40Ar/39Ar ages of Eocene dikes and the age of extension in a Tertiary magmatic arc Idaho and Montana

Steen W. Simonsen

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\textit{\textsuperscript{40}Ar/\textsuperscript{39}Ar} AGES OF EOCENE DIKES AND

THE AGE OF EXTENSION IN A TERTIARY MAGMATIC ARC,

IDAHO AND MONTANA

by

Steen W. Simonsen

B.A., University of Montana, 1992

Presented in partial fulfillment of the requirements

for the degree of

Master of Science

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1997

Approved by

\underline{\textit{Dr. Donald Hyndman, Committee Chair}}

\underline{\textit{Dean, Graduate School}}

\underline{30-October, 1997}
Early Tertiary northwest-directed extension in the Idaho and Montana portions of the western North American continental magmatic arc produced 300 km of northeast-trending structures, igneous centers, and geophysical anomalies, defined herein as the Idaho-Montana Extensional Belt (IMEB). I present 11 new \(^{40}\text{Ar}/^{39}\text{Ar}\) mineral ages from 10 early and mid-Eocene intermediate and felsic dikes, sampled in four swarms situated parallel to dike trend over the southwestern 225 km of the IMEB. These swarms are located 1) in the hanging wall of the Bitterroot metamorphic core complex, 2) in Proterozoic rocks of the Salmon River Arch, near the Challis volcanic field, 3) in the Idaho Atlanta batholith, at the northeastern limit of the Idaho Porphyry Belt, near the Challis volcanic field, and 4) at the southwestern limit of the Idaho Porphyry Belt, north of the Boise Basin. In each locale, I sampled the oldest northeast-trending dikes, as identified through field determinations and previous mapping. As such, \(^{40}\text{Ar}/^{39}\text{Ar}\) ages from these dikes reflect the earliest yet identified manifestations of extensional strain that affected central Idaho and western Montana between 55 and 50 Ma. Previous studies determine the timing of initial extension indirectly, through deformation and cooling dates extrapolated back to the onset of extension. To provide a more precise geochronology, I use the \(^{40}\text{Ar}/^{39}\text{Ar}\) mineral ages to demonstrate that the extensional stress field was in place by 53 Ma over much of the IMEB, with initial strain possibly delayed as much as 3 Ma near the Boise Basin. In addition, dikes indicate the initial extensional direction was approximately 110° over much of the IMEB. This provides regional support for previous extensional determinations derived from structural analyses of the Bitterroot mylonite and Challis-area faulting. I utilize dike ages, previous mapping, and major, trace, and rare earth element analyses with respect to local coeval structural and magmatic events to determine emplacement settings and temporal and magmatic relationships. On a northern Cordilleran scale, the new precise dates illustrate a short delay in the progression of strain from the southern Omineca Belt, north of the Lewis and Clark Line, south into central Idaho and northern Nevada, between 58 and 50 Ma.
ACKNOWLEDGEMENTS

Funding for this research was supplied by Amoco Production Company grants to Steen Simonsen and Dr. Don Hyndman of the University of Montana. Thanks to Ms. Sara Foland at Amoco Production Co., for helping to arrange funding. It would not have happened without her aid. Thanks to my advisor, Dr. Don Hyndman, for all his teaching, support, and most of all, patience. Thank you to my thesis committee, Dr. Steven Sheriff and Dr. Randolph Jeppesen, and to Dr. Peter Copeland and staff at the University of Houston for patience and help with the argon analysis. Thanks to the Mansfield Library staff, and to Judy Fitzner and Loreene Skeel, who throughout my rather extended college career did their utmost to make my work easier. Thanks to my field help and moral supporters, Ms. Kelly Brunt (Good one, hon), Mr. Stephen Porder, and Mr. Pat Collins. Special thanks to Dr. Ted ‘shotgun geologist’ Doughty, who, in between beating up goalies, provided geologic counsel and made me keep things in perspective. Thanks, everyone. Mom, Dad, and Mike, guess I lost the bet. But here ya go.

Thankyouverymuch. Ladies and Gentlemen, Steen has left the building.
DEDICATION

I dedicate this thesis to the memory of Dr. May Wright, who passed away in May, 1995. A remarkable woman, Dr. Wright challenged societal norms by completing her M.S. in plant genetics, following with a Ph.D. in plant ecology at the University of Chicago in 1930. She was a scientist driven by a love of and curiosity towards the natural world, and enjoyed a career that included developing the biology department and teaching plant genetics at Northern Montana College in Havre in the 1930's, research and teaching at the University of Minnesota, Minneapolis, and raising a family. In her later years Dr. Wright strove to advance our understanding of the role of native plants in natural ecosystems, and worked for the preservation of rare and sensitive species in northern woodlands, including, at the age of 77, co-founding the Minnesota Native Plant Society. Most importantly though, my grandmother instilled in her clan an intense curiosity and a love of learning that has permeated the following generations. She set the standard.
Table of Contents

<table>
<thead>
<tr>
<th>CHAPTER</th>
<th>PAGE</th>
</tr>
</thead>
<tbody>
<tr>
<td>Abstract</td>
<td>ii</td>
</tr>
<tr>
<td>Acknowledgements</td>
<td>iii</td>
</tr>
<tr>
<td>Dedication</td>
<td>iv</td>
</tr>
<tr>
<td>List of Tables</td>
<td>vi</td>
</tr>
<tr>
<td>Introduction</td>
<td>8</td>
</tr>
<tr>
<td>Geologic Setting</td>
<td>8</td>
</tr>
<tr>
<td>Idaho-Montana Extensional Belt</td>
<td>10</td>
</tr>
<tr>
<td>Statement of Problem</td>
<td>11</td>
</tr>
<tr>
<td>Pertinent Studies</td>
<td>13</td>
</tr>
<tr>
<td>Methods</td>
<td>16</td>
</tr>
<tr>
<td>Data, Results, and Local Interpretations</td>
<td>19</td>
</tr>
<tr>
<td>East Fork Dike Swarm</td>
<td>20</td>
</tr>
<tr>
<td>Shoup</td>
<td>43</td>
</tr>
<tr>
<td>Seafoam Dike Swarm</td>
<td>54</td>
</tr>
<tr>
<td>Boise Basin Dike Swarm</td>
<td>60</td>
</tr>
<tr>
<td>Regional Interpretations</td>
<td>72</td>
</tr>
<tr>
<td>Chronology</td>
<td>72</td>
</tr>
<tr>
<td>Integrating IMEB extension with regional Eocene tectonics</td>
<td>78</td>
</tr>
<tr>
<td>Summary and Conclusions</td>
<td>82</td>
</tr>
<tr>
<td>References</td>
<td>84</td>
</tr>
<tr>
<td>Appendix A: Outcrop location and sample descriptions</td>
<td>94</td>
</tr>
<tr>
<td>Appendix B: Argon release spectra and Isotope correlations</td>
<td>98</td>
</tr>
<tr>
<td>Appendix C: Geochemical Analyses and CIPW Normative minerals</td>
<td>116</td>
</tr>
<tr>
<td>Appendix D: Recommended future work</td>
<td>120</td>
</tr>
</tbody>
</table>
List of Tables and Figures

