Constraints on the Formation of the Bitterroot Lobe of the Idaho Batholith, Idaho and Montana, from U-Pb Zircon Geochronology and Feldspar Pb Isotopic Data

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By MARGO I. TOTH and JOHN S. STACEY

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By Margo I. Toth *and* John S. Stacey

Abstract

Zircons from Cretaceous and Tertiary plutonic rocks of the west-central portion of the Bitterroot lobe of the Idaho batholith contain abundant inherited cores that make precise U-Pb age determination difficult. Zircons from a tonalite emplaced along the western periphery of the batholith yield an almost concordant age of 94±1.4 Ma. Monzogranite and granodiorite plutons were intruded along the northern edge of the lobe from 75 to 71 Ma. Subsequently, the volumetrically more dominant plutons in the central and western parts of the lobe were emplaced between 59 and 54 Ma; most of the intrusive activity in this area probably occurred at about 55 Ma.

The most accurately determined upper-intercept age for the Cretaceous plutons is 1,779±20 Ma. This is similar to the 1,751±63 Ma age that was obtained for a gneissic muscovite-biotite monzogranite of probable Proterozoic age. For Tertiary rocks, upper-intercept ages are more variable, ranging from 1,812±21 Ma to 1,625±46 Ma. These results, combined with Pb isotopic data from feldspars, confirm that the magmas of the Bitterroot lobe were derived mainly from an Early Proterozoic lower continental crust. Our findings are consistent with studies in the northern and northeastern parts of the Bitterroot lobe by other workers.

INTRODUCTION

The Bitterroot lobe of the Idaho batholith has a complex geologic history that has made interpretation of geochronologic and isotopic data quite difficult. Important aspects of this geologic history include: diverse ages and compositions of source rocks, variable assimilation of country rock, presence of restite minerals (specifically xenocrystic zircon), intrusive activity over a period of at least 40 m.y., widespread hydrothermal aureoles generated by Eocene plutons, rapid uplift, and mylonitization of the central and eastern parts of the lobe.

Many workers have attempted different methods to resolve the geochronology of the Bitterroot lobe. Fleck (1990), Criss and Fleck (1987), Fleck and Criss (1985a, 1985b), Chase and others (1978), Armstrong and others (1977), Tripp (1976), and McDowell and Kulp (1969) have attempted to solve the problems through the use of Rb-Sr, K-Ar, and oxygen isotopic studies; Ferguson (1975) used fission-track dating. These studies indicated that the plutons of the Bitterroot lobe are mostly Paleocene in age. The studies also document a younger thermal event, about 45–40 Ma, that disturbed the isotopic systems in many of the older rocks concurrent with the formation of a mylonitic zone at about 48–42 Ma along the eastern border of the Bitterroot lobe.

To understand the generation of the mylonitic rocks, Garmezy (1983) and Garmezy and Sutter (1983) relied on Ar-Ar dating and detailed petrographic study. Further work on the mylonitic rocks was done by Chase and others (1983) and Arruda (1981) using U-Pb methods.

U-Pb studies of the Bitterroot lobe include those by Shuster and Bickford (1985), Chase and others (1983, 1978), Bickford and others (1981), Nelson (1981), Tripp (1976), and Grauert and Hoffman (1973). These studies have dealt mostly with the more accessible northeastern border of the Bitterroot lobe and the eastern zone of mylonites. Although Grauert and Hoffman (1973) were the first workers to recognize the presence of a component of old zircon in the granitic rocks, the work of Bickford and others (1981) is the most detailed study of the Proterozoic zircon components.

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The objective of our work is to study the evolution of the plutonic rocks of the Bitterroot lobe by using U-Pb data from zircon and monazite; we also present new data from interior, less accessible parts of the lobe. Common-Pb data from feldspars in the same samples provide additional information on source-rock characteristics. We compare our results with those of other workers and summarize the constraints on the ages and origins of the plutons of the Bitterroot lobe.

Granitic rocks are classified in this paper according to Streckeisen (1973). Structure, petrology, and petrogenesis of the Bitterroot lobe is described in detail by Hyndman (1984), Hyndman and Foster (1987), Foster and Hyndman (1990), Garmezy (1983), Reid (1987), and Toth (1987, 1983a, 1983b).

GEOLOGIC SETTING

The Idaho batholith is one in a series of subduction-related, Mesozoic, granitoid batholiths extending from Alaska to Mexico along the Cordillera of western North America. The Idaho batholith is similar to the other batholiths in age and petrology, but it was emplaced into mostly Proterozoic miogeoclinal metasedimentary rocks instead of Phanerozoic eugeoclinal rocks. The batholith is 390 km long and 80 to 160 km wide and extends from north-central Idaho and western Montana to southern Idaho, where it is covered by basalts of the Snake River Plain. An apparent arch of Proterozoic metasedimentary rocks divides the batholith into two parts: the Atlanta lobe in the south and the smaller and younger Bitterroot lobe in the north.

The Bitterroot lobe comprises a plutonic complex of diverse ages and compositions that crops out over about 14,000 km^2 (fig. 1). The lobe forms a northwest-trending tabular body that intrudes Proterozoic metasedimentary rocks to the north, south, and east and that intrudes allochtonous metavolcanic and metasedimentary rocks to the west. Small, isolated Proterozoic bodies of augen gneiss occur along and to the north of the Lochsa River (not shown in fig. 1); these bodies have not been mapped in any detail. Three compositionally different plutonic phases are contained within the Bitterroot lobe. From oldest to youngest these are: hornblende-biotite tonalite and quartz diorite plutons (unit Kt, fig. 1); muscovite-biotite granodiorite and monzogranite plutons (unit TKg, fig. 1); and hornblende-biotite syenogranite, quartz syenite, and monzogranite plutons (unit Tg, fig. 1).

Cretaceous mesozonal hornblende-biotite tonalite and quartz diorite plutons form the western border of the Bitterroot lobe (fig. 1); similar rocks also occur along the northeastern border as small, isolated plutons (one of these is shown in fig. 1). In hand specimen, the rocks are gneissic to well foliated, equigranular, medium grained, and contain accessory sphene, epidote, apatite, zircon, allanite, magnetite, and thorite. Major- and trace-element geochemistry (Toth, 1987) suggests that the plutons have assimilated some metasedimentary material; this conclusion is supported by field evidence.

Paleocene mesozonal plutons of muscovite-biotite granodiorite and monzogranite form the major part of the Bitterroot lobe. These plutons intrude tonalite and quartz diorite on the western, southwestern, and northern borders of the lobe (fig. 1). Complex migmatitic terranes sometimes separate different plutons and involve septa of Middle(?) Proterozoic, and possibly older, metasedimentary rocks. Inclusions of metasedimentary rock are also common along margins of plutons and consist of calc-silicates, quartzite, schist, and augen gneiss. The appearance, chemistry, and mineralogy of the granitic rocks are fairly homogeneous, but the work of Williams (1977), Toth (1983b, 1987), and Hyndman (1984) have outlined at least ten large plutons with different textural or chemical variations; more bodies may be present but are, as yet, unidentified.

