URANIUM-LEAD ISOTOPIC SYSTEMATICS IN THE NORTHEASTERN BORDER ZONE OF THE IDAHO BATHOLITH, BITTERROOT RANGE, MONTANA

by

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B. A., Carleton College, 1977

Submitted to the Department of Geology and the Faculty of the Graduate School of the University of Kansas in partial fulfillment of the requirements for the degree of Master of Science.

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ABSTRACT

The Idaho batholith is a composite calc-alkaline plutonic complex that formed as part of a continuous belt of intrusions that extended along the western margin of Mesozoic North America. The limited available data suggest the batholith is mainly of Cretaceous age with later Eocene additions. Further, there is isotopic evidence which suggests that the batholith magmas originated from or were significantly contaminated with Precambrian crustal material. Through two experiments, this study confirms this hypothesis, places age constraints on the old source, and clarifies some of the intrusive age relationships of the northeastern border zone of the batholith.

In the first experiment U-Pb (zircon) isotope systematics of a suite of inclusions from the batholith were analyzed. The suite, which includes xenoliths ranging from little altered to strongly digested, defined a mixing line between young and old material with ages of 46 +/- 2 and 1636 +/- 43 million years, respectively. Many zircons from granite of the batholith contain old xenocrystic cores, so a second experiment was designed to analyze the old cores and young rims separately. Whereas the stepwise dissolution procedure employed here did not completely accomplish this, the U-Pb data did define a discordia with

a lower and upper intercept age of 43 +/- 2 and 1523 +/- 90 million years.

Although it is questionable that ages obtained from the stepwise dissolution experiment have geological significance, they agree within experimental error with the results of the first experiment. Age data from Precambrian terranes near the Idaho batholith indicate that it is reasonable to postulate the existence of a 1640 million year old source terrane for the old zircon component. Data from this study also indicate the occurrence of previously unrecognized Tertiary intrusions in the northeastern part of the Idaho batholith.

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INTRODUCTION

In studying the evolution of the continental crust, the genesis and emplacement of major batholiths is a significant and perhaps dominant process to consider. During the last decade geochemists have gained considerable insight into the problems of the origin and formational history of granitic batholiths from the study of Rb-Sr, Sm-Nd and U-Pb isotopic systematics. In addition to indicating ages of events in the history of a batholith, isotopic variations also reflect characteristics of the source region and of the country rock assimilated by ascending magmas.

The Idaho batholith represents a major addition of continental crust to the North American plate. Thus, detailed petrogenetic study of the Idaho batholith will lend insights to the more general problem of crustal evolution. Utilizing U-Pb isotope systematics in zircons from rocks of the northeastern border zone of the batholith, I have addressed several problems: 1) was there contribution of old crustal material to the batholithic magmas?; 2) if there was, what is the age of this old component and what constraints does this place on possible source regions?; and 3) what is the crystallization history of some rocks of the northeastern part of the batholith?

The northeastern border zone of the Idaho batholith was chosen for this study because it has already been mapped in detail (Chase, 1973; Chase and Talbot, 1973). Little U-Pb isotopic work has been done previously in the area. The metamorphic and deformational history of the country rock intruded by this part of the batholith is also understood in some detail (Chase 1973, 1977; Nold, 1974; Cheney, 1975; Wehrenberg, 1972). Finally, considerable topographic relief and nearly continuous exposure allow for adequate sampling and field observation.

PREVIOUS WORK

Tectonic Setting and General Features:

The Idaho batholith is a composite group of calc-alkaline plutons that crops out through central Idaho and extends into western-most Montana (Fig. 1). It can be subdivided into two lobes on the basis of outcrop pattern: the Atlanta lobe in the south and the Bitterroot lobe in the north. This study focuses on the northeastern border zone of the Bitterroot lobe (Fig. 2).

The Idaho batholith formed as part of a long, perhaps originally continuous, belt of calc-alkaline intrusions. It was situated above and genetically related to an east dipping slab of subducting lithosphere. Hamilton (1969, 1978) places the batholith within a great volcano-plutonic belt that extended along the entire western margin of Mesozoic North America, roughly 200 to 250 km inland from the top of the continental slope. The batholith was later moved to its present location during the middle and late Tertiary by tectonic rotation and wrench faulting (Livaccuri, 1979). Thus, it represents a southern continuation of the Shuswap-Okanagan complex of northeastern Washington and British Columbia, and a northern continuation of the Sierra Nevada.

Figure 1.

Tectonic setting of the Idaho batholith. Sapphire block indicated by stippled pattern. Dash patterns indicate regions dominated by plutons. Zone of cataclasis is along western edge of the Sapphire block.

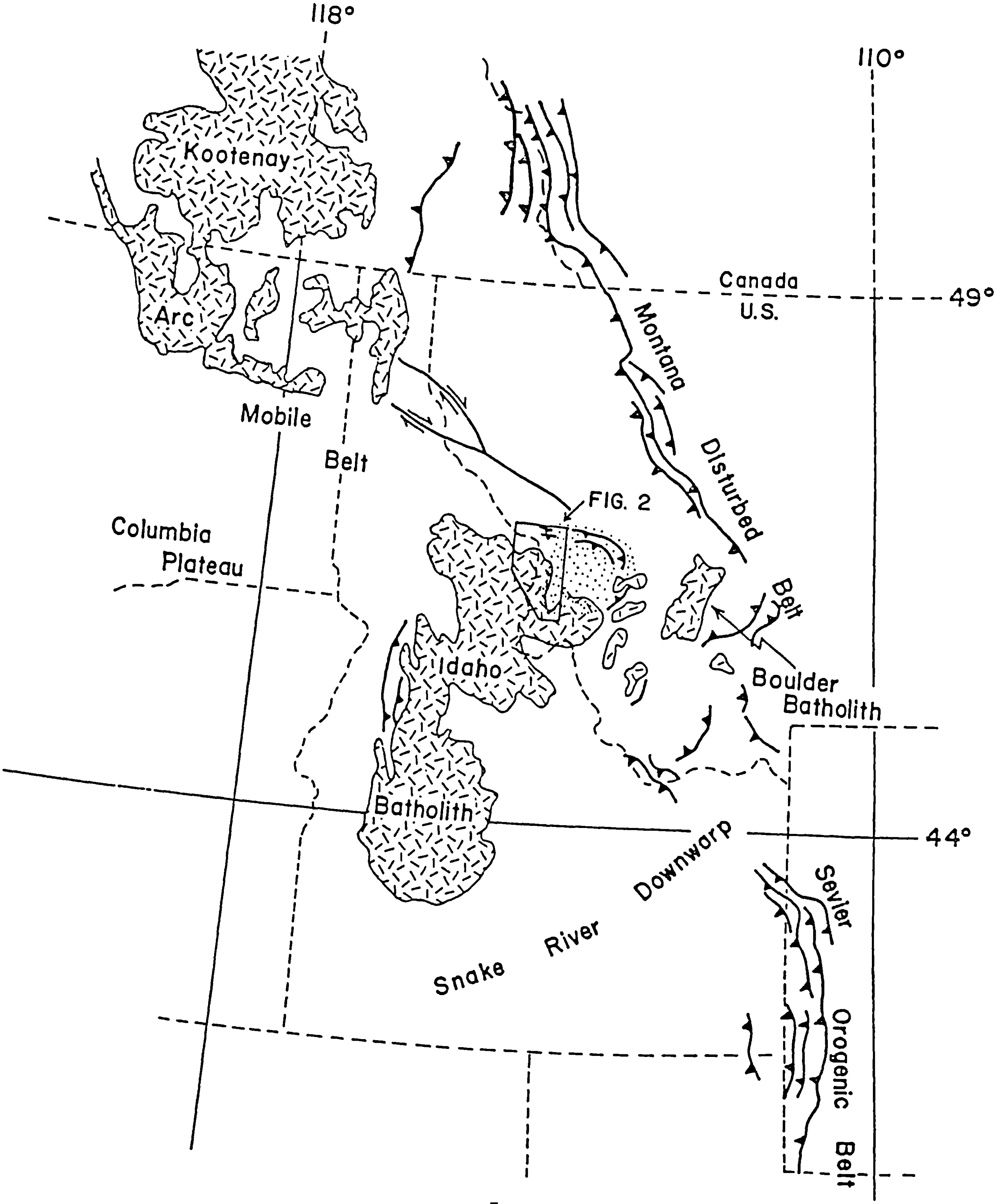
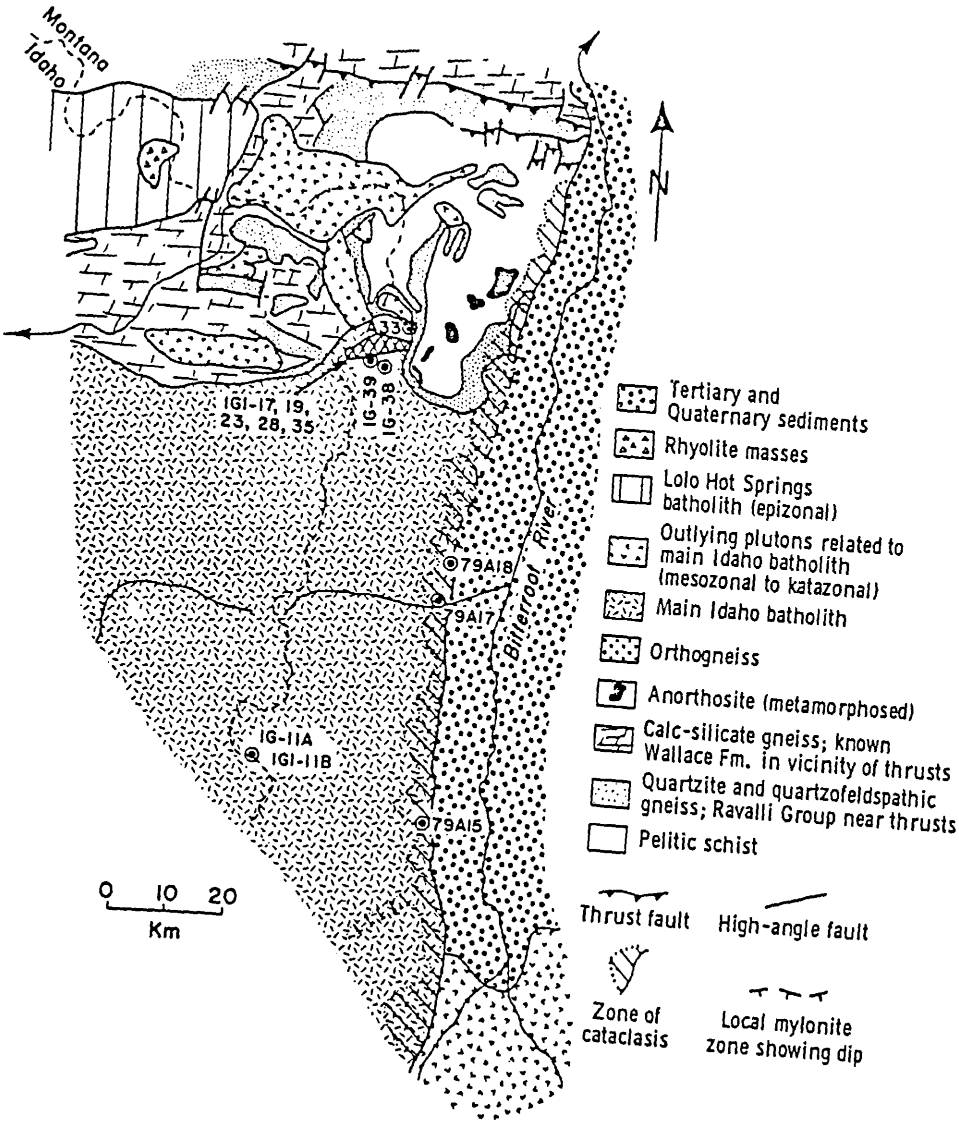

118°
110°
Kootenay
Arc
Mobile
Belt
Canada
U.S.
49°
Montana
Disturbed
FIG. 2
Columbia
Plateau
Belt
Idaho
Boulder
Batholith
Batholith
44°
Downwarp
River
Snake
Sevier
Orogenic
Belt

Figure 2.

