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RADIOMETRIC AGE DETERMINATIONS, BIGHORN MOUNTAINS, WYOMING*

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ABSTRACT. Field and petrographic studies suggest that a foliated granitic complex in the northern half of the Bighorn Mountains evolved essentially contemporaneously with gneiss and associated rocks in the southern half of the area during a major episode of metamorphism and metasomatism. Small ultramafic bodies within both complexes may have been emplaced before metamorphism in some cases and after it in others. At least two generations of granitic pegmatites are present, one presumed to have been introduced during the metamorphism and metasomatism and the other afterward. All major lithologies are transected by younger, Precambrian, dolerite dikes.

K-Ar apparent ages of biotite from the gneiss fall in the range 2.31 to 2.78 billion years. Apparent ages of biotite from the granitic complex, which consists of quartz diorite and quartz monzonite, show a similar spread. Ages for quartz monzonite are internally consistent and cluster at 2.78 billion years. Ages for quartz diorite samples are grouped at 2.79 and 3.14 billion years. Miscellaneous individual K-Ar ages include the following: schist skialith, 3.06 billion years (biotite); peridotite, 2.99 billion years (whole-rock); dolerite, 2.04 billion years (feldspar).

Based on their wide variation, due to differential argon loss, the K-Ar ages can be considered minimum ages only. They suggest the possibility of contemporaneity for the two major complexes in the Bighorn Mountains and indicate, to a first approximation, that the major metamorphic-metasomatic episode occurred on the order of 3 billion years ago in this area. This approximate date is supported by zircon-monazite age determinations from a single sample of the quartz monzonite, which yield a concordia intercept age of 2.89 billion years for this unit.

INTRODUCTION

The Bighorn Mountains, located in north-central Wyoming (fig. 1), represent a crustal block uplifted initially during the Laramide Orogeny and subsequently eroded. This activity has resulted in the exposure of roughly 1100 square miles of Precambrian rocks along the axis of the uplift. The results of field mapping and petrographic studies of portions of these rocks have been published in recent years by the following: Osterwald (1949, 1955, 1959), Hoppin (1961), Hoppin, Palmquist, and Williams (1965), Hoppin and Palmquist (1965), and Palmquist (1965, 1967).

Heimlich (1965) has completed reconnaissance mapping and petrographic studies of the entire Precambrian exposure. The present paper reports the results of a preliminary radiometric age program initiated to support the field work and other laboratory studies currently in progress.

GEOLOGIC RELATIONSHIPS

Lithologies.—Seven major lithologic units, other than dikes, have been mapped in the Bighorn Mountains (fig. 1). A gneiss complex which includes amphibolite, schist, and quartzite underlies the southern half of the area. The gneiss is quartzofeldspathic, consisting primarily of

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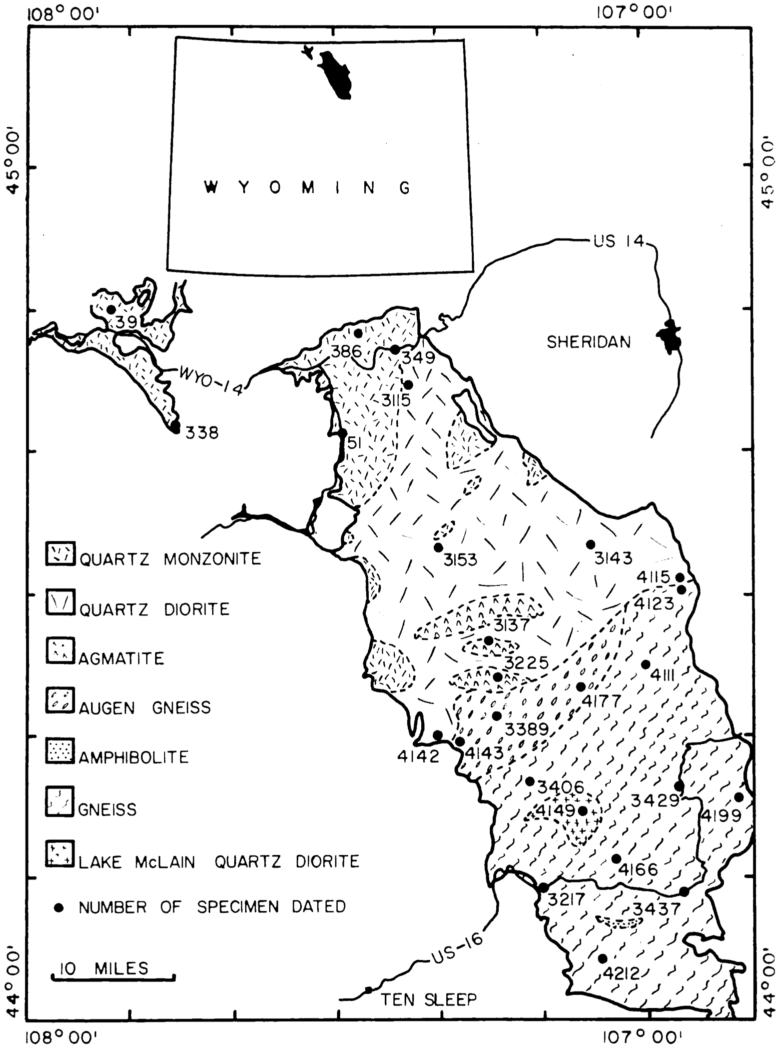