The plutonic rocks are typically white to light gray and slightly porphyritic with subhedral to euhedral feldspar phenocrysts 6–8 mm long in a groundmass of medium-grained feldspar, quartz, biotite, and muscovite; medium-grained equigranular rocks are less abundant. Accessory minerals consist of magnetite, ilmenite, zircon (some with anhedral cores and euhedral overgrowths), monazite, apatite, and rare epidote. Major- and trace-element geochemistry suggests that the plutons may belong to a single comagmatic suite (Toth, 1987). The plutons were most likely derived from a homogeneous source rock (probably Proterozoic metasedimentary rocks) and underwent very little differentiation or assimilation before emplacement (Toth, 1983b; Hyndman, 1984).

Large, Eocene, epizonal bodies of subalkalic syenogranite, quartz syenite, and monzogranite intrude both country rock and all earlier igneous phases of the Bitterroot lobe along the northern, southwestern, and southern sides of the lobe (fig. 1). Rhyolitic to andesitic tuff and lava flows are preserved in two places in down-dropped fault blocks and locally overlie parts of the epizonal plutons on the south side of the Bitterroot lobe (Lund and others, 1983). The epizonal plutons are unfoliated, medium to coarse grained, and contain miarolitic cavities that commonly contain crystals of fluorite, smoky quartz, and microcline. Major minerals of the epizonal plutons are perthite, plagioclase, quartz, biotite, and hornblende; allanite, apatite, baddeleyite, cassiterite, fluorite, monazite, sphene, thorite, zircon, and iron-titanium oxide minerals are accessories. Major- and trace-element geochemistry (Toth, 1987) suggests that the plutons were probably derived from partial melting of a depleted upper-mantle source and assimilated some crustal material during emplacement.

A 200- to 500-m-thick mylonite zone, which dips shallowly to the east, borders the eastern and southeastern sides of the Bitterroot lobe, adjacent to the Bitterroot

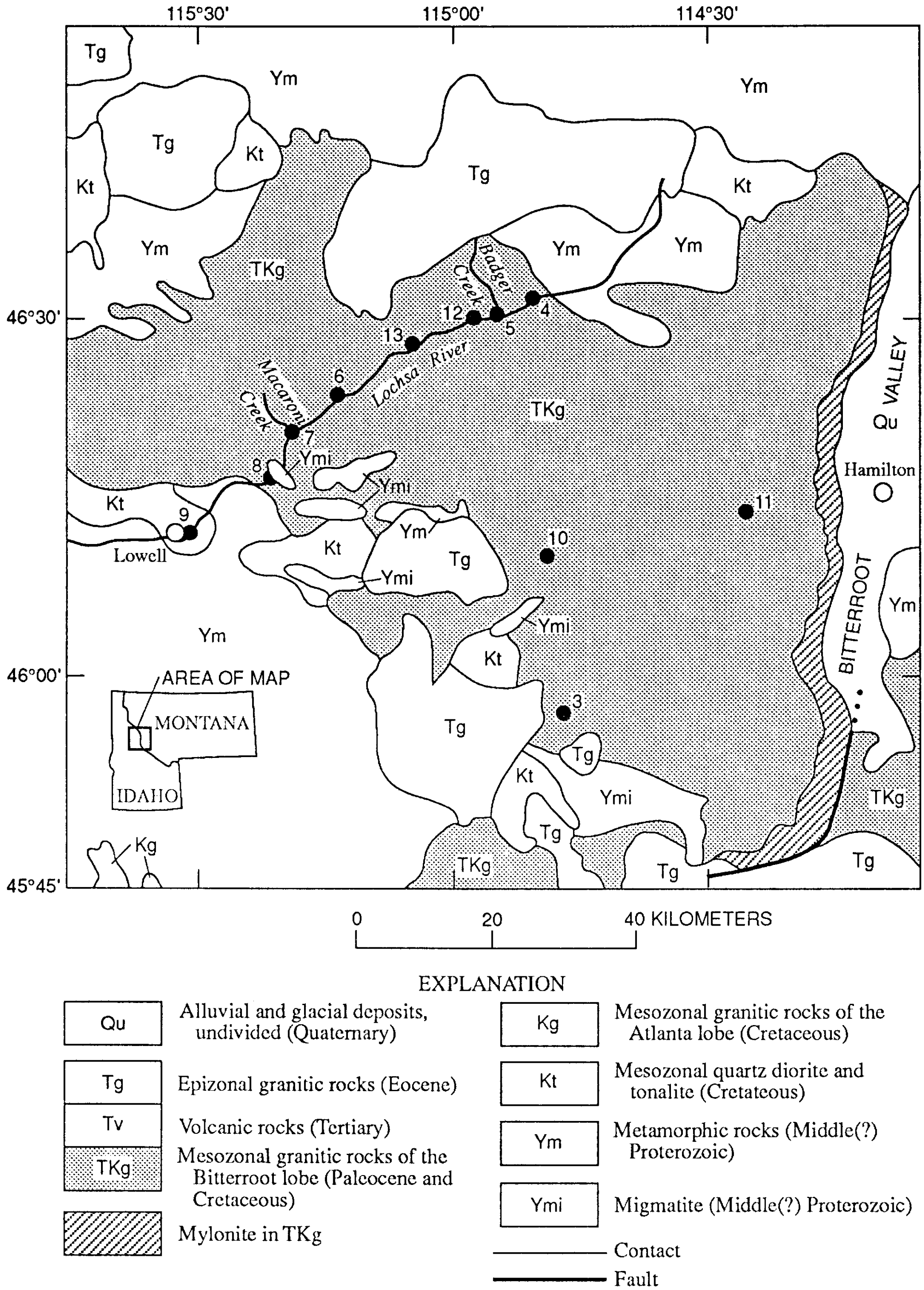

Figure 1. Geologic map of the Bitterroot lobe of the Idaho batholith showing sample locations and BRM sample numbers (from table 1). Geology from Bond (1978), Toth (1983a), Lund and others (1983), and Nold (1968).

Valley, Montana (fig. 1). The mylonite zone is younger than all other rock units except the Eocene epizonal plutons, although its formation may have been contemporaneous with the intrusion of the younger of the Eocene plutons. The formation of the mylonite zone has been attributed to gravity sliding of an overlying block (Hyndman, 1977) and to emplacement-related phenomena combined with extensional tectonism or crustal thinning (Garmezy, 1983;

Garmezy and Sutter, 1983; Chase and others, 1983). Mylonitic textures in the granodiorite and monzogranite plutons extend into the center of the lobe.