Geologic map of the northeastern border zone of the Idaho batholith (enlargement of area outlined in Fig. 1) showing location of samples studied. Map is adapted from Chase and others (1978).



Montana
Idaho
IGI-17, 19,
23, 28, 35
IG-39
IG-38
33
79A18
79A17
IG-11A
IGI-11B
79A15
Bitterroot River
N
0 10 20
Km
Tertiary and Quaternary sediments
Rhyolite masses
Lolo Hot Springs batholith (epizonal)
Outlying plutons related to main Idaho batholith (mesozonal to katazonal)
Main Idaho batholith
Orthogneiss
Anorthosite (metamorphosed)
Calc-silicate gneiss; known Wallace Fm. in vicinity of thrusts
Quartzite and quartzofeldspathic gneiss; Ravalli Group near thrusts
Pelitic schist
Thrust fault
High-angle fault
Zone of cataclasis
Local mylonite zone showing dip

Detailed mapping of the northeastern border zone (Chase, 1973; Chase and Talbot, 1973) has shown that the granitic-metamorphic complex of the deepest level of the batholith is exposed in this area. The exposures are of granitic rocks from the basal and root zone of the plutons and subjacent high grade metamorphic country rocks originally formed as deep as 3000 meters below the projected original floor of the batholith. Analogous katazonal environments within the Cordilleran core are exposed in the Omineca complex of British Columbia and areas of the Peninsular Ranges batholith in southern California and Baja California.

Early workers (e.g. Anderson, 1942) realized that the Idaho batholith is a complex multiple intrusive. Hyndman and Williams (1976, 1977) recognized two major granitic units within the Bitterroot lobe. The greatest proportion of the intrusion consists of a granodiorite-granite suite, which they estimate is of late Cretaceous age. The remaining 20 percent of the Bitterroot lobe is a younger granite suite associated with an Eocene event. Miarolitic cavities, intergradations with porphyritic volcanics, and aphanitic texture are characteristic epizonal features of the Tertiary plutons. This contrasts with the mesozonal or katazonal features typical of the older suite. Other Tertiary plutons distinct from, but in the vicinity of, the

Idaho batholith also show similar characteristic epizonal petrographic features (Nold, 1974).

Geologic Setting of the Northeastern Idaho Batholith:

Igneous rocks in the northeastern sector of the batholith (Fig. 2) have intruded quartzofeldspathic gneiss, pelitic schist, and calc-silicate gneiss which have been subjected to conditions of upper amphibolite-grade metamorphism and multiphase deformation, and remetamorphosed under lower-pressure conditions (Chase, 1973, 1977; Nold, 1974; Cheney, 1975; Wehrenberg, 1972). The host rocks were penetratively deformed at least twice before batholithic intrusion.

The main portion of the batholith contains many plutons whose contact relations are not known in detail. In the northeastern border zone the earliest intrusive phase was quartz diorite on the periphery of the igneous complex, which was deformed and metamorphosed along with the host rocks prior to more voluminous intrusion of the main plutonic complex (Chase, 1973). There are two principle non-foliated granitic types, an earlier medium-grained, gray granite and a later porphyritic, gray granite distinguished by large potassium feldspar megacrysts. The emplacement sequence of these three plutonic types can be determined in the field by crosscutting and xenolith distribution relationships.

The contact between batholith rocks and their metamorphic sheath is characterized by a wide zone of granitic sills and isolated blocks of country rock. Xenoliths are very abundant near the contact and decrease southwestward into the batholith over a zone that is about 2 km wide. Inclusions of quartzofeldspathic gneiss, pelitic schist, and calc-silicate gneiss are readily identified as fragments of the invaded country rocks. Other inclusions that are less abundant consist of biotite-rich schist and amphibolite. Amphibolite occurs in the country rocks as isolated boudins, but biotite schist is not common and may be present as exotic blocks from a greater depth than is represented by present exposures. Xenoliths of quartzofeldspathic gneiss and pelitic schist can be observed in various stages of digestion within igneous rocks, the extent of digestion commonly increasing into the batholith until only wisps of aligned mafic minerals and faint hints of lithologic layering indicate the position of an inclusion. Calc-silicate gneiss is evidently far less reactive, for such xenoliths typically show sharply angular borders and little evidence of reaction with magmatic rocks.

The northeastern part of the Idaho batholith and its metamorphic envelope has been regionally uplifted to form a gneiss dome (Chase, 1977; Chase and Talbot, 1973). A zone of cataclasis which is up to 850 meters or more thick has developed along the eastern margin of, and locally within, this dome. Isolated parts of the dome have been affected by faulting, with mylonitization, retrograde metamorphism, and intrusion of felsic dikes in the fault zone. The large number of Eocene or younger K-Ar and Rb-Sr (biotite) and fission track (apatite) ages may be related to the thermal (Armstrong and others, 1977), hydrothermal (Criss and Taylor, 1978), or uplift (Ferguson, 1975) events associated with the doming, or some combination of these (Chase and others, 1978). Erosion to the present level has exposed the inner sector of the dome and the lower part of the batholith as well as the zone of cataclasis.

Regional Geologic History:

In the last few years evidence has been found for an extensive pre-Belt, Precambrian terrane in Idaho. Armstrong (1975a) determined a 1500 million year Rb-Sr age for gneissic rocks from the Salmon River Arch which separates the Atlanta lobe from the Bitterroot lobe. He further proposed that other regional exposures of quartzite and argillite, as well as other metasedimentary, gneissic and granitic rocks are also part of a wide-spread Precambrian terrane that formed the basement of the Belt depositional basin. A more detailed description of known pre-Belt events in the area will be presented below along with the discussion of results.

Most of northern Idaho and western Montana is underlain by the Precambrian Belt series, which consists of a thick low-grade metasedimentary sequence of rocks that were originally fine-grained clastic sediments with minor carbonates, and mafic igneous rocks. The basin of deposition was the northern-most of three major reentrants of the sea into the western edge of the North American craton (Harrison, 1972). Deposition of this thick sequence, of which at least 20 km of structurally intact section remains, occurred in the Belt basin from approximately 1500 to 850 million years ago (Harrison and others, 1974). Sedimentation was not continuous but probably was interrupted by several major hiatuses of at least 20 million years (Obradovich and Peterman, 1968).

As a result of the East Kootenay orogeny in southeastern British Columbia about 800 million years ago (Gabrielse, 1972), sedimentary rocks of the Belt basin were uplifted, slightly deformed and exposed to erosion. The only widespread Precambrian sedimentation following this uplift occurred seaward of the Belt basin and is represented by the Windermere formation (and equivalents) which unconformably overlap the western-most Belt sedimentary rocks.

The Paleozoic record is not well preserved, but indicates that much of the Belt basin and surrounding areas was covered by the sea from Middle Cambrian through Middle Jurassic. Harrison and others (1974) concluded that during this time some parts of the Belt basin did not receive any significant amounts of marine sediment, and that overall the basin underwent only minor vertical movement.

Beginning in the Jurassic, tectonism and magmatism occurred discontinuously throughout the area until the Eocene. The Belt sedimentary rocks were uplifted, and sedimentation was restricted to volcaniclastic and nonmarine deposits along the eastern edge of the basin (McMannis, 1965). The Idaho batholith was emplaced south of and into the southern part of the Belt basin. As a result, the Belt rocks are separated from pre-Belt metasedimentary rocks to the south by the batholith. Metasedimentary rocks exposed in the northeastern border zone of the batholith may either be a fragment of pre-Belt metasedimentary rocks that was preserved north of the batholith, or represent Belt sediments metamorphosed in the border zone.

Harrison and others (1974) could not identify any tectonic features or crustal weakness that might have controlled the emplacement of the batholith. Rocks as young as Early Cretaceous on the eastern side and northeastern end have been shouldered aside by emplacement of the batholith. This places at least an upper bound on the age of the pluton.

Finally, after a period of tectonic quiesence in the early Eocene, there was a major mid-Eocene volcanic-plutonic event with associated block-faulting (Armstrong, 1975a). As suggested in the following section, this event had a widespread effect in and around the batholith.

Geochronology of the Idaho Batholith:

The few geochronologic data that are available for the Idaho batholith present an inconclusive and complicated picture. The earliest work utilized the Pb-alpha technique (Chapman and others, 1955; Larsen and others, 1958; Jaffe and others, 1959; Larsen and Schmidt, 1958; Nelson and Ross, 1968). Results that were obtained indicated ages ranging from Eocene to Jurassic, averaging around 100 million years. This was the first work to suggest that the Idaho batholith is a Cretaceous intrusive. The Pb-alpha method, however, is inherently inaccurate because the assumptions one must make to calculate an age are largely invalid (Faure, 1977, p. 197). Furthermore, Armstrong (1974) pointed out that the zircon population in rocks of the Idaho batholith is contaminated by older xenocrysts. This will consistently result in Pb-alpha ages that are greater than the actual emplacement age. Therefore, the Pb-alpha method should be restricted to reconnaissance work.

Most of the dates calculated for the Idaho batholith are K-Ar ages. The first K-Ar ages obtained by Hayden and Wehrenberg (1960), Percious and others (1967), McDowell and Kulp (1969), and McDowell (1971), range from 38 to 156 million years. Later work by Armstrong (1975b) and Williams (1979) confirmed this pattern. Armstrong (1975) noted that all the ages obtained from the Bitterroot lobe are less than 70 million years, whereas maximum ages from within the Atlanta lobe approach 100 million years. This indicates that the Atlanta lobe probably is actually older than the Bitterroot lobe.

A strong episode of magmatism and heat flow, termed the Challis event, occurred from 40 to 49 million years ago throughout the region of the Idaho batholith (Armstrong, 1974). Partial to complete resetting of the K-Ar isotopic systems by this event may explain the apparently anomalously young mineral ages of 38 to 60 million years. Williams (1979) selected two samples from the Bitterroot lobe, apparently far removed from any Tertiary intrusions or tectonic movements, specifically to obtain K-Ar ages of crystallization unaltered by younger events. In spite of this he still obtained Eocene ages of 44 and 45 million years. He interpreted the young ages as resulting from heating effects, and concluded that heating must have been considerably more widespread than the extent of known Tertiary intrusives would indicate. On the basis of oxygen and hydrogen isotopic studies, Taylor (1977, 1978) concluded that these late epizonal intrusives induced numerous overlapping meteoric-hydrothermal convective systems that were pervasive throughout the Atlanta lobe and somewhat less intense in the Bitterroot lobe. He concluded that it was this hydrothermal activity that caused partial to complete resetting of K-Ar isotopic systems.

