1986

Synplutonic dikes of the Idaho batholith Idaho and western Montana and their relationship to the generation of the batholith

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SYNPLUTONIC DIKES OF THE IDAHO BATHOLITH, IDAHO AND WESTERN MONTANA, AND THEIR RELATIONSHIP TO THE GENERATION OF THE BATHOLITH

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B.A., SUNY Potsdam

Presented in partial fulfillment of the requirements

for the degree of

Master's of Science

University of Montana

1986

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Dean, Graduate School

May 16, 1986

Date

Foster, David A., M.S., June 1986 Geology

Synplutonic Dikes of the Idaho Batholith, Idaho and Western Montana, and their Relationship to the Generation of the Batholith (79 pp.)

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The Bitterroot Lobe of the Idaho batholith contains numerous synplutonic mafic dikes which are oriented ENE-WSW. These dikes tend to be near vertical, roughly planar, sharply bounded, metamorphosed, and have an asymmetric foliation oblique to their walls. They average 2.4 meters in thickness and make up about 20 percent by volume of a representative section of the batholith. These dikes show a wide range in degree of metamorphism and deformation by the host granite, which suggests they were injected when the batholith was still hot. The least metamorphosed mafic dikes are basaltic andesites and andesites which are high alumina in character. The more highly metamorphosed dikes are highly sheared, fully recrystallized, and more dacitic. A wide range of mutually intrusive relationships indicate these mafic dikes were injected into the granite when it was still partly liquid but solid enough to fracture. Mafic inclusions, inclusion trains, and schlieren which in places grade into dikes represent similar mafic magma that was being emplaced in the batholith before the dikes were injected. Some of the dikes are very inhomogeneous and consist of pillow shaped globs of andesite surrounded by thin rims of aplite. In and around these dikes there is strong evidence for mechanical mixing of magma between the andesite and the host granite. These mafic dikes probably represent the latest stages of subduction generated mafic magmas which were injected into the continental crust and contributed to melting to form the batholith.

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I. 1 INTRODUCTION

The generation of voluminous granitic magma to form major continental margin batholiths is controversial. In most cases lower crustal temperatures produced by decay of radioactive elements are several hundred degrees below the solidus of granite for reasonable continental margin crustal thicknesses. In one area of the prebatholith Sierra Nevada (locality SJ), Lachenbruch (1968) calculated temperatures at the base of the crust of only about 250°C before magma generation, assuming mantle heat flow was the same as present and heat producing elements were uniformly redistributed through the crust.

In areas that lack geologic evidence for extreme crustal thickening, elevated lower crustal temperatures can be caused by the upward transport of mafic magma into the crust, which promotes melting out of lower melting fractions (e.g., Lachenbruch, 1968; Presnal and Bateman, 1973; Younker and Vogel, 1976; Lachenbruch, and others, 1976; Eichelberger and Gooley, 1977; Hildreth, 1981; Cobbing and Pitcher, 1983; Shaw, 1985; Mukasa, 1986). Numerous criteria suggest that basalt injection aids in the formation of batholiths: 1) the need for additional heat to cause melting, 2) nearly all magmatic systems contain some basalt (Hildreth, 1981), and 3) heat flow calculations by Lachenbruch and others (1976) suggest that large silicic systems require continued intrusion of basaltic magma to support the system for their lifetimes of a few million years. Basalt injection into the crust may be in the form of dikes and sills (e.g., Lachenbruch and others, 1976; Younker and Vogel, 1976) underplating at the crust- mantle boundary (e.g., Cobbing and Pitcher, 1983), or a combination of these (e.g., Hildreth, 1981).

Granitic batholiths are commonly associated in space and time with basaltic and intermediate rock types. According to Bateman (1983) this mafic material is present as satellitic gabbroic bodies, mafic clots and inclusions, and synplutonic dikes. Direct evidence of the mafic magma that aided in the formation of a batholith has been elusive, but it may be recorded in the widespread mafic to intermediate dike swarms common to large batholiths (e.g., Hildreth, 1981; Reid and others, 1983; Hyndman, 1984, 1985b; Toth, 1985). In most batholiths, synplutonic dikes range in composition from basalt to granodiorite. The most mafic of these dikes in a particular batholith are generally high alumina basalt or basaltic andesite, suggesting they are related to subduction. Most of these dikes are to some extent recrystallized and foliated.

A review of the literature of synplutonic dikes reveals a few older papers that provide good descriptions of the petrography, metamorphism, and deformation of these dikes (e.g., Roddick and Armstrong, 1959; Roddick, 1965; Moore and Hopson, 1961). In more recent papers, synplutonic dikes have been mentioned in regional reports about large batholiths, or in the context of a model involving the generation of granitic magma. A list of some of the major papers that discuss synplutonic dikes in some detail is in Table 1.

Table 1, References describing synplutonic dikes

associated pluton	author(s)
Sierra Nevada batholith	Moore and Hopson (1961); Bateman (1983); Chen and Moore (1979); Reid and others (1983); and Bateman and others (1963); Furman and Spera (1985); Hibbard and Watters (1985).
Coast Plutonic Complex, B.C.	Roddick and Armstrong (1959) Roddick (1965)
Coastal batholith of Peru	Cobbing and Pitcher (1972).
Idaho batholith	Williams (1977); Hyndman (1984)
Topsails igneous terrane, Western Newfoundland	Whalen and Currie (1984)
Mount Desert Island, ME	Taylor and others (1980)
Wadi Um-Mara area, Sinai	Eyal (1980)

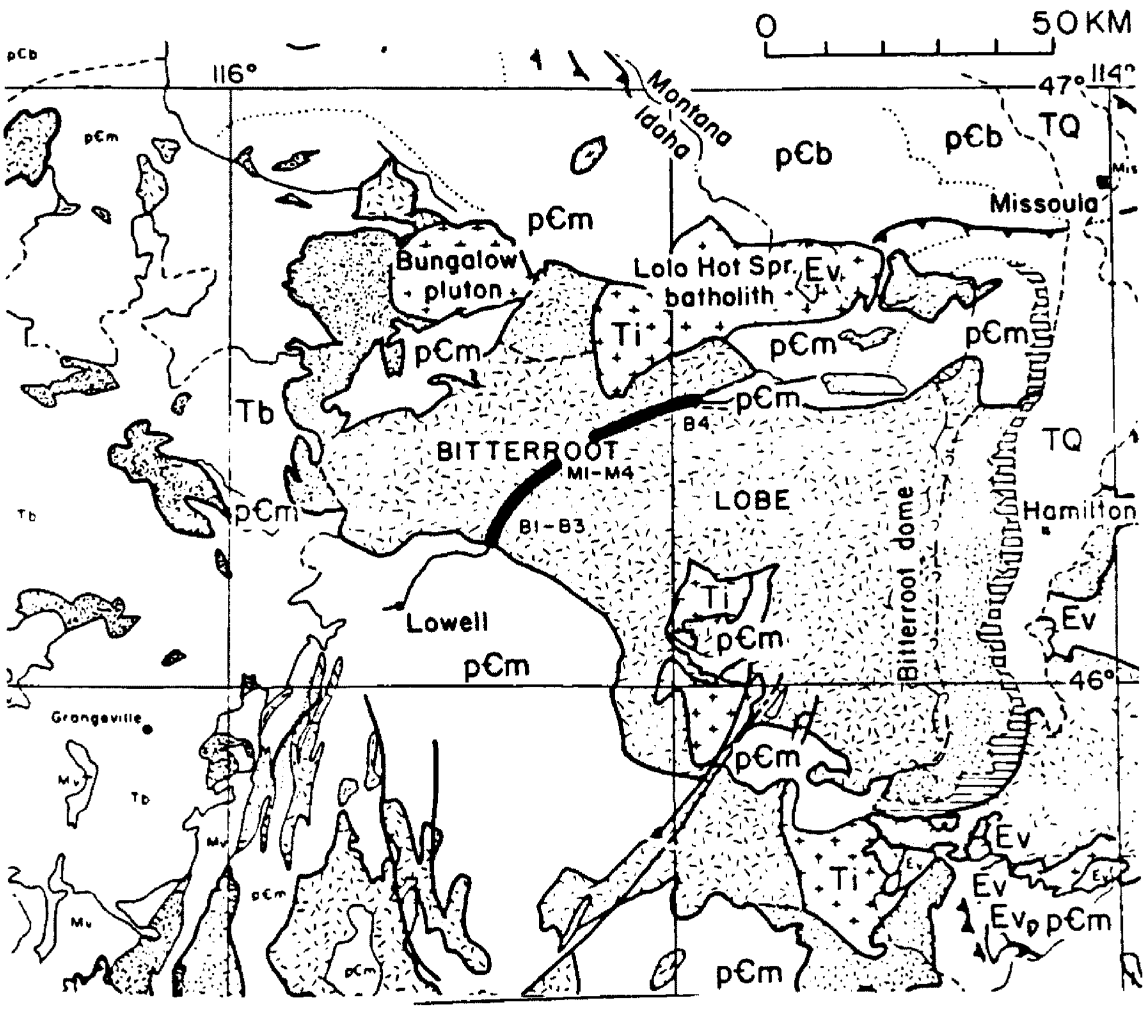
I. 2 GEOLOGIC SETTING

As background to the following discussion of the role the synplutonic dikes played in generation of the Idaho batholith, I briefly introduce the tectonic setting and petrology of the batholith.

The late Cretaceous Idaho batholith occupies much of central Idaho and part of western Montana. It is 390 km north-south and averages 100 km east-west, giving a total area of 39,000 km^2 (Hyndman, 1983). Geographically, the batholith is in two parts, separated by the southeast-trending Salmon River Arch. The southern, or Atlanta, lobe is elongate north south and covers 25,000 km^2. The northern Bitterroot lobe is elongate from northwest to southeast and covers 14,000 km^2 (figure, 1).

The Idaho batholith, like other muscovite-bearing plutons of the Cordillera, is in the hinterland of the Cordilleran belt. It lies within a belt that underwent Mesozoic regional metamorphism (Miller and Bradfish, 1980). Country rock for the Bitterroot lobe includes amphibolite facies Belt Supergroup schists and paragneisses and pre-Belt basement orthogneisses. Country rocks for the Atlanta lobe include: on the north and northeast, Belt and pre-Belt basement schists and gneisses; on the east, Devonian to Permian marine clastics; and along

Figure 1
Map of the Bitterroot Lobe of the Idaho batholith, showing position of U.S. 12 (thick line) (after Hyndman, 1984).



the western contact, submarine and arc volcanics of Permian to Triassic age (Hyndman, 1983). This sequence of rock suggests that the majority of the batholith is in old continental crust, the western border is emplaced in sediments and volcanics of a converging ocean-continent plate boundary.

The Idaho batholith lies east of a Permian to Triassic subduction complex and apparently allochthonous terranes of westernmost Idaho and eastern Oregon. These terranes include the Seven Devils volcanic arc and the Wallowa-Blue Mountains terrane (Vallier, 1977; Talbot, 1977; Hamilton, 1978; Hyndman 1983,1984; Brooks and Ferns, 1985; Follo, 1985). The Permian and Triassic Seven Devils group is exposed in the Snake River Canyon and in the Seven Devils Mountains. Vallier (1977) indicates these rocks consist of flysch sediments and volcanic rocks formed in the vicinity of a volcanic arc at a converging plate boundary. The easternmost accreted island arc is in the Riggins region. It is metamorphosed and consists of dacite, andesite, arc basalt, tholeiitic basalt, serpentinite, serpentinite matrix melange and sediments. Hamilton (1978) interprets the serpentinites and tholeiitic basalts as ophiolite materials.

Major structures in the western part of the island arcs are east to southeast-dipping thrusts. Eastward toward Riggins are multiple thrusts and recumbent folds. Farther east toward the batholith metamorphic grade increases and contemporaneous stuctures are tightly appressed (Lund, 1984; Hyndman, Alt, and Sears, 1986). This structural pattern is the expression of a subduction complex, that was brittle in its outer, colder zones and ductile in the inner hotter zones (Talbot, 1977).

Most of the Late Cretaceous main phase of the Idaho batholith is medium grained, massive to moderately foliated, muscovite-biotite granite to granodiorite (Hyndman, 1983, 1984). A 12 tO 16 km western border zone is made up of foliated hornblende and biotite rich-tonalite and quartz-diorite. Granodiorite is more common towards the border of each lobe, granite is more common toward the center. Emplaced into the granite and granodiorite are mafic synplutonic dikes with compositions ranging from basaltic andesite to dacite.

The boundary between the tonalite and quartz diorite to the west and granodiorite and granite to the east approximately coincides with a 0.704 $^{87}Sr/^{86}Sr$ inital ratio boundary defined by Armstrong and others (1977). This change probably marks the boundary between Precambrian crust in the east and Phanerozoic oceanic material in the west

(Armstrong, and others, 1977; Hyndman, 1983). In the batholith K_2O increases toward the craton and MgO and CaO increase toward the west. This trend is consistent with derivation above a subducting slab and is probably due to an increasing contribution of old continental materials to the magmas toward the craton. (Hamilton, 1978; Hyndman, 1985a, p. 260-262)

Low strontium isotope ratios and compositions of the tonalites and quartz diorites of the western border zone of the Idaho batholith suggest derivation by partial melting of the mantle above subducted ocean crust without much contamination from old continental crust. The higher strontium ratios and compositions of the main phase granites and granodiorites suggest derivation primarily from old continental crust. For the Idaho batholith this could include Belt metasedimentary rocks or pre-Belt basement rocks. (Hyndman, 1984; Armstrong, and others, 1977)

Hyndman (1984) suggests that most of the characteristics of the mainphase Idaho batholith can be explained by a model involving the melting of lower continental crust aided by the upward movement of subduction generated mafic magma from the mantle. Some of this mafic magma may now exist as the mafic synplutonic dikes that intrude the batholith. This model can explain: the low, intermediate, and high initial Sr isotope ratios; the felsic, intermediate, and mafic rocks in the batholith; and the attainment of temperatures needed for regional partial melting of the crust.