GEOCHRONOLOGY

Sample Selection

State Highway 12, which roughly parallels the Lochsa River (fig. 1), transects the northern end of the Bitterroot lobe of the Idaho batholith (the highway is not shown on fig. 1), and a cross section of the lobe is conveniently exposed along the highway. Relationships between the rocks exposed along Highway 12 are described in Reid and others (1979). Eight samples were collected from this area and included a tonalitic gneiss from the western periphery, six granodiorites or monzogranites from the mesozonal plutons, and one sample of Proterozoic augen gneiss that caps many of the plutons exposed along the highway. Farther south, in the central and western parts of the Bitterroot lobe, three samples of monzogranite or granodiorite were also collected. One of these samples was collected close to a younger epizonal pluton, but unaltered, unsheared rocks were otherwise chosen. Sample locations are shown in figure 1; analytical data are shown in table 1; and brief sample descriptions are given in table 2. Samples of Eocene epizonal plutons were not collected.

Analytical Methods

Methods for chemical separation and mass spectrometry of Pb and U in zircons were slightly modified from Krogh (1973). Pb blank levels were 0.3 to 1.0 nanograms.

Feldspar concentrates were washed with hot 7*N* HNO_3 and 6*N* HCl and then leached for 20 minutes in warm 5 percent HF to minimize any radiogenic Pb derived from in situ U and Th decay (Ludwig and Silver, 1977). Leaching was followed by microwave dissolution in a closed Teflon capsule using a solution of 48 percent HF, 15*N* HNO_3, and 50 percent $HClO_4$ (Fischer, 1986). Pb was extracted using a chloride-form anion-exchange column.

Isotope ratios were measured with a 30-cm radius, 90° sector, computer-controlled mass spectrometer. Uranium analyses utilized a triple filament assembly, and Pb was run with silica gel on a single rhenium filament. Corrections for Pb isotopic fractionation were determined from replicate analyses of a standard sample (NBS SRM–981) to be 0.12±0.025 percent per mass unit. For feldspar Pb, reproducibility of isotope ratios with respect to ^{204}Pb is within 0.10 percent as determined from duplicate filament loadings of feldspar Pb separates. All errors are quoted at the 95 percent level of confidence.

Table 1 shows that most of the ^{206}Pb/^{204}Pb ratios in the zircons were greater than 1,000; the effects of common-Pb corrections on our concordia-intercept ages are, therefore, generally small. Common-Pb corrections were made by using the composition of the appropriate feldspar Pb or by using model ratios of Stacey and Kramers (1975) corresponding to the intrusion age of the sample. Analytical precision is within approximately 1.0 percent for ^{206}Pb/^{238}U and ^{207}Pb/^{235}U (at the 95 percent confidence level) as determined from replicate analyses of a standard concordant zircon sample.

Regression lines for the data on concordia plots are calculated using the procedure of York (1969). Concordia-intercept errors are computed from error envelopes generated around the regression lines at the 95 percent confidence level. For almost all samples in this study, scatter in the data around the regression lines exceeds that expected from analytical errors alone. Unlike analytical error, this extra "geologic" scatter may vary between samples, and, therefore, a Student's T factor must be used to estimate errors for the slopes of the regression lines and their intercepts on concordia (Ludwig, 1980). This statistical limitation was not applied in some earlier zircon studies of the Idaho batholith (e.g., Bickford and others, 1981; Chase and others, 1983). For example, a Student's T factor for four fractions is 4.3, and such a factor would significantly increase the errors claimed for a number of the intercept-model ages in those studies. Nevertheless, although different methods have been adopted for error estimation, ages obtained here are in good agreement with previous studies.

Krogh (1982) demonstrated that abrasion could be used to remove the more metamict outer portions from Archean zircon crystals, thereby yielding more concordant U-Pb data. Most samples in our study contain zircons with Early Proterozoic cores and younger Mesozoic-Cenozoic overgrowths; these zircons are similar to those from farther east in the Idaho batholith that were described in detail by Bickford and others (1981). For a number of cases in our study, abrasion of such composite zircons has removed a portion of the younger material surrounding older cores and has resulted in data with older U-Pb model ages. Consequently, when data from all zircon fractions of a sample are combined, longer chords are obtained that usually yield more precisely determined concordia-intercept ages. For our work, abrasion was carried out in an air-driven cell developed by J.N. Aleinikoff and reported by Goldich and Fischer (1986).

Table 1. U-Pb analytical data for zircons and monazite from the Bitterroot lobe of the Idaho batholith

[Atomic ratios are corrected for common Pb as described in the text. Abbreviations following mesh-size designations are: M, magnetic; NM, nonmagnetic; A, abraded; no abbreviation indicates that the sample was not split magnetically and was not abraded]