Fig. 1. Reconnaissance lithologic map of the Bighorn Mountains, Wyoming, showing the location of samples dated.

quartz, plagioclase, and biotite. Its microcline content ranges from 0 to 25 percent. Over broad areas the gneiss is a typically layered rock with predominantly equigranular texture. Locally, however, it is inequigranular due to the presence of augen or tabular porphyroblasts. Although inequigranular texture is present at scattered locations throughout the main gneiss terrain, a large relatively continuous mass of augen gneiss crops out along the northwest margin of the complex (fig. 1).

Intercalated with the gneiss are layers of amphibolite which range in thickness from 1 to 150 feet. One exceptional mass, which crops out on the north flank of Hazleton Peak in the south-central part of the gneiss terrain, has a maximum thickness on the order of 2500 feet and is the only exposure of amphibolite of mappable dimensions. Minor rock types of the gneiss complex in addition to amphibolite include hornblende-biotite schist, muscovite schist, quartzite, and garnetiferous gneiss. Studies in the Horn area (Palmquist, 1965) at the southernmost extremity of the main Precambrian exposure have revealed the presence of marble, calc-silicate rock, and iron formation interlayered with the gneiss.

A discrete body of granitic rock lies immediately east and north of Lake McLain within the west-central part of the gneiss complex. The rock is a relatively homogeneous, coarse-grained quartz diorite. It is structurally and lithologically distinct from the surrounding gneiss which, although quartz dioritic in composition, contains roughly 10 percent less plagioclase and 15 percent more quartz. The body is broadly concordant relative to the structure of the enclosing gneiss, and there is a suggestion that the strike of the gneiss foliation "wraps" around it.

The northern half of the Bighorn Mountains is underlain by a granitic complex consisting of two major units, quartz diorite and quartz monzonite, which grade into each other mineralogically and texturally. Both units exhibit substantial variations in color, texture, mineral percentages, type of foliation, and amount of foreign material. The classification of these units is based on the average or dominant ratio of plagioclase to microcline, but in detail there is an almost complete gradation in percentages of these minerals between the two rock types. For example, in quartz diorite the microcline content ranges from 0.6 to 8.3 percent, whereas in quartz monzonite it ranges from 12.0 to 50.4 percent.

As figure 1 shows, major bodies of quartz monzonite occur in the northwest portion of the main Precambrian exposure and in an isolated area northwest of it. However, quartz monzonite is also present in scattered patches throughout the main quartz diorite area. Both units possess well-developed foliation (parallelism of platy biotite aggregates, alignment of tabular microcline crystals, and local compositional layering). Elongate masses of biotite schist, gneiss, amphibolite, and agmatite occur within the granitic complex and are commonly conformable with its structure.

The mapped bodies of agmatite consist of angular to subangular blocks, thick lenses, and discontinuous streaks of black, fine to medium-grained, mafic material "soaked", veined, and surrounded by gray felsic material. The mineral composition is qualitatively the same for both types of material. Dark phases of the agmatite are richer in biotite and hornblende and poorer in plagioclase and quartz than are the lighter phases.