<table>
<thead>
<tr>
<th>FIGURE</th>
<th>PAGE</th>
</tr>
</thead>
<tbody>
<tr>
<td>Figure 1: Regional geologic map and delineation of the Idaho-Montana Extensional Belt</td>
<td>9</td>
</tr>
<tr>
<td>Figure 2A-J: Argon step heating spectra and isochrons</td>
<td>22-25, 46, 47, 55, 62-65</td>
</tr>
<tr>
<td>Figure 3: Map of Bitterroot Dome and Skalkaho Slab, and location of the East Fork Dike Swarm</td>
<td>27</td>
</tr>
<tr>
<td>Figure 4A,B: Major-element trends; East Fork dikes, Biotite Granodiorite suite, Paradise pluton, Bear Creek pluton, and mid-Eocene syenogranitic plutons</td>
<td>30-33</td>
</tr>
<tr>
<td>Figure 5A,B: REE and trace-element comparison plots, Bear Creek pluton and East Fork dikes</td>
<td>35, 36</td>
</tr>
<tr>
<td>Figure 6A,B: REE and trace-element comparison plots, Paradise pluton and East Fork dikes</td>
<td>37, 38</td>
</tr>
<tr>
<td>Figure 7A,B: REE and trace-element comparison plots, synplutonic dikes and East Fork dikes</td>
<td>39, 40</td>
</tr>
<tr>
<td>Figure 8: Eu/Eu* comparison plot</td>
<td>41</td>
</tr>
<tr>
<td>Figure 9: Map of the Idaho portion of the Idaho-Montana Extensional Belt, with three Idaho sampling locales (adapted from Kiilsgaard et al., 1986)</td>
<td>44</td>
</tr>
<tr>
<td>Figure 10A,B: REE and trace-element comparison plots, Paradise pluton and Shoup dikes</td>
<td>50, 51</td>
</tr>
<tr>
<td>Figure 11A,B: REE and trace-element comparison plots, early Challis volcanic rocks and Shoup dikes</td>
<td>52, 53</td>
</tr>
<tr>
<td>Figure 12A,B: REE and trace-element comparison plots, early Challis volcanic rocks and Seafoam dikes</td>
<td>57, 58</td>
</tr>
</tbody>
</table>
### Tables and Figures, cont.

<table>
<thead>
<tr>
<th>FIGURE</th>
<th>PAGE</th>
</tr>
</thead>
<tbody>
<tr>
<td>Figure 13A,B: REE and trace-element comparison plots, early Challis volcanic rocks and Boise Basin dikes</td>
<td>68, 69</td>
</tr>
<tr>
<td>Figure 14: Major element trends; Idaho dike swarm samples and the Quartz Monzodiorite suite</td>
<td>70, 71</td>
</tr>
<tr>
<td>Figure 15: Ages and directions of documented early Tertiary extension in Idaho and western Montana</td>
<td>76</td>
</tr>
<tr>
<td>Figure 16: Northern Cordillera comparison of latitude vs. earliest extensional deformation and length of early and mid-Tertiary extension</td>
<td>77</td>
</tr>
<tr>
<td>Table 1: Summary of $^{40}$Ar/$^{39}$Ar ages</td>
<td>19</td>
</tr>
<tr>
<td>Table 2: Comparative compositional summary of Idaho igneous suites vs. Idaho-Montana Extensional Belt dikes, in time sequence</td>
<td>49</td>
</tr>
</tbody>
</table>
Introduction

I report 11 new $^{40}$Ar/$^{39}$Ar age spectra and chemical analyses from 10 northeast-trending Eocene dikes in 4 localities of the Idaho-Montana Extensional Belt (IMEB) (Fig. 1). These dikes resulted from crustal extension, starting in early to mid-Tertiary times, and related pre- and syn-extensional arc magmatism. While previous studies have approximated the date of initial extension as ~50 - 60 Ma, there have been no measurements that provide a precise geochronological account over a broad area. The new age dates provide the most precise estimate yet documented for the extensional paleostress field. They show that regionally, and locally, large-scale dike emplacement is the earliest identified manifestation of the Eocene extensional stress field. These data, combined with studies by others, indicate that the extensional regime was in place by 55 - 50 Ma, over at least the southwestern 250 km of the IMEB. The new ages also provide a temporal basis for inferring IMEB magma affiliation with other local and regional Eocene magmatism. In addition, I place the precise extensional timing determined herein in the context of regional and northern Cordilleran Eocene continental arc extension, from the Omineca Belt of southern British Columbia south across the Snake River Plain and into northern Nevada.

Geologic Setting

During late Cretaceous time, nearly all of western North America underwent significant east-west shortening. The duration, direction, and extent of shortening varied
Figure 1: Selected Idaho-Montana Extensional Belt and contemporaneous features. Adapted from Olson, 1968; Hyndman, 1983; O'Neill and Lopez, 1985; Doughty and Sheriff, 1992.
locally and regionally along the Cordilleran arc. Shortening in the Idaho platelet, bounded by the Snake River Plain to the south and the Lewis and Clark Line to the north (Fig. 1), ranged between 30 and 50% (Wernicke et al., 1987). By early to mid-Tertiary times, and following the emplacement of the mesozonal 94 - 55 Ma Idaho batholith, this roughly E-W compression had relaxed or rotated throughout the the Idaho platelet. Subsequently, this region experienced a nearly immediate stress field change to directionally and temporally variable northwest-southeast extension. Activation of the extensional strain field in central Idaho and western Montana, variously dated between 60 Ma and 45-55 Ma, resulted in and influenced the development of diverse local and regional features that together comprise the Idaho-Montana Extensional Belt.

**Idaho-Montana Extensional Belt**

The Idaho-Montana Extensional Belt (IMEB) is a broad band of northeast-trending high-angle normal faults, shear zones, graben, as well as substantial early and mid-Tertiary volcanism, plutonism, dike swarms, and tectonically-induced coarse sedimentation. The IMEB is nearly continuous over 350 km northeast from Boise into at least southwest Montana. It is defined herein as containing all or part of the Salmon-Anaconda tectonic zone (O’Neill and Lopez, 1985), the Bitterroot metamorphic core complex (Hyndman, 1980; Garmezy 1983; Garmezy and Sutter, 1983; Hyndman and Myers, 1988; Doughty and Sheriff, 1992; House et al, 1993; Hodges and Applegate, 1993; House and Hodges, 1994; House, 1995; Foster and Fanning, 1997), the 270 km
trans-Challis fault system (Bennett and Knowles, 1986; Kiilsgaard and Lewis, 1986; Kiilsgaard et al., 1986; Bennett, 1986; Moye, 1990; Kiilsgaard and Fisher, 1995), the Challis volcanic field (McIntyre et al., 1982; Janecke 1991; Norman and Mertzman, 1991; Janecke and Snee, 1993), and the Idaho porphyry belt (Olson, 1968), along with numerous other named and un-named Eocene extensional features. Previous workers have generally treated these features individually or in small groups; however, I believe that the extensional relationship shared by all of them warrants an all-inclusive designation.