Sample no. (mesh size)	Sample weight (milligrams)	U (ppm)	Pb (ppm)	Atomic ratios $^{206}Pb/^{238}U$	Atomic ratios $^{207}Pb/^{235}U$	Atomic ratios $^{207}Pb/^{206}Pb$	Apparent ages (Ma) $^{206}Pb/^{238}U$	Apparent ages (Ma) $^{207}Pb/^{235}U$	Apparent ages (Ma) $^{207}Pb/^{206}Pb$	Measured $^{206}Pb/^{204}Pb$
BRM-3										
(+100)	12.01	724	17.6	.02370	.2878	.08807	151	257	1384	2467
(-400)	15.15	834	8.68	.01021	.0783	.05558	66	77	4372	183
BRM-4										
(+150)NM	11.04	853	63.6	.07224	.9902	.09941	450	699	161	12476
(-150+250)M	7.78	994	65.4	.06259	.8130	.09420	391	604	1512	4334
(-325+400)M	10.97	971	60.2	.05982	.7715	.09354	375	581	1499	16042
(-400)	12.36	931	57.9	.06019	.7824	.09428	377	587	1513	6050
BRM-5										
(+100)	7.43	1386	59.5	.04224	.4956	.08509	267	409	1318	3458
(-400)NM	8.66	1554	68.8	.04421	.5215	.08556	279	426	1328	5444
BRM-6										
(+200)NM	14.3	997	30.9	.02940	.3650	.09002	187	316	1426	2035
(+200)NM,A	17.06	922	32.3	.03314	.4186	.09159	210	355	1459	3500
(-400)M	12.06	1391	28.3	.01958	.2190	.08109	125	201	1224	4380
BRM-7										
(+100)	32.59	485	52.4	.10019	1.4431	.10447	616	907	1705	1118
(+100)NM,A	1.29	557	87.9	.15291	2.2420	.10634	917	194	1737	4683
(-150+200)	9.36	602	53.5	.08468	1.1956	.10239	524	799	1668	7126
(-400)	36.28	687	35.1	.04816	.6440	.09698	303	505	1567	12520
(-400)NM	4.23	633	41.8	.06202	.8497	.09937	388	624	1612	5540
(-400)M	8.10	684	39.3	.05386	.7250	.09764	338	554	1580	4668
BRM-8										
(+100)NM	11.25	1116	133.6	.13921	2.0530	.10696	840	1133	1748	23970
(+100)A	10.57	636	159.0	.24210	3.5900	.10750	1398	1547	1758	17140
(-250)NM	19.96	972	190.0	.19010	2.7872	.10630	1122	1352	1737	54640
(-325)M	5.96	1062	183.6	.16641	2.4170	.10534	992	1248	1720	10280
(-400)NM	6.57	956	137.6	.16720	2.4515	.10634	997	1258	1737	6650
BRM-9										
(+100)NM	31.74	219	3.6	.01467	.0972	.04806	94	94	102	531
(-100+150)NM	33.25	246	3.6	.01403	.0958	.04955	90	93	174	1284
(-150+200)	31.54	269	4.4	.01495	.1017	.04934	95	98	164	630
(-325+400)	32.59	339	5.4	.01481	.0995	.04871	95	96	134	876
BRM-10										
(-100+150)	8.53	840	29.1	.03311	.4307	.09433	210	364	1515	1806
(-150+200)	14.58	882	24.3	.02653	.3256	.08900	169	286	1404	2574
(-325+400)	32.59	849	14.0	.01394	.1427	.07428	89	135	1049	334
(-400)	31.27	1062	11.0	.00980	.0862	.08900	70	84	734	2994
BRM-11										
(+150)NM	22.17	646	16.2	.02477	.2918	.08543	158	260	1325	4340
(-400)NM	11.62	734	10.2	.01326	.1201	.06569	85	115	796	855
BRM-12										
(+150)	7.98	1282	28.7	.02226	.2474	.08060	142	224	1212	4149
(-250+325)M	8.05	1258	36.2	.02773	.3244	.08485	176	285	1312	4077
(-400)	18.27	1443	39.4	.02689	.3117	.08408	171	276	1295	8528
BRM-13										
(+200)NM	17.94	762	39.0	.04964	.6646	.09710	312	517	569	8260
(-250)M	5.99	1005	33.7	.03182	.3882	.08849	202	333	1353	7973
Monazite	5.11	670	102.4	.00951	.0605	.04613	61	61	60	222

Table 2. Summary of U-Pb zircon ages, U-Pb model ages, and sample descriptions from the Bitterroot lobe

[Errors are at the 95 percent level of confidence. MSWD, mean standard weighted deviates of Brooks and others (1972)]

Sample no.	Coordinates (lat) (long)	Zircon fractions regressed	Intercept-model ages Lower (Ma)	Intercept-model ages Upper (Ma)	MSWD	Sample description
				Proterozoic		
BRM-5	46°28'30"N. 114°54'40"W.	2	*	*	*	Proterozoic augen gneiss at Badger Creek, approx 90% Pb loss at approx 55 Ma.
BRM-8	46°13'42"N. 115°28'49"W.	3	24±140	1,751±63	53.2	Gneissic muscovite-biotite monzogranite, Pb loss effectively at approx 115 Ma.
				Jurassic		
BRM-4	46°29'06"N. 114°52'15"W.	4	170±52	2,000±200	2.5	Medium-grained, muscovite-biotite granodiorite from near roof of pluton.
				Cretaceous		
BRM-7	46°17'06"N. 115°23'06"W.	6	71±9	1,779±20	6.2	Medium-grained, muscovite-biotite monzogranite.
BRM-9	46°08'24"N. 115°36'06"W.	3	94±1.4	Concordant	0.003	Medium-grained, gneissic biotite tonalite; some (primary?) muscovite.
BRM-13	46°27'06"N. 115°05'30"W.	2	75±5	1,801±23	0	Medium-grained biotite granodiorite; slightly mylonitized.
				Tertiary		
BRM-3	45°56'53"N. 114°46'23"W.	2	57±1.3	1,812±21	0	Medium-grained biotite monzogranite.
BRM-6	46°21'30"N. 115°17'12"W.	3	54±2.5	1,722±20	1.6	Medium-grained, foliated, muscovite-biotite monzogranite.
BRM-10	46°09'42"N. 114°47'12"W.	4	47±11	1,720±130	89.4	Medium-grained, weakly foliated, biotite granodiorite.
BRM-11	46°13'12"N. 114°27'06"W.	2	59±1.7	1,738±25	0	Foliated, muscovite-biotite granodiorite porphyritic gneiss.
BRM-12	46°28'30"N. 114°56'06"W.	3	55±5.4	1,625±46	0.67	Slightly foliated, medium-grained, muscovite-biotite granodiorite; capped by Proterozoic augen gneiss.

* Almost coincident data; no regression possible

U-Pb Zircon Data

Sample locations are shown on figure 1 and are described in table 2. Analytical data are presented in table 1, and the results of age computations are shown in table 2. The samples are discussed below in order from Proterozoic to Tertiary.

Proterozoic Samples

BRM–5: Some of the Cretaceous and Paleocene plutons clearly intrude a Proterozoic augen gneiss unit that is well exposed at several points along Highway 12 (Reid and others, 1979). Xenoliths of petrologically similar augen gneiss occur along pluton margins in the center of the lobe. The augen gneiss was sampled near the northeastern edge of the lobe at Badger Creek (BRM–5, fig. 1). Unfortunately, the two data points are very discordant and almost coincident because the zircons in both fractions have lost almost 90 percent of their Pb. An attempt was made to abrade the remainder of these highly metamict zircons, but they completely disintegrated in the process. The proximity of this sample to the younger intrusive rocks of the batholith probably accounts for this discordant data. A concordia plot is not presented for BRM–5, and the data are useless for geochronology. However, similar augen gneisses have been dated elsewhere in central and northern Idaho at 1,370±10 Ma and 1,576±13 Ma by Evans and Fischer (1986) and Evans and Zartman (1990).

BRM–8: This sample is a muscovite-biotite monzogranite from a pluton along the northwestern edge of the batholith and was collected between Proterozoic metamorphic country rock and a Proterozoic migmatite unit. Although the rock has some similarity to some other phases of the Bitterroot lobe in hand specimen, it is strongly gneissic. In thin section, the rock shows several features that contrast to the main plutons of the Bitterroot lobe. The sample is strongly foliated; contains large, euhedral grains of muscovite; and has a higher concentration of euhedral, accessory zircon and monazite, most of which form large radioactive haloes in the biotite and muscovite. Our data show that Early Proterozoic zircons in this sample are the least discordant of any we have analyzed and have an upper-intercept age of 1,751±63 Ma (fig. 2*A*, table 2). The MSWD (mean standard weighted deviates of Brooks and others (1972)), equal to 53, indicates that the zircons of this pluton must have had a complex Phanerozoic history (the lower-intercept model age of 24±140 Ma is meaningless). BRM–8 is most likely a Proterozoic remnant of either wallrock or source rock for the Bitterroot lobe and was severely affected during emplacement of the batholith.