Alternatively, the young ages may reflect the time of major tectonic uplift. Chase and Talbot (1973), and Hyndman and others (1975) presented evidence for a domal uplift of the northern Bitterroot area. They suggested that batholithic emplacement began as early as 100 million years ago, and subsequent tectonic unloading of an overlying 10 to 15 km-thick block began approximately 80 million years ago. Criss and others (1980) found a positive linear correlation between K-Ar (biotite) ages and elevation in the Atlanta lobe. They concluded this reflects closure of K-Ar systems by cooling during regional Eocene doming.

Attempts to determine the age of the Idaho batholith by Rb-Sr isotopic analysis have produced equally undefinitive answers. Because of variability in initial 87Sr/86Sr ratios and low Rb/Sr ratios (Armstrong, 1975b), the use of the whole rock technique to obtain Rb-Sr ages has been precluded. Essentially all interpretable analyses are from minerals, and most of these are from biotite separates. Similar to K-Ar results, the Rb-Sr ages reflect partial to complete resetting of the isotopic system, again probably caused by the Eocene Challis event. The maximum Rb-Sr ages reported also indicate that the Atlanta lobe is older than the Bitterroot lobe (Armstrong, 1975b). Hofmann and Grauert (1973) attempted to measure the recrystallization age of sedimentary rocks metamorphosed by intrusion of the batholith. They applied the Rb-Sr method to small contiguous slabs cut from different compositional bands of a single metasedimentary rock. There was enough variation in the Rb-Sr ratio among the slabs to make isochron measurements. One isochron yielded an age of 64 million years, which is a reasonable age for the intrusion of the batholith. But they also obtained two other isochrons of 340 and 530 million years which cannot be associated with any known geologic events. Rb-Sr analyses by Chase and others (1978) on rocks from the northeastern border zone did not yield an isochron. Their data did suggest, though, that a quartzofeldspathic gneiss affected by heat related to emplacement of the batholith became a system closed to Rb and Sr approximately 85 million years ago.

Compared to Rb-Sr and K-Ar systems, U-Pb isotopic systems preserved in zircon are not as easily disturbed by metamorphic or thermal events subsequent to crystallization. U-Pb isotopic analysis of zircons from the Idaho batholith could help resolve the uncertainties and ambiguities presented by Rb-Sr and K-Ar data alone. To date there have been only a few preliminary studies. Reid and others (1973) reported two analyses of single bulk zircon populations. Both were too discordant to allow any assignment of a minimum or maximum age to the batholith.

Grauert and Hofmann (1973a, 1973b) studied a cogenetic suite of zircons and obtained a chord on a concordia diagram that has a lower intercept of 53.5 million years and an upper intercept of 1300 to 2100 million years. These data could indicate that the batholith is a Precambrian intrusive which underwent a metamorphic event that caused significant Pb loss from the zircons approximately 54 million years ago. Because field evidence indicates the batholith is Cretaceous and detrital zircons in surrounding metasedimentary rocks yield a Precambrian age, Grauert and Hofmann chose instead to interpret the chord as a two component-mixing line. The lower intercept represents the younger component, presumably the age of zircons crystallized from batholith magmas, and the upper intercept represents the apparent age of an older inherited zircon population, perhaps derived from assimilated country rock. In such a situation one might expect to observe at least two crystal morphologies in the zircon concentrate, one euhedral and perhaps elongate, characteristic of an igneous origin, and the other subhedral and rounded, characteristic of detrital or metamorphic origins. One might also expect that zircons assimilated into the magma would serve as nucleation sites for new zircons and now be present as xenocrystic cores within the crystals. Grauert and Hofmann did not report on either the crystal morphology or on the presence of cores.

Chase and others (1978) analyzed a single bulk zircon population from each of three rocks in the batholith: a quartz diorite orthogneiss, which on the basis of structural evidence is an early phase of the batholith; a granodiorite from the margin; and a quartz monzonite from the interior of the pluton. Assuming the granodiorite and quartz monzonite are cogenetic, the data yielded a two point discordia with a lower intercept of 66 million years and an upper intercept of 2075 million years, in general agreement with Grauert and Hofmann's (1973) data. Chase and others (1978) also interpreted their data as a mixing line. They concluded that the batholith assimilated Precambrian material from its source or from the country rock into which it intruded, and then crystallized approximately 66 million years ago. Furthermore, they cited the observation in thin section of at least one zircon with a rounded core and euhedral overgrowth.

Because of the limited geochronologic data and later disturbance of the isotopic systems, the geochronology of the Idaho batholith, and the Bitterroot lobe in particular, is still not clear. Evidence for a Late Cretaceous age comes mainly from an average of three fission track ages from sphene (Ferguson, 1975), one U-Pb discordia determined by Grauert and Hofmann (1973a, 1973b), and the preliminary age of Chase and others (1978) which is based on the assumption that the granodiorite and quartz monzonite are cogenetic. The K-Ar and Rb-Sr data show wide scatter that is tentatively ascribed to disturbance by widespread Eocene volcanic-plutonic activity and associated hydrothermal-meteoric systems. Though many of the Eocene intrusives have been identified and characterized by their epizonal features, Armstrong (1974) pointed out that there may be many more of these Tertiary intrusives than have been recognized. From modeling heat flow data from the Idaho batholith, Swanberg and Blackwell (1973) also concluded that it is a temporally composite body, and that Eocene igneous activity was probably more extensive than has been recognized.

Initial 87Sr/86Sr Data:

The presence of a significant component of inherited zircon material suggests that the granitic magmas were derived directly from old crust, or formed in the lower crust or mantle and were significantly contaminated with Precambrian crustal material. Initial 87Sr/86Sr ratios also require a similar interpretation.

The Idaho batholith lies east of a boundary delineated by Armstrong and others (1977) that represents a sharp contrast in values for the initial 87Sr/86Sr ratios of the rocks. Oceanward of this line, which extends north-south just west of the Idaho batholith, rocks have initial ratios that are low (less than 0.7043), whereas on the continental side initial ratios are distinctly higher (greater than 0.7055). Kistler and Peterman (1973), and Early and Silver (1973) recognized a similar boundary in the Sierra Nevada batholith and the Southern California batholith. Armstrong and others (1977) postulate that this boundary represents the juxtaposition of Precambrian crust against younger material accreted onto the craton during late Paleozoic to early Mesozoic time. Plutons east of this boundary, such as the Idaho batholith, must be partially derived from Precambrian crustal material. From the Atlanta lobe of the Idaho batholith, Armstrong and others (1977) measured initial 87Sr/86Sr ratios ranging from 0.7072 to 0.7378, with one as high as 0.8159. Chase and others (1978) obtained initial ratios from 0.7103 to 0.7128 for samples from the Bitterroot lobe. Both these high, variable initial Sr isotopic ratios and the inherited radiogenic Pb component are reasonably explained by a crustal source or crustal contamination model.

MODEL FOR THE PETROGENESIS OF THE IDAHO BATHOLITH

The U-Pb isotopic data summarized above from Grauert and Hofmann (1973a, 1973b), and Chase and others (1978) represent only preliminary work but are the only useful U-Pb data that have been published for the Idaho batholith. Combined with initial 87Sr/86Sr ratios (Armstrong and others, 1974), the isotopic data suggest a testable working hypothesis for the origin of the Idaho batholith: the magmas were produced either by anatexis of old crustal material or by the assimilation of significant amounts of this material into mantle-derived magmas. Using two different experimental designs, I completed a detailed study of the U-Pb isotopic systematics preserved in the zircons in order to test this hypothesis.

One interpretation of available data is that within the granite there is an inherited zircon population of a poorly defined Precambrian age. Unfortunately there does not appear to be any simple characteristic such as crystal size, morphology, magnetic susceptibility, or mineral association that would allow one to discriminate between the old and young populations (M.E. Bickford, pers. com., 1979). But Chase and others (1978) observed in thin section that some zircons contain an anhedral core with a euhedral overgrowth, suggesting that the old zircons served as nucleation sites for crystallization of overgrowths and are preserved as the cores of the zircons. As a first experiment, then, I developed a technique for sequential dissolution of zircon crystals in order chemically to strip the young rims from the old cores and analyze more directly both the crystallization age of the batholith preserved in the rim and the age of the inherited zircon preserved in the core. Another motivation for this experiment was to develop a controlled stepwise dissolution technique that allows one to investigate distribution patterns of U and Pb isotopes within zircon crystals. Limited success was obtained by Silver and Deutsch (1963) who performed a two step dissolution experiment. But, for reasons discussed in a following section, consecutive stripping of discrete layers from zircon crystals does not occur. There is no simple relationship between U and Pb distribution in a crystal and removal of these atoms during HF stepwise dissolution.

Evidence of inherited Precambrian material is also preserved within xenoliths of metasedimentary rocks and schlieren that are abundant in some areas of the batholith. Studies on other granitic bodies have suggested that such inclusions may be samples of the source region which have been plucked out and entrained in the magma as it was produced (White and Chappell, 1977), or may be country rock that has been incorporated during ascent and emplacement. For my second experiment I obtained U and Pb ratios on representative zircon populations from a suite of xenoliths that range from those only slightly altered and easily identifiable as surrounding country rock, to xenoliths which are strongly digested schlieren whose original mineralogy cannot be identified. This suite of rocks defines a series which represents progressive degrees of mixing of the old (Precambrian) and young (Tertiary) components. The sample closest to the young component end member of the mixture is a granite sample free of observable inclusions. Samples of unaltered xenolithic material represent the closest approach to the old component end member of the series. The rest of the suite consists of inclusions that have undergone various degrees of reaction with the granitic magma. These samples represent mixtures of various proportions of the two end members. U-Pb data from these endpoints and samples that represent their intermediate mixtures must necessarily form a linear mixing curve when plotted on a graph of 206Pb/238U against 207Pb/235U. All xenoliths analyzed in this study fall on the same linear mixing curve and thus must all be from a common source or different but coeval sources. Those samples with a greater proportion of inherited (xenolithic) material plot closer to the upper intercept of the mixing line with the concordia curve.

Furthermore, the data suggest the inherited zircon component in apparently uncontaminated granite has the same or a coeval source as the xenoliths. U-Pb data from sized and magnetically split zircon populations obtained from the granite define a chord that is colinear with the mixing line obtained from analysis of the xenolith suite. Because of the greater abundance of newly crystallized zircon in the uncontaminated granite, these splits plot near the lower intercept of the mixing line with concordia.

The data from this research 1) show that all inclusions analyzed in this suite are of a similar apparent age, 2) establish a range of values for this age, 3) confirm the presence of an inherited zircon population in the granite, 4) place constraints on the age and, therefore, the source of the inherited material, and finally, 5) remove some of the ambiguities discussed earlier surrounding the primary crystallization age of this part of the Idaho batholith.