I. 3 FIELD AREA AND METHODS

I studied mafic synplutonic dikes that intrude the Idaho batholith to evaluate the model that these dikes represent the later stages of rising basalt which aided crustal melting to form the batholith (Hyndman, 1985b). Most of my field work was along the Lochsa River in central Idaho in the Bitterroot Lobe of the Idaho batholith, where there are good exposures of the batholith in the road cuts of US 12 (figure, 1). To evaluate how far this study could be extrapolated throughout the batholith I did reconnaissance work in numerous areas of the Bitterroot Lobe. These areas include: 1. along the North Fork of the Clearwater River between Kelly Forks and Bungalow, Idaho, 2. the Elk Summit area south of the Powell Ranger Station, Idaho, 3. Lost Horse Canyon and Bear Creek in the Bitterroots of Montana and Idaho, 4. Blodgett Canyon, Montana.

In the field I described outcrops and intrusive relationships, sampled dikes and granite for chemical analysis and thin sections, and measured the percentage of dikes in well exposed sections of the batholith. About one half of the dikes were sampled for thin sections. In addition, I carefully selected fresh dikes to sample for chemical analysis. Of these, I sampled several "typical" dikes, a suite of the most mafic and most felsic dikes, and a suite that appeared to show evidence for mechanical magma mixing. All dike samples along the Lochsa section were keyed to mile posts along US 12.

I estimated the volume percentage of dikes in the batholith along the Lochsa section by measuring perpendicular to the strike of the dikes with a 30-meter tape. I measured both the thicknesses of the dikes and the thicknesses of the granite between the dikes. I did not measure poorly exposed outcrops. Measurements were taken at road level only to be sure that everything was measured from an equal datam. At the end of each outcrop 1/2 the distance from the last dike was measured to maintain proportions of dike and granite. I also, made scale measurements of photographs taken of large cliffs of the batholith exposed in Blodgett Canyon.

II. 1 OUTCROP AND MEGASCOPIC DESCRIPTIONS

Mafic dikes in the Idaho batholith are black to grayish or greenish black basalts, andesites, and dacites that range from about 5 cm to 20 m thick. Most of the dikes trend ENE-WSW and have steep dips, a smaller group on the west side of the batholith trend N-S (figure, 2a+b). Dikes constitute 20% by volume of a representative section of the batholith. My reconnaissance work suggests that these dikes exist throughout the batholith. Everywhere I examined the batholith there were mafic dikes or dioritic complexes. Toth (1983) noted diorite complexes and synplutonic dikes throughout the Selway-Bitterroot Wilderness area.

Intensity of dike deformation ranges from nearly undeformed, massive dikes to highly sheared and dismembered dikes. Because there is a full gradation, I will describe trends of how these dikes appear in outcrop rather than grouping distinct types.

The fresher mafic dikes have distinct very fine grained borders, straight contacts, and have intruded along fractures. These tend to be dense, massive, medium to dark gray or black, and range from basaltic andesite to dacite. Most appear nonmetamorphosed in the field. Many have 1-2mm phenocrysts of plagioclase and pyroxene, and some of the larger dikes are fine to medium grained in their cores. A large number of

Figure 2a
Poles to dike strikes, n=158.

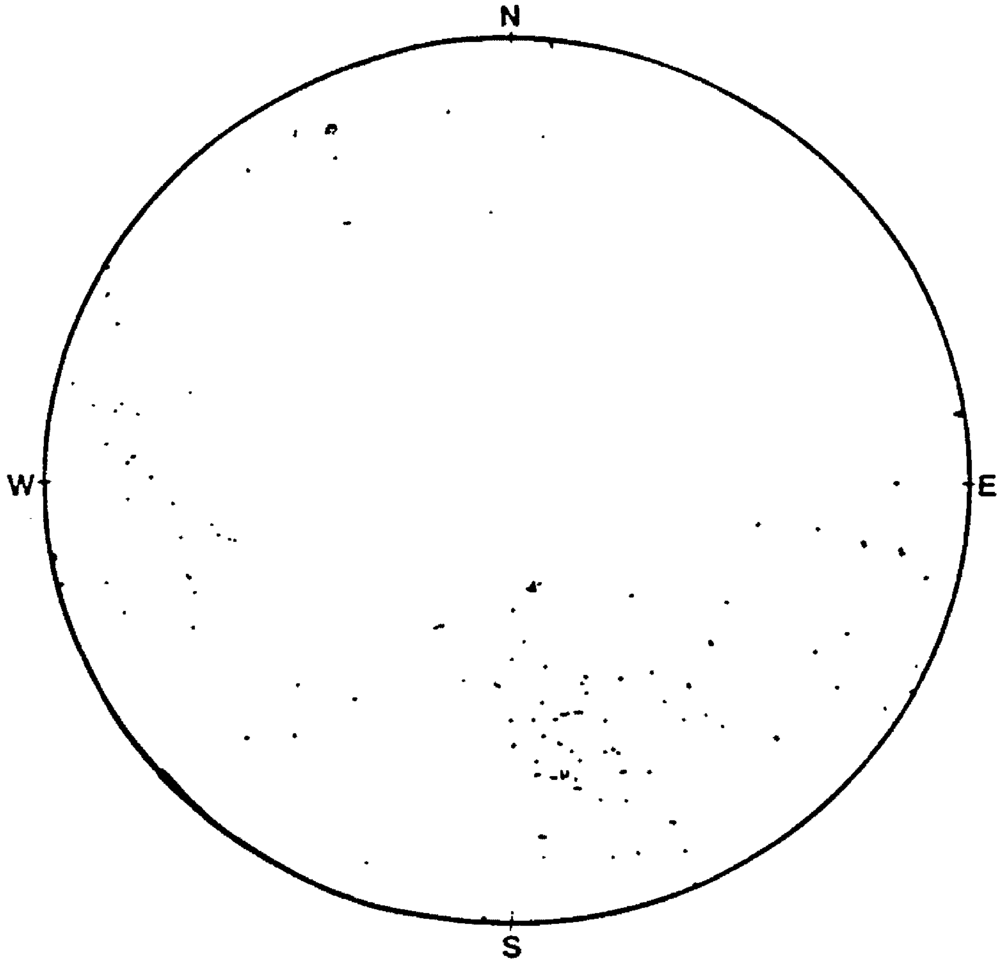
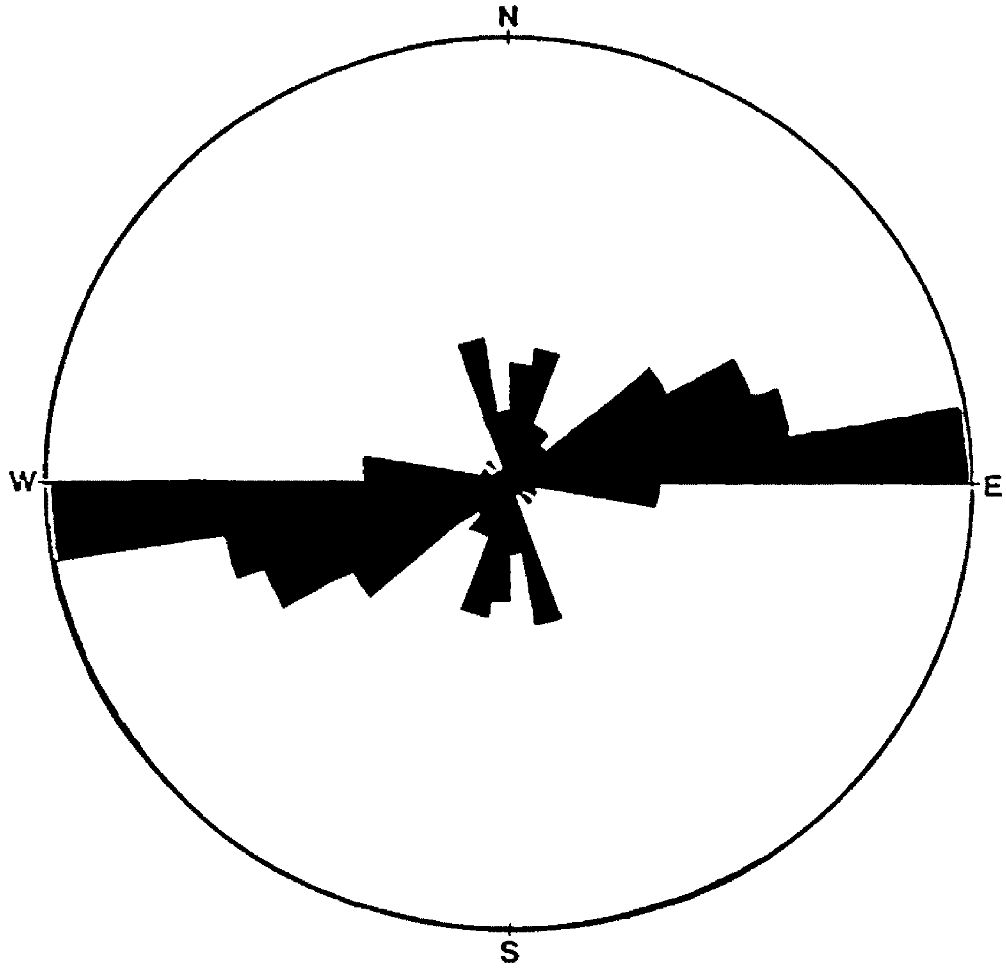


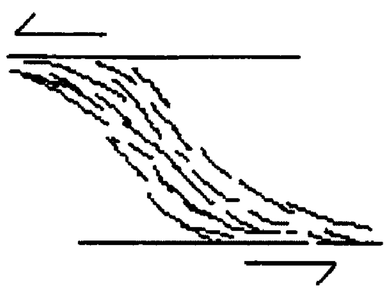
Figure 2b
Rose diagram of dike strikes, n=158.



dikes contain clumps of light colored fine to medium grained euhedral plagioclase and pyroxene crystals that range in size from 2mm to 2cm. These dikes nearly always contain rounded very fine grained black inclusions of more mafic basaltic rock of various sizes, and many contain angular and rounded xenoliths of granite. About two thirds of these dikes are cut by tabular granitic dikes and veins.

Moderately deformed dikes are greenish black or gray basaltic andesite to dacite. They are tabular, and sharply bounded with slighly irregular contacts. These dikes are foliated and in some cases lineated. The foliation is defined by biotite; the lineations are defined by hornblende and actinolite, laths of plagioclase, and elongate mafic clots. The foliation in many of these dikes is asymmetric and oblique to the walls, suggesting that it was imposed by movements of the host granite (figure, 3) (Williams, 1977). Other dikes have foliation parallel to their borders. Some dikes of this suite occur as complex sheets of variable thickness that branch and recombine around large angular blocks of granite. Dikes of this suite are almost always cut by granitic dikes and veins, some of which are irregular in form. These dikes commonly contain fragments of granite of various sizes, and in some cases, rounded inclusions of basaltic rock. One dike contains a 15 cm angular piece of biotite schist similar to the country rock around the batholith.

Figure 3, Oblique foliation



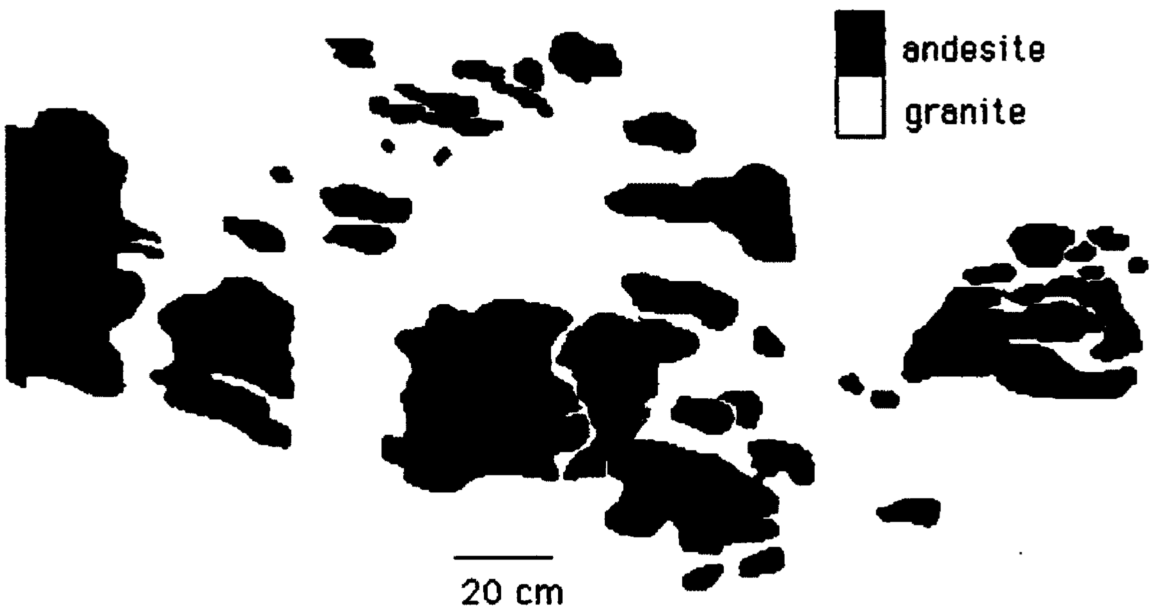
Highly deformed dikes are very fine grained, green to brownish green, schistose, and range from andesite to dacite. Many pinch and swell and have irregular to diffuse borders and irregular shapes. A number are folded, disjointed, and boudinaged. All are foliated and lineated with the foliations defined by biotite and lineations defined by aligned hornblende, plagioclase, and quartzofeldspathic lenses. The lineations curve with the foliations. In some dikes a crenulation cleavage cuts the earlier schistosity. This schistosity and crenulation is rarely tracable into the enclosing granite. The foliation tends to be sub-parallel to the borders of the dikes, but in some the foliation is oblique and asymmetric to the borders. Some of the most deformed dikes are mylonites that show well defined S and C surfaces. These dikes are nearly always cut by granitic dikes and veins, many of which are irregular and ptygmatic. Williams (1977) noted that small offsets of the borders of mafic dikes do not cut the granite, suggesting the granite in those cases was not deformed during the development of the schistosity. The deformation and metamorphism in this dike suite grades into that seen in mafic xenolith trains. These inclusion trains consist of angular blocks of dike-like rock which are probably broken up, dismembered, and rotated portions of former dikes (figure 4).

FIELD DESCRIPTIONS OF INHOMOGENEOUS DIKES AND DIKE COMPLEXES

A number of dikes are inhomogeneous, consisting of various proportions of mafic and felsic parts. The number of these dikes and their importance in the evolution of the batholith warrants a separate description. Some of are nearly undeformed and fresh whereas others are moderately deformed. I saw no inhomogeneous dikes that were highly deformed, probably because at this state of deformation the textures described below would be unrecognizable.