Both the IMEB and the Great Falls tectonic zone (O’Neill and Lopez, 1985) seem to comprise similar sets of features with overlapping geographic boundaries (Fig. 1). However, the IMEB is defined as strictly a temporal and tectonic relationship, the grouping of structures and geologic features directly related to the Eocene extensional event. In contrast, the Great Falls tectonic zone contains not only these features, but Paleozoic depositional patterns as well as folds, igneous bodies, and physiographic features that arguably may be traced to Proterozoic, Paleozoic, Mesozoic, and possibly even late Cenozoic origins. The two designations apply to regional geologic features that are the result of different formative and tectonic environments.

Statement of Problem

Northeast-trending dike swarms are an integral part of the Idaho-Montana Extensional Belt as a whole (Hyndman et al., 1977), and are most numerous in the Idaho
Porphyry belt (Olson, 1968) and the trans-Challis fault system (Kiilsgaard and Lewis, 1985). Despite this, the implications of dike swarms have generally received only cursory treatment in studies that have mapped and analyzed most other significant portions of the IMEB. While local aspects of early Tertiary dikes are included in work concerning various igneous and metamorphic complexes, (Olson, 1968; Hyndman et al., 1977; Badley, 1978; Desmarais, 1983; Batatian, 1991; Doughty and Sheriff, 1992; Hodges and Applegate, 1993; Schmidt, 1994; Jordan, 1994; Halter, 1995), they are not generally considered in a regional scope. In addition, attempts to determine magmatic affiliations are often indeterminate due to the lack of geochronologic and trace element data for the dikes. Therefore, geochronologic and geochemical analyses of dikes should provide new insights into regional and local Eocene stress field and magmatic events. The new data provides a base for interpretations regarding local and regional structural, tectonic, and magmatic relationships.

The intent of the study is to identify the age and geochemical characteristics of the oldest dikes of the early Eocene stress field, and with this in mind, I suggest the following:

1) Previously, the onset of NW-SE directed extensional strain has been dated indirectly through extrapolation from deformation and plutonic cooling ages. On thin, hypabyssal dikes, $^{40}\text{Ar}/^{39}\text{Ar}$ dike geochronology will approximate emplacement dates. Dates of the
oldest northeast-trending dikes will provide a precise estimate of the onset of extensional
deformation, both regionally and locally.

2) Dike swarm trends, taken at a ~270km scale, should with some limitations, indicate
the initial regional stress field orientation (Delaney et al., 1986). These should compare
favorably with extensional trend measurements from Bitterroot mylonite shear and
stretching lineations and Challis-region fault data.

3) Local magmatic relationships and regional affinities of the dikes may be inferred from
new dates and major, trace, and rare-earth element analyses.

4) Extensional timing is more precisely known both to the north and south of the central
Idaho region. Published data indicate that Tertiary E-W extension started and ended
slightly earlier north of the Lewis and Clark Line than in central Nevada (Fig. 16), but
there is a gap in precise data over the IMEB region. \(^{40}\text{Ar}/^{39}\text{Ar}\) data for this gap provides a
basis for temporal comparisons in the northern third of the United States Cordillera.

5) Conclusions reached herein provide a more concrete regional estimation of the timing
and orientation of Eocene extension.

Pertinent Studies

I utilize isotopic dates and chemical analyses of early Tertiary dikes to provide an
absolute geochronology and a regional perspective on the initiation and subsequent
modification of Eocene extensional strain in central and eastern Idaho and southwestern
Montana. Previous studies in other regions utilize dikes to document and date both
compressive and extensional regimes. For instance, isotopic dating can provide an absolute chronology of the initial emplacement and rotation or modification of stress fields (Feraud et al., 1987). Rehrig and Heidrick (1976) used this method, combined with vein and pluton orientations, to document and date the Laramide and early Tertiary paleostress fields in southern Arizona. Zoback and Thompson (1978) and Laughlin et al. (1983) utilized K-Ar geochronology and trend analysis to document and date early east-northeast Basin and Range extension, and to analyze temporal variations in stress direction, in Nevada and the southwest United States. In Trans-Pecos, Texas, Henry and Price (1986) utilized dikes to describe low strain rates synchronous with initial magmatism, with dikes emplaced in pre-existing fractures trending variably sub-perpendicular to extension. Upon initiation of higher strain rates and accompanying large-scale faulting, extensional forces overcome the influence of pre-existing fractures and dikes are emplaced normal to \( \sigma_3 \).

In addition, previous studies in other regions have determined that dike trends in tectonically active environments can accurately portray \( \sigma_3 \) on a regional scale, and indeed, are more likely to reflect the principal stress directions than those emplaced into tectonically inactive regions. However, as described above, Henry and Price (1986) illustrate a potential problem in large-scale dike trend analysis, that of pre-existing fractures. Some workers suggest that emplacement trends of dikes in the IMEB were controlled by a pre-existing fracture set (O’Neill and Lopez, 1985; Kiilsgaard and Fisher,
1995). They describe Precambrian movement on Shoup-area faults and paleodepositional trends as evidence of these fractures. Other workers, however, dispute the determination of Precambrian movement on these faults (T. Doughty, 1997, personal communication), and there has been little published documentation of pre-dike fracture patterns or joint sets in host rock for most portions of the IMEB (see Allen and Hahn, 1994). This study identified no shear across dikes or erratic trends as might be expected with pre-existing fractures. In addition, although Luthy (1982) suggested that pre-existing fractures controlled the emplacement of epizonal Eocene plutons and associated dikes in one portion of the IMEB, petrologic descriptions therein indicate that the dikes in question are from later, mid-Eocene plutonism that occurred following the development of large scale trans-Challis faulting.

Tempering this concern, however, Feraud et al. (1987) suggest that the direction of dike emplacement is not significantly influenced by pre-existing fractures, if those fractures vary more than a few degrees off parallel to the greatest horizontal force. As a result, even in regions that have been tectonically active for long periods before initial dike emplacement and thus contain significant fracturing, dikes will be in planes approximately normal to the least principal stress direction (Delaney et al., 1986; Pollard, 1987).

Thus, it seems likely that the orientation of IMEB dikes reflects the direction of least compressive stress. Indeed, IMEB dikes in this study are approximately normal to
extension directions measured independently in the Bitterroot mylonite and Challis
scatter is expected and present, resulting from interaction of the regional stress field with
a variety of local pre-existing structures (Henry and Price, 1986).

For clarity, I use the terminology of Janecke (1992) to distinguish 3 successive
Tertiary extensional stress fields in the Idaho magmatic arc. Extensional episode 1, the
first documented extension, was NW-SE directed and is the primary concern of the
present work. A subsequent, rapid change to ENE-WSW to NE-SW-directed extension,
apparently controlling deformation in much of central Idaho by 46-48 Ma, is termed
episode 2. Episode 3, generally known as Basin and Range type extension, refers to
Miocene and younger NE-SW directed extension.