Jurassic Samples

BRM–4: This sample is from a muscovite-biotite granodiorite pluton from the northeastern edge of the batholith. Four zircon fractions yield a significant scatter in the concordia diagram and a very poorly defined "age" of 170±52 Ma (fig. 2*B*, table 2). This sample is within several miles of Proterozoic metamorphic wallrocks. The sample apparently assimilated material of different ages from the source region or wallrocks, resulting in the scatter in the zircon data. The U-Pb systems in the zircons may also have been disturbed by later events in the batholith. These data demonstrate the difficulty of zircon dating for many of the rocks of the batholith.

Cretaceous Samples

BRM–7: This muscovite-biotite monzogranite was collected near the northwestern edge of the batholith at Macaroni Creek where it enters the Lochsa River. Five data points from BRM–7 form a chord with intercept-model ages of 72±16 and 1,782±39 Ma. Data from abrasion of the plus-100-mesh fraction extends the chord and improves the precision of the intercept ages to 71±9 and 1,779±20 Ma (fig. 2*C*, table 2).

BRM–9: This sample is from one of the older biotite tonalite gneisses at the western edge of the Bitterroot lobe (Toth, 1987). Three zircon fractions, each having small inherited components, yield a lower-intercept age of 94±1.4 Ma (fig. 2*D*, table 2).

BRM–13: This biotite granodiorite is from the northern part of the Bitterroot lobe. A chord for two fractions gives concordia intercepts of 75 Ma and 1,801 Ma with analytical errors of 5 and 23 Ma, respectively—these errors are surely underestimated (fig. 2*C*, table 2). These data plot close to the chord for BRM–7 and are nearer to the concordia lower intercept. It seems likely that BRM–13 was intruded during the same event as BRM–7, approximately 72–71 Ma.

Tertiary Samples

Four widely spaced Tertiary samples have lower-intercept ages that are almost identical, in the range 59 to 54 Ma; a fifth sample may be somewhat younger: 47±11 Ma.

BRM–3: This biotite monzogranite is in the southern part of the Bitterroot lobe, close to several Eocene plutons. Only two zircon fractions have been analyzed from this sample; these fractions yielded concordia-intercept ages of 57±1.3 and 1,812±21 Ma (fig. 3*A*, table 2). For two fractions, the computed uncertainties are for analytical errors only and are probably underestimates. Data for the minus-400-mesh fraction lie very close to concordia, and any geologic scatter in the data from analysis of additional fractions could have only a minimal effect on the precision of the lower-intercept age. Therefore, in the absence of further data, we estimate the emplacement age of this pluton to be about 57 Ma.

BRM–6: This sample is from one of the monzogranite plutons in the north-central part of the Bitterroot lobe. The data for three zircon fractions that were analyzed from this monzogranite plot farther from the concordia than the other samples and give intercept ages of 54±2.5 and 1,722±20 Ma with an MSWD of 1.6 (fig. 3*B*, table 2).

BRM–10: This is a sample of a biotite granodiorite from the central part of the lobe, a few km from the epizonal Whistling Pig pluton of Lund (1980). Four zircon fractions from this weakly foliated granodiorite exhibit scattered data about a chord on the concordia with an MSWD of 89.4 (fig. 3*C*, table 2). The intercept ages are 47±11 and 1,720±130 Ma. The Whistling Pig pluton was dated by K-Ar methods on hornblende at 51.3±1.0 Ma (Toth, 1983b) and clearly intrudes the pluton from which BRM–10 was collected. Significant hydrothermal alteration effects extend into the country rocks for several km around the Whistling Pig pluton, and it is quite likely that Pb loss occurred in sample BRM–10; this yielded the scatter in the data and the anomalously young age.

BRM–11: This muscovite-biotite granodiorite is in the eastern part of the Bitterroot lobe, close to the mylonite front, and shows mylonitic textures. Two zircon fractions give concordia-intercept-model ages with unrealistic errors