Inhomogeneous dikes are black to light gray basaltic andesites to dacites that show: 1) intimate small scale intermigling of granitic and mafic material, 2) very felsic patches in a mafic matrix, and 3) 1-2cm alkali feldspar megacrysts similar to the surrounding granite. Typically, these dikes contain 3-50 cm rounded, pillow-like masses of mafic rock

Figure 4,Xenolith train from Lochsa mile 127



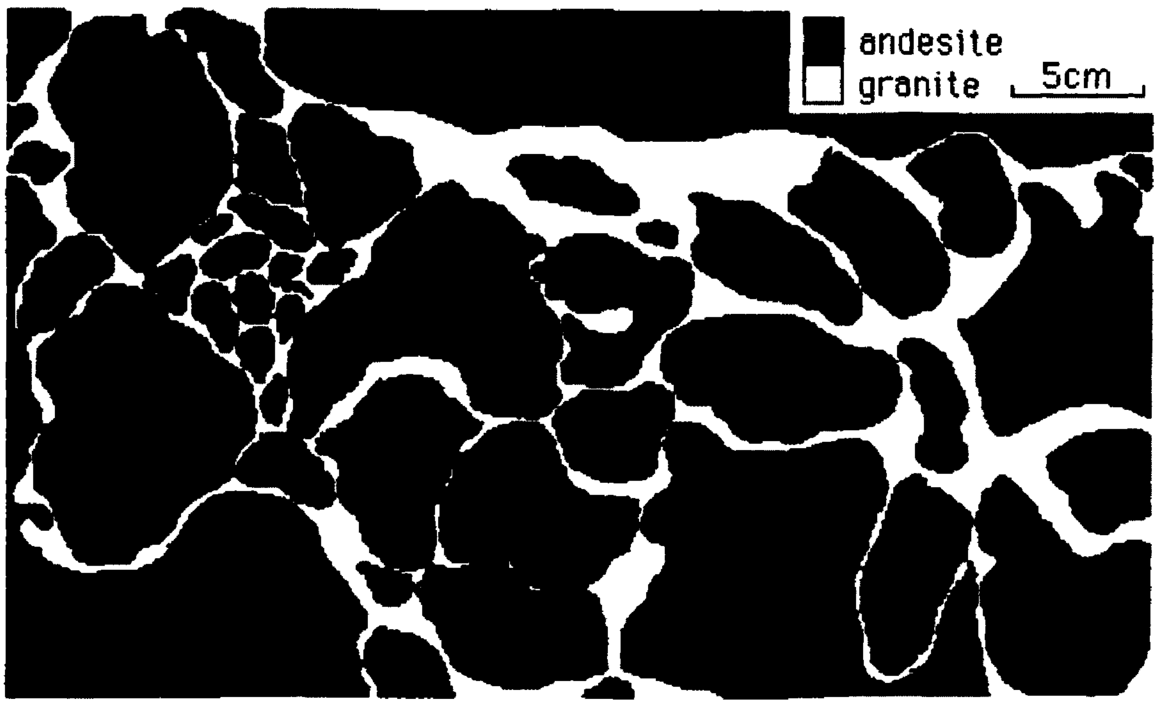


Figure 5
Sketch of a portion of an inhomogeneous dike that has pillows of andesite surrounded by fine grained leucogranite.

with 1-2 mm elongate mafic clots encircled by thin films (0.1-1 cm) of fine grained granitic material (figure 5), all in a matrix of andesite that contains 1-2 mm mafic clots. Some of these dikes appear streaky due to the alignment of elongate felsic pods of various sizes. Some felsic streaks and pods are up to 40 cm in diameter and contain chilled pillows of mafic rock. The fine grained more felsic dikes of this suite, granodiorite in composition, contain 1-5cm round inclusions of basalt.

The inhomogeneous dikes commonly occur in complexes with rocks more mafic than ordinary Idaho batholith granite (quartz diorite to granodiorite). One of the complexes is about 1/2 km long. Examples include complexes along U.S. 12 at miles 139.3 and 134.2, at Elk Summit, and along Bear Creek. These complexes are extremely inhomogeneous with the predominate rock type a mafic hornblende biotite quartz diorite to granodiorite with round inclusions of very fine grained andesite or basalt. In some of these complexes I saw dikes that are tabular low in the outcrop, but spread upwards or laterally into various sized mafic pillow-form blobs and schlieren in a matrix of mafic rich granitic rock (figure 6), as though the mafic magma injected almost solid granite low in the outcrop and partially molten granite above. I describe two typical complexes, one from mile 139.3, and another from mile 134.3.

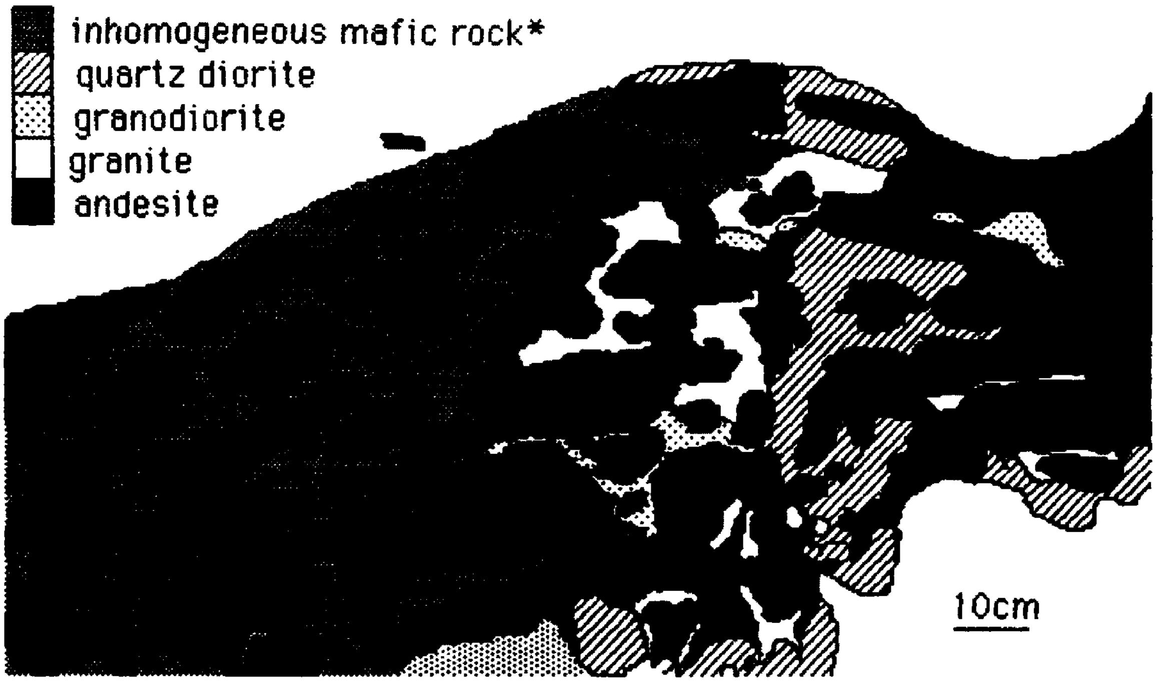


Figure 6
Sketch of a dike that grades into mafic inclusions and schlieren from Lochsa mile 124.95.

*a very inhomogeneous schlieren-rich, fine grained rock of granodiorite to quartzdiorite composition with diffuse blocks and pillows of andesite and streaks of granitic rock.

The complex at mile 139.3 consists of inhomogeneous mafic rich granodiorite cut by discontinous sheets and dikes of andesite up to 10m long and 2.0m wide, and granitic pegmatite. In this inhomogeneous mafic rich granitoid there are: 1) numerous angular and rounded inclusions of andesite, 2) areas of irregular granitic pegmatite, pods of andesite up to 2m in diameter, one of which is cut by an andesite dike, 3) "marble cake-like" swirling of granodiorite and diorite with patches of granite and andesite, 4) a fine grained mafic rich granodiorite dike which cuts an andesite dike, and 5) a large bulbous mass of andesite high in the outcrop which is fed by sheets of andesite (figure 7).

The complex at mile 134.3 includes a dike that curves and steepens higher in the outcrop where it spreads out and vanishes in a sill-like sheet. This sheet exibits large scale comingling of the dike rock and granitic rock including granitic pegmatite swirled in with andesite. Ptygmatic pegmatites which pinch and swell are extremely contorted and range in size from 1-50cm. Some of these pegmatites and aplites seem to surround "pillows" of the mafic rock. Surrounding this contaminated sheet is an area of hornblende biotite quartz diorite which contains numerous rounded mafic inclusions in various stages of resorption.

INTRUSIVE RELATIONSHIPS OF SYNPLUTONIC DIKES

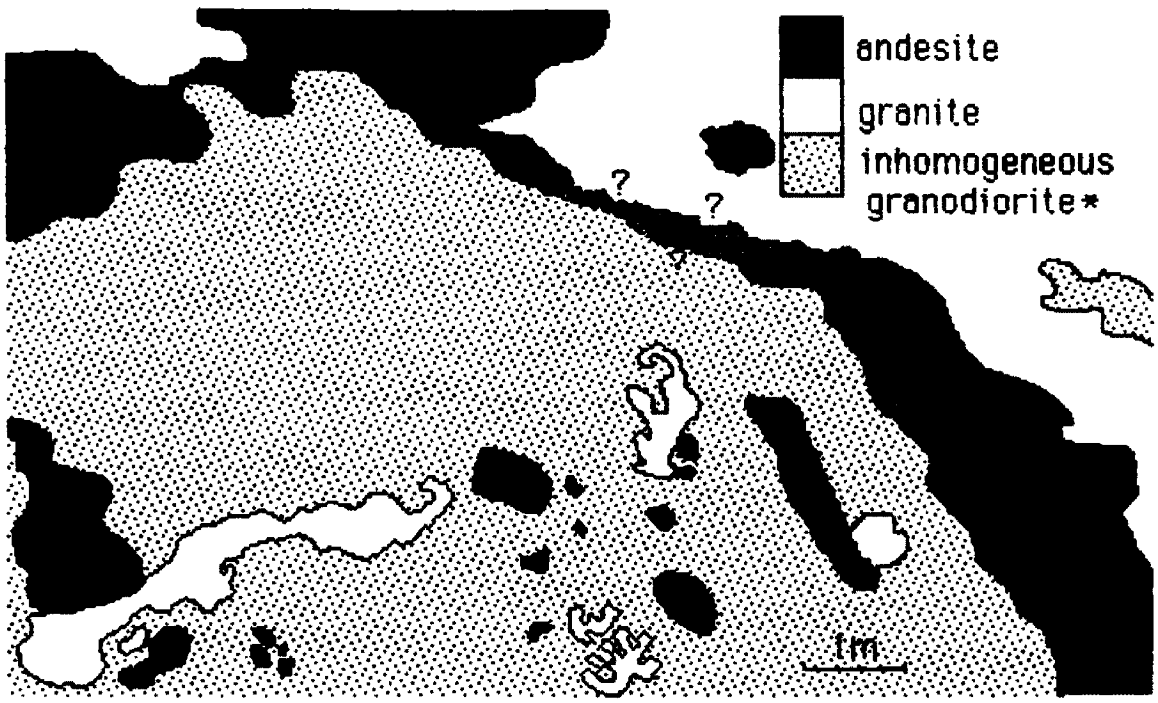


Figure 7
Generalized sketch of a dike complex from Lochsa mile 139.3.
*the inhomogeneous granodiorite contains andesitic and granitic portions, including marble-cake like swirling of granite and granodiorite

Various intrusive relationships suggest that these dikes were emplaced at the same time as their surrounding granitic rock. Examples of mutually intrusive relationships include:

1) Foliations that are oblique or subparallel to the walls of the dike, a foliation imposed by post-emplacement movements of the granite. Other deformational textures imposed by the host pluton include: lineations, crenulated foliations, and mylonitic textures.

2) Mafic dikes cut by tabular or ptygmatic granitic pegmatite and aplite dikes of the host granite.

4) Trains of segmented and dismembered dikes in undeformed granite.

5) Dikes that grade into mafic pillow shaped globs or mafic schlieren where they enter a more felsic, in some cases pegmatitic and presumably later crystallized part of the pluton. A few dikes grade to pillows then to schlieren.

6) Dikes metamorphosed by the heat of the host granite indicating the dikes were intruded before the pluton cooled.

7) Isolated angular to rounded mafic xenoliths, xenolith trains, and mafic schlieren of rock very similar to the dike rock may be early intruded mafic rock.

8) Small scale intermingling of aplite and andesite in inhomogeneous

dikes, some of which contain pillow shaped blobs of basalt within thin aplite rims.

9) Mafic dikes which contain late alkali feldspar megacrysts like those in the granite.

10) The granite near some of the dikes is more mafic than normal Idaho batholith granite, suggesting contamination by the dike material.

11) Dikes that are folded and boudinaged in non-foliated, seemingly undeformed granite.

12) Intermediate dikes with rounded and angular inclusions of granite and basalt.

13) A wide range in degree of metamorphism and deformation suggests a range in times of intrusion.

II. 2 PETROGRAPHY

Synplutonic dikes in the Idaho batholith include numerous rock types, textures, and mineral assemblages that show a wide range in degree of metamorphism, recrystallization and deformation. These differences may be grouped into three loosely defined sets: 1) dikes that show only minor metamorphism and have near igneous textures, 2) dikes that are recrystallized but are not metamorphosed beyond recognition, and 3) dikes that are completely recrystallized, are highly sheared and retain no primary igneous textures.

The least metamorphosed dikes are very fine to fine grained, massive, and commonly porphyritic with plagioclase and mafic phenocrysts. These probably are near original rock types including basaltic andesite, andesite, dacite, and lesser, medium grained, massive dikes of hornblende- biotite quartz diorite and biotite granodiorite. These dikes generally contain zoned plagioclase (An_{75-30}) , hornblende, biotite, with or without augite, pigeonite or quartz. Accesory minerals include apatite, sphene, zircon, and iron oxides.

These dikes are slightly metamorphosed and grain boundaries are slightly ragged and cuspate. The metamorphism mainly involves recrystallization of primary mafic minerals and some breakdown of plagioclase. Primary plagioiclase has fine granular plagioclase of lower An content around the boundaries of once euhedral laths, and strong alteration of phenocryst cores to sericite and epidote. Augite phenocrysts are partly to fully recrystallized to fine grained fibrous actinolite, hornblende, biotite, epidote, and sphene. Much of the primary hornblende has recrystallized to form finer grained hornblende, fine grained actinolite, fine grained biotite, and epidote. Some biotite is primary and may be intergrown with actinolite. A late low-temperature alteration is characterized by plagioclase altered to epidote and sericite, and biotite and hornblende partially replaced by chlorite.