Methods

To provide both local and regional insights, I obtained age and geochemical data
from dike swarms that represent a wide geographic sampling of the IMEB. The IMEB
probably contains several dozen individual swarms, and as much of the IMEB is located
in unroaded wilderness, only a small portion is mapped in detail. I sampled previously
mapped dikes, specifically to represent the oldest group of northeast-trending dikes in
their respective swarms. Dikes were sampled in the East Fork Bitterroot River valley,
southwestern Montana (EFDS), near Shoup, Idaho (SH), near the Seafoam Ranger
Station north of Stanley, Idaho (SF), and along the South Fork Payette River north of the
Boise Basin, Idaho (BB) (Fig. 1). Samples were chosen based on dike orientation, relative dike age as documented by previous workers and field relations, and rock freshness. A common problem with chemical and isotopic analyses of Eocene igneous rocks in central Idaho is hydrothermal alteration and mineralization, and finding fresh samples proved difficult. Even with careful sampling and optical examination of suitable mineral specimens, some $^{40}$Ar/$^{39}$Ar analyses proved unusable. The sampled areas lie within the Dillon, MT, Elk River, ID, and Challis, ID 1$^\circ$ x 2$^\circ$ quadrangles. All sampled areas are accessible by main roads or four-wheel drive trails (Appendix A).

Chemical analyses are for crushed bulk rock samples. All analyses were done at the Washington State University GeoAnalytical Laboratory under the supervision of Diane Johnson. Major oxide and trace-element analyses presented here (Appendix C) were obtained by X-ray fluorescence techniques on a Rigaku automated wavelength spectrometer. Rare-earth elements analyses (Appendix C) were obtained using a Sciex inductively-coupled plasma mass spectrometer. Analytical specifications for the Washington State Laboratory are noted in Hooper et al. (1993).

Mineral suitability for the $^{40}$Ar/$^{39}$Ar process was determined optically from thin sections. True to the nature of many Tertiary rocks in IMEB, all K-feldspar grains were altered beyond use for $^{40}$Ar/$^{39}$Ar analysis. Generally, biotite and hornblende grains were in more pristine condition, although in most rocks only one mineral proved suitable for dating. $^{40}$Ar/$^{39}$Ar analysis was supervised by Dr. Peter Copeland at the University of Houston Thermochronology Lab (see McDougall and Harrison, 1988). Fourteen pure
biotite and pure hornblende mineral separates were produced from eleven whole-rock samples, using standard mechanical and heavy liquid techniques. Samples were wrapped in Al-foil and along with controls were irradiated at the Ford reactor at the University of Michigan, to produce $^{39}$Ar from $^{39}$K by neutron bombardment. Irradiation control was sample number 92-176, a 27.9 Ma Fish Canyon sanidine (Cebula et al., 1986). Following irradiation, the samples were wrapped in Sn-foil and step heated in a double-vacuum resistance-heated furnace of the type described by Harrison and Fitzgerald (1986), with a temperature control of ± 4°. Gas clean-up was performed with turbo and ion pumps, and SAES getters. Extraction line blank is typically $6 \times 10^{-15}$ mol $^{40}$Ar. Argon isotope ratio analysis was done with a Mass Analyzer Products mass spectrometer model 215-50, and the data analyzed and plotted by MassSpec software, written by Alan Deino of the University of California, Berkeley Thermochronology Laboratory.
Data & Results

Age spectrum diagrams for many of the analyzed samples are shown in Figure 2 and a summary of dates is given in Table 1. The remainder of the plots and pertinent data tables are in Appendix B.

Table 1

<table>
<thead>
<tr>
<th>Fig.</th>
<th>Sample number and mineral</th>
<th>Interpreted age (Ma)</th>
<th>Plot method</th>
</tr>
</thead>
<tbody>
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<td>2A</td>
<td>EF9514a, Biotite</td>
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<tr>
<td>2B</td>
<td>EF9513, Biotite</td>
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<tr>
<td>2C</td>
<td>EF9505, Hornblende</td>
<td>52.5 ± 0.48</td>
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<td>2D</td>
<td>EF9301, Hornblende</td>
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<td>2E</td>
<td>SH02a, Biotite</td>
<td>50.2 ± 0.4</td>
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<tr>
<td>2F</td>
<td>SH01, Hornblende</td>
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<td>2G</td>
<td>SF9517a, Hornblende</td>
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<td>2H</td>
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</tr>
<tr>
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</tr>
<tr>
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</tr>
</tbody>
</table>
**EAST FORK DIKE SWARM**

The dike swarm exposed in the East Fork of the Bitterroot River valley, east of Sula, MT, is in the northeastern section of well-mapped dike swarms that are readily identified as part of the Idaho-Montana Extensional Belt (Hyndman et al., 1977; Badley, 1978). The East Fork Dike Swarm was emplaced into rocks that subsequently became the Skalkaho slab, the hanging wall of the northwest-translocating Bitterroot metamorphic core complex (Doughty and Sheriff, 1992). The dikes are emplaced in a complex mosaic of Cretaceous and Paleocene Idaho-Bitterroot batholith granodiorite, and quartzo-feldspathic gneisses and migmatites of the regionally metamorphosed mid-Proterozoic Belt Supergroup. Trending almost exclusively northeast with an average trend of N47°E, the East Fork swarm intrudes an area larger than 300 sq. km. Compositions range from rhyolite to basaltic andesite, although the majority of the dikes are porphyritic rhyolite and rhyodacite. Comprehensive petrological, geochemical, and geological descriptions for the dike swarm and host rocks as a whole are presented in Badley (1978), Desmarais (1983), and Halter (1995).

Significant geochronologic data is available for portions of the nearby Bitterroot metamorphic core complex, and for the main stage of the Idaho-Bitterroot batholith. Data from relevant dikes is scarce, although previous mapping of the dikes illustrates crosscutting relationships between older rhyolite and rhyodacite dikes and younger, more mafic dikes (Badley, 1978). Desmarais (1983) provides several age dates (see below) for
intrusive and extrusive Eocene rocks in the East Fork area, including K-Ar and fission track (zircon and apatite) dates for two dikes, a biotite granodiorite in the southeast margin of the swarm, and an altered pyrite-bearing rhyolite from the west margin. In addition, Hodges and Applegate (1994) provide a $^{40}\text{Ar}/^{39}\text{Ar}$ date of $46.4 \pm 0.8$ Ma from a dike cutting mylonitic textures of the Bitterroot mylonite zone, approximately 50 km west of the East Fork swarm.

I provide four new $^{40}\text{Ar}/^{39}\text{Ar}$ incremental release age spectra from the East Fork dike swarm. Samples EF9514a and EF9513 are from porphyritic biotite high-K rhyodacite dikes trending N45°E and N44°E respectively. The biotite exhibits minor amounts of mineralogical alteration and no apparent mechanical deformation. Grains were chosen for analysis based on lack of mineralogical alteration, determined optically in thin section. Biotite argon release spectra from EF9514a correspond to an age of $55.5 \pm 0.5$ Ma, defined by 70% of total released $^{39}\text{Ar}$ (Fig 2A). Argon release spectra from EF9513 correspond to an age of $53.6 \pm 0.5$ Ma, defined by approximately 90% of total released $^{39}\text{Ar}$ (Fig 2B).