Both complexes contain small ultramafic bodies, mostly altered pyroxenites and peridotites. Their contacts with the country rock are commonly concealed. Some are discordant dikes; others appear to be concordant or discordant lens-shaped bodies. Granitic pegmatites are ubiquitous within the granitic and gneiss complexes. Both discordant and concordant, sharply-bounded and gradational types are present. Other units, occurring mostly as cross-cutting dikes or veins, include aplite, quartz porphyry, epidote, and quartz. All major lithologies of both complexes are transected by dolerite dikes, which range in thickness from 1 inch to over 500 feet and many of which crop out for several miles along strike.

Age relations.—The field and petrographic observations noted in the foregoing, and to be reported more fully elsewhere (Hodgson and Heimlich, ms), suggest the following working hypothesis for the origin of the two major complexes: during an interval of regional deformation, a largely sedimentary rock sequence was folded and recrystallized to gneiss and associated rocks and, in the northern part of the area, was converted by metasomatism to granitic rocks. According to this hypothesis the two major complexes acquired their present state essentially contemporaneously. Some difference in time of formation undoubtedly exists. For example, the metasomatism may have occurred during the latter stages of the regional metamorphism. However, this difference may not be resolvable by current geochronologic techniques.

Meager field data suggest that the Lake McLain quartz diorite is a discrete pluton that intruded the gneiss or its pre-metamorphic equivalent, causing a structural dislocation in the wallrocks. Petrographic study of the rock indicates the presence of narrow mosaic zones between the large grains (which give the rock its overall xenomorphic-granular texture). On the basis of these limited data, the Lake McLain quartz diorite could be pre-metamorphism, late-metamorphism, or post-metamorphism.

The large bodies of agmatite within the granitic complex may represent disaggregated portions of a broad stratigraphic unit within the original gneiss complex which, at an early stage in the evolution of the terrain, is thought to have extended farther north as recognizable gneiss. Mechanical deformation of this unit to produce agmatite occurred, presumably, during the major metamorphic event affecting the area.

The chronologic position of the ultramafic bodies within the two complexes is conjectural from the point of view of field and petrographic studies. Most of the few ultramafic bodies examined show evidence of mineralogic alteration which may be the result of deuteric processes or of metamorphism. It is probable that some of these bodies were emplaced prior to regional metamorphism and some afterward.

The pegmatitic material, prevalent throughout the area, is thought to represent the alkalic fluids that caused the conversion of the gneiss to granitic rocks. Much of it displays irregular, replacive relationships with its host rocks and is probably contemporaneous with their metamorphism. Sharply-bounded pegmatite dikes which transect some of the granitic rocks may reflect a later generation of alkalic fluids.

The dolerite dikes are major features within the area. They cut all rock types, and nearly all are fresh under the microscope. Their emplacement was the last major event in the Precambrian history of the region. A small percentage of the dikes, altered to amphibolite, were intruded prior to the major regional metamorphic event.

PREVIOUS GEOCHRONOLOGIC STUDIES

To our knowledge, only two samples of Precambrian rock from the Bighorn Mountains have been dated thus far by other laboratories. Gast, Kulp, and Long (1958) dated a biotite sample at 2.76 billion years by the Rb-Sr method and at 2.54 billion years by the K-Ar method. Judging from their generalized map, the biotite was separated from quartz monzonite in the northern part of the Bighorns. Giletti and Gast (1961) report a Rb-Sr age of 2.55 billion years on biotite from gneiss in the Powder River Pass area. Rb-Sr age determinations in both studies were calculated on the basis of a Rb^{87} half-life of 50 billion years.

With respect to ages in nearby areas, Giletti and Gast (1961) also report two Rb-Sr dates from the Owl Creek Mountains to the southwest. A muscovite sample was dated at 2.72 billion years, and a microcline sample at 2.64 billion years. Gast, Kulp, and Long (1958) obtained a Rb-Sr age of 2.72 billion years on biotite from pegmatite in Shoshone Gorge near Cody, Wyoming, due west of the Bighorns.