Many of these nearly fresh dikes contain gabbroic inclusions 1 to 2 cm in diameter and clumps of crystals. The clumps contain medium grained euhedral plagioclase and augite which are out of equilibrium with the groundmass of the dikes. The plagioclase is usually irregularly zoned (eg. from core to rim: An_{45} to An_{28} to An_{70}), and considerably altered to sericite, epidote, and possible clay minerals. The augite is usually rimmed or altered to actinolite and biotite.

Dikes of group 2 are the most common. They lack primary igneous textures but are not sheared beyond recognition. They are lightly or moderately to strongly foliated or lineated by are not penetratively sheared and altered. These dikes tend to be slightly more felsic than dikes of group 1, rock types include hornblende andesite, and dacite. Most are fine grained and consist of dominantly subhedral and anhedral grains. 1 to 2 mm plagioclase phenocrysts may still be recognizable and retain a lath shape but the groundmass tends to be recystallized to very fine grained granular grains. Lightly zoned plagioclase (An_{50-25} depending on rock type) is the most common felsic mineral, quartz is common, alkali feldspar is rare. In these dikes augite and primary hornblende are fully recrystallized. Where phenocrysts of augite once existed there are now aggregates of interlocking fine grained actinolite, hornblende, biotite, epidote, sphene, and chlorite. Accessory minerals include: apatite, epidote, sphene, allanite, zircon, and iron oxides. Most of the phases are anhedral, interlocking, and granular to elongate.

The more highly metamorphosed dikes of group 3 are generally very fine grained, schistose, and have granular textures. These dikes tend to be the most felsic with higher quartz contents and less biotite and hornblende. The plagioclase is generally of lower An content (An_{25-12}), the pyroxene has disappeared, hornblende is recrystallized to finer grained hornblende, actinolite, and biotite; chlorite and epidote are also common. Quartz and alkali feldspar may have incresed in abundance. Accessory minerals include: apatite,sphene, tourmaline, calcite, sericite, and iron oxides. Group 3 dikes show schistosity defined by biotite, hornblende or actinolite, and lenses of quartz and plagioclase. The schistosity is commonly crenulated. Large moderately zoned plagioclase grains with crenulate boundaries may have been original plagioclase phenocrysts. Many of these dikes are penetratively cleaved, with chlorite and epidote along cleavage. Some show mylonitic textures with well developed S and C surfaces.

A subset of the fresh and moderately deformed dikes are those inhomogeneous dikes described in the previous section. Most of these are fine grained, massive or lightly foliated or lineated, inhomogeneous andesites. Intruding streaks and dikelets of granitic material are commonly aplitic. These aplites are commonly 1- 2mm to 2cm in size, but some are as much as 50 cm. They are typically ptygmatic, and surround blobs or pillows of andesite at all scales. From the centers of the veinlets to the andesite there is a full spectrum of rock types intermediate between granite and andesite.

Granitic streaks and dikelets are generally very fine grained, granular, and massive. They contain slighly zoned plagioclase (An_{38-18}), which is anhedral to blocky, and lesser anhedral quartz, with or without alkali feldspar. Myrmekitic and graphic textures exist in some. Mafic minerals include biotite and hornblende. The color index ranges from 1 to 15, increasing toward the borders of the veinlets. Accessory minerals include epidote, iron oxides, sphene, apatite, and zircon. The mafic portions of these dikes are typical of those described above except that the color index may be slighly higher than other similar dikes (C.I. of 45-65). The hornblende to biotite ratios are higher away from the contacts with the aplite veinlets (10:1) and lower near the aplites (C.I. 4:1 to 1:1).

Another important rock type is the hornblende biotite quartz diorites and granodiorite of the mafic complexes described in the previous section. These rocks are very inhomogeneous, typically foliated, medium grained, equigranular, and hypidiomorphic. Their major minerals include plagioclase, quartz, alkali feldspar, biotite and hornblende. The plagioclase tends to be blocky, wheras quartz and alkali feldspar are granular and interstitial. Accessory phases include epidote, iron oxides, chlorite, apatite, zircon, sphene, and allanite. The distribution of grains, color index and grain size is highly variable, with more dioritic portions tending to have a finer grain size, a higher color index and more hornblende then biotite. The opposite is true for the more granodioritic portions. Most of these rocks are deformed as evidenced by broken plagioclase euhedra, bent plagioclase twins, granulated quartz and alkali feldspar, bent biotite books, and strong undulose extinction in most phases.

Microscopic disequilbrium textures exist in many dikes, these include:

1) Strongly zoned plagioclase phenocrysts with vastly dissimilar zoning histories in the same thin section. Zoning types include normal, reverse, and oscillatory. Cores of the strongly zoned plagioclase are strongly altered to sericite and epidote, suggesting they are out of equlibrium.

2) Strongly altered clumps of euhedral phenocrysts of both gabbroic and granitic character in dikes of andesite composition.

3) Variable hornblende to biotite ratios in similar rock types.

4) Disequilbrium phenocrysts of augite and calcic plagioclase in felsic granodiorite.

The wide range in degree of recrystallization and deformation may be a function of residence time in the pluton. Metamorphism of the dikes after emplacement is the result of their remaining at the temperature of the pluton for a long time while the pluton cooled. During this time, the minerals re-equilibrated to an assemblage stable at a temperature near the temperature of the enclosing batholith. Williams (1977) indicated metamorphic assemblages in the mafic dikes represent metamorphism near the greenschist- amphibolite facies transition representing temperatures of 500 to 550^{0}C depending on pressure.

II. 3 CHEMICAL ANALYSIS

For chemical analysis of these rocks I sampled a suite from the most mafic dikes to the more felsic dikes. In addition to these I sampled some of the surrounding granitic rocks, both the mafic rich ganitiods of the dike complexes and the more normal granite in the areas of some of the dikes. Of these I sampled three suites in areas that showed field evidence for magma interaction. The samples were analysed in Peter Hooper's lab at Washington State University using X-ray fluorescence for major elements and at a cooperating lab using induction coupled plasma spectrometry for selected trace elements. Norms were calculated using "IGPET" computer software written by Michael J. Carr. A brief description of the analysed samples is given in appendix 1.

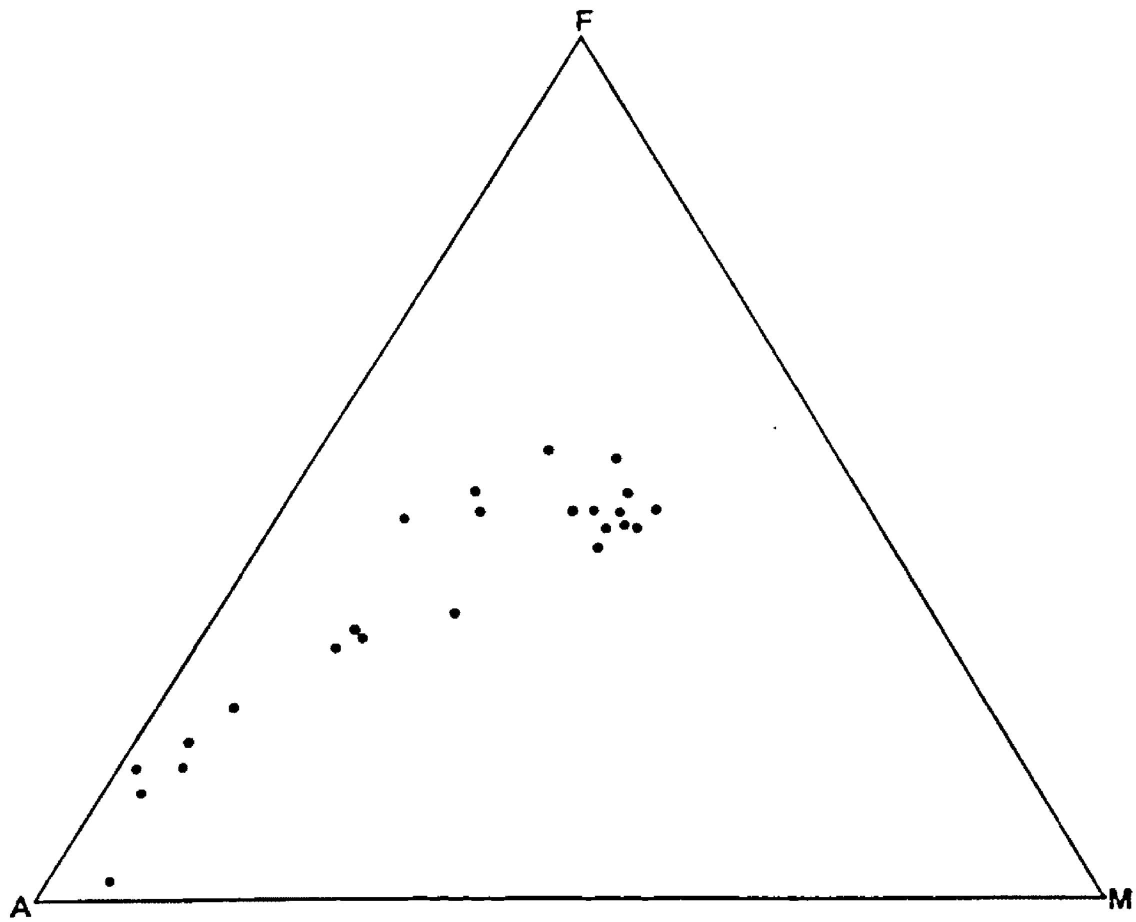
The most mafic dikes are similar in composition to basaltic andesite (table 2). The felsic dikes have compositions in the range of hornblende andesite or dacite. Silica contents of the dikes range from 53.61-72.22%. The basaltic andesites are hypersthene and quartz normative and slightly high alumina with Al2O3 ranging from 14.89-17.28%. They have high large ion lithophyle element contents for their SiO2 content. K2O contents are high, ranging from 1.39-2.78% with an average of 2.08%, this average is much higher than average basaltic andesite as given by Hyndman (1985a, p. 256) and thus the rocks should be called high-K basaltic andesite. Sr contents range from 477-681ppm with an average of 580ppm. The FeO*/MgO ratio is less than 2.25 at 57% SiO_2. A calc-alkaline differentiation trend is shown when rocks of this suite are plotted on an AFM diagram (figure 8). All of the above are characteristics of calc-alkaline series, arc volcanics.

The andesites and dacites resemble the basaltic andesites in that they are hypersthene and quartz normative, are fairly high alumina and have high Sr, K_2O and other large ion lithophyle element contents. The andesites and dacites are darker in hand sample than their SiO_2 contents would suggest.

Mafic granitiods sampled from the dike complexes (eg. no.'s 134.3C, 139.3B, 139.3A) are similar in composition to tonalite or quartz diorite, granodiorite, and granite. These rocks tend to have lower K_2O and higher Na_2O contents than rock of this suite with similar SiO_2 contents (e.g., compare no. 139.3A with the other granites).

The granites are similar to typical Idaho batholith compositions as found by Hyndman (1984) throughout this same cross-section. 139.3D may be a late stage alkali feldspar rich phase of the batholith as suggested by its very high SiO_2 and K_2O contents.

Figure 8
AFM plot of all samples.



Harker diagrams for the total suite from basaltic andesite to granite show good linear trends for major elements when plotted against SiO2 (figure 9). Element-element diagrams for trace elements for the whole suite do not show any visible trends, but rather are scattered over the graphs (figure 10).

The rare earth elements show relatively similar trends for all the rock types (figure 11). As a suite, these rocks are enriched in light rare earth elements, depleted in heavy rare earth elements. Andesites tend to be relatively less enriched in light rare earth elements than the granites, and conversely the granites tend to be more depleted in heavy rare earth elements than the andesites. Thus, there is a crossover trend in relative enrichment occurring around europium; this crossover shows up best in the heavy rare earth elements. It should be noted that the mafic rich granitoids lie in almost all cases between the andesites and granites.

TABLE 2, Chemical analysis of mafic dikes, intermediate rocks, and granites from the Idaho batholith, major elements in weight percent, trace elements in parts per million.*

	124.95B	141.95	136.05A	135.9A	125.05	136.05B	134.3A1	134.3A2	139.3C
SiO2	53.61	53.75	53.89	54.19	54.94	55.07	55.43	56.72	58.08
TiO2	0.91	1.36	0.82	2.28	0.93	0.82	0.75	0.66	0.98
Al2O3	14.89	17.28	15.88	15.21	15.77	15.92	16.19	16.49	17.29
Fe2O3	4.42	4.56	4.19	4.53	4.11	4.15	3.67	3.34	3.56
FeO	5.06	5.22	4.80	5.19	4.71	4.75	4.20	3.83	4.08
MnO	0.17	0.15	0.15	0.14	0.15	0.14	0.14	0.13	0.11
MgO	7.15	5.08	6.70	6.46	5.58	6.32	6.35	5.63	4.60
CaO	9.12	8.26	8.18	6.87	8.35	7.79	8.82	8.08	6.36
Na2O	2.62	2.61	2.56	2.64	2.21	2.45	2.17	2.39	3.19
K2O	1.62	1.39	2.49	1.94	2.78	2.28	2.04	2.49	1.57
P2O5	0.440	0.334	0.346	0.545	0.457	0.324	0.244	0.249	0.185
Ba				645		673	778	957	535
Rb				56		85			56
Co				38		35	35	35	35
Cr				211		188	307	247	134
Cu				54		83	19	21	42
Li				24		32	12	13	39
Nb				42		10	15	13	13
Ni				146		36	25	25	44
Sc				17		22	28	22	16
Sr				681		584	477	529	536
V				152		155	194	147	102
Y				23		17	16	16	17
Zn				97		71	73	65	80
Zr				106		21	21	20	17
Norms									
AN	52.07	58.65	53.89	52.00	57.11	55.36	60.74	57.10	51.11
Q	5.13	8.78	3.97	8.98	7.63	7.32	8.61	9.22	12.58
or	9.57	8.21	14.72	11.46	16.43	13.47	12.06	14.72	9.28
ab	22.17	22.09	21.66	22.34	18.70	20.73	18.36	20.22	26.99
an	24.09	31.33	24.49	23.92	24.90	25.71	28.41	26.91	28.22
C	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
di	14.57	5.97	11.03	5.21	10.81	8.63	10.94	9.19	1.69
hy	15.33	13.65	15.71	15.92	12.73	15.83	14.32	13.07	13.79
mt	6.41	6.61	6.08	6.57	5.96	6.02	5.32	4.84	5.16
il	1.73	2.58	1.56	1.26	1.77	1.56	1.42	1.25	1.86
ap	1.02	0.77	0.80	0.00	1.06	0.75	0.57	0.58	0.43
FeO*	9.04	9.32	8.57	9.27	8.41	8.49	7.50	6.84	7.28
Rare earth elements									
La				37.98		13.41	14.38	15.94	18.60
Ce				79.10		29.67	28.29	30.49	35.68
Pr				9.01		4.00	3.58	3.66	6.30
Nd				37.15		17.87	14.34	14.38	16.89
Sm				7.42		4.07	3.22	3.09	3.52
Eu				2.27		1.30	1.08	1.03	1.14
Gd				6.38		3.53	3.13	2.98	3.26
Dy				4.57		2.72	2.67	2.50	2.85
Ho				0.82		0.52	0.53	0.51	0.56
Er				2.31		1.51	1.43	1.35	1.64
Yb				1.77		1.34	1.48	1.44	1.54
Lu				0.24		0.20	0.25	0.25	0.22