Subsequently, dikes substantially more mafic than the early rhyodacites were emplaced within approximately 3 million years of EF9513 and EF9514a. EF9505, a clinopyroxene-hornblende granodiorite trending N60°E, shows some individual hornblende alteration to green hornblende or chlorite, and many grains are fractured. Analyzed grains, chosen for a lack of alteration and fracturing, yield spectra that show an excess of argon in the first heating step, corresponding to approximately 4% of total $^{39}\text{Ar}$.
EF9514A, Biotite

Fig. 2A

Apparent Age (Ma)

Cumulative % 39Ar released

55.5 ± 0.5 Ma
Calculated age: 52.5 ± 0.48 Ma
Fig. 2D

EF9301, Hornblende

51.8 ± 0.8 Ma

Cumulative % 39Ar released
released. Approximately 96% of total $^{39}$Ar released corresponds to an age of 52.5 ± 0.5 Ma (Fig. 2C).

The youngest East Fork dike identified in this study is EF9301, a hornblende basaltic andesite trending N40°E. Large, minimally fractured and unaltered hornblende grains from EF9301 yield an argon release pattern of 90% total $^{39}$Ar corresponding to 51.8 ± 0.8 Ma (Fig. 2D). These data agree with the crosscutting relationships.

East Fork Tectonic Relationships

The East Fork dikes currently exposed in the southern Skalkaho slab have an average trend of N47°E. This, however, is not the original emplacement trend. As the footwall of the Bitterroot core complex moved northwest relative to the Skalkaho slab, the trend of the swarm changed with synextensional rotation (Fig. 3). Paleomagnetic data from East Fork dikes and footwall dikes of the Bitterroot core complex demonstrate 25° of clockwise hanging wall rotation around a pole near the northern edge of the hanging wall slab, following the emplacement of the East Fork swarm (Doughty and Sheriff, 1992). This suggests some 40 km of southeast-relative translation following an initial emplacement trend of approximately N22°E. In addition, Doughty and Sheriff (1992) demonstrate 3° of counterclockwise synextensional rotation in the footwall. Measured trends, adjusted for the hanging wall and footwall rotation, agree with the 110° extensional direction indicated by stretching lineations in the Bitterroot mylonite.
Figure 3: Map of the Bitterroot metamorphic core complex and related features and structures. Adapted from Hyndman, 1983; Toth, 1987; Doughty and Sheriff, 1991.
Most studies concerning the onset of extension in southwestern Montana utilize dates from the Bitterroot core complex. Various workers have inferred this onset as between 54 - 43 Ma. For example, Garmezy and Sutter (1983) utilize $^{40}\text{Ar} / ^{39}\text{Ar}$ spectra from mylonitized granites to describe peak mylonitization between 45.5 and 43 Ma. Most chronologies document peak extension between 52-45 Ma, approximating the chronology from Chase et al. (1983) who describe three Pb-U dates from sheared granites in the mylonitic zone, $52 \pm 1$ Ma, $49 \pm 1$ Ma, and $48 \pm 1$ Ma. They suggest that these rocks underwent a major shearing event at least as young as 50 Ma, although they do not rule out older movement, either from the same or a different event. Toth (1987) suggests the transition from east-west compression to northwest-southeast extension is first evidenced with the emplacement of the 54 Ma Paradise pluton, approximately 45 km northwest of the East Fork swarm. Most recently, Foster and Fanning (1997) use new and composite data from other workers to suggest that early Tertiary compression gave way to extension at approximately 50 Ma, with peak extension and cooling at 45-47 Ma. These methods necessitate extrapolation back to initial extension, as the time lag between initial extensional stress and dike emplacement, and cooling to the mineral closure temperature may be significant. In the thermally elevated Bitterroot mylonite zone, which at peak conditions was at approximately 600° C (Kerrich and Hyndman, 1986), cooling lag is estimated at 1-3 million years (House and Hodges, 1994; House, 1995). However, the new data presented here provide a 55.5 Ma estimate for the onset of extension, suggesting a total delay as long as 4-7 m.y. This undoubtedly reflects not only the time between
unroofing and cooling to closure, but also time between initial extensional strain and significant unroofing which until now have not been bracketed.

**East Fork Geochemical Comparisons**

The East Fork dikes of this study fall into two compositional groups, peraluminous \( \frac{Al_2O_3}{CaO + Na_2O + K_2O} > 1 \) rhyodacites at ~55-54 Ma and more mafic, metaluminous \( \frac{Al_2O_3}{CaO + Na_2O + K_2O} < 1 \) compositions at ~52 Ma. East Fork dike geochemistry is compared with contemporaneous compositions from the East Fork area (Desmarais, 1983), the ~57 Ma Idaho batholith main-phase Bear Creek pluton, the ~54 Ma arguably main-phase Paradise pluton (Toth, 1983; Toth, 1987), and ~49 Ma syenogranitic epizonal plutons and volcanic rocks (Lund, 1980; Snyder, 1997).

The East Fork rhyodacites are approximately contemporaneous with late main-phase plutons of the Idaho-Bitterroot batholith, as dated by Chase et al. (1978), Bickford et al. (1981), Toth and Stacey (1991), House and Hodges (1994), House (1995), and Foster and Fanning (1997). Considering the volume of magmatism at this time, it seems plausible that the early dikes were directly related to this activity. The igneous unit closest in time and space to the East Fork dikes is described by Desmarais (1983). His 51-53 Ma Biotite Granodiorite suite (BGS), outcropping as a pluton and isolated dikes in the East Fork area, plot as high-K rhyodacites in the petrological classification scheme of Le Maitre (1989). A first-order major-element comparison between both compositional groups from the East Fork and the BGS is a satisfactory match, with compositions, plotting along a \( SiO_2 \)-dependent trend (Fig. 4A).
Figure 4A: Trend line analysis determined for all data points. Biotite granodiorite suite, Desmarais, 1983; Paradise pluton, Toth, 1987; Bear Creek pluton, Toth, 1987.
Figure 4A (cont.): Trend line analysis determined for all data points. Biotite granodiorite suite, Desmarais, 1983; Paradise pluton and Bear Creek pluton, Toth, 1987.
Figure 4B: Trend line analysis determined for all data points. Biotite granodiorite suite, Desmarais (1983); Paradise pluton, Toth (1987); Darby volcanics and epizonal plutons, W. Motzer, from Toth (1987), and Snyder (1997).
A: Indicates some data is reported at lower detection limits.
Figure 4B (cont.): Trend line analysis determined for all data points. Biotite granodiorite suite, Desmarais (1983); Paradise pluton, Toth (1987); Darby volcanics and epizonal plutons, W. Motzer, from Toth (1987), and Snyder (1997).