ISOTOPIC STUDY OF U-PB SYSTEMS IN ZIRCONS FROM INCLUSIONS AND GRANITE

Inclusion suite:

The suite of inclusions consists of five samples of xenoliths plus a sample of apparently uncontaminated granite. The samples are arranged in Table 1 from the least to the most extensively assimilated xenoliths. Petrographic descriptions of the samples are given in Table 2. Zircon populations from three of the samples, IGI-28, IGI-17 and IG-38, were split into two size fractions, -100 +200 and -200 +400 mesh. This provided more samples with a wider spread of isotopic ratios and helped define the mixing line more accurately. Only a bulk zircon population of the other three samples (IGI-35, IGI-23, IGI-19) was analyzed. Additionally, two aliquots were taken from four of the samples and both were analyzed to document reproducability of the results. These second aliquots are samples IGI-35B, IGI-28B (-100 +200), IGI-19B, and IG-38B (-100 +200).

Each of the zircon separates was observed in high refractive index (n = 1.94) oil. In this medium cores, inclusions, and other features are clearly visible. Sample IG-38, the xenolith free granite, showed that zircons with

TABLE 1

SUMMARY OF ROCK TYPES IN INCLUSION SUITE

IGI-35	Only slightly digested inclusion of quartzofeldspathic gneiss.
IGI-28	Fresh, only slightly digested inclusion of quartzofeldspathic gneiss.
IGI-23	Sample of partly digested inclusion of quartzofeldspathic gneiss.
IGI-19	Quartzofeldspathic gneiss inclusion, more reacted with magma than IGI-23.
IGI-17	Substantially digested inclusion, has undergone considerable reaction with the magma.
IG-38	Granite without observable inclusions.

TABLE 2

PETROGRAPHY OF INCLUSION SUITE ROCK SAMPLES

IGI-35	Texture: allotriomorphic-granular, strained quartz grains with sutured boundaries, zoned feldspar, schistose with alternating biotite and quartz-rich layers. Mineralogy: quartz, 60%; biotite, 30%; K-feldspar, 7%; plagioclase, 3%; opaques, apatite, zircon, muscovite, chlorite, and sphene in trace quantities.
IGI-28	Essentially same texture and mineralogy as IGI-35, though with slightly more feldspar.
IGI-23	Texture: hypidiomorphic-granular, strained quartz with sutured boundaries, zoned feldspar, parts of rock unreacted with granitic magma essentially same as IGI-35. Mineralogy: plagioclase, 50%; quartz 25%; K-feldspar 10%; biotite, 10%; opaques, 5%; apatite, zircon, chlorite, sphene and muscovite in trace quantities.
IGI-19	Texture: hypidiomorphic-granular, complexly zoned feldspar, secondary alteration of biotite to chlorite and sericitization of plagioclase, strained quartz. Mineralogy: quartz, 45%; plagioclase, 35%; K-feldspar, 10%; biotite, 10%; chlorite, rutile, apatite, sphene, zircon and opaques in trace quantities.
IGI-17	Texture: hypidiomorphic-granular, secondary alteration of biotite and plagioclase, rare myrmekite, sutured quartz grains, few zones of mylonitization. Mineralogy: plagioclase, 40%; quartz, 28%; K-feldspar (perthite) 24%; biotite, 8%; apatite, sphene, zircon, opaques, chlorite and rutile in trace amounts.
IG-38	Texture: hypidiomorphic-granular, some sutured quartz grains, myrmekitic, complexly zoned feldspars with minor secondary alteration. Mineralogy: plagioclase, 35%; K-feldspar, 30%; quartz, 30%; biotite, 5%; apatite, monazite, allanite, chlorite, sphene, zircon and opaques in trace amounts.

cores are prevalent. Forty to fifty percent of the crystals in the coarse (-100 +200) fraction have cores, whereas 25 to 35 percent in the fine (-200 +400) fraction have cores. Both anhedral and euhedral cores are present, with the anhedral variety predominating. Unidentifiable inclusions within the grains are abundant and commonly more concentrated within the core than rim. Most of the inclusions are rod-shaped crystals; the remainder are opaque substances. The zircon crystals are predominantly euhedral or have broken terminations.

Sample IGI-17 is an extensively digested xenolith of quartzofeldspathic gneiss. Zircon crystals are similar to those from IG-38, though the percentage with cores is smaller; 10 to 20 percent of the fine fraction and 30 to 45 percent of the coarse fraction have cores. In addition, both the rims and cores are commonly zoned. Observation of Becke lines indicates that the index of refraction of successive zones increases toward the edge of the crystal, suggesting a chemical zonation. Inclusions are abundant within grains, and in a few crystals the cores are choked with inclusions, whereas the rims are relatively inclusion free.

Samples IGI-19 and IGI-23 represent partly digested inclusions of quartzofeldspathic gneiss. In each sample there are two types of zircon: a rounded, nearly spherical population and a euhedral elongate population. The rounded zircons generally constitute 70 to 90 percent of the sample. They do not contain cores, and typically contain abundant opaque inclusions. These are probably the zircons that were originally in the xenolith and have not extensively interacted with the granitic magma. The euhedral population, which is inferred to be the young zircons crystallized from the granitic magma, is elongate (length to width ratios of 1:5 to 1:8) and uniformly shows rounded, inclusion filled cores with clear overgrowths. Long, rod-shaped crystals and equant bubbles are also present as inclusions within grains.

Samples IGI-28 and IGI-35 are very slightly digested xenoliths of quartzofeldspathic gneiss. IGI-35 and the granite sample, IG-38, are the two end members of the xenolith suite. Zircons from IGI-35 range from subhedral to nearly equant. Crystals with cores are rare. Euhedral zoning is present but also rare. Inclusions within crystals consist of: 1) small round bubbles, 2) rod-shaped crystals, and 3) black 'dusty' opaques. Zircons from IGI-28 are somewhat more elongate and euhedral than those from IGI-35. The only other difference is that approximately one-half to two-thirds of the zircons from IGI-28 have cores. The cores are predominantly anhedral, equant, and embayed.

Analytical Procedures:

Zircon samples were separated and concentrated using standard heavy liquid and magnetic separation procedures. Isotopic measurements were performed by standard techniques described by Bickford and others (1969) and Bickford and Mose (1975). Isotopic analysis was obtained on an automated, nine-inch, single filament mass spectrometer with on-line data acquisition. For calculations, I used the natural decay constants recommended by Steiger and Jager (1977): 238U = 1.55125 x 10(-10)/yr; 235U = 9.8485 x 10(-10)/yr; 238U/235U (atomic ratio) = 137.88. Laboratory blanks measured during the experiments indicated 2 nanograms of total Pb contamination, requiring insignificant corrections in the final results. The NBS radiogenic Pb standard was measured four times during the period of this study. The data, presented in Appendix I, show excellent precision and accuracy; inter-run variations are not a consideration for interpretation of the data.

The isotopic data must be corrected for nonradiogenic Pb incorporated into the zircon at the time of crystallization. For these samples there are two apparent episodes of zircon crystallization; one at about 1636 million years and one at 46 million years. Therefore, it was necessary to consider the effect that using older or younger common Pb isotope ratios has on the corrections to the data. For all results in this study calculations show that the choice of 1636 or 46 million year old common Pb, or some mixture of the two, results only in variations in the fourth significant figure of the final value for the radiometric ages. This is well beyond experimental uncertainties.

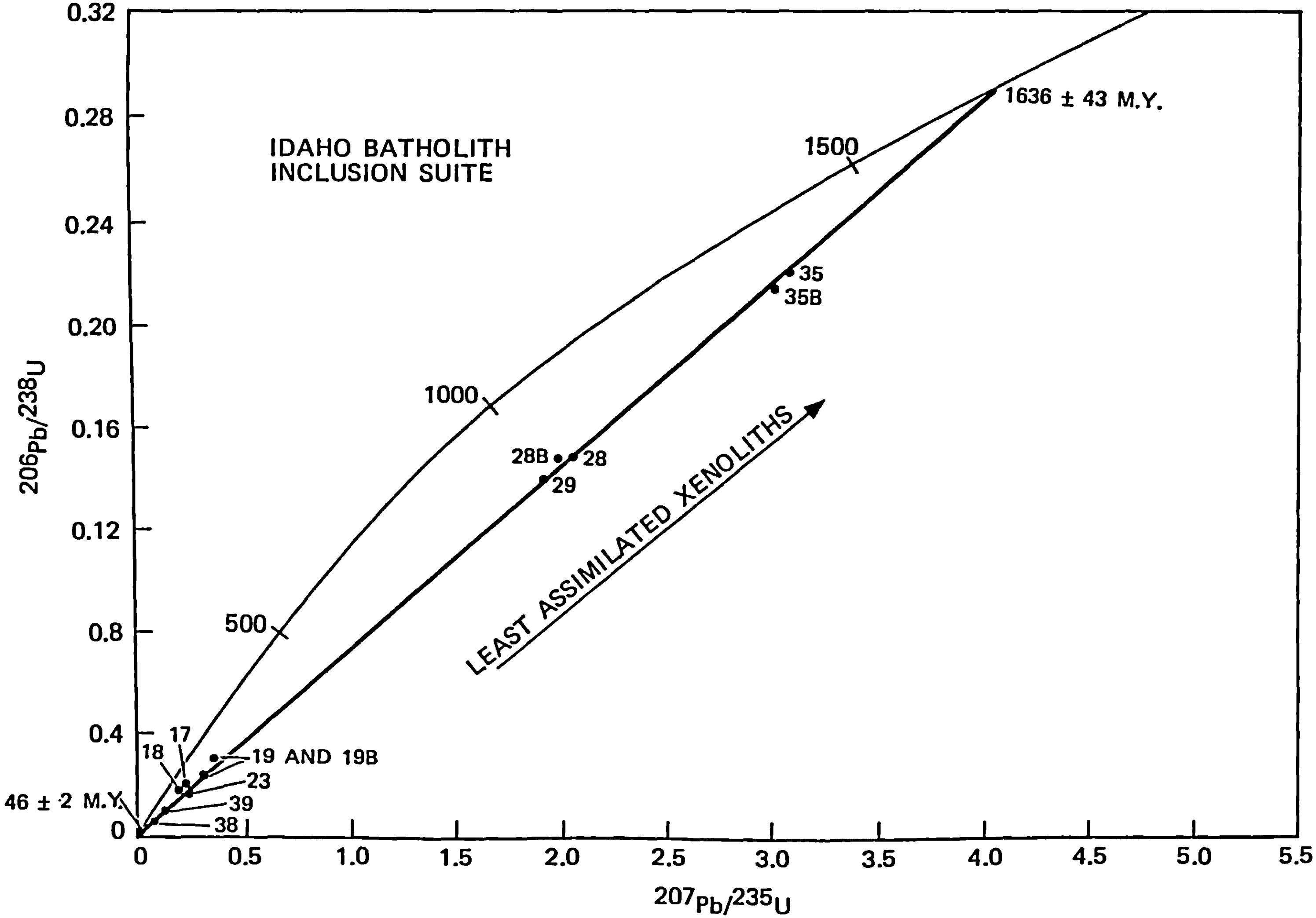
The analytical results for the suite of inclusions are listed in Appendix II and plotted on a concordia diagram in Fig. 3. When these data are fit to a line using York's (1966) method, they define a line with a slope equal to 0.0711 +/- 0.0017, which represents an uncertainty of 2.4 percent. The line intercepts the concordia curve at model ages of 45.9 +/- 2.2 million years and at 1636 +/- 43 million years.

Stepwise Dissolution:

The only published studies that had any bearing on stepwise dissolution techniques for zircons were those similar to Krogh and Davis' (1974) study, which only involved a cold HF acid leach of zircons, and Silver and Deutsch's (1963) attempt at stepwise dissolution in a

Figure 3.