	132.9	134.3C	116.7	116.0	124.0	124.95A	118.6	139.38	135.0
SiO2	58.33	59.37	60.43	61.32	64.82	68.22	68.24	68.36	68.72
TiO2	0.82	0.89	1.11	1.23	1.17	0.52	0.57	0.53	0.58
Al2O3	15.94	17.15	17.32	16.85	15.88	15.84	16.05	16.65	16.08
Fe2O3	3.47	3.95	3.14	3.34	2.87	1.77	1.57	1.55	1.49
FeO	3.98	4.52	3.60	3.82	3.29	2.02	1.80	1.78	1.71
MnO	0.13	0.13	0.11	0.08	0.08	0.06	0.05	0.04	0.05
MgO	5.49	3.32	2.79	2.56	1.71	2.55	1.70	1.50	1.49
CaO	7.17	6.07	5.71	4.88	3.79	3.62	3.61	3.45	3.27
Na2O	2.25	2.45	2.94	2.91	2.96	3.32	2.68	3.20	2.72
K2O	2.18	1.93	2.56	2.69	3.11	1.95	3.57	2.77	3.71
P2O5	0.247	0.247	0.285	0.331	0.323	0.131	0.174	0.176	0.187
Ba		1002	513	944			619	1148	
Rb			89	96			147	94	
Co		35	29	23			33	37	
Cr		78	42	32			60	71	
Cu		33	25	23			17	14	
Li		21	21	31			27	38	
Nb		19	16	19			13	11	
Ni		26	12	16			18	22	
Sc		17	12	10			7	6	
Sr		571	404	616			320	586	
V		147	71	114			49	46	
Y		25	18	17			10	9	
Zn		92	73	75			58	70	
Zr		7	63	90			45	6	
Norms									
AN	58.61	57.89	51.55	47.24	39.99	37.84	42.52	37.09	39.46
Q	14.04	18.74	16.85	19.68	25.08	29.57	28.68	29.47	29.52
or	12.88	11.41	15.13	15.90	18.38	11.52	21.10	16.37	21.93
ab	19.04	20.73	24.88	24.62	25.05	28.09	22.68	27.08	23.02
an	26.96	28.50	26.47	22.05	16.69	17.10	16.77	15.97	15.00
C	0.00	0.59	0.01	1.07	1.53	2.00	1.63	2.54	2.09
di	5.57	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
hy	14.35	12.08	9.34	8.75	6.15	7.85	5.39	4.92	4.75
mt	5.03	5.73	4.55	4.84	4.16	2.57	2.28	2.25	2.16
il	1.56	1.69	2.11	2.34	2.22	0.99	1.08	1.01	1.10
ap	0.57	0.57	0.66	0.77	0.75	0.30	0.40	0.41	0.43
FeO*	7.10	8.08	6.43	6.83	5.87	3.61	3.21	3.18	3.05
Rare earth elements									
La		30.92	29.05	39.84			38.72	34.42	
Ce		58.63	57.80	77.69			72.81	61.85	
Pr		6.62	6.56	8.60			7.67	6.30	
Nd		27.20	26.45	32.64			28.52	23.18	
Sm		5.50	5.25	5.58			4.75	3.72	
Eu		1.52	1.46	1.64			1.09	1.00	
Gd		4.86	4.50	4.38			3.28	2.60	
Dy		3.91	3.55	3.10			2.05	1.59	
Ho		0.76	0.68	0.60			0.39	0.30	
Er		2.10	2.00	1.72			1.20	0.94	
Yb		2.07	1.72	1.40			0.96	0.74	
Lu		0.35	0.25	0.19			0.13	0.10	

	139.3A	135.9B	134.38	125.0	134.15	139.3D
SiO2	71.18	72.22	72.79	72.80	73.38	75.09
TiO2	0.21	0.47	0.24	0.22	0.26	0.15
Al2O3	17.70	15.14	16.39	15.57	15.29	15.01
Fe2O3	0.49	1.02	0.58	0.79	0.63	0.24
FeO	0.56	1.17	0.67	0.91	0.72	0.28
MnO	0.01	0.03	0.02	0.03	0.02	0.00
MgO	0.28	0.70	0.21	0.43	0.57	0.20
CaO	2.42	1.96	2.15	1.98	1.97	1.17
Na2O	4.78	2.81	3.63	2.59	2.80	2.49
K2O	2.31	4.35	3.27	4.61	4.25	5.27
P2O5	0.054	0.146	0.060	0.070	0.090	0.084
Ba	610	975	1477			1406
Rb	57	179				109
Co	35	26	29			45
Cr	31	33	32			17
Cu	11	13	8			7
Li	23	33	12			15
Nb	9	15	10			7
Ni	7	11	8			7
Sc	2	4	2			2
Sr	932	352	1050			645
V	12	26	17			9
Y	6	10	5			6
Zn	42	60	45			35
Zr	4	49	4			4
Norms						
AN	22.37	26.95	25.07	29.94	27.94	19.96
Q	28.95	34.13	34.23	35.08	35.82	37.84
or	13.65	25.71	19.32	27.24	25.12	31.14
ab	40.45	23.78	30.72	21.92	23.69	21.07
an	11.65	8.77	10.27	9.37	9.19	5.26
C	3.07	2.60	3.11	2.89	2.72	3.28
di	0.00	0.00	0.00	0.00	0.00	0.00
hy	0.99	2.33	0.92	1.78	1.83	0.57
mt	0.71	1.48	0.84	1.15	0.91	0.35
il	0.40	0.89	0.46	0.42	0.49	0.28
ap	0.13	0.34	0.14	0.16	0.21	0.19
FeO*	1.00	2.09	1.19	1.62	1.29	0.50
Rare earth elements						
La	27.08	36.48	45.96			19.75
Ce	50.21	69.91	81.34			37.20
Pr	5.58	7.33	8.19			4.18
Nd	19.21	27.19	28.43			14.84
Sm	3.19	4.60	3.84			2.38
Eu	0.85	0.95	1.00			0.73
Gd	2.08	3.13	2.35			1.70
Dy	1.55	1.90	1.26			1.23
Ho	0.33	0.35	0.26			0.26
Er	1.22	1.05	0.78			0.79
Yb	0.96	0.81	0.75			0.72
Lu	0.13	0.10	0.15			0.09

* sample numbers are keyed to mile postes along U.S. 12.

Figure 9
Major element SiO_2 variation diagrams.

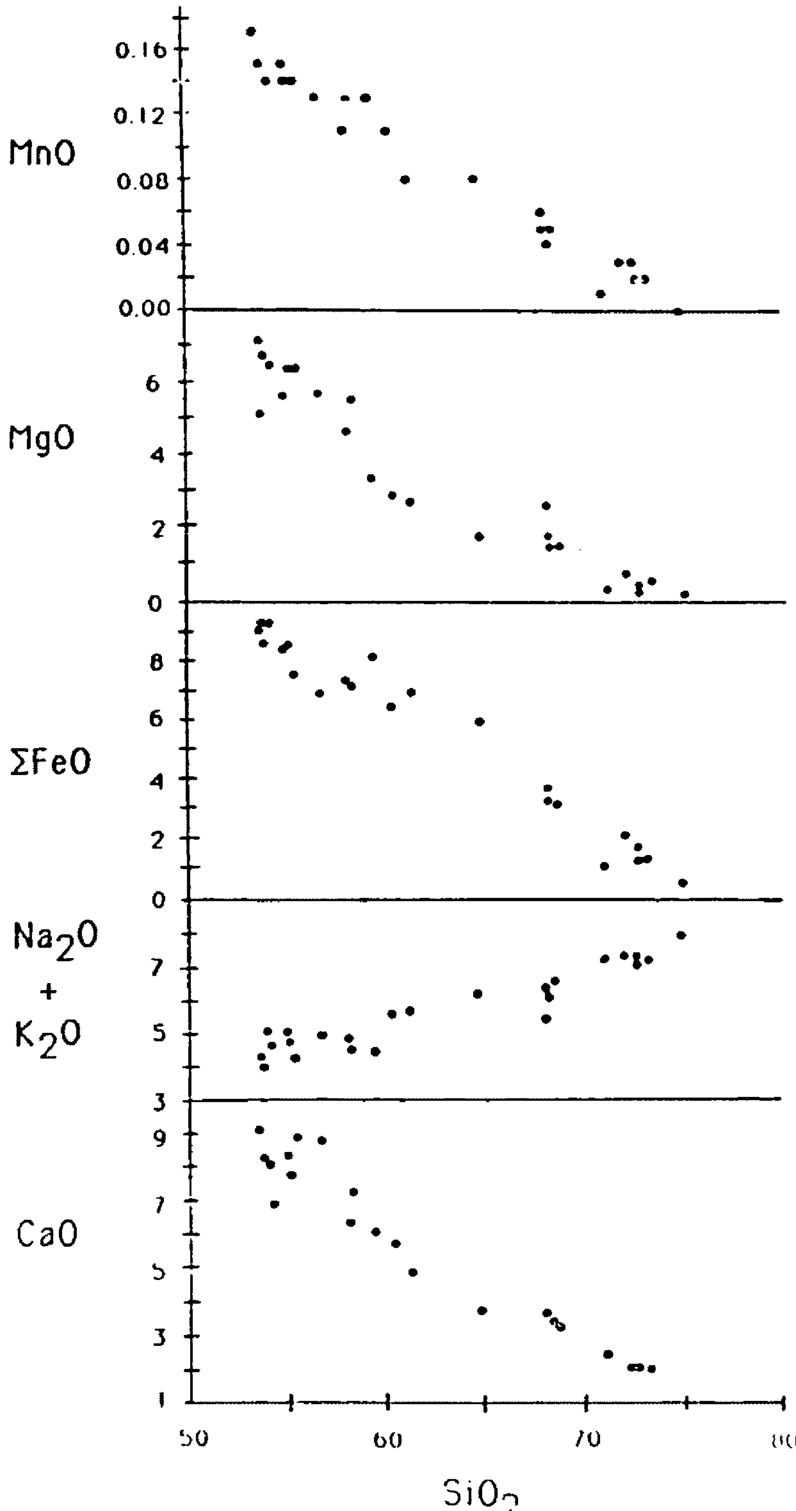
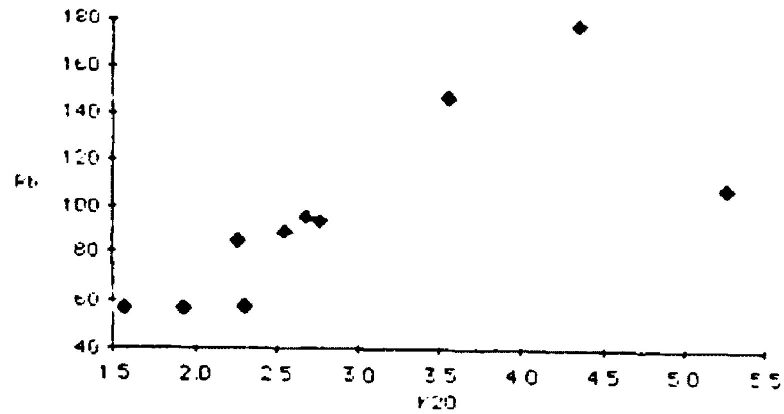
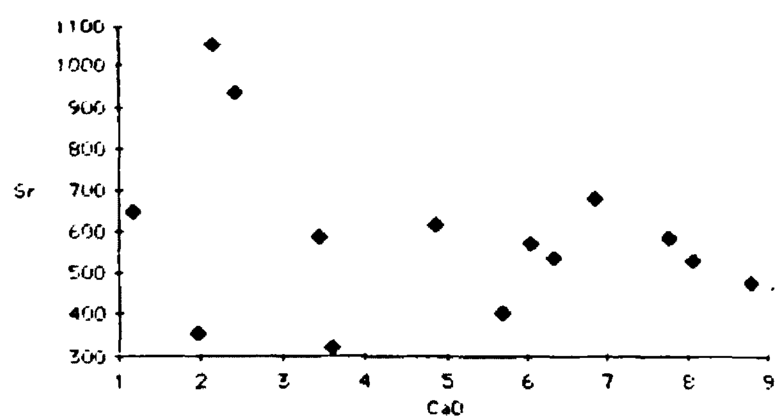