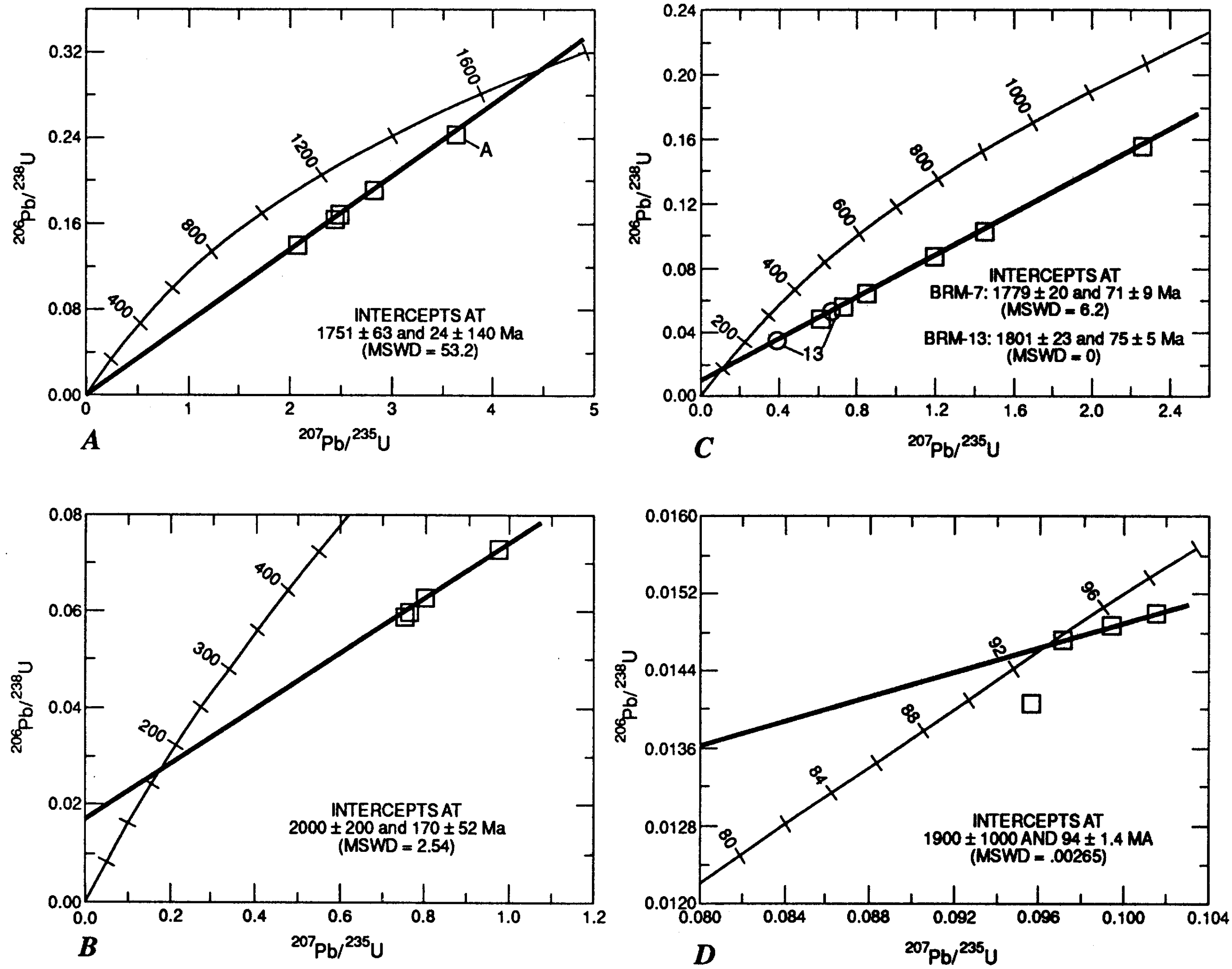

Figure 2. Concordia plots for zircon data that have Cretaceous lower intercepts. Data points labeled "A" indicate abraded zircon fractions. *A*, Concordia plot for sample BRM–8; *B*, Concordia plot for sample BRM–4; *C*, Concordia plots for samples BRM–7 and BRM–13 (open squares indicate data points for sample BRM–7; open circles indicate data points for sample BRM–13); *D*, Concordia plot for sample BRM–9.

of 59±1.7 and 1,738±25 Ma (fig. 3*A*, table 2). The point for the smallest sized fraction also plots fairly close to concordia, and, in the absence of further data, we conclude that the age is approximately 59 Ma.

BRM–12: This sample is from one of the muscovite-biotite plutons in the northeastern part of the Bitterroot lobe and is capped by Proterozoic augen gneiss. Data from three zircon fractions form a linear array (MSWD=0.67) with intercept ages of 55±4 Ma and 1,625±46 Ma (fig. 3*D*, table 2).

BRM–12 is from one of the plutons that intrudes the Proterozoic augen gneiss of BRM–5, and the sample was taken about 10 m below the roof of the pluton. The upper-intercept age of 1,625±46 Ma is the youngest that we have found in this part of the Bitterroot lobe; most of the other upper-intercept ages are in the range from 1,780 to 1,720 Ma (table 2). It is possible that zircons that were assimilated from the approximately 1,500-Ma augen gneiss are responsible for the younger upper-intercept age of this sample.

Feldspar Pb Data

Pb isotopic data for feldspars are presented in table 3 and are plotted in figure 4.

In the previous section, we concluded that monzogranite BRM–8 is probably a remnant of one of the Early Proterozoic source rocks that was significantly affected by emplacement of the batholith. The Pb isotopes are considerably more radiogenic in this sample than in any of

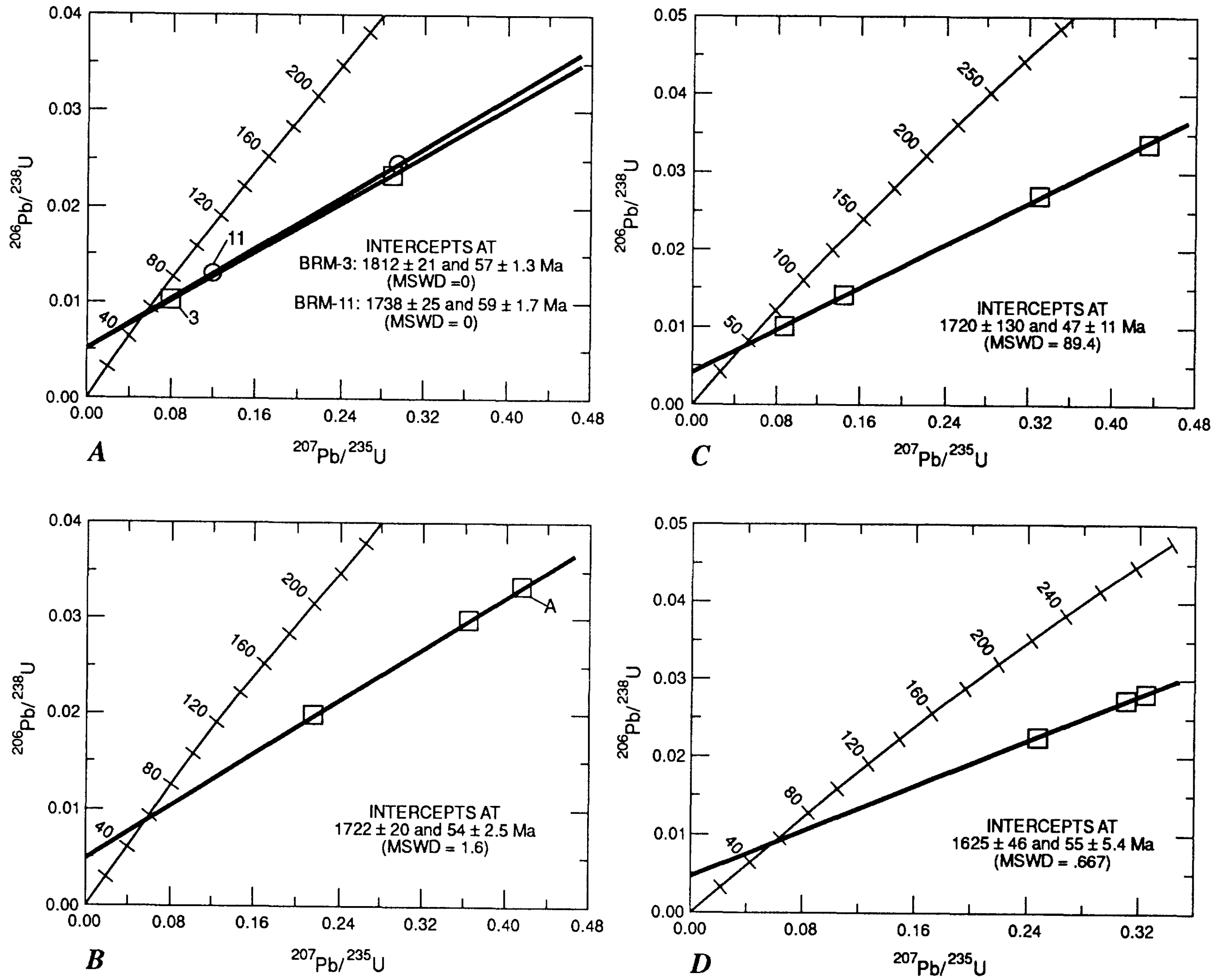

Figure 3. Concordia plots for zircon data that have Tertiary lower intercepts. Data points labeled "A" indicate abraded zircon fractions. *A*, Concordia plots for samples BRM–3 and BRM–11 (open squares indicate data points for sample BRM–3; open circles indicate data points for sample BRM–11); *B*, Concordia plot for sample BRM–6; *C*, Concordia plot for sample BRM–10; *D*, Concordia plot for sample BRM–12.