A: Indicates some data is reported at lower detection limits.
First-order major-element comparisons between the 57 Ma Bear Creek pluton, a main-phase pluton outcropping in the eastern edge of the Bitterroot metamorphic core complex, and the East Fork dikes provide positive trend correlations (Fig. 4A) (Toth, 1983; Toth, 1987; Foster and Fanning, 1997). Normative corundum values of > 1% and $\text{Al}_2\text{O}_3/(\text{CaO} + \text{Na}_2\text{O} + \text{K}_2\text{O})$ values ≥ 1.1 from the Bear Creek pluton (Toth, 1983) resemble those of the ~54 Ma East Fork rhyodacites, but in both cases are greater than values from the ~52 Ma dikes. However, trace-element comparisons between the Bear Creek pluton and both groups of East Fork dikes (limited data available for EF9301) do not provide satisfactory matches, and REE comparisons are indeterminate (Fig. 5A,B).

The 54 Ma Paradise pluton compares well with the ~52 Ma East Fork dikes (Toth, 1983; Toth, 1987). Major-element trend comparisons indicate a plausible magmatic relationship between the Paradise pluton, the East Fork dikes, and by inference, the BGS and possibly the Bear Creek pluton (Fig. 4A). Normative corundum values are < 1% (commonly 0%) and $\text{Al}_2\text{O}_3/(\text{CaO} + \text{Na}_2\text{O} + \text{K}_2\text{O})$ values ≤ 1.1 in the Paradise pluton and the intermediate dikes. Trace-element diagrams correlate more closely than with the Bear Creek data, although again, REE plots are somewhat indeterminate (Fig. 6A,B). In addition, it appears that restored to their original position (Fig. 3), the East Fork dikes were approximately aligned with the Paradise pluton. Trace-element and REE trends of dikes radial to the Bear Creek and Paradise plutons (Toth, 1983) generally resemble those of the Bear Creek pluton (Fig. 7A,B), contrary to the East Fork comparisons. However, although East Fork dike Europium anomalies are within the range of the Bear Creek
Figure 5A: Bear Creek pluton (Toth, 1983) and East Fork dikes, selected trace elements, Chondrite C1 normalized.
Figure 5B: Bear Creek pluton (Toth, 1983) and East Fork dikes, rare-earth elements, Chondrite CI normalized.
Figure 6A: Paradise pluton (Toth, 1983) and East Fork dikes, selected trace elements, Chondrite Cl normalized
Figure 6B: Paradise pluton (Toth, 1983) and East Fork dikes, rare-earth elements, Chondrite C1 normalized
Figure 7A: Fine grained granitic dikes (Toth, 1983)
and East Fork dikes, selected trace elements, Chondrite Cl normalized
Figure 7B: Fine grained granitic dikes (Toth, 1983) and East Fork dikes, rare-earth elements, Chondrite C1 normalized
Figure 8: Europium anomalies, measured Eu / expected Eu (Eu*); IMEB dikes, Challis volcanics (Norman and Mertzman, 1991), Paradise pluton, Bear Creek pluton, and fine grained granitic dikes (Toth, 1983)
pluton itself, the radial dikes tend to have significant negative anomalies (Eu/Eu* ≤ 0.7) (Fig. 8), much below that seen in the other groupings, and actually approaching those of the ~49 Ma syenogranitic plutons (c.f.: Lund, 1980). Toth (1983) does not offer an explanation for these anomalies in her study on the pluton and related dikes, and I hesitate to make correlations with no geochronometry on the radial dike set.

These comparisons indicate that the ~54 Ma dikes most closely match Paradise pluton trends, despite sharing some compositional characteristics with the Bear Creek pluton. The ~52 Ma dikes also compare most favorably with the Paradise pluton. These relationships suggest that magmatism associated with the Paradise pluton continued for as much as 4 m.y., lasting from ~55.5 to ~51.5 Ma. The limited data available do not identify a definite compositional progression with time in the East Fork, but in a compositional transition reflected regionally in the early Challis intermediate volcanic rocks (described below), differentiation in a parental magma may have produced the Paradise pluton, the BGS, and both East Fork compositional groups. Although Toth (1983, 1987) discusses the relationship of the Bear Creek and Paradise plutons, she reaches no definite conclusions.

Approximately 3 million years later, magmatic compositions changed to syenogranite, reflecting a regional trend. A biotite ⁴⁰Ar/³⁹Ar age from a biotite-rhyodacite of the Darby volcanics, 30 km northwest of the EFDS, and dating of granodiorite on which the volcanic rocks sits, indicate a transition back to felsic
magmatism by 48.8 ± 0.4 Ma (Snyder, 1997). This is precisely contemporaneous with the widespread hypabyssal plutonic episode of the Lolo Hot Springs batholith and Whistling Pig batholith Pb-U dated at 48.7 ± 0.3 Ma and 48.9 Ma, respectively (House, 1995). These represent a departure from compositional trends seen in earlier magmatism, most notably in a significant drop in Al₂O₃, MgO, and K₂O concentrations (Fig. 4B). Another distinguishing characteristic of the syenogranitic rocks, compared to most previous Tertiary igneous rocks in the Bitterroot region, is a marked negative Eu anomaly (Lund, 1980), suggesting feldspar fractionation at some point in the magmatic process.

**SHOUP DIKES**

Shoup, Idaho is located approximately 70 km southwest of the center of the East Fork dike swarm, and 30 km north of the line generally interpreted as the northern edge of the trans-Challis fault zone (Fig. 9). Country rocks in this area consist of a complex assemblage of Precambrian basement gneisses, schists, and the Proterozoic Yellowjacket Formation, as well as porphyritic and non-porphyritic Proterozoic granites (Davidson, 1928), that lie in the Salmon River Arch between the Atlanta and Bitterroot lobes of the Idaho batholith. It is a highly faulted area, with significant northeast-trending Tertiary normal faults, including the Shoup, Hot Springs, and Copper Mountain faults (Hughes, 1990). Some workers suggest that these may be re-activated Precambrian basement faults, that in Tertiary times contributed to the formation of the trans-Challis fault zone and the Great Falls tectonic zone (O’Neill and Lopez, 1985), although the interpretation
Figure 9: Selected schematic features of the Idaho-Montana extensional belt. Map adapted from Olson, 1968; Bennett, 1980; Kiilsgaard et al., 1986; Lewis and Kiilsgaard, 1991; Allen and Hahn, 1994; Janecke et al., 1997.
of Precambrian movement is disputed (T. Doughty, 1997, pers. comm.). The area hosts large numbers of dikes of various compositions, including a mineralized rhyolite swarm of Squaw Creek (S. Fuller, unpublished M.S. thesis mapping), and individual dacite and andesite dikes identified in generalized mapping (Davidson, 1928). As there have been no detailed studies in the area, it is likely that more Tertiary dikes outcrop in this area than have been mapped. In addition, there is little data from pre-50 Ma igneous activity in this area.