Figure 10
Trace element variation diagrams





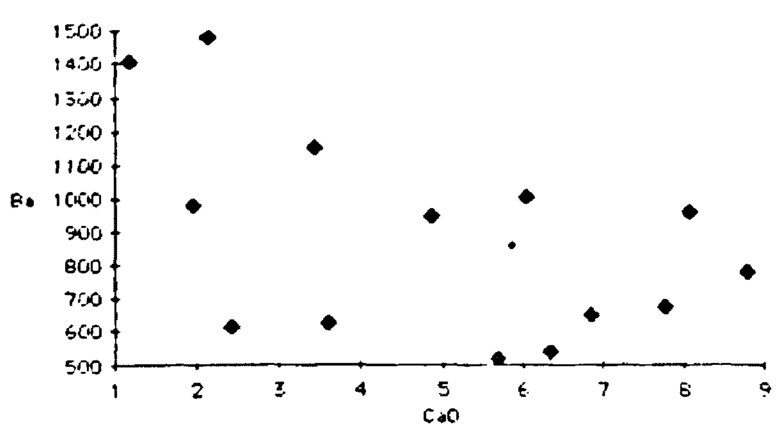
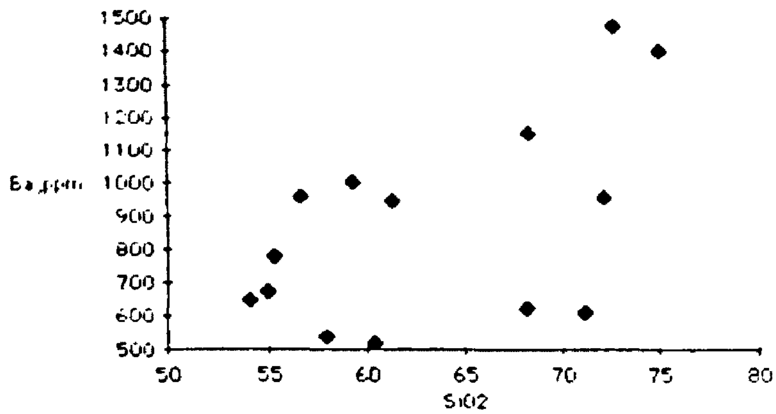
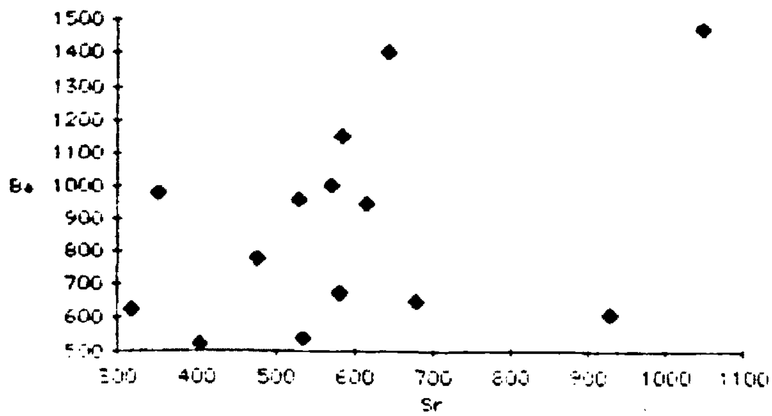


Fig. 10, cont.


1500
1400
1300
1200
1100
Ba ppm
1000
900
800
700
600
500
50
55
60
65
70
75
80
SiO2


1500
1400
1300
1200
1100
Ba
1000
900
800
700
600
400
500
600
700
1000
1100
Sr

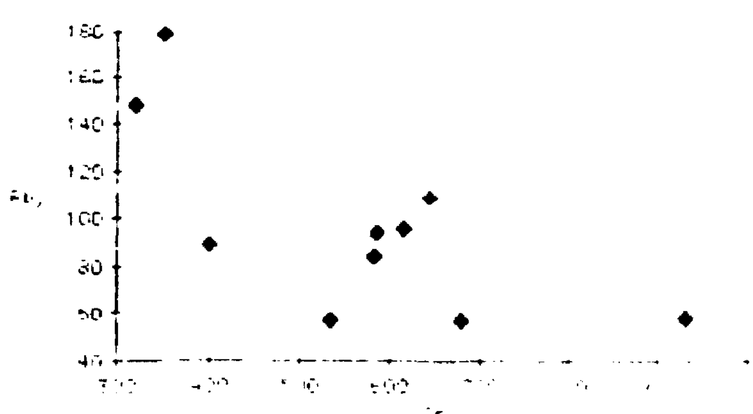
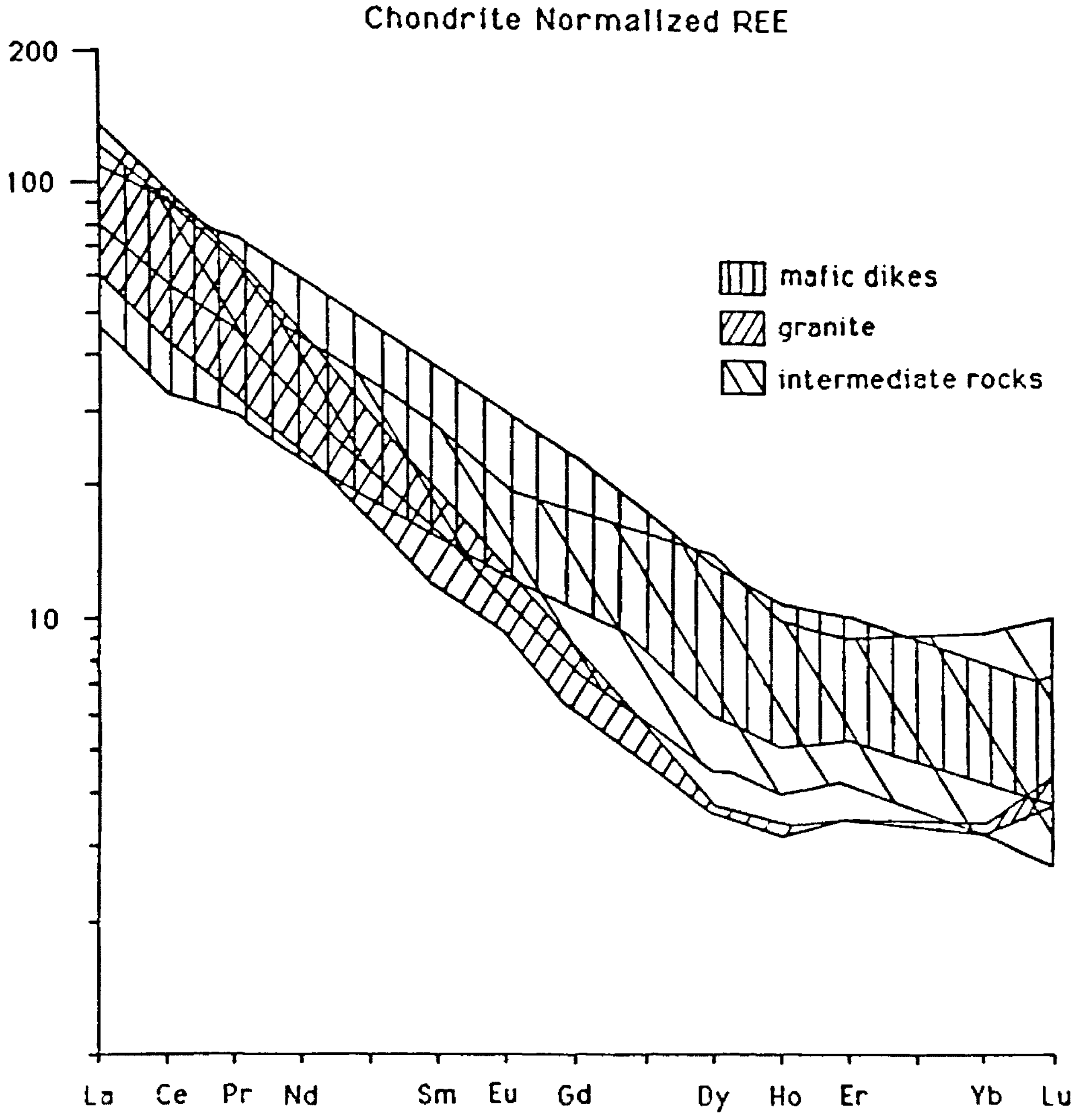

140
100
80

Figure 11
Chondrite normalized REE plot of all samples.



III.1a SYNPLUTONIC RELATIONSHIPS AND TIMING OF MAFIC MAGAMA INJECTION

Before I discuss the importance of the dikes in the history of the batholith I show that the dikes were present when the batholith was partly liquid. The mutually intrusive relationships described in detail in section II-1 demonstrate that the mafic dikes were injected into the batholith when it was still partly liquid. In addition, although the mafic dike-pegmatite or aplite contacts appear sharp, in detail, interlocking crystals bridge the contact, and many lie across the contact. Metamorphism of the dikes by the host granite, also suggests the dikes were injected while the batholith was still hot and mobile.

Most of the dikes intruded along fractures; evidently they were intruded late in the development of the batholith when the proportion of crystals to melt in the granite was large enough to sustain a fracture. Pitcher (1979) noted that magma with a higher crystal content will fracture if a rapidly applied point force is directed against it. Since dikes must be injected rapidly or they will solidify (Delaney and Pollard, 1982), fast injection of mafic magma may open the fracture the dikes intruded.

Several geologists have attempted to calculate the percent crystals need to allow fracturing of a magma and emplacement of dikes (e.g., Furman and Spera, 1985; Hibbard and Watters, 1985). They conclude that a magma must be at least 70% crystalline for mafic magma to inject as dikes. At 70% crystals the system is capable of transmitting compressive and shear stresses via a crystalline framework.

Using viscous flow laws Furman and Spera (1985) calculated that mafic magma injected into a granitic pluton that is less than 70% crystalline will rapidly disperse the quenched mafic inclusions even with very slow convection. Table 4 of Furman and Spera (1985) (table 3) explains very well what will become of mafic magma injected into a partly molten granitic pluton. This suggests that in volcanic arc environments where granitic magmatism is open to basalt injection from the mantle, a continuum of disequilbrium features can be expected from hybrid magmas to mafic dikes. Thus, a pluton that contains late synplutonic mafic dikes should also show evidence that mafic magma was emplaced before it was injected as dikes. In the Idaho batholith I saw the full range of characteristics described in table 3. I believe that the hybrid magmas (quartz dioritic) of some of the dike complexes and some of the mafic schlieren , inclusions, and inclusion trains, especially those that grade into dikes, are evidence of this early mafic magma.

Table 3
Generalized magma mixing in a closed-system intermediate-composition pluton (from Furman and Spera, 1985).

Time after intrusion (yr)	Convective velocities	Host magma crystallinity (vol. percent)	Spatial Scale (m)	Physical features
0	5 km yr^{-1}	0-10	$<10^{-3}$, microscopic	isolated xenocrysts, multiple melt inclusion population, melt immiscibility
		10-30	10^{-3} to 10^{-2}	wispy layering, multiple phase clots, colonial xenocrysts, salic or mafic mineral heterogeneity
		30-50	10^{-2} to 10^{-1}	mesoscopic schlieren, mafic inclusions
		50-70	10^{-1} to 1	long wavelength schlieren, large mafic inclusions, small swarms
10^6	10 cm yr^{-1}	>70	>1, outcrop	dike (inclusion) trains, large inclusion swarms, outcrop scale schlieren bands, late-stage dike swarms, composite dikes

III. 1b STRUCTURE, ORIENTATON AND INJECTION OF DIKES

It is a matter of some controversy whether dikes and sills are injected along preexisting joints or whether pristine rock is fractured during the injection process. There is extensive evidence that the direction of injection of most dike swarms has a preferred orientation. This preferred orientation could be associated with a preexisting set of joints, in which case the dikes would have a trend parallel to the regional joint set. Or, the preexisting state of stress could influence the direction of new fractures. In this case, injection would occur perpendicular to the least principal stress (Spence and Turcotte, 1985). Hales (1982) notes that directions of dikes intruded at greater depths, such as those of the Idaho batholith, tend to be influenced more by regional stress fields.

Numerous questions arise as to why the dikes are intruded along parallel fractures and why the ENE-WSW trend is predominant. The strong parallelism of the dikes indicates extension perpendicular to ENE-WSW. Hamilton (1983) believes that major continental margin batholiths are intruded in extensional environments. Tobish and others (1986) believe that the structures in and around the Sierra Nevada batholith are best explained in an extensional model, with extension produced by subduction zone heating, magmatism and emplacement of the batholith. There are numerous possibilities for the extensional environment associated with the Idaho batholith; those included below I believe are the most logical.

1) Diapiric rise of the batholith may promote parallel shear zones and joints that could be intruded by dikes. Additional diapiric movement may be concentrated along dike boundaries and inside the dikes because of rheology differences. The oblique asymmetric foliation may have been caused by movements such as these. Also, the dikes in the center are

near vertical and the dikes on the borders are subhorizontal, especially on the East side, giving a mushroom shape dome.

2) Shrinkage of the batholith due to cooling may cause parallel fractures, but this would probably produce numerous directions unless it was coupled with a prevailing tectonic environment.

3) Subduction at the time of formation of the batholith was most likely oblique, directed NE. Oblique subduction may have caused an extensional environment between 75 and 65 m.y. According to Sears (1985, personal communication) there seems to be a period of quiesence in the thrusting between 75 m.y. and 65 m.y. during which a different tectonic environment may have existed. Oblique subduction may have been accompanied by northward strike-slip motion which could form E-W extentional fractures (e.g., Young, and others, 1985). (figure 12a)

4) With ENE directed oblique subduction the greatest stress, sigma 1, would be ENE-WSW, and the least stress, sigma 3, would be NNW-SSE thus, dikes would be injected ENE-WSW (Spence and Turcotte, 1985) (figure 12b).

5) Extension caused by subduction zone heating, magmatism and emplacement of the batholith (e.g., Tobish and others (1986).

Since the dike magmas were subduction generated, and thus

Figure 12a
Cartoon of oblique subduction produced strike-slip movement causing fractures in the Idaho batholith.