Concordia plot of data from inclusion suite zircons.


IDAHO BATHOLITH
INCLUSION SUITE
0.32
0.28
0.24
0.20
0.16
0.12
0.8
0.4
0
0.5
1.0
1.5
2.0
2.5
3.0
3.5
4.0
4.5
5.0
5.5
206Pb/238U
207Pb/235U
1636 ± 43 M.Y.
1500
1000
500
35
35B
28B
28
29
LEAST ASSIMILATED XENOLITHS
17
18
19 AND 19B
23
39
38
46 ± 2 M.Y.

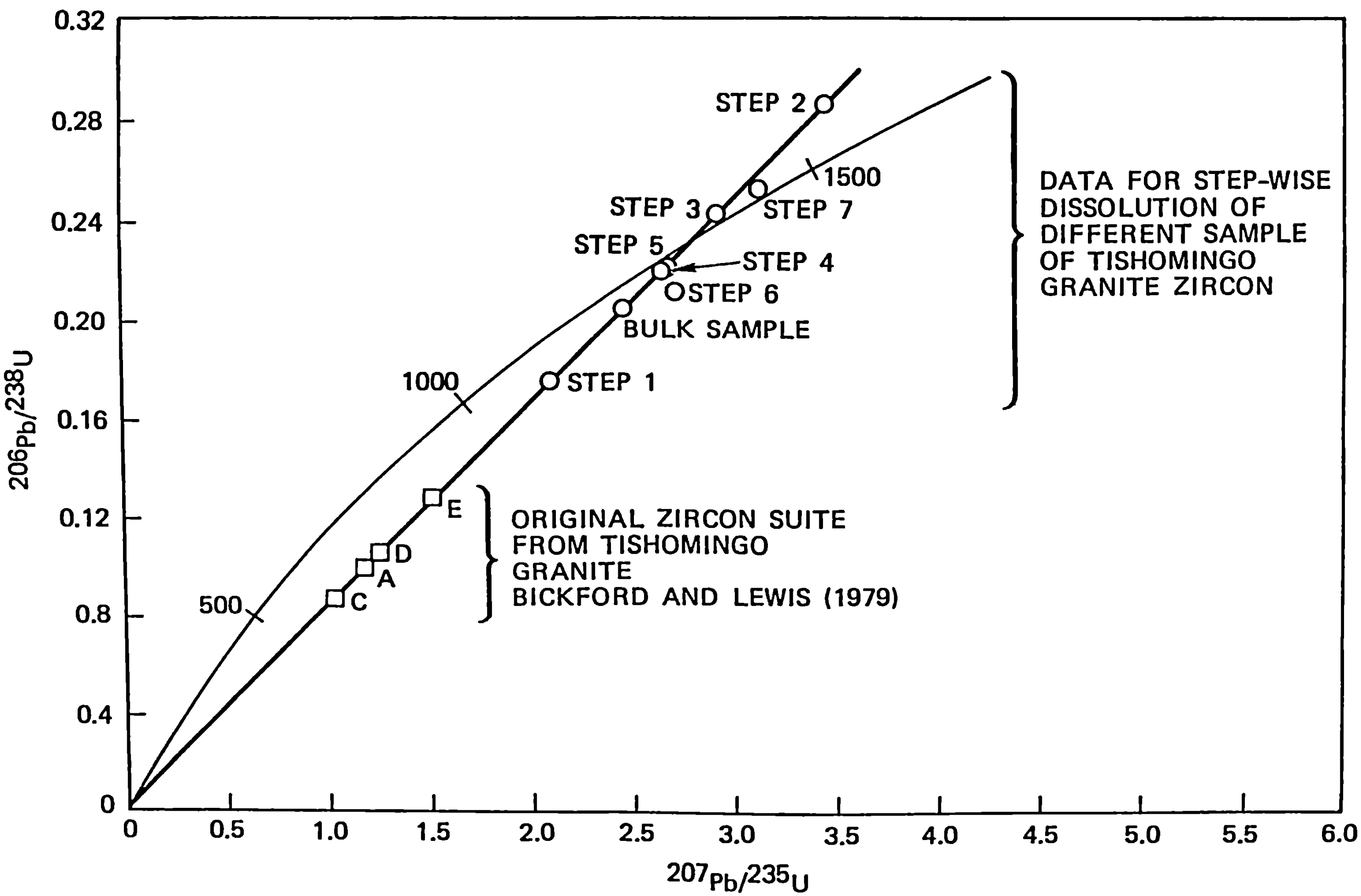
sodium tetraborate flux. Because of the difficulty they encountered controlling the rate of dissolution, and complications in the sodium tetraborate procedure, I developed a technique using sequential dissolution in HF acid at high temperatures. The procedure is outlined in Appendix III.

In order to test the technique, I first applied it to a population of zircons from the 1374 million year old Tishimingo granite of Oklahoma (Bickford and Lewis, 1979). These zircons do not have cores, and are optically unzoned. The zircons were sized to -100 +140 mesh, and an aliquot was taken for a standard bulk analysis. The data for the bulk sample and seven dissolution steps are tabulated in Appendix II. The stepwise dissolution data, and the four analyses and 1374 million year chord determined by Bickford and Lewis are plotted in Fig. 4.

The bulk sample and steps one through five plot directly on the 1374 million year chord. Steps six and seven plot slightly below the chord. This may be a result of their relatively small sample size, the cumulative blank level after five steps, and the dissolution in the last step of minor phases with different U-Pb systems that are present as inclusions in the zircon. Steps three, seven and two plot significantly above concordia. These data

Figure 4.

Concordia diagram showing data from zircons from the Tishomingo granite. Lower points (C, A, D, E) are from a suite of zircons analyzed by Bickford and Lewis (1979). The discordia line is plotted using these four points. Upper points show results of stepwise dissolution of a zircon separate from a different sample of the granite.


STEP 2
1500
STEP 3
STEP 7
STEP 5
STEP 4
STEP 6
BULK SAMPLE
1000
STEP 1
DATA FOR STEP-WISE DISSOLUTION OF DIFFERENT SAMPLE OF TISHOMINGO GRANITE ZIRCON
E
D
A
C
500
ORIGINAL ZIRCON SUITE FROM TISHOMINGO GRANITE BICKFORD AND LEWIS (1979)
0.32
0.28
0.24
0.20
0.16
0.12
0.8
0.4
0
0.5
1.0
1.5
2.0
2.5
3.0
3.5
4.0
4.5
5.0
5.5
6.0
206Pb/238U
207Pb/235U

show that a sample analyzed by stepwise dissolutions will define a chord representing its U-Pb age. The experiment also indicates that HF acid dissolution preferentially leaches out U or Pb atoms from different zones of the zircon crystals. This is especially evident for the three steps which plot above concordia.

For this procedure to work, layers must be chemically removed from the crystals without disturbing the radioactive equilibrium among the U and Pb isotopes in the remainder of the zircon. The analytical data suggest this did not occur. Observation of grain mounts of the sample after each dissolution step also supports this conclusion. The first evidence of HF attack was enlargement and preferential etching of fission tracks. With successive steps outer layers of the zircon were removed, but the fission tracks in the undissolved portion continued to enlarge until they formed an interconnnecting honeycomb of tunnels throughout the crystal. In the case of steps two, three and seven, for example, radiogenic Pb certainly must have been leached out without all of the supporting U.

Even though this technique does not provide a method to analyze the distribution of U and Pb isotopes in successive layers of a model composite zircon 'crystal', I did stepwise dissolutions on a zircon population from Idaho batholith granite for two reasons. First, the experiment on the Tishimingo zircons demonstrates that the data can define a line coincident with a chord determined by standard procedures. Such a line obtained from stepwise dissolution of Idaho batholith zircons would be valuable for comparison with the mixing line obtained from the inclusion suite.

Secondly, it is possible that the physical presence of a core-rim boundary in the zircon would tend to inhibit leaching of Pb or U from the core before most of the rim was dissolved. Thus it might be possible to analyze the U-Pb systems of the core and the rim separately.

The stepwise dissolution was performed on the -100 +200 size fraction of zircons from sample IG-38. This size fraction has the greatest proportion of crystals with cores. An aliquot also was taken for a bulk sample analysis.

Following each dissolution step, I immersed several grains in a high index oil in order to observe the progressive microscopic effects of the HF attack. A typical IG-38 zircon with core is shown before HF treatment in Fig. 5. After the first HF attack at 140 C for six hours, the margins had been preferentially dissolved leaving only remnants of rims attached to the core. But there was also extensive etching of fission tracks throughout the core causing it to become nearly opaque (Fig. 6). Following step two, another six hour attack at 140 C, fewer remnants of the rim remained, and in some crystals the core was strongly embayed (Fig. 7). Fig. 7 also shows that inclusions protruding into the embayment must have been more resistant to HF dissolution. Step three, a twelve hour attack, left behind only rounded embayed crystal fragments. Pieces of rim still attached to the core were much more rarely observed, but in a few cases, as shown in Fig. 8, it appeared that the core was being preferentially dissolved, leaving behind the rim. After 21 more hours of HF attack only ragged, strongly digested crystals remained (Fig. 9). Though most of the remaining grains appeared to be cores, traces of rims persist even through this dissolution with a cumulative length of 45 hours. With step five all remaining zircon was dissolved.

Figure 5.

Typical IG-38 zircon with core before HF acid attack.

Figure 6.

Zircon after six hour HF acid attack at 140 C.

Figure 7.

Zircon after step 2 - a second six hour Hf acid attack at 140 C.

Figure 8.

Zircon after step 3 - a twelve hour HF acid bath at 140 C.

Figure 9.

Zircon after step 4 - a 21 hour HF acid bath at 140 C.