Several radiometric age studies have been made to the northwest in the Beartooth Mountains. Rb-Sr and K-Ar ages of six samples including biotite, muscovite, and microcline from pegmatite and the associated gneiss and migmatite suggest an age of 2.75 billion years for the last metamorphic event in the area (Gast, Kulp, and Long, 1958). In a subsequent study, Catanzaro and Kulp (1964) focused their attention on zircon that occurs in the gneiss and migmatite. These rocks are thought to have formed during regional metamorphism of an older sequence of sedimentary rocks (Eckelmann and Poldervaart, 1957). The zircons are predominantly rounded and are argued to be detrital in origin. Pb^{207}-Pb^{206} ages determined for seven samples from several of the metamorphic units range from 2.58 to 3.08 billion years (Catanzaro and Kulp, 1964). Catanzaro and Kulp conclude that the area was subjected to an earlier metamorphism at least 3.1 billion years in age. In their summary at the end of the paper they refer, without explanation, to unpublished biotite dates that support the 3.1 billion year figure.

Further support for an event of such antiquity in this region comes from other localities in southwestern Montana. Giletti (1966) obtained K-Ar ages of 3.22 and 3.27 billion years on biotite from granitic gneiss in Gallatin Canyon. Using a Rb^{87} half-life of 47 billion years, he also obtained a whole-rock Rb-Sr age of 3.1 billion years on pre-Cherry Creek gneiss south of Dillon, Montana. In the Jardine-Crevasse Mountain area, Brookins (personal communication) obtained a strictly preliminary K-Ar age of 3.2 billion years for metamorphic rocks intruded by the Crevasse Mountain granite.

POTASSIUM-ARGON DATA

Potassium-argon data for samples analyzed in this study appear in table 1, and the geographic location of the samples is given in figure 1. The mineral dated in most samples is biotite. Feldspar or whole-rock determinations were made for a few samples from which it was difficult to separate sufficient quantities of biotite.

Apparent ages of 12 samples of gneiss have a broad spread from 2.78 to 1.40 billion years. Potassium-argon ages of feldspars are commonly low by as much as 30 to 40 percent (Zartman, 1964; Tilton and Hart, 1963; Wetherill, Aldrich, and Davis, 1955; Goldich, Baadsgaard, and Nier, 1957) due to argon loss, and therefore, the particularly low whole-rock and feldspar ages may have no geologic significance. The remaining biotite ages of seven samples of gneiss range from 2.78 to 2.31 billion years; the spread clearly exceeds the experimental error (5 percent).

Apparent ages of biotite from the granitic rocks, including both quartz diorite and quartz monzonite, show the same substantial spread (0.47 b.y.) as those of biotite from the gneiss. Ages for quartz monzonite are internally consistent and average 2.78 billion years. Ages for two quartz diorite samples agree well with this figure (2.79 b.y.); however, ages for two other quartz diorite samples are almost identical and average 3.14 billion years.

The age determinations on single specimens of agmatite, Lake McLain quartz diorite, peridotite, dolerite, and a schist skialith (sample 3115) within quartz diorite are of interest but probably not very meaningful because of the limited sampling of these lithologic units. However, the antiquity of the peridotite (2.99 b.y.) and the skialith (3.06 b.y.) bears nothing.

LEAD-URANIUM DATA

The location of a sample of quartz monzonite (#51) dated by the Pb-U method is plotted on figure 1. Zircon, monazite, and microcline were separated from the sample by standard procedures and analyzed for Pb isotope composition and Pb and U concentrations. Zircon fractions were washed for one hour in hot concentrated HNO_3 prior to dissolution in borax flux. Monazite, handpicked for purity, was dissolved in HF and $HClO_4$ without previous treatment. The $-100 + 200$ mesh fraction of microcline was washed in hot concentrated HNO_3, pulverized to -200 mesh in a clean mortar, and washed in hot acid again before being dissolved.

Analytical results for the zircon and monazite fractions are shown in table 2 and for the feldspar in table 3. We assign the following uncertainties to the observed Pb isotope ratios:

Ratio	Radiogenic Pb	Common Pb
206/204	± 1.5 percent	± 0.4 percent
206/207	± 0.4 percent	± 0.15 percent
206/208	± 0.6 percent	± 0.25 percent

Pb/U concentration ratios have an uncertainty of approximately 1.5 percent.