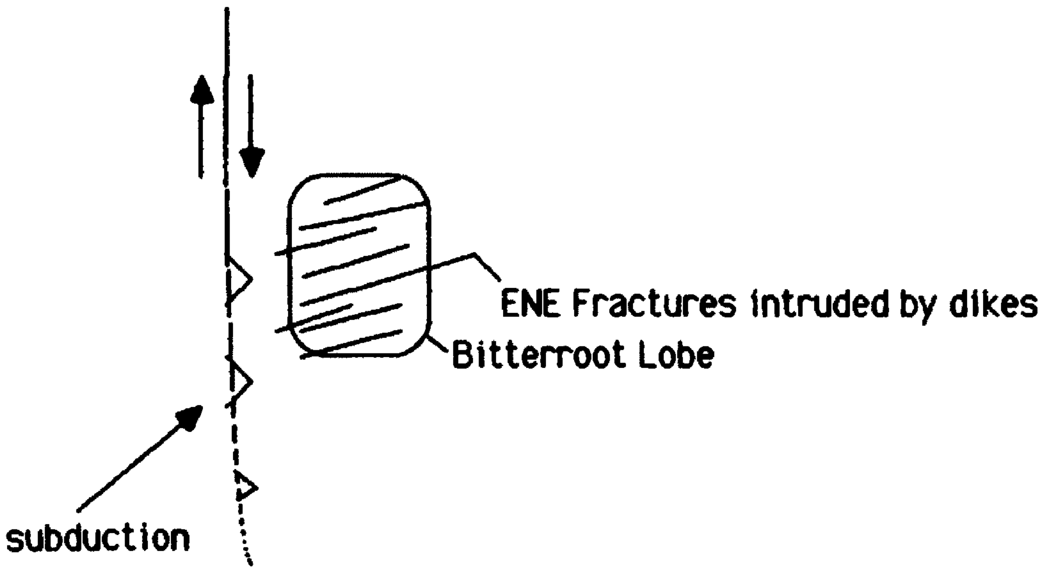
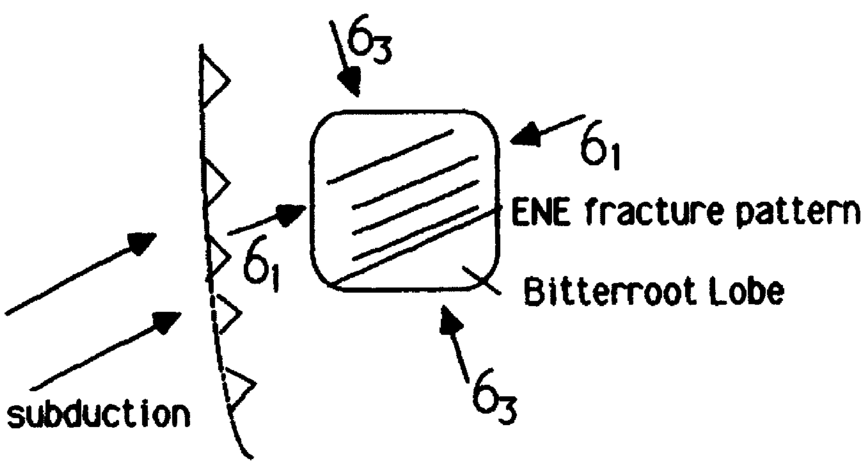


Figure 12b
Cartoon of possible stresses due to oblique subduction



produced by plate interactions, it seems probable that crustal scale stress regimes would control dike injection and orientation. This might be especially true in a partly molten batholith where preexisting fractures may not have existed before the injection of the dikes. Of this list of suggestions I think that models 3, 4 and 5 are the most plausible. Models 3 and 4 accurately describe the orientation of the dikes as well as their parallelism and fit best with plausible plate tectonics active at the time of intrusion. I would suggest that the dike injection was limited to the batholith because of the dike's injection above a restricted zone of mafic magma generation at a depth of 100-150 km over a subducting slab (Hyndman, 1985a).

A possible problem with any of these thoughts is that on the west side of the cross-section the dikes trend N-S rather than ENE-WSW. Hales (1982) notes that where two dike trends are present in a swarm they are commonly orthogonal to one another or within about 30 degrees of orthogonality.

III. 2 MIXING OF MAGMA

This section discusses the evidence for mechanical magma mixing between end members of basaltic andesite and granite. This evidence is found at all scales, outcrop, hand sample, thin section, and chemical composition. Mixing in and around the dikes was dominated by mechanical mixing, that in some cases was efficient enough to allow diffusional processes to homogenize the liquids.

Vogel (1982) explains that in mixing systems, liquid homogenization is a function of the efficiency of mechanical mixing whereas the presence of disequilibrium phenocrysts is a function of the residence time in the hybrid magma. With enough time even the disequilibrium phenocrysts can be dissolved. According to Furman and Spera (1985) the efficiency of magma mixing, as quantified by the spatial scale of heterogeneity, is dependent on both the vigor of convection and the viscosity ratio of the mixing components. Recent results on magma chamber convection theory (e.g., Spera and others, 1982) suggest convection rates are quite large so that map scale evidence for magma mixing should be very rare. Only mixing that occurred when a pluton was nearly soldified, and thus had slow convection, will be preserved as outcrop scale evidence. Thus, the dikes in the Idaho batholith represent a good area to study mixing because they were injected late in the history of the bathlith. Any evidence for mafic magma that intruded before the batholith was able to fracture is presumably erased by internal convection (table 3).

Kouchi and Sungawa (1985) have experimentally determined with forced convection that: 1) basaltic and dacitic melts can be easily mixed by forced convection forming a homogeneous andesitic melt and a banded texture in a dacitic melt, 2) basaltic melts change their composition more readily to andesite than dacitic melts, 3) this is because the convective movement preferentially occurs in the more fluid basalt. This work suggests that mechanical magma mixing is realistic even though diffusion across an interface is slow (e.g., Watson, 1982).

FIELD EVIDENCE

Field evidence for mixing is best portrayed in the inhomogenous dikes where pillow shaped blobs of andesite are surrounded by thin rims of leucogranite. These dikes exhibit a complete series of compositions across the contacts between the leucogranite and andesite. Contacts between the granitic and basaltic components in the inhomogeneous dikes are cuspate and sharp on both a megascopic and microscopic scale suggesting that two liquids were in contact. These textures are very common in other igneous systems that are considered to have formed by commingling of magmas (e.g., Taylor and others,1980; Reid and others, 1983; Vogel and others, 1984; Mezeger and others, 1985). According to Taylor and others (1980) commingling of magmas invloves the failure of two miscible magmas to mix due to high viscosities and rapid crystallization, which in this case is caused by quenching of andesite against cooler granite. Textures in the inhomogeneous dikes are very similar to textures produced experimentally by Kouchi and Sungawa (1985) by forced mechanical mixing of basalt and dacite.

The mafic "complexes" described above are another good example of mixing. I believe the biotite-rich granodiorite and hornblende biotite quartzdiorite are mixed rocks. These rock types are restricted spatialy to the areas near inhomogenous dikes, they contain rounded and angular partly resorbed inclusions of both andesite and granite, and they are very inhomogeneous at all scales.

In places I observe dikes that grade into round inclusions and eventually into schlieren. I would suggest this represents an intermediate stage of mixing between simple dikes and the hornblende biotite granodiorites, and that mafic inclusions and schlieren that exist in the more mafic phases of the batholith were in the process of making that mafic phase more mafic.

I see a broad range in degree and type of mixing from: the inhomogeneous dikes where mixing is very incomplete and the andesite was chilled into pillow forms, to nearly homogeneous dikes of dacitic composition with microscopic disequilibrium textures and xenocrysts, to nearly granitic dikes with streaks and blebs of basaltic material, and finally to the mafic hornblende biorite quartz diotite with inclusions of both granite and basalt.

THIN SECTION EVIDENCE

Numerous authors report that disequilibrium xenocryst assemblages can be used as evidence for mixing processes (e.g., Bell, 1983; Hibbard and Watters, 1985). Frozen disequlibrium textures are seen in many of the phenocrysts in these rocks, especially plagioclase in the dacite dikes, the hornblende biotite quartzdiorites, and the inhomogeneous dikes. The mixed liquid was not in equilibrium with the phenocrysts from either the andesite or granite parents. These disequilbrium plagioclase phenocrysts show: 1) very strong normal, and reverse zoning, 2) partly resorbed cores which are altered to secicite and epidote with unaltered rims, and 3) adjacent grains that have dissimilar cores and zoning history. I did not carefully measure all zoning profiles of disequilbrium plagioclase phenocrysts because of complications due to reequilibration of the rims during metamorphism and the problem of phenocyst dissolution within the hybrid magma (e.g., Roberson and Carlson, 1985). But, those I did measure showed large oscillations from core to rims and large total changes in total An content (e.g., 1. core to rim: An_{70} to An_{25}; 2. core to rim: An_{25} to An_{65}).

The fact that the plagioclase phenocrysts have different core compositions and zoning histories suggest that some of the plagioclases have had different origins than others. The average compositions of the disequilibrium plagioclase phenocryts differs greatly from the composition of the plagioclase in the groundmass. The discontinuous nature of plagioclase zoning and irregular shapes of the cores implies resorption and a secondary origin of the rims. I believe the disequlbrium plagioclase phenocrysts in the mixed rocks are original phenocysts of the basalt and in some cases of the granite present at the time of intrusion. The outer rims may be the result of partial equilibrium with the silicic host following intrusion.

In some incompletely mixed rocks, such as the hornblende biotite quartz diorites and granodiorites, mineral compositions and properties of the same species are vastly different, even in the same thin section. These rocks even contain small spherical inclusions of basaltic material that have different compositions in the same sample. These are two textures that Taylor and others (1980) use as evidence for incompete mixing. These rocks also contain alkali feldspar megacrysts similar to those in the nearby granite. The alkali feldspar megacrysts have resorbed borders suggesting that they, too, are out of equlibrium. In the mafic rich granodiorite dike rocks, mixing of a hotter mafic magma with cooler felsic magma has caused rapid crystallization of relatively calcic plagioclase. The calcic plagioclase occurs as many small crystals and as calcic growth zones ("mega-oscillations", Hibbard and Watters, 1985), on more sodic plagioclase already existing in the granodiorite at the time of mixing. An example of this is plagioclase phenocrysts in no. 135 with cores of An_{30}, a zone of An_{60}, and a rim of An_{25}.

Other geologists report that in mixing systems the augite phenocysts in the mafic rocks are partly to fully replaced by fibrous actinolite, biotite and chlorite (e.g., Taylor and others, 1980; Ried and others, 1983; Hibbard and Watters, 1985; Mezger and others, 1985). Whalen and Currie (1984) suggest that the replacement of augite by amphibole and biotite indicates the late introduction of potassium and water from the granitic rocks. Experimental work by Watson (1982, 1984) indicates that potassium diffuses rapidly across an interface into the basaltic liquid making the liquid more andesitic. This phenomenon is precisely what I see in the mafic dikes in the Idaho batholith. In the weakly metamorphosed dikes, those that appear to have interacted to a great extent at a level of exposure below what is exposed now have no remaining pyroxene. However, the dikes that are incompetely mixed, the inhomogeneous dikes, still contain augite which is in various stages of breakdown to actinolite, biotite, epidote, and sphene.

CHEMICAL EVIDENCE

I believe that the chemical compositions of the dikes and associated rock indicate a mixing origin for the intermediate rock types. I think that most of this mixing was mechanical but some component of it must have been diffusive. The fact that these dikes, which were probably originally basalt, have higher K2O and SiO2 than basalt suggests there has been some contamination of basalt to produce andesite. Experimental work by Watson (1982,1984) indicates that the uptake of SiO2 by a well mixed basalt is moderately rapid and that K2O and Na2O are extremely mobile, resulting in considerable uptake of potassium by the basaltic melt and the eventual loss of sodium from the basalt to the granite by up-gradient diffusion. The rate of uptake of SiO2 by the basalt is slower than potassium, causing potassium contamination well beyond the zone of SiO2 contamination (Watson, 1984). This explains why the more highly deformed dikes which have been in the batholith longer than the undeformed dikes have much higher SiO2 and K2O contents, and why the Na2O and K2O contents of the dikes and mixed rocks are highly variable and do not show the linear trend expected by mixing.

Mixing of mafic and felsic magmas will produce linear trends on element- element variation plots (e.g., Vogel and others, 1983). The compositions of the dikes and the granitic rocks plot good linear trends when major elements are ploted against SiO2. For most elements, the compositions of the intermediate rocks lie on straight mixing lines between mafic and felsic rocks. Unfortunately, these trends are similar to those produced by crystal-liquid separation. Vogel and others (1983) state that mixing produces hyperbolic-type trends in element- ratio plots (element-ratio a/b verses element-ratio c/d). However, element-ratio plots and element-element plots of trace elements for the dike to granite suite in the Idaho batholith are ambiguous. This may be due to local contamination of the trace elements in the batholith by variable country rock material.

Vogel and others (1983) point out that rare earth element plots are particularly sensitive to distinguishing simple crystal fractionation from liquid- only fractionation. In the case of mixing, the inverse fractionation in light rare earth elements will lead to a "crossover" pattern. Samples lowest in light rare earth elements will be highest in heavy rare earth elements. This is an inverse relationship in contrast with simple crystal-fractionation where there is an upward shift in all rare earth elements except europium (Vogel and others, 1983). Rare earth element patterns of Idaho batholith rocks clearly show this "crossover" pattern, suggesting that liquid mixing has been important in producing the different rock types seen in and around the dikes. Likewise, the similar-in-magnitude, parallel trends of the rare earth elements suggest the andesites and granite are not related by crystal fractionation.

The streaky alternation of the leucogranite and andesite in the inhomogeneous dikes might suggest that the leucogranite evolved through liquid immisciblity. However, it is more likely that this texture was produced by magma mixing. The rare earth element data suggests that the two magmas did not evolve through liquid immisciblity. Experimental work by Watson (1982) indicates that in immiscible systems the basic component would show an enrichment in rare earth elements over the acid component by a factor of 5 to 10; this is not seen in the Idaho batholith data.

DISCUSSION

The evidence presented above strongly suggests mixing of the basaltic and granitic magmas. The clearest example of mixing is shown in the inhomogeneous dikes and dike complexes where the pillow-shaped andesite surrounded by aplite represents liquid-liquid commingling similar to that in many complexes throughout the world (e.g., Vogel and others,1984; Taylor and others, 1980; Vogel, 1982). It is possible that the most mafic dikes in the batholith do not repesent the original mafic magma composition that was intruded from the mantle. The original mafic magma may have been a more calcic basalt that was contaminated by SiO2 and alkalis during its residence time in the batholith and its rise through the lower crust.

III. 3 HEAT FLOW MODELS

In this section I attempt to numerically evaluate models of granitic magma generation by the injection of basaltic dikes into the crust as discussed in the introduction. In these models I calculate the thermal rise due to conductive transfer of heat from dikes into dry country rocks. Previous calculations of this type have been performed by: Furman and Spera (1985), Lachenbruch and others (1976), and Younker and Vogel (1976).

EQUATIONS

Equations for calculating conductive heat transfer from a cooling dike or sill are given in numerous sources (e.g., Jaeger, 1957,1967; Turcotte and Schubert, 1982; Furman and Spera, 1985) all of which use essentially the same approach. I used the equations of Jaeger (1967) to calculate the heat introduced into the crust by the injection of dikes.