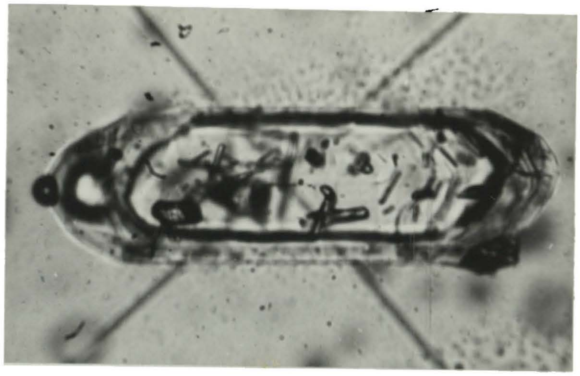


Figure 5

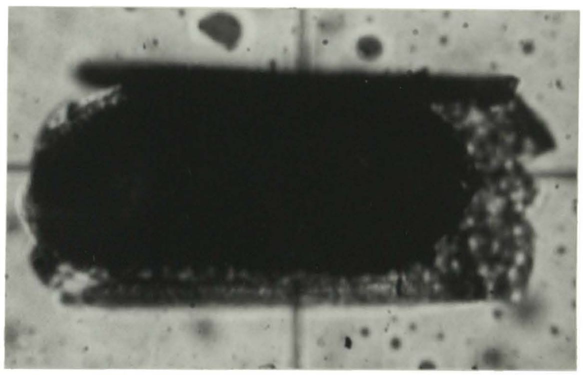


Figure 6

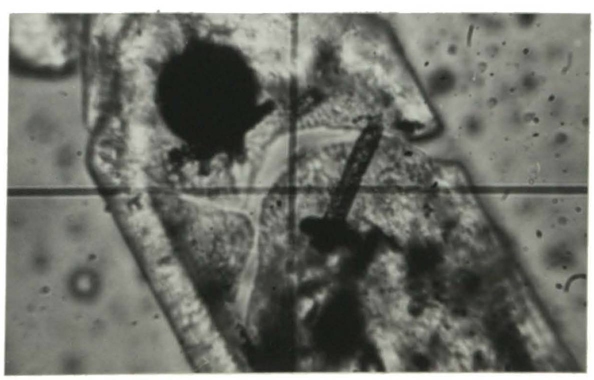


Figure 7

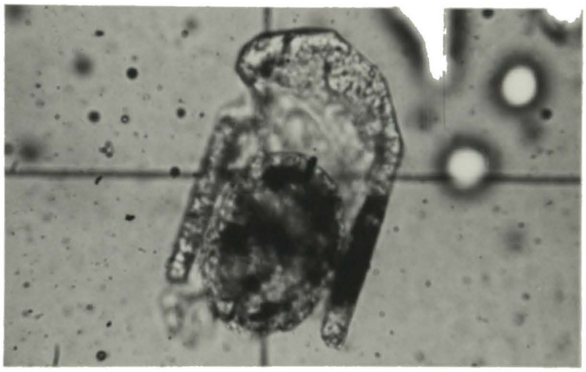


Figure 8

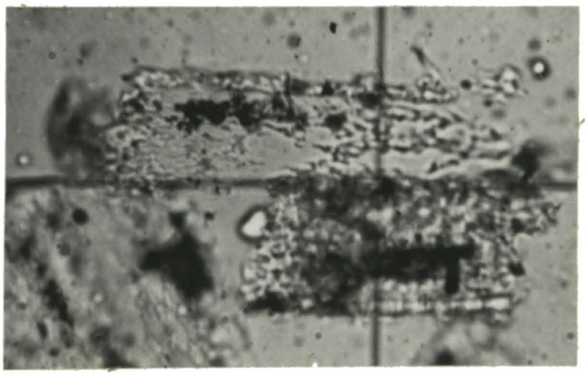


Figure 9

The isotopic data for IG-38 steps one through five, bulk sample aliquot, and the -100 +200 and -200 +400 size fraction analyses are plotted on a concordia diagram in Fig. 10. The mixing line obtained from the study of the inclusion suite is also plotted for reference. The isotopic values are tabulated in Appendix II.

Pb and U concentrations measured in the bulk sample are in excellent agreement with the summed concentration data for all five steps (Table 3). Pb concentrations agree within one percent and U concentrations within 0.7 percent. This is a good indication that the experiment was completed without loss of U or Pb.

If the stepwise dissolution procedure worked succesfully and the working hypothesis for the origin of the zircons is correct, the data for steps one through five should plot on the same line defined by the inclusion suite. Furthermore, step one, which predominantly sampled the rims, should plot nearest the lower intercept, and steps two through five should plot successively closer to the upper intercept as more core material was exposed to HF attack. Fig. 10 shows this expected pattern, with the exception of step two.

Figure 10.

Concordia diagram showing bulk analysis and five dissolution steps for zircons from an Idaho batholith granite (IG-38). Discordia line is fit to five of these points (excluding step 2 data). Data from those samples of the inclusion suite that fall within this field of view are also plotted for reference.

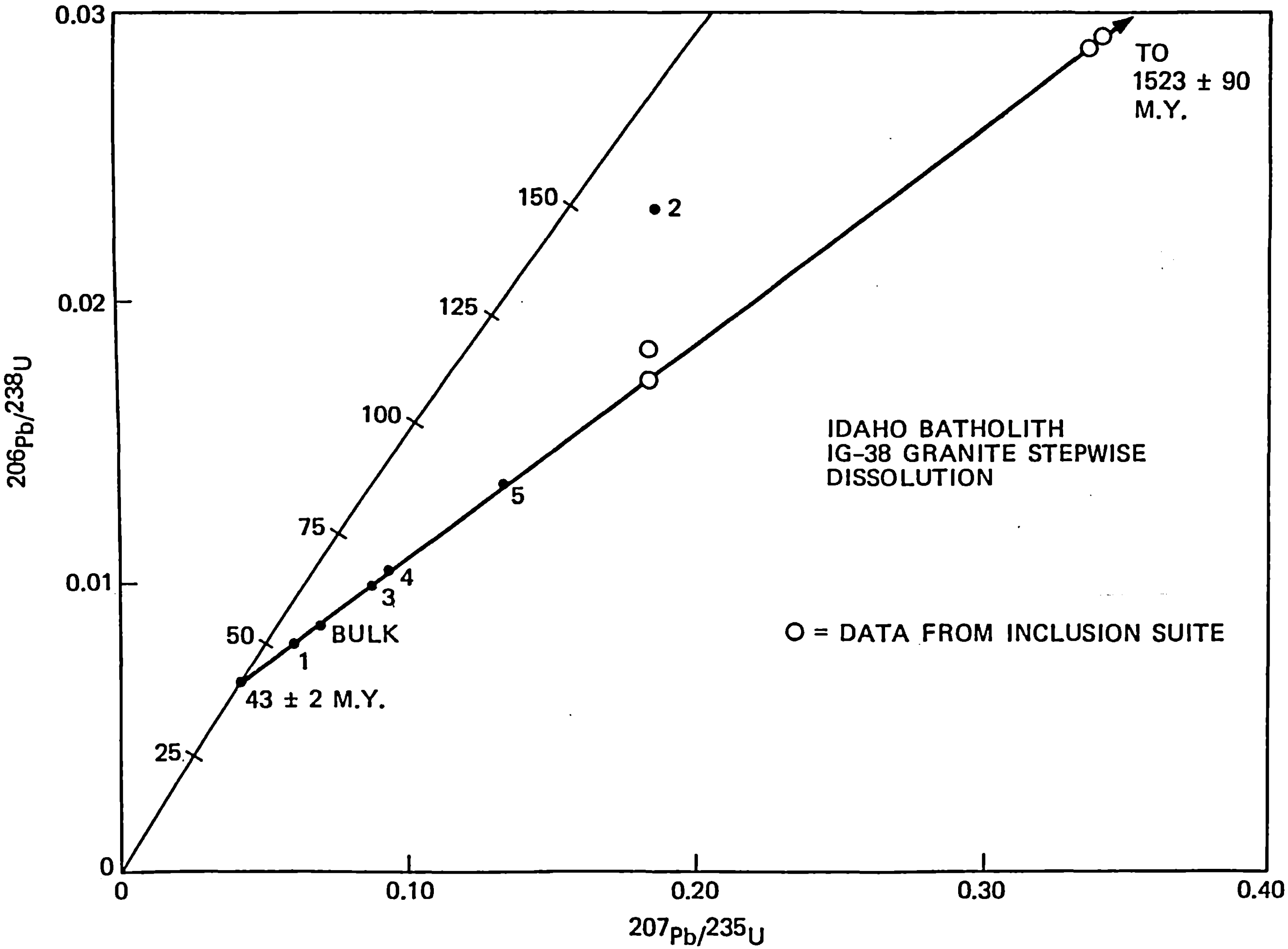

0.03
0.02
0.01
0
$^{206}Pb/^{238}U$
0
0.10
0.20
0.30
0.40
$^{207}Pb/^{235}U$
25
50
75
100
125
150
TO
1523 ± 90
M.Y.
2
5
4
3
BULK
1
43 ± 2 M.Y.
IDAHO BATHOLITH
IG-38 GRANITE STEPWISE
DISSOLUTION
O = DATA FROM INCLUSION SUITE

TABLE 3

TOTAL U AND Pb CONCENTRATIONS OBTAINED FOR THE COMPARISON OF A BULK SAMPLE ALIQUOT WITH THE WEIGHTED AVERAGE OF DISSOLUTION STEPS 1 - 5 (SAMPLE IG-38)

	LEAD		URANIUM	
STEP	ppm	micrograms	ppm	micrograms
1	19.50	0.3568	2449.0	44.82
2	11.42	0.0605	448.0	2.37
3	6.07	0.0789	544.5	7.08
4	6.04	0.0767	380.3	4.83
5	11.27	0.0282	473.7	1.18
1 - 5	11.89		1285.4	
Bulk Aliquot	12.02		1294.2	

The displacement of the step two data from the inclusion suite line cannot be explained by possible amounts of U or Pb contamination, or errors in weighing or isotopic spiking during the experimental procedure. It must be a result of either the presence of a different U-Pb system, perhaps preserved in an included phase and only dissolved during step two, or differential leaching of Pb or U from various zones of the crystal that are not in radioactive equilibrium. This last is the most likely explanation, considering the results of stepwise dissolution of Tishimingo zircons and the microscopic observations made after each step of the IG-38 dissolutions. The stepwise procedure did not strip successive discrete layers from the zircon crystals. I conclude that it may have been fortuitous that points from four of the steps appear to define a line and, therefore, that little information is available from the data. Using the same stepwise procedure, M. E. Bickford (pers. com., 1980) recently obtained similar scatter in the data on other zircon populations.

DISCUSSION OF RESULTS

Age of Inherited and Primary Zircons:

U-Pb isotope data from both the inclusion suite and the stepwise dissolution experiment demonstrate that the zircon population of the Idaho batholith comprises a young and an old component. The chord obtained from the inclusion suite yields an upper intercept age of 1636 +/- 43 million years and a lower intercept age of 45.9 +/- 2.2 million years. Disregarding step two, data from the stepwise dissolution procedure define a chord with intercept ages of 1523 +/- 90 and 42.8 +/- 2.1 million years. Within experimental error the two studies yielded the same ages.

However, the ages obtained from the stepwise procedure are only tentative because of the arbitrary exclusion of step two from the calculations. The variance of at least one of the steps from a chord defined by the rest of the data was observed in this study and confirmed in a subsequent experiment (M. E. Bickford, pers, com., 1980). This suggests that data obtained from a stepwise dissolution experiment are not necessarily constrained to a straight line on a 206Pb/238U versus 207Pb/235U plot. Therefore, fitting a line through the data may not give ages with any geological meaning. It is certainly clear that the stepwise dissolution process is more complex than a simple analysis of U and Pb in consecutive layers stripped from a crystal. The following discussion will, therefore, only deal with ages obtained from the inclusions.

I interpret the 46 million year age as the minimum time of primary crystallization of this phase of the Idaho batholith, and the 1640 million year value as the apparent age of zircons entrained either from the source region or from contaminating crustal material during ascent of the magma. These are actual ages only if the following two assumptions are correct: 1) the inherited zircons came from a single source or sources of the same age, and 2) there has been no U or Pb loss from any of the zircons, that is, the old and young component would each yield a concordant U-Pb age.