I present three new $^{40}\text{Ar}/^{39}\text{Ar}$ mineral dates from two dikes. Dike SH02a is located 0.75 km east of the Pine Creek road, west of Shoup. It is a N44°E-trending high-K hornblende-biotite dacite porphyry, emplaced into a Precambrian augen gneiss. Biotite and hornblende grains from SH02a were dated. Biotite is large, subhedral, and generally unaltered, although some individual grains show some pleochroic inhomogeneity, and sphene, rutile, and opaque inclusions that may indicate minor alteration. Hornblende is generally euhedral, and both green and brown grains are present. Unaltered brown grains were the primary dating target. For both minerals, grains containing inclusions and those exhibiting inhomogenous pleochroism were avoided for argon analysis.

The hornblende argon release spectra do not form a significant plateau. However, an inverse isotope correlation diagram indicates a cooling date of 53.4 ± 0.4 Ma. In contrast, biotite release spectra correspond to 50.2 ± 0.4 Ma, determined by 97% of total $^{39}\text{Ar}$ released (Fig 2E). These dikes were emplaced into a region characterized by continued, long term, voluminous magmatism. The resultant hot host-rock environment
Cumulative % $^{39}$Ar released

Apparent Age (Ma)

50.2 ± 0.4 Ma
Fig. 2F

SH01, Hornblende

Cumulative % 39Ar released

48.4 ± 0.5 Ma
slowed cooling to the biotite closure temperature. Thus, the hornblende age is considered closer to the emplacement date.

Sample SH01 is from a N18°E-trending fine-grained hornblende high-K andesite dike located in a large roadcut 1.3 km west of Shoup. The analyzed mineral sample is fine-grained subhedral brown hornblende. It is not significantly altered, although some grains contain inclusions of optically opaque minerals, and these grains were avoided in the argon analysis. This sample provides argon release spectra corresponding to a cooling date of 48.4±0.5 Ma, determined by 78% of total 39Ar released (Fig. 2F).

At 53.4 Ma, SH02a is one of the two oldest analyzed northeast-trending dikes in the Idaho portion of the IMEB. Previous workers have suggested that crustal extension in east-central Idaho and the development of the trans-Challis fault zone began just prior to or coeval with the Challis volcanics, approximately 51-50 Ma (McIntyre et al., 1982; Kiilsgaard et al., 1986; Bennett, 1986; Moye, 1988; Janecke, 1991; Janecke, 1992). However, these new dates indicate that brittle extensional deformation started in the Shoup area 2 - 3 m.y. prior to the Challis volcanic activity, a significant time period when referenced to total eruptive time of about 5 m.y.

**Shoup Geochemical Comparisons**

Eocene plutonic rocks in central Idaho are divided into two primary, genetically unrelated compositional groups, a compositionally variable 47-50 Ma quartz monzodiorite suite and a 44-48 Ma pink granite suite (Bennett and Knowles, 1985;
<table>
<thead>
<tr>
<th></th>
<th>Main-phase Idaho batholith</th>
<th>IMEB (this study)</th>
<th>Early phase Challis volcanics</th>
<th>Tertiary quartz monzodiorite, incl. Paradise pluton</th>
<th>Tertiary pink granite, incl. epizonal plutons</th>
</tr>
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<tbody>
<tr>
<td><strong>General Comp.</strong></td>
<td>two-mica (primary muscovite present in some phases)</td>
<td>rhyodacite to basaltic andesite,</td>
<td>intermediate to mafic lavas and tuffs</td>
<td>variable compositions, primary muscovite absent</td>
<td>one mica, hypersolvus granites</td>
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<td><strong>Ages</strong></td>
<td>Bitterroot: 95-53 Ma</td>
<td>55.5 - 45 Ma</td>
<td>47-51 Ma</td>
<td>47-50 Ma</td>
<td>44-48 Ma</td>
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<td>Atlanta: 112-70 Ma</td>
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<td><strong>A/C+N+K (mole%)</strong></td>
<td>&gt; 1.00, peraluminous</td>
<td>early EF = 1.1</td>
<td>&lt; 0.7, metaluminous</td>
<td>generally &lt; 1.00 (metaluminous to weakly peraluminous)</td>
<td>1.03 - 1.14 (peraluminous)</td>
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<td></td>
<td></td>
<td>(peraluminous)</td>
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<td></td>
<td></td>
<td>all others = .78-.97 (metaluminous)</td>
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<tr>
<td><strong>Eu/Eu</strong>*</td>
<td>mean = 0.86</td>
<td>mean = .91</td>
<td>mean = 0.85</td>
<td>mean = 0.85 (PP)</td>
<td>~ 0.50</td>
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<tr>
<td><strong>Rb/Sr</strong></td>
<td>mean ~ 0.12</td>
<td>mean = .236</td>
<td>mean = 0.18</td>
<td>0.08 - 0.20</td>
<td>mean = 0.89, all &gt; 0.38</td>
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<tr>
<td></td>
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<td>median = .144</td>
<td>median = 0.17</td>
<td>mean = 0.15</td>
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</tbody>
</table>

References: Fisher, 1995; McIntyre et al., 1982; Foster and Fanning, 1997; Bickford et al., 1981; Toth and Stacey, 1992; Norman and Mertzman, 1991; See Lewis and Kilsgaard, (1991) for detailed chemical comparisons

**Table 2**
Figure 10A: Paradise pluton (Toth, 1983) and Shoup-area dikes, selected trace elements, Chondrite C1 normalized
Figure 10B: Paradise pluton (Toth, 1983) and Shoup-area dikes, rare-earth elements, Chondrite C1 normalized
Figure 11A: Mafic and intermediate-composition Challis volcanic rocks (Norman and Mertzman, 1991) and Shoup-area dikes, selected trace elements, Chondrite Cl normalized
Figure 11B: Mafic and intermediate-composition Challis volcanic rocks (Norman and Mertzman, 1991) and Shoup-area dikes, rare-earth elements, Chondrite C1 normalized
Bennett and Lewis, 1990; Lewis and Kiilsgaard, 1991). The quartz monzodiorite suite is further divided into mafic and intermediate subgroups. No temporal relationships between the two subgroups have been suggested (Lewis and Kiilsgaard, 1991).

Both analyzed dikes from the Shoup area exhibit characteristics of the quartz monzodiorite suite (Table 2), and major-element characteristics of the 53.4 Ma SH02a and the 48.4 Ma SH01 correlate well with the intermediate subgroup and mafic subgroup, respectively (Fig. 14). REE and trace-element data for the quartz monzodiorite suite are sparse, so definitive chemical comparisons are difficult. However, comparison with the contemporaneous 54 Ma Paradise pluton, which may be the earliest identified body of the quartz monzodiorite suite (Lewis and Kiilsgaard, 1991), shows that indeed they may be related. As is characteristic of the mafic subgroup, SH01 underwent a lesser degree of REE differentiation (Fig. 10A,B). REE and trace-element comparisons of both dikes to 50-51 Ma Challis intermediate volcanic rocks are somewhat closer (Norman and Mertzman, 1991) (Fig 11A,B), yet magmatic associations any more specific than the quartz monzodiorite subgroups are difficult to determine.