If a sheet-like body, dike or sill, of thickness 2a is instantaneously injected beneath deep cover, the temperature T at time t at a distance x from the midplane of the sheet is given by:

$$\frac{T}{T_o} = \varnothing(x',t') = \frac{1}{2}\left(\mathrm{erf}\,\frac{x'+1}{2\sqrt{t'}} - \mathrm{erf}\,\frac{x'-1}{2\sqrt{t'}}\right)$$ (see notation,Table 4)

where x' and t' are dimensionless parameters defined by: $x' = \frac{x}{a}$; $t' = \frac{kt}{a^2}$

and erf u is defined by: $\mathrm{erf}\ u = \frac{2}{\sqrt{\pi}} \int_0^u e^{-z^2}\, dz$

For x'>1, the country rock outside the intrusion, the maximum temperature (T_{max}) at a position x is attained when $t' = t'_{max}$ given by:

$$t'_{max} = x' / \ln[(x'+1)/(x'-1)]$$

When $t' = t'_{max}$ is known, T_{max} can be found from:

$$\frac{T_{max}}{T_0} = \frac{1}{2} - \frac{1}{2}\ \mathrm{erf}\ \frac{x}{2\sqrt{(kt)}}$$

Some values of t'_{max} and T_{max}/T_0 are given in Table 5.

Table 4 Notations for heat flow equations

T	temperature at any point
T_0	initial temperature of magma
T_1	initial temperature of country rock
T_{max}	maximun temperature attained in country rock
t	time after intrusion
K	thermal conductivity of country rock
d	density of country rock
c	specific heat of country rock
$k = K/dc$	thermal diffusivity of country rock
x	distance from the midplane of the dike
$2a$	thickness of the dike
$\emptyset = T/T_0$	dimensionless temperature parameter
$x' = x/a$	dimensionless position ratio
$t' = kt/a^2$	dimensionless time parameter

Table 5
Maximum temperature (T_{max}) attained in country rock (from Jaeger, 1967).

x'	1	1.2	1.5	2	3	4	5
t'_{max}	0.00	0.500	0.932	1.820	4.33	7.83	12.33
T_{max}/T_0	0.5	0.407	0.324	0.212	0.161	0.121	0.097

CALCULATIONS OF SOME MODELS

The first model calculates the maximum amount of partial melting of the pre-Idaho batholith crust due to the injection of dikes. I use the data I measured in the Bitterroot Lobe of the batholith. The number and size of dikes I measured may be less than the number and size of the dikes that were injected into the crust before melting. Since these are the only data obtainable they will be used and evaluated. The average thickness of the dikes in this section is 2.4 m. Average distance between the dikes is 11.8m, and the volume of dikes is 20% that of the batholith.

In all of these calculations I calculate T_{max} for various values of x. For this model the time it takes to attain Tmax is assumed not important, because Tmax is the value which will determine whether or not melting will take place. However, if too much time is need to attain

ΔTmax the material melted at the contact may solidify.

Model 1: Will the injection of 20% dikes into the lower to middle crust at the same time (t=0) raise temperatures high enough to form a batholith? For this first-order problem, attained temperatures must be higher than the solidus for granite at one half the distance between any two dikes to produce the wholesale melting needed to form a batholith.

Since T_{max} is being calculated table 5 may be used.

Case A:
To=1050°C; a=1.2m

x (m)	T_{max}	final Tmax if T1=250°C	final Tmax if T1=400°C	final Tmax if T1=500°C
1.2*	525	775	925	1025
2.0	325	575	725	825
2.4	243	493	643	743
3.0	211	461	611	711
6.0	102	352	502	602

* 1.2m is the dike contact

Case B:
To=1400°C; a=1.2m

x	Tmax	final Tmax if T1=250°C	final Tmax if T1=400°C	final Tmax if T1=500°C
1.2	700	950	1100	1200
2.0	454	704	845	945
2.4	339	589	739	839
3.0	282	532	682	782
6.0	136	386	536	636

Values of To and T1 were chosen to represent realistic temperatures of mafic magmas, in this case basalt or basaltic andesite, and lower crustal temperatures. T1=250°C represents the lowest temperature at the base of the crust estimated by Lachenbruch (1968) for the pre-Sierra Nevada crust. T1=400°C is an intermediate value, and

$T1=500^{\circ}C$ is a higher value expected in continental margins before melting (Lachenbruch, 1968). Mid-crustal temperatures higher than this are realistic only in high heat flow provinces where crustal melt already exists. Case A has $To= 1050^{\circ}C$; this is a lower value for water saturated mafic magmas (Furman and Spera, 1985). Case B with $To=1400^{\circ}C$ is a realistic value of the temperature mafic magma generated above a subduction zone and rising adiabatically into the crust in a state of super heat (e.g., Presnall and Bateman, 1973).

A temperature of about 750-850°C is needed to produce a melt of granitic composition in the lower to middle crust with normal water content (Hyndman, 1985a). Temperatures near this are met in Case A out as far as the dike contact for $T1=250^{\circ}C$, $x=2.0m$ for $T1=400^{\circ}C$, and $x=2.4m$ for $T1=500^{\circ}C$. In Case B these temperatures are found out to the contact dike for $T1=250^{\circ}C$, $x=2.4m$ for $T1=400^{\circ}C$, and $x=3.0m$ for $T1=500^{\circ}C$. None of these cases has melting temperatures out as far as the $x=7.1m$ distance needed to cause melting one half way between two average dikes. Even if temperatures are taken into consideration from the two dikes on either side of the midpoint, temperatures are still not high enough for any of these cases (e.g., combined $T_{max} = 2(<100) + 500= <700^{\circ}C$). If these assumptions are correct it is evident that in this first-order model that temperatures needed to partially melt the country rock are not attained half way between any two dikes. Also, this model does not take into consideration the heat of fusion of the crustal rocks which would serve to increase the heat needed to cause melting.

Multiple injection of mafic dikes into the lower continental crust can cause melting of the country rock at the dike contacts and even a couple of meters away from the contact, but not enough to melt the volume needed to form a batholith. This model is obviously oversimplified; there is no reason to assume that all of the dikes needed to melt the crust were injected at one time. The post-melting dikes that I see in the Idaho batholith clearly show a wide range in times of intrusion. Also, the volume of dikes in the batholith may give no indication of that at deeper levels where the batholithic magmas were generated. It might be added that from model one it is plausible that some of the granite around some of the large dikes could have been remelted by the injection of the dikes.

Model 2: Another way to visualize model one is to model the temperature increase for the total mass of mafic magma intruded at the level of exposure. This second model estimates the maximum thermal rise in the pre-Idaho batholith crust due to the injection of the total mass of mafic magma as a single tabular body. This model is made realistic because at the depth of intrusion, the crust is sufficiently insulated that heat intruded over a relativly short time would remain at this level in the crust until other dikes were injected. Thus, in a way this model estimates the injection of a mass of mafic magma as wide as 20% of the batholith.

Along the Lochsa cross-section the thickness of the batholith is about 64 km, 20% of which is mafic dikes, giving a combined thickness of mafic diks of 12.8km. In this model I use the same initial temperatures as used in model one except that I also use $T_0=1200^{o}C$.

Case C:

To=1050°C; a=6.4 km

x (km)	T_{max}	final Tmax if T1=250°C	final Tmax if T1=400°C	final Tmax if T1=500°C
6.4*	525	775	925	1025
12.8	250	500	650	750
19.8	170	420	570	760
Case D:				
To=1200°C; a=6.4km				
6.4	600	850	1000	1100
12.8	290	540	690	790
19.2	190	440	590	690
Case E:				
To=1400°C; a=6.4km				
6.4	700	950	1100	1200
12.8	340	590	740	840
19.2	225	475	625	725
25.6	170	420	570	670

*6.4km is the dike contact

For this model to work as written, temperatures high enough to cause melting must be attained 36 km from the center of the combined mafic dikes, assuming that all the dikes are at the center of the batholith. Temperatures this high are met in Case C only out to the dike contact for T1=250^{o}C and 400^{o}C, and x=12.8 km for T1= 500^{o}C. In case D melting temperatures are attained as far as the dike contact for T1=250 and 400^{o}C, and x= 12.8 km for T1=500^{o}C. Finally in case E temperatures near melting are met at the contact for T1=250^{o}C, x=12.8 km for T1= 400^{o}C, and x=19.2 km for T1=500^{o}C (note that x=19.2 km is only 12.8 km from the dike contact). In no case are melting temperatures attained out as far as the required x=36 km needed to cause melting of the whole batholith cross-section. Using To= 1200^{o}C and T1= 400^{o}C, about 60% of the batholith must be mafic dikes to melt the batholith cross-section.

DISCUSSION

The first order models given here do not fully explain the generation of batholiths, and the assumptions I made in using the data collected in the Idaho batholith are incomplete. A more realistic model involves the prolonged injection of magma from the mantle which preheats the crust and eventually causes melting (e.g., Hildreth, 1981). The region of generation of the granitic magmas would have been deep enough and thus insulated so that the heat gained from an injected dike would not leave the system for many years. The heat built up over a long period of dike injections would be enormous and probably sufficient to form a batholith. Shaw (1985) estimates that mafic input into the crust at realistic rate must be concentrated over at least 30,000 and as much as 300,000 years to develop batholiths the size of the Sierra Nevada. In this case the dikes seen in the Idaho batholith may just be representative of material that maintained the high temperatures in the crust needed to keep the batholith active, and are not really the material that initially melted the crust.

There are numerous reasons why the amount of dike material seen at the present level of exposure should be a minimum, including:

1) The injection and through flow of basalt to higher levels, feeding arc type volcanos, would heat crustal rocks enroute. This would introduce more heat because the country rock would be in contact with molten basalt for longer periods of time. The deep level of exposure of the Idaho batholith may explain why no late Cretaceous volcanic mafic rock is found around the batholith. It is possible that much of this volcanic material is now present in the thick sections of late Cretaceous volcanogenic sediments east of the batholith.

2) Dikes inflate with the pressure of intrusion, then deflate

somewhat as magma pressure is relieved.

3) Some of the basaltic magma injected was used in the contamination of granitic melt to form intermediate rocks.

4) The granitic magma of the batholith was generated at somewhat deeper levels where there may be a greater proportion of dike material.

5) The dikes present in the batholith were emplaced later in the history of the batholith when the granite could sustain a fracture.

A model that combines the heat introduced from a combination of dike injection and mafic underplating could bring about substantial increases in crustal temperature. The underplate would act to preheat the crust and diking would act as a means of convecting heat higher into the crust. Cobbing and Pitcher (1983) use a combination of these models to explain batholith generation and the extreme thickness of the crust in the Andes.

In the continental volcanic arc environment of major batholiths such as the Idaho batholith it seems that the formation of voluminous granitic magmas would be unavoidable. Large volumes of mafic magma are generated due to subduction. These mafic melts move into and through the continental crust at temperatures of 1050-1400^{o}C for long periods of time. They preheat and cause local and large scale melting of the lower and middle crust which culminates in large-scale diapiric mobilization of partially molten zones from which batholithic magmas separate (Hildreth, 1981).

CONCLUSIONS

1) The mafic dikes that intrude the Idaho batholith are synplutonic as evidenced by, mutually intrusive relationships, and metamorphism of the dikes by the host granite.

2) The dikes make up 20 percent by volume of a representative section of the batholith.

3) The dikes range in composition from high alumina basaltic andesite to dacite and are calc-alkaline.

4) The ENE-WSW orientation of most of the dikes suggests extension in the batholith at the time of intrusion in a NNW-SSE direction.

5) Evidence is abundant for magma mixing between the basaltic andesites and the host granite, producing intermediate rock types with disequilibrium features.

6) The mafic dikes probably represent latest stages of subduction-generated mafic magmas that at slightly earlier stages and deeper levels were injected into the continental crust and contributed to melting to form the batholith.

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APPENDIX 1

116.0 is a very fine grained, black, massive andesite with 1-2mm euhedral plagioclase and mafic phenocrysts.
116.7 is a very fine grained, black, massive andesite with 1-3mm euhedral plagioclase and augite phenocrysts.
118.6 is a fine grained, medium brown, massive dacite with 1-2 cm clumps of euhedral interlocking plagioclase and augite.
124.0 is a moderatly fine grained, medium gray and massive dacitic rock that is very inhomogeneous, containing 2mm laths of plagioclase, 3mm xenocrysts of alkali feldspar± quartz, angular chuncks of granite up to 1.5 cm and round basaltic inclusions.
124.95A is a fine grained dark green gray dacite that is strongly foliated.
124.95B is a very fine grained, dark green gray basaltic andesite which contains numerous 1-2mm elongate mafic clots of fine grained hornblende, actinolite and biotite that define a strong lineation.
125.0 is a two mica granite.
125.05 is a fine grained, foliated, dark gray basaltic andesite with 1-2mm hornblende and actinolite aggregates.
132.9 is a very fine grained, dark gray to black, slightly foliated andesite.
134.3A is a fine grained, dark gray very inhomogeneous basaltic andesite to andesite, that consists of rounded areas of basaltic andesite surounded by thin films of aplite. 134.3A1 is of the mafic portion of this dike only and 134.3A2 is of the whole rock.
134.3B is a two mica granite sampled near 134.3A.
134.3C is a medium grained, medium gray and white, very inhomogeneous, foliated hornblede biotite quartz diorite, sampled from near 134.3A.
135.0 is a very fine grained, medium gray, massive dacite with 1mm plagioclase phenocrysts and numerous 0.5-1 cm clumps of euhedral of plagioclase and augite.
135.9A is a very fine grained, dark gray, massive basaltic andesite, with 1-4mm euhedral plagioclase and augite phenocrysts.
135.9B is a fine to medium grained, medium gray, massive, inhomogeneous biotite rich granitic dike that contains numerous small subangular basaltic inclusions.
136.05A is a very fine grained, dark gray, massive basaltic andesite.
136.05B is a very fine grained, dark gray, slightly lineated basaltic andesite with 1-2mm laths of plagioclase.
139.3A is a medium grained, medium gray foliated biotite granodiorite.

139.3B is a fine to medium grained, medium gray foliated biotite granodiorite.
139.3C is a fine grained dark gray, foliated andesite.
139.3D is a medium to coarse grained two mica granite with alkali feldspar megacrysts.
141.95 is a very fine grained, lightly foliated basaltic andesite.