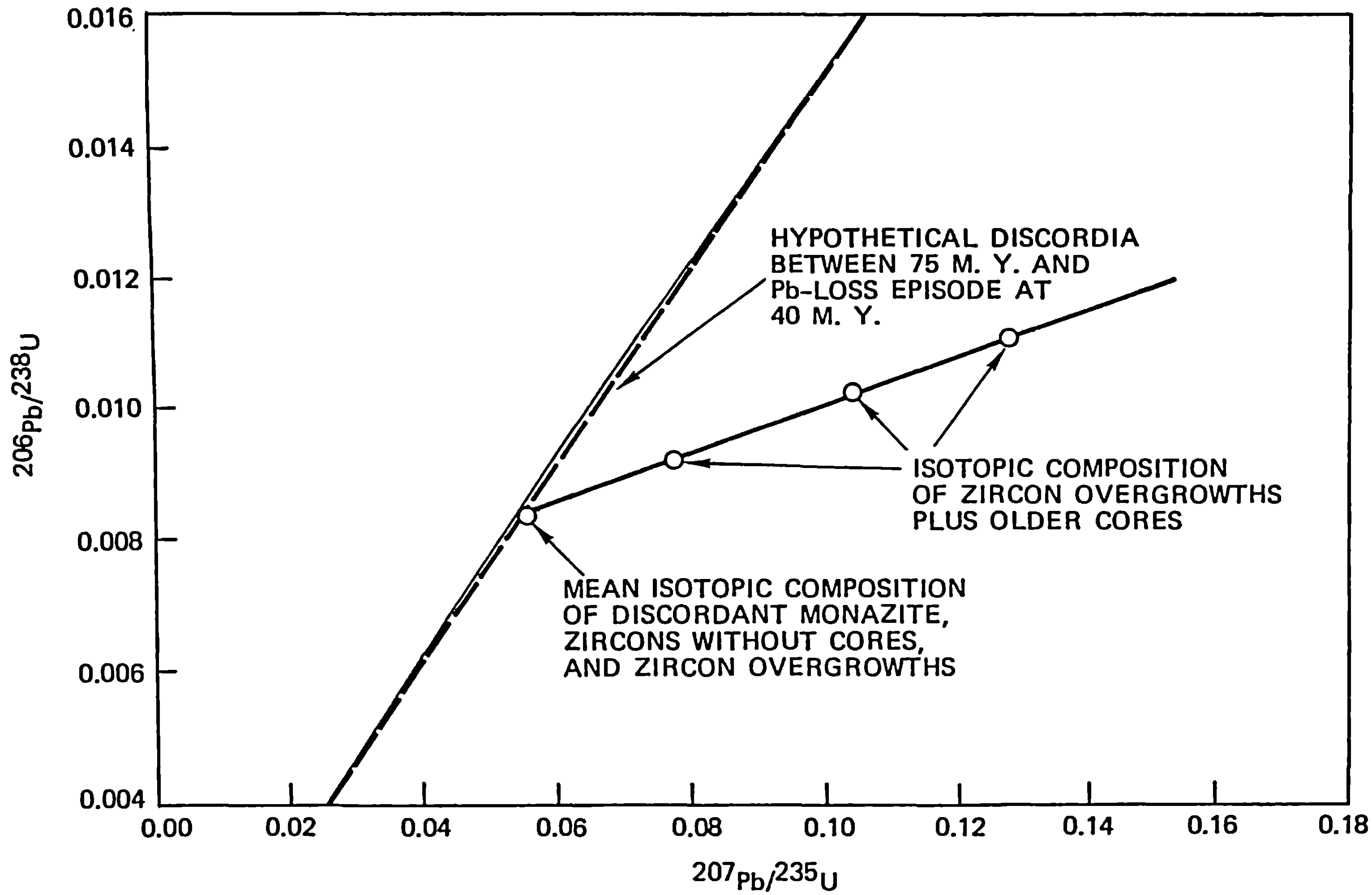
There is no direct way to evaluate uncertainties introduced by the first assumption. In the following section I will review evidence that shows there are possible source terranes for the inherited zircons that would account for the 1640 million year age. Still, the value should only be considered as an apparent age for the source. The zircons may actually have been inherited from several noncoeval sources.

The second assumption is probably not valid because zircons do not generally behave as closed systems with respect to U and Pb. If both the old and young zircons were concordant, then a mixing line through these two points would intersect the U-Pb growth curve at the actual ages of the two components. But if either or both end members lay on a discordia, the mixing line they define would indicate ages that are too low (refer to Fig. 11).

By assuming reasonable ages for the two U-Pb systems of this particular study (46 and 1640 million years), one can calculate to what extent different amounts of Pb loss will lower the ages that would be calculated from a mixing line. This provides some idea of the errors that may be introduced by assuming closed system behavior. The value of the upper intercept is not very sensitive to Pb loss in either the old or young component. For example, the upper intercept does not change beyond experimental error for Pb loss as extensive as 50 percent in the young and 80 percent in the old component. On the other hand, the lower intercept of the mixing curve is very sensitive to Pb loss from the younger component. Below 75 million years the concordia curve closely approximates a straight line and any Pb loss would cause a directly proportional decrease in the model age. Thus, loss of only five percent of the Pb from the young component would cause the calculated age to

Figure 11.

Hypothetical diagram to show the effect of mixing discordant zircons from a young and an old population.



0.016
0.014
0.012
0.010
0.008
0.006
0.004
$^{206}Pb/^{238}U$
0.00
0.02
0.04
0.06
0.08
0.10
0.12
0.14
0.16
0.18
$^{207}Pb/^{235}U$
HYPOTHETICAL DISCORDIA BETWEEN 75 M. Y. AND Pb-LOSS EPISODE AT 40 M. Y.
ISOTOPIC COMPOSITION OF ZIRCON OVERGROWTHS PLUS OLDER CORES
MEAN ISOTOPIC COMPOSITION OF DISCORDANT MONAZITE, ZIRCONS WITHOUT CORES, AND ZIRCON OVERGROWTHS

vary beyond the limits of experimental uncertainty. While the 1640 million year value probably represents a good estimate of the apparent age of the inherited zircons, 46 million years is a minimum age for the crystallization of the batholith.

Source of Inherited Zircons:

Precambrian rocks of pre-Belt age (greater than approximately 1400 million years) are recognized in many areas throughout the northern Rocky Mountains. These exposures provide evidence for events that generated the source terranes of the zircons incorporated into the Idaho batholith. Although available isotopic data are few and subject to large uncertainty, at least some data indicate a 1600 million year event.

The most direct evidence for a pre-Belt terrane subjacent to the Idaho batholith comes from Armstrong's (1975) study of the Salmon River arch. This structural high separates the Bitterroot lobe from the Atlanta lobe. Although an isochron is not well defined, four Rb-Sr analyses of gneissic granites suggest a minimum age of 1450 million years. This may indicate a widespread, previously unrecognized region of approximately 1500 million year old, pre-Belt rocks that has been erroneously mapped as Cretaceous batholith or as Belt terrane. Near Elk City, Idaho, also in the vicinity of the batholith, Reid and others (1970) obtained a 207Pb/206Pb age of 1525 +/- 83 million years from zircon in an augen gneiss. This also represents a minimum age for a previously unknown pre-Belt terrane.

The major recognized Precambrian exposures nearest the Idaho batholith are the Tobacco Root, Beartooth, Ruby, Jefferson, and Madison ranges of southwest Montana, and the Bighorn, Owl Creek, Wind River, and Granite Mountains of Wyoming. Early igneous and metamorphic events occurring about 2700 million years ago have been recognized in all of these areas by several workers (Gast and others, 1958; Gilleti and Gast, 1961; Bassett and Giletti, 1963; Catanzaro and Kulp, 1964; Mueller and Cordua, 1976; Garihan, 1979; James and Hedge, 1980).

Except for some poorly defined K-Ar dates from Wyoming (Condie and others, 1969), 1600 million year ages in which any confidence can be placed have only been reported from the ranges of southwest Montana. Wooden and others (1978) obtained Rb-Sr data from the Dillon granite gneiss that confirmed the earlier work by Giletti and Gast (1961) and Giletti (1966) indicating a major heating event between 1600 and 1700 million years ago specifically in the Tobacco Root Mountains and generally throughout much of southwestern Montana. Other workers (Garihan, 1979; James and Hedge, 1980) have pointed out that these ages are equivocal. From studies on the Dillon granite gneiss in the Ruby Range they suggest instead a 2700 million year old event followed by little or no activity at 1600 million years. The relationship of the granite gneiss with surrounding metamorphic rocks is complex (Garihan, 1979) and the conflicting 1600 and 2700 million year age determinations may simply reflect the noncorrelative nature of the 'Dillon granite gneiss' between different areas. The general pattern is for rocks toward the southwest to yield ages on the order of 2600 million years whereas identical rocks farther northwest yield ages of 1600 to 1800 million years. The westernmost granitic rocks of the Dillon type consistently yield 1600 million year ages (King, 1976). Although evidence is not conclusive, the data from the literature to date make it reasonable to postulate the presence of a 1600 million year old basement terrane as the source for the inherited zircons in the Idaho batholith.

Age of the Idaho Batholith:

The Eocene age obtained in this study lends additional support to the suggestion that, because of inadequate or ambiguous age determinations, many Tertiary intrusives within the batholith are unrecognized and mistakenly mapped as Cretaceous (Armstrong, 1974; Swanberg and Blackwell, 1973). The known Tertiary intrusives are dispersed along a general north-south trend on the east side of the batholith. Other, smaller Tertiary intrusions and dikes are known but unmapped. Contours of K-Ar ages show a general relationship to mapped Eocene plutons. By studying oxygen and hydrogen isotope data, Taylor (1978) demonstrated that resetting of the K-Ar ages was caused by widespread meteoric-hydrothermal activity induced by the Eocene plutonic events.

Previous workers (Williams, 1979; Hyndman and Williams, 1977) have noted some general mineralogic and textural differences between the Tertiary and Cretaceous granitic rocks. The older intrusives generally are two mica rocks, with abundant plagioclase, strained quartz, lobate or sutured grain boundaries, and myrmekitic textures. The Tertiary granites have no primary muscovite, but abundant alkali feldspar, miarolitic cavities, sharp, straight grain boundaries, undeformed quartz, and rare myrmekite. As a general observation, the Tertiary plutons show epizonal features, whereas the Cretaceous bodies are meso- or katazonal.

Even though the granite examined in this study is apparently of Tertiary age, it has most of the features attributed to Cretaceous granites. With the exception of the absence of primary muscovite, sample IG-38 has all the features attributed to Cretaceous granite, but yields a minimum age of 46 million years. Thus, in addition to indicating the Idaho batholith may be as young as Eocene, this study also suggests that these young intrusives may be easily overlooked because of their mineralogical and textural similarities to the Cretaceous intrusions.

As argued in the preceeding section, one should recognize that the crystallization age found for the pluton in this study represents a minimum, and a Cretaceous age cannot be ruled out. Nevertheless, it is clear that the Idaho batholith is a complex multiple intrusive consisting of Cretaceous plutons invaded and modified by Eocene plutons and their attendant meteoric-hydrothermal systems. A considerable amount of field and isotopic data will be necessary to unravel the field and age relationships among the plutons.

Further Studies:

The U-Pb isotope data obtained from this study clearly demonstrate significant involvement of Precambrian crustal material in the origin of the Idaho batholith; but, one cannot determine the relative contributions of mantle and crustal material solely from the data presented here. In addition, chemical data, especially on the trace and rare earth elements, are necessary to place constraints on the source region and possible mantle-crust mixing models. Data from other radiogenic isotope systems will also be important in answering these questions. Correlated Nd and Sr isotopic variations have been valuable in understanding the result of mixing magmas from different sources (DePaolo, 1979; DePaolo and Wasserburg, 1979). Data of this nature, in conjunction with mathematical modeling of partial melting, fractional crystallization and assimilation, will have important applications to characterizing the source regions and genesis of major granitic terranes such as the Idaho batholith.