the other samples. The data are consistent with this being a Proterozoic crustal rock in which the feldspar Pb was equilibrated with the whole rock at the time of batholith emplacement during the Cretaceous. The isotopic resetting of Pb in feldspars is a common phenomenon and occurs at relatively low temperatures, as demonstrated in the study of the Eldora stock in Colorado by Doe and Hart (1963). For Proterozoic augen gneiss BRM–5, the feldspar Pb was also probably reset at the time of batholith intrusion. Although BRM–5 may well have been derived from the same crust (approximately 1,780 Ma) as the younger plutons, the feldspar Pb isotopic data indicate a different history: perhaps there was an additional U-Pb fractionation at about 1,500 Ma when the augen gneiss was metamorphosed. From the zircon data, we have postulated some contamination of BRM–12 by BRM–5; the similar Pb isotopic compositions in both graphs of figure 4 may be evidence for this.

If closed system conditions are assumed for Pb evolution in BRM–8 between 1,780 and approximately 100 Ma, then the $^{238}U/^{204}Pb$ (μ) value was 15.8 (average crustal $\mu \cong 9.7$). During the same period, for evolution of Pb in the source rock of the least radiogenic sample (BRM–10), μ is computed to be 8.4—this is consistent with Pb evolution at a deep level in the Proterozoic crust (Zartman, 1965). Intermediate μ values were computed for the other samples, and all are shown in table 3. If the linear trend evident in the $^{207}Pb/^{204}Pb$ versus $^{206}Pb/^{204}Pb$ plot of figure 4 is given an age interpretation and if closed system evolution from a common isotopic source is assumed, then for a common crystallization age of 60 Ma, the average age of the source

Table 3. Feldspar Pb isotope data, arranged by descending $^{206}Pb/^{204}Pb$ values

[$^{238}U/^{204}Pb$ (μ) values are values in source rocks that are calculated assuming a closed system between 1,760 Ma and time of crystallization]

Sample no.	$^{206}Pb/^{204}Pb$	$^{207}Pb/^{204}Pb$	$^{208}Pb/^{204}Pb$	Zircon age (Ma)	$^{238}U/^{204}Pb$
BRM-8	20.552	15.808	41.100	≅1,750	15.8
BRM-9	19.330	15.677	39.042	94	11.9
BRM-5	19.307	15.637	39.051	≅1,500	11.8
BRM-12	19.247	15.644	38.976	55	11.6
BRM-7	19.063	15.651	38.984	71	11
BRM-6	18.805	15.629	38.667	54	10.2
BRM-11	18.478	15.597	38.343	59	9.1
BRM-13	18.294	15.574	38.144	75	8.5
BRM-10	18.270	15.562	38.335	47	8.4

is computed to be 1,650±120 Ma. This encompasses the entire age spectrum of basement rocks that we have found in this part of the batholith. The trend in figure 4 is consistent with derivation of Phanerozoic rocks by remobilization of the Proterozoic basement rocks that they intrude.

Because of their relatively low values of $^{207}Pb/^{204}Pb$ with respect to $^{206}Pb/^{204}Pb$, the least radiogenic data plot just below the average growth curve of Stacey and Kramers (1975) for continental crust (fig. 4). Thus, the Early Proterozoic crust from which the batholithic rocks were derived was formed from comparatively unevolved material, probably in an island arc setting. Because Archean upper crustal rocks of the Wyoming craton have very high $^{207}Pb/^{204}Pb$ ratios (Wooden and Mueller, 1988; Stacey and Hedlund, 1983), little contribution to this Early Proterozoic crust could have been made from evolved continental crust of significantly older age—otherwise the data would plot above the growth curve in figure 4*B*. Our data, therefore, indicate that, during formation of the Proterozoic source rocks of this part of the lobe, there was little recycling of material from the Archean craton that is presently located no more than 100 km to the southeast. Shuster and Bickford (1985) presented some Pb isotope data from batholithic and Proterozoic host rocks in the vicinity of the northeastern boundary of the Bitterroot lobe. Feldspar data from granites show significant variation in $^{207}Pb/^{204}Pb$ (Shuster and Bickford, 1985), and several points form a parallel trend that plots above our data. This situation indicates that the continental crust along the eastern edge of the lobe incorporated a component of more radiogenic, and perhaps older, crustal material when it was formed 1,700–1,800 Ma.

Among the regions bordering the Wyoming Archean province for which Pb isotope data are available, data for the Bitterroot lobe are most similar to those from the rocks and ores of southwestern Colorado, southern New Mexico, and Arizona (Doe and others, 1979; Stacey and Hedlund, 1983; Wooden and others, 1988). Galena data from Precambrian and Mesozoic-Cenozoic ore deposits in the Belt Supergroup, to the north of the Idaho batholith, show similarly unevolved Pb isotope characteristics for underlying basement rocks (Zartman and Stacey, 1971). These observations are evidence that, during the period 1,800–1,600 Ma, considerable additions of new crust were made by accretion of extensive island arc systems to the North American continent (Wooden and others, 1988; Nelson and DePaolo, 1985; Stacey and Hedlund, 1983).

DISCUSSION

The oldest rocks of the Bitterroot lobe are tonalite and quartz diorite plutons that were intruded along the western and northeastern margins of the lobe. A tonalite on the northwestern periphery of the batholith is dated at 94 Ma in this study, and four samples from the northeastern border zone of the batholith have been dated by other workers at: 73 Ma, using U-Pb methods (Bickford and others, 1981); 73 Ma, using U-Pb methods (Chase and others, 1978); 81 Ma, using fission-track methods (Ferguson, 1975); and 87 Ma, using U-Pb methods (Tripp, 1976). Ages of the tonalite and quartz diorite plutons thus span the period from 94 to 73 Ma; younger phases were emplaced near the eastern edge of the lobe.

Very few data exist concerning sources for the tonalitic and quartz dioritic rocks in this region. However, zircon U-Pb and feldspar Pb isotopic data indicate that the tonalite and quartz diorite magmas ingested some Proterozoic crustal material during emplacement (Toth, 1987).