TABLE 1
Potassium-argon data, Bighorn Mountains, Wyoming

Specimen number	Rock unit	Material dated	% K	% Radiogenic argon	Grams A^{40}/ gram mineral	A^{40}/K^{40}	Age in b.y.*
3389	Gneiss	Biotite	6.50	99.1	2.87×10^{-6}	0.371	2.78
3437	Gneiss	Biotite	5.74	99.0	2.44×10^{-6}	0.358	2.72
4166	Gneiss	Biotite	6.19	99.2	2.60×10^{-6}	0.353	2.70
4123	Gneiss	Biotite	5.97	99.0	2.48×10^{-6}	0.349	2.68
3217	Gneiss	Biotite	6.60	99.1	2.48×10^{-6}	0.316	2.55
4199	Gneiss	Biotite	3.34	76.8	1.15×10^{-6}	0.290	2.43
4212	Gneiss	Biotite	6.62	98.6	2.09×10^{-6}	0.266	2.31
4111	Gneiss	Feldspar	0.82	97.5	2.00×10^{-6}	0.206	1.98
4177	Gneiss	Feldspar	0.98	97.3	2.37×10^{-7}	0.204	1.97
3406	Gneiss	Feldspar	2.05	98.2	4.40×10^{-7}	0.180	1.83
4143	Gneiss	Feldspar	3.20	98.4	4.61×10^{-7}	0.121	1.40
3429	Gneiss	Whole-Rk	1.89	97.9	5.36×10^{-7}	0.238	2.17
3137	Agmatite	Biotite	3.93	99.4	1.96×10^{-6}	0.420	2.96
4115	Quartz Diorite	Biotite	6.93	98.7	4.00×10^{-6}	0.485	3.18
3153	Quartz Diorite	Biotite	6.42	99.5	3.52×10^{-6}	0.461	3.10
4142	Quartz Diorite	Biotite	6.42	87.1	2.97×10^{-6}	0.389	2.85
3143	Quartz Diorite	Biotite	5.49	99.7	2.37×10^{-6}	0.363	2.74
338	Quartz Monzonite	Biotite	6.50	99.2	2.98×10^{-6}	0.386	2.83
349	Quartz Monzonite	Biotite	4.67	98.4	2.09×10^{-6}	0.377	2.80
386	Quartz Monzonite	Biotite	6.68	98.6	2.82×10^{-6}	0.360	2.71
3115	Schist Skialith	Biotite	5.89	99.0	3.15×10^{-6}	0.449	3.06
4149	Lake McLain Quartz Diorite	Whole-Rk	6.31	86.6	2.64×10^{-6}	0.351	2.69
3225	Peridotite	Whole-Rk	0.45	95.8	2.28×10^{-7}	0.428	2.99
39	Dolerite	Feldspar	1.22	95.6	3.12×10^{-7}	0.215	2.04

* Constants used: $\lambda\beta = 4.72 \times 10^{-10}\ yr^{-1}$, $\lambda_e = 0.585 \times 10^{-10}\ yr^{-1}$.

TABLE 2

Pb-U analytical data for zircon and monazite from quartz monzonite

Mineral	Size fraction	Weight (mg)	Observed Pb isotope ratios			Radiogenic Pb isotope composition (%)			Concentration (ppm)		Atom ratios			Apparent ages (m.y.)		
			$\frac{206}{204}$	$\frac{206}{207}$	$\frac{206}{208}$	206	207	208	Pb^r	U	$\frac{206}{238}$	$\frac{207}{235}$	$\frac{207}{206}$	$\frac{206}{238}$	$\frac{207}{235}$	$\frac{207}{206}$
Zircon	P100R200	88.1	1067.5	4.861	7.569	77.37	15.04	7.59	287.8	580.4	0.4461	11.96	0.1944	2400	2635	2820
Zircon	P200R285	38.8	1041.1	4.853	7.210	77.02	14.98	8.00	300.4	595.1	0.4521	12.12	0.1945	2425	2650	2820
Zircon	P285	47.8	1237.1	4.882	6.600	75.89	14.81	9.30	284.7	533.0	0.4712	12.68	0.1951	2510	2690	2830
Monazite	P100R200	1.8	206.31	3.833	0.04563	4.036	0.818	95.15	9984	829.7	0.5597	15.65	0.2027	2890	2890	2890

Decay constants: $\lambda_{238} = 1.54 \times 10^{-10}\ yr^{-1}$; $\lambda_{235} = 9.72 \times 10^{-10}\ yr^{-1}$.