ACKNOWLEDGEMENTS

I wish to acknowledge the guidance of Dr. M. E. Bickford in selecting and carrying out this study, Drs. W. R. VanSchmus and A. W. Walton for critically reviewing the manuscript, Dr. R. B. Chase for guidance in the field, and K. Chase for carrying more than his share of the rocks. Financial support for this work was provided by National Science Foundation Grant EAR 78-13687.

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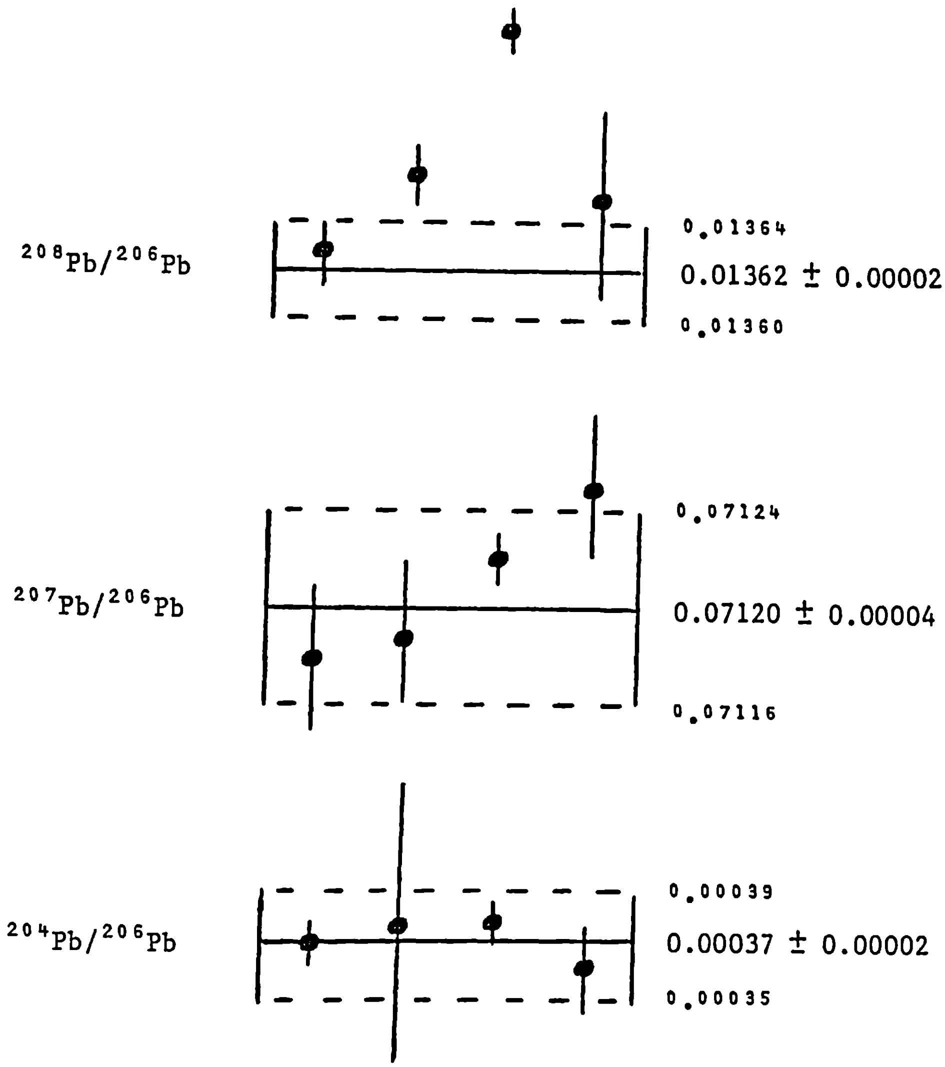
APPENDIX I

ANALYSES OF NBS RADIOGENIC LEAD STANDARD



Measured values of NBS radiogenic lead standard from June through August 1979. Data points are plotted with 2 sigma error bars. Accepted NBS values are given by the reference line and its error envelope. With the exception of one $^{208}Pb/^{206}Pb$ value, the data show reproducibility and, within error, agreement with NBS values.

APPENDIX II

URANIUM-LEAD (ZIRCON) DATA FOR STEPWISE DISSOLUTION OF IG-38, INCLUSION SUITE, IDAHO BATHOLITH GRANITE, AND TISHIMINGO GRANITE STEPWISE DISSOLUTION

ZIRCON DATA: STEPWISE DISSOLUTIONS

SAMPLE [a]		observed Pb isotope ratios [b]			concentration(ppm)		calculated atom ratios [c]		
		208/206	207/206	204/206	$^{206}Pb*$	^{238}U	$\frac{207*}{235}$	$\frac{206*}{238}$	$\frac{207*}{206*}$
IG-38	step 1	0.0913	0.0633	0.00056	16.81	2432	0.0607	0.0080	0.0552
IG-38	step 2	0.1575	0.0875	0.00196	9.02	444.8	0.1907	0.0234	0.0590
IG-38	step 3	0.1574	0.0833	0.00142	4.83	540.6	0.0894	0.0103	0.0628
IG-38	step 4	0.4184	0.1800	0.00798	3.44	377.6	0.0943	0.0105	0.0650
IG-38	step 5	0.5606	0.2191	0.01023	5.53	470.4	0.1370	0.0136	0.0731

* designates radiogenic lead

a samples with a B suffix are a second aliquot of zircons from the preceeding sample

b observed ratios have been corrected for mass fractionation

c U/Pb ratios are estimated to be correct to ± 2 percent.
Pb isotopic composition used for blank correction is
(208:207:206:204) = (39.0:15.8:18.6:1.0)
All samples are corrected for incorporation of 42 m.y. common Pb.

ZIRCON DATA: INCLUSION SUITE

SAMPLE [a]		observed Pb isotope ratios [b]			concentration(ppm)		calculated atom ratios [c]		
		208/206	207/206	205/206	$^{206}Pb^*$	^{238}U	$\frac{207^*}{235}$	$\frac{206^*}{238}$	$\frac{207^*}{206^*}$
IGI-35	-100+400	0.1043	0.1112	0.00013	88.35	493.2	3.125	0.2070	0.1097
IGI-35B	-100+400	0.1047	0.1105	0.00009	87.81	495.3	3.088	0.2049	0.1093
IGI-28	-100+200	0.1945	0.1172	0.00129	57.73	474.2	1.929	0.1407	0.0995
IGI-28B	-100+200	0.2150	0.1234	0.00176	58.85	461.5	2.014	0.1474	0.0991
IGI-28	-200+400	0.1754	0.1060	0.00051	77.97	600.7	2.047	0.1500	0.0990
IGI-23	-100+400	0.1218	0.0930	0.00097	24.62	1656	0.1878	0.0172	0.0793
IGI-19	-100+400	0.1139	0.0885	0.00030	45.52	1788	0.3418	0.0294	0.0842
IGI-19B	-100+400	0.1171	0.0893	0.00037	48.69	1907	0.3419	0.0295	0.0841
IGI-17	-100+200	0.1127	0.0844	0.00076	15.09	947.4	0.1868	0.0184	0.0736
IGI-17	-200+400	0.2254	0.1216	0.00344	15.23	1016	0.1731	0.0173	0.0725

ZIRCON DATA: IDAHO BATHOLITH GRANITE

SAMPLE [a]	observed Pb isotope ratios [b]			concentration(ppm)		calculated atom ratios [c]		
	208/206	207/206	204/206	$^{206}Pb^*$	^{238}U	$\frac{207^*}{235}$	$\frac{206^*}{238}$	$\frac{207^*}{206^*}$
IG-38 -100+200	0.1654	0.0908	0.00195	9.92	1286	0.0769	0.0089	0.0625
IG-38B -100+200	0.1572	0.0873	0.00188	9.50	1285	0.0707	0.0085	0.0600
IG-38 -200+400	0.1492	0.0690	0.00114	7.22	1079	0.0558	0.0077	0.0523

ZIRCON DATA: TISHIMINGO GRANITE STEPWISE DISSOLUTIONS

SAMPLE [a]	observed Pb isotope ratios [b]			concentration(ppm)		calculated atom ratios [c]		
	208/206	207/206	204/206	$^{206}Pb^*$	^{238}U	$\frac{207^*}{235}$	$\frac{206^*}{238}$	$\frac{207^*}{206^*}$
bulk	0.2394	0.1086	0.00153	9.22	51.72	2.4747	0.2060	0.0871
step 1	0.5223	0.2560	0.01215	45.70	298.0	2.0934	0.1772	0.0857
step 2	0.3372	0.1661	0.00560	41.56	168.4	3.4515	0.2852	0.0878
step 3	0.2890	0.1285	0.00289	22.40	107.4	2.9277	0.2410	0.0881
step 4	0.2899	0.1231	0.00254	10.22	52.99	2.6914	0.2228	0.0876
step 5	0.5322	0.2198	0.00945	7.28	37.39	2.7155	0.2249	0.0876
step 6	0.6976	0.2833	0.01361	8.04	44.12	2.7405	0.2106	0.0944
step 7	1.3932	0.5662	0.03422	0.84	3.86	3.0180	0.2512	0.0871

All samples are corrected for incorporation of 1,375 m.y. common lead.

APPENDIX III

Stepwise dissolution analytical procedure

Zircons were separated from the samples using standard crushing, heavy liquid, and magnetic separation techniques. Before any chemical processing the zircons were sieved to four or five size fractions and any impurities were picked out by hand. The sample was then washed several times in acetone and deionized water.

In order to correct for adsorption of moisture on to the sample, all weighings before and after each step were done only after the sample and its teflon container had been heated in a drying oven and then cooled to ambient temperature in a dessicator. Approximately 50 milligrams of zircons were used for the stepwise procedures. First the sample was boiled in distilled 7N nitric acid for approximately one hour. The acid was decanted and the zircons were rinsed with deionized water several times and reweighed.

The sample was then subjected to an HF acid leach at room temperature. Ten ml of distilled HF were added and one hour later pipetted off into a clean teflon beaker. A clean, disposable pipette tip was used for each step.

Before use each tip was rinsed twice in deionized water and once in distilled HF. After removing the acid, the zircons were washed twice with 10 ml of deionized water. This rinse water was combined with the HF, dryed down, converted to chlorides, and processed for subsequent U and Pb isotopic analysis. The zircons and teflon container were dryed in an oven, cooled in a dessicator, and reweighed. At this point in each step 10 to 20 zircon crystals were removed with a clean spatula for microscopic observation. The zircon sample was then reweighed before the subsequent step.

The high temperature dissolution steps were performed in the same manner, though dissolution took place in a teflon bomb at temperatures of 180 to 200 degrees C for various lengths of time. After the bomb was removed from the oven, it was allowed to cool for four hours before the acid was removed. Three to 4 ml of distilled HF and 3 drops of distilled 7N nitric acid were used in each of the dissolution steps. The steps were continued until the entire sample had been dissolved. A blank was run simultaneously with every step in order to monitor lab contamination.