. Bull.* **22,** 32–37.
Switzer, G., J. M. Axelrod, M. L. Lindberg and E. S. Larsen 3d (1948) Tables of *d* spacings for angle 2θ *U. S. Geol. Surv. Circ.* **29.**
Vistelius, A. B. and V. J. Hurst (1964) Phosphorus in granitic rocks of North America. *Geol. Soc. Amer. Bull.* **75,** 1055–1092.
Wones, D. R. (1963) Physical properties of synthetic biotites on the join phlogopite-annite. *Amer. Mineral.* **48,** 1300–1321.
——— and H. P. Eugster (1965) Stability of biotite: experiment, theory, and application. *Amer. Mineral.* **50,** 1228–1272.
Wright, N. P. (1966) Mineralogic variation in the Stone Mountain Granite, Georgia. *Geol. Soc. Amer. Bull.* **77,** 207–210.

THE AMERICAN MINERALOGIST, VOL. 52, MAY–JUNE, 1967

ECLOGITES AND JADEITE FROM THE MOTAGUA FAULT ZONE, GUATEMALA

Alexander McBirney, *Center for Volcanology, University of Oregon, Eugene, Oregon*
Ken-ichiro Aoki, *Institute of Mineralogy, Petrology, and Economic Geology, Tohoku University, Sendai, Japan*
and
Manuel N. Bass, *Department of Earth Sciences, University of California, La Jolla, California.*

Introduction

Jadeite occurs as inclusions in serpentinite in the Motagua fault zone. A boulder of remarkable size and purity was found near Manzanal, Guatemala, and chemical, optical, and X-ray studies have been made of a separated sample of the jadeite. Elsewhere in the Motagua fault zone, omphacite-garnet eclogites and glaucophane-lawsonite-omphacite-garnet rocks have been found in similar environments. Only at the Manzanal locality, however, have sodic pyroxenes been found in rocks which are nearly monomineralic.