TABLE 3

Pb-U analytical data for potassium feldspar from quartz monzonite

Weight (gm)	Pb isotope ratios $\frac{206}{204}$	$\frac{206}{207}$	$\frac{206}{208}$	$\frac{207}{204}$	$\frac{208}{204}$	Concentration (ppm) Pb	U
1.875	13.68	0.9187	0.4058	14.89	33.71	21.9	0.08
Corrected*	13.57	0.9127	—	14.87	—		

* Corrected for decay of U.

The common lead correction for the zircon analyses was assumed to have the isotopic composition of Pb extracted from the borax fusion flux, which is very nearly the same as that of modern common lead (Chow and Patterson, 1962). The calculated apparent ages for these samples are virtually unaffected by the particular choice of common lead correction, so that the maximum uncertainty in any of the individual apparent ages is about 1.5 percent. The common Pb correction for the monazite sample was assumed to consist of equal parts of feldspar lead and modern common lead. The Pb^{207}/Pb^{206} apparent age of this sample is not strongly affected by the choice of correction and therefore has an uncertainty of about 1 percent. The Pb^{206}/U^{238} apparent age, however, is somewhat affected by the choice of correction, so that we assign to it an uncertainty of 2 percent.

Figure 2 summarizes the radiometric data on a concordia diagram (Wetherill, 1956). A straight line has been drawn through the analytical points to emphasize their coherent relationship but is not based on any particular model of isotopic disturbance in the zircons. Wasserburg (1963) has shown that a variety of diffusion loss models can be constructed to "explain" a linear relationship near the upper intersection with the concordia curve, episodic loss being a limiting case of rapid diffusion during a short time interval. Because any interpretation of the disturbance mechanism for this single rock sample is non-unique and because considerable uncertainty accompanies the long extrapolation of the chord to its lower intersection with the concordia curve, the lower part of the diagram has been omitted. The data are internally consistent with the interpretation that the U-Pb systems in these minerals originated or were completely reset 2890 ± 25 million years ago and that the zircons subsequently lost a fraction of their radiogenic Pb by a process or processes not specified by the data. The uncertainty assigned to the resulting age for the rock is an estimate of our ability to reproduce this figure and does not take into account possible errors in half-lives. et cetera.

The isotopic composition of Pb in the feldspar, uncorrected for uranium, has a closed system model age of 2.9 billion years, which agrees very well with the radiometric age of the rock. Correction for included uranium introduces a considerable uncertainty because it is not known

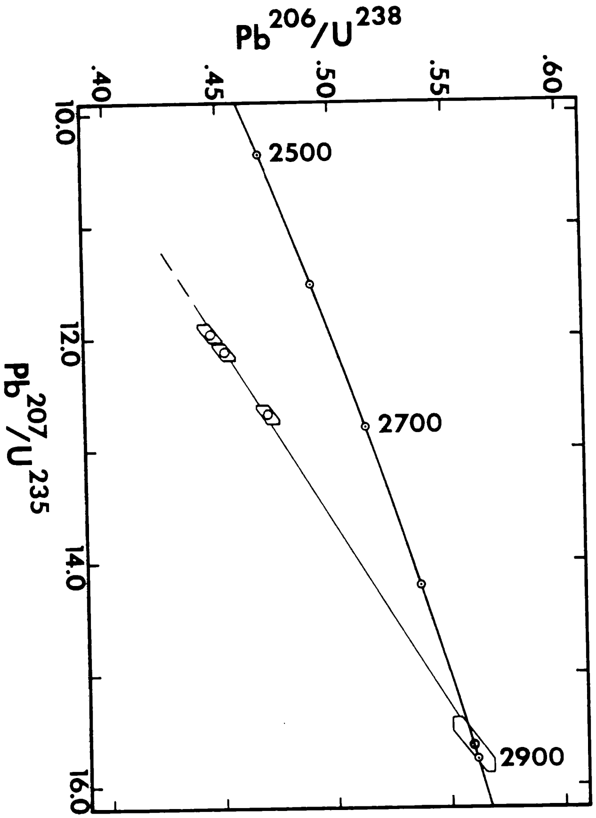

Fig. 2. Concordia plot of analytical results for zircon and monazite from quartz monzonite.

whether U and radiogenic Pb are leached in constant proportion during acid washing. In the present case the corrected lead has a model age of 3.0 billion years, which is still in good agreement with the radiometric age. The amount of Pb^{207} in the feldspar lead is somewhat higher than that in most leads of this general age range (see for example Doe, Tilton, and Hopson, 1965) and yields a value of about 9.5 for the apparent U^{238}/Pb^{204} ratio (μ_0) of the source region. In the absence of more extensive data from this and other localities we cannot draw any significance from this factor.