Two granodiorite and monzogranite plutons from this study were intruded during the period from 75 to 71 Ma along the northern edge of the lobe, and we have dated four plutons in the central and western parts of the lobe between 59 and 54 Ma. In the eastern part of the lobe, Shuster and

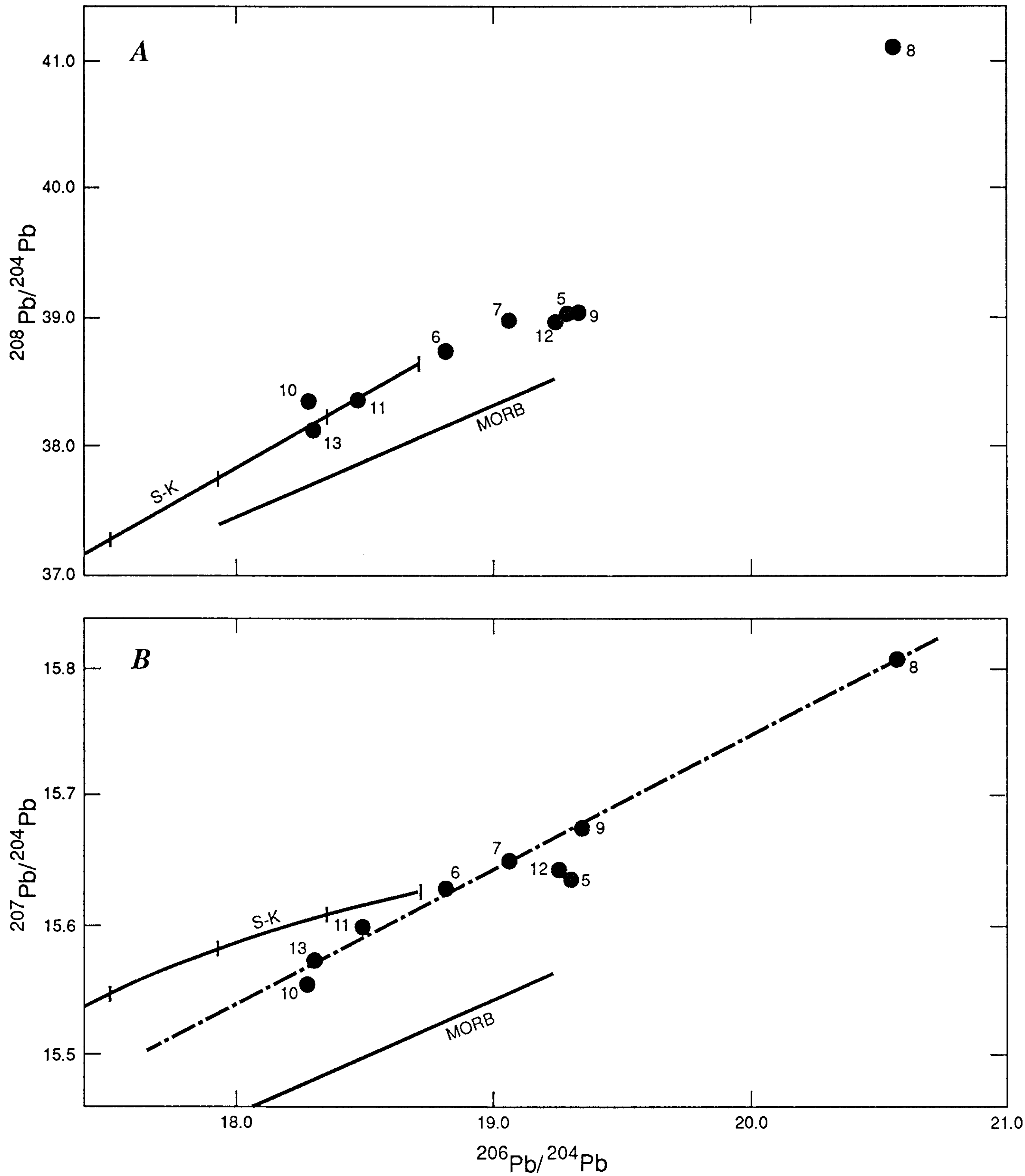

Figure 4. Pb isotope data for feldspars from most of the samples in this study. Data are labeled with BRM sample numbers (from table 1). Model growth curves (labeled "S-K") are for average crust (Stacey and Kramers, 1975). MORB trend lines are from Tatsumoto (1978). *A*, Plot of $^{208}Pb/^{204}Pb$ versus $^{206}Pb/^{204}Pb$; *B*, Plot of $^{207}Pb/^{204}Pb$ versus $^{206}Pb/^{204}Pb$.

Bickford (1985) give a crystallization age for one sample of 60 Ma; Bickford and others (1981) cite ages of 55 to 51 Ma for crystallization of the nonfoliated granites; Chase and others (1978) cite 66±10 Ma as the age of emplacement of the granites in their study; Nelson (1981) gives crystallization ages of 56 to 46 Ma; and Tripp (1976) reports a crystallization age of 67 Ma. In summary, U-Pb dating by many workers indicates that the granitic rocks of the

Bitterroot lobe were first emplaced along the northern edge of the lobe at 72 Ma; the volumetrically more important granitic rocks in the central and western parts of the lobe were emplaced between 60 and 51 Ma. In addition, Chase and others (1983), Arruda (1981), and Garmezy and Sutter (1983) indicate that, during the period from 48 to 42 Ma, mylonitization that was approximately contemporaneous with emplacement of epizonal plutons in the Bitterroot lobe occurred along the eastern border of the lobe.

The presence of older, xenocrystic zircons in all of the studied plutonic units indicates that the magmas were derived, at least in part, by partial melting and remobilization of older continental crust. Fleck (1990), Criss and Fleck (1987), Fleck and Criss (1985b), and Shuster and Bickford (1985) demonstrate from their Rb-Sr, Pb-Pb, and rare-earth-element (REE) data that the main source for granitic rocks in the lobe is not the exposed country rock, but is instead depleted lower crust. However, in the western and central parts of the lobe that we have studied, the data indicate that Pb is more radiogenic than in areas along the eastern border of the lobe investigated by Shuster and Bickford (1985). It is also possible that we have identified one of the Early Proterozoic (approximately 1,750 Ma) source rocks on the western edge of the batholith.

Upper-intercept ages for the Cretaceous rocks in table 2 are consistent with a common age of 1,780±20 Ma for the Proterozoic basement in this northern part of the Bitterroot lobe. Upper-intercept ages for the Tertiary plutons are more variable, but most lie in the range from 1,812 to 1,720 Ma. Assimilation of small components of country rock during emplacement may have further reduced some upper-intercept model ages (as discussed for BRM–12). Our results are similar to those of Bickford and others (1981), who found a range of 1,780 to 1,720 Ma in upper-intercept model ages for most of their zircon samples. However, the presence of older material along the eastern edge of the lobe is indicated by a few upper-intercept-model ages in the range of 2,340 to 2,000 Ma. The presence of older material is also evident in some of the Pb isotope data from the same region (Shuster and Bickford, 1985).

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