DISCUSSION

Because of their wide variation, the K-Ar apparent ages reported here can be considered minimum ages only, the oldest apparent ages being closer to the time of formation of the rock. This is suggested, particularly, by the biotite ages of gneiss samples. These ages show a progressive variation rather than a distinct clustering about one or more ages.

The narrow spread of quartz monzonite biotite ages and the great spread but grouping of quartz diorite ages about two values may well reflect the limited sampling of these granitic units as compared with that of the gneiss. In this connection it is interesting to note that the zircon-monazite data for quartz monzonite indicate an age that is older than any of the biotite ages for this unit. In short, partial argon loss probably affected all samples used in this study. Petrographic analyses indicate that, in all samples dated, at least some biotite displays incipient chloritization. However, no direct relationship exists between degree of chloritization and apparent age.

There is nothing definitive about the data that would indicate an age difference between the gneiss and the granitic rocks. From the U-Pb data, the quartz monzonite has a best-value age of 2.89 billion years. K-Ar data for this unit suggest a minimum age of 2.83 billion years which agrees well. The quartz diorite, which is gradational with quartz monzonite in the field and under the microscope, has a minimum K-Ar age of 3.18 billion years, surprisingly old, although analytical errors have possibly biased this figure toward the high side. However, the K-Ar minimum age of the gneiss, 2.78 billion years, agrees rather well with the 2.89 billion years age of the quartz monzonite. In summary, one might conclude that the three lithologies crystallized or were recrystallized during a widespread event that occurred on the order of 2.9 to 3.1 billion years ago in this area. The individual agmatite (2.96 b.y), schist skialith (3.06 b.y.), and peridotite (2.99 b.y.) dates tend to confirm this. Whether the spread of 200 million years or so reflects analytical errors, uncertainties in half-lives, and the like, or whether there truly are resolvable events separated by a time interval of this magnitude, must await further study. In view of the prevailing lack of resolution in identifying Precambrian orogenic events, we consider it a useful first approximation to state that the event or events that formed the crystalline rocks of the Bighorn Mountains took place during a relatively short time interval (that is, roughly of the order of magnitude of the resolving power of the radiometric methods) centered approximately 3.0 billion years ago.

Petrographic observations were made of the zircons to determine their origin relative to that of the quartz monzonite in which they occur. Individual zircons are typically purplish-brown in color. Some show well-developed zoning. Their length/width ratio is on the order of 2. A small but significant percentage of the zircons have clearly euhedral shapes; however, a distinct majority in all size fractions show some degree of (detrital?) rounding. Statistical study following the method of Poldervaart (1950) and Larsen and Poldervaart (1957) produced the following results:

Size fraction	Rounded grains
–100 + 200	67 percent
–200 + 285	76 percent
–285	79 percent

Rounded grains range from well-rounded to subrounded and include a small percentage that may well have undergone resorption. The shapes of some "resorbed" grains, which possess curved surfaces convex toward their interiors, may have developed by rounding of composite grains consisting of partial overgrowths, the faces of which make obtuse angles with the prism faces of the nucleus grain. By the same token, so-called rounded grains may appear so, in part at least, due to indistinct development of the second and third order dipyramids and the basal pinacoid. Some grains appear to consist of a subhedral or euhedral overgrowth built on a well-rounded core. On the other hand, some of the zoned zircons have simulated overgrowths because the outermost zones possess a shape different from that of the central portion. The point is that unequivocal interpretations based on zircon morphology cannot be made relative to this single sample.

Field and petrographic evidence accumulated by Osterwald (1955), Hoppin (1961), and Heimlich (1965) suggest that the quartz monzonite is of metasomatic origin. Therefore, study of additional zircon suites may well demonstrate that the zircons are of detrital origin. If so, it must be concluded that they were reset during the major metamorphic-metasomatic event in this area.

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