**Seafoam Dike Swarm**

The Seafoam dike swarm outcrops approximately 40 km northwest of Stanley, Idaho, in the heart of the trans-Challis fault zone. I sampled Seafoam dikes in the northeastern corner of the Idaho Porphyry Belt, approximately 112 km southwest of the Shoup sample area. The swarm is emplaced into 72-65 Ma leucocratic granite, a late-
Calculated age: 53.4 ± 0.5 Ma
Stage differentiate of the Idaho batholith biotite granodiorite (Kiilsgaard and Bennett, 1995), and consists of a variety of rhyolite, dacite porphyry, and andesitic dikes with an average trend of N30°E (Olson, 1968). The dikes are located 20 km northwest of the northeast-trending 48-47 Ma Custer graben (Allen and Hahn, 1994) (Fig. 9).

I present two $^{40}$Ar/$^{39}$Ar mineral analyses from the Seafoam swarm. Sample SF9517a is a hornblende high-K rhyodacite dike trending N23°E, located 3 km west of the Seafoam Ranger Station. Hornblende grains from this sample are brown, euhedral, and commonly fractured. These grains do not provide a suitable argon release spectral plateau. However, an inverse isotope correlation diagram defines a cooling date of 53.4 ± 0.4 Ma (Fig. 2G). A second hornblende $^{40}$Ar/$^{39}$Ar date from the Seafoam swarm, sample SF9518b, is from a high-K hornblende dacite dike trending N15°E. This sample provides neither a good plateau on an age spectrum plot nor a definitive isochron plot, and must be discarded.

Seafoam Tectonic Relationships

At 53.4 Ma, dike SF9517a is the oldest identified extensional feature in the central portion of the trans-Challis fault zone. A significant nearby extensional structure, the Custer graben, formed in response to magma expulsion and normal faulting, with voluminous K-rich andesitic Challis volcanic rocks erupting from numerous vents ~ 51-50 Ma (McIntyre et al., 1982). These were followed by quartz latite, rhyolite, and pyroclastic debris deposited during actual graben subsidence 48-47 Ma (McIntyre and Johnson, 1985; Allen and Hahn, 1994). Measured directions for extension vary. Allen
Figure 12A: Mafic and intermediate-composition Challis volcanic rocks (Norman and Mertzman, 1991) and Seafoam dikes, selected trace elements, Chondrite Cl normalized.
Figure 12B: Mafic and intermediate-composition Challis volcanic rocks (Norman and Mertzman, 1991) and Seafoam dikes, rare-earth elements Chondrite C1 normalized.
and Hahn (1994) describe initial extensional faulting and dike emplacement in the eastern Custer graben as N25°-40°E, followed by counterclockwise rotation to the northwest and subsequently a return to the northeast, but do not provide precise dates. Janecke (1992) describes ~N45°E-trending faults in the Lost River Range 35 km east of the Custer graben site which were active between 49-48 Ma (Janecke and Snee, 1993). Allowing for some local structural influence, the ages and trends of the Seafoam dikes indicate that some 4-6 m.y. passed between the onset of northwest-directed extensional forces in this area and coeval peak-rate extension and Challis volcanism.

Seafoam Geochemical Comparisons

Early volcanic rocks in the Custer graben are part of the Challis volcanic group, and along with the undated Warm Springs, Red Mountain, and Marsh Creek stocks, belong to the Idaho quartz monzodiorite suite (Table 2). Major-element characteristics and the metaluminous $\frac{Al_2O_3}{(CaO + Na_2O + K_2O)} < 1$ ratio of SF9517a and SF9518b also place them in the quartz monzodiorite suite; specifically, in the intermediate subgroup (Fig. 14).

Overall REE and trace-element patterns of SF9517a compare very well with Challis volcanic rocks (Fig. 12A,B), although the dike’s Eu/Eu* = 0.96 is slightly higher than the Challis range (Eu/Eu* = 0.74-0.85, see Fig. 8) (c.f.: Norman and Mertzman, 1991). I believe that SF9517a represents an early phase of quartz diorite suite magmatic
activity in this area, while the parent magma body continued to undergo some minor feldspar fractionation.

Trace element content of SF9518b compares very well to the early Challis rocks, and although overall REE patterns indicate a slightly lesser degree of rare-earth differentiation than early phase Challis volcanics, a value of Eu/Eu* = 0.85 is correlative to the volcanic rocks (Norman and Mertzman, 1991). These similarities suggest that SF9518b was probably emplaced during main intermediate Challis volcanism, which peaked at about 49 Ma (Fig. 12A,B).

Other local intrusive rocks, such as the Casto pluton, the Banner Creek stock, and the Sawtooth pluton are all part of the later pink granite suite (Lewis and Kiilsgaard, 1991), and are not directly related to the early dikes (Table 2).

BOISE BASIN DIKE SWARM

The Boise Basin dike swarm is well exposed along state highway 21, on the South Fork of the Payette River 15 km west of Lowman, Idaho (Olson, 1968). It is approximately 75 km southwest of the Seafoam dike swarm, at the southwestern end of the Idaho Porphyry Belt (Fig. 9). The large, dense, northeast-trending Boise Basin swarm is 2 to 3 miles wide, and consists of rhyolite, dacite, and isolated mafic dikes (Olson, 1968). The host rock is an assemblage of biotite granodiorite and quartz monzonite of the Idaho batholith and early Eocene plutons, including the Long Gulch, Birch Flat, and Boise Basin stocks (Anderson, 1947; Lewis and Kiilsgaard, 1991).
I present 5 new $^{40}\text{Ar}/^{39}\text{Ar}$ mineral dates from 3 northeast-trending dikes in the Boise Basin dike swarm. Sample BB9521b is a hornblende-biotite high-K dacite dike trending N45°E, located 1.6 km west of the Deadwood River along Highway 21 (Fig. 9). Host rock is biotite granodiorite of the Idaho-Atlanta batholith. Both biotite and hornblende from BB9521b were analyzed. Many biotite grains show some chloritization, although essentially unaltered grains are available for analysis. The biotite provides an argon release pattern corresponding to a cooling date of 50.1 ± 0.9 Ma, defined by 94% of total $^{39}\text{Ar}$ (Fig. 2H). Similarly, much of the hornblende in BB9521b is altered. However, brown optically-homogenous hornblende, without inclusions and with minor green rims was analyzed. This hornblende did not provide a sufficient argon release plateau, although 42% of total $^{39}\text{Ar}$ released corresponds to an emplacement date at 51.0 ± 1.5 Ma, within the 95% confidence range of the biotite analysis. The 1 Ma discrepancy resembles that seen in many Eocene intrusive rocks of this region and other IMEB dikes. Due to the statistical limitations, however, the hornblende date is not considered reliable.

Sample BB9525a is a biotite-hornblende high-K dacite dike sampled approximately 5 km west of the Deadwood River along highway 21 (Fig. 9). The dike trends N18°E and is emplaced into biotite granodiorite of the Idaho-Atlanta batholith. Biotite and hornblende from BB9525a were analyzed. Much of the biotite is substantially chloritized or contaminated with optically opaque minerals. However, some biotite grains are optically homogenous and show no visible alteration, and these provide an argon release spectra corresponding to a cooling date of 47.0 ± 0.6 Ma., defined by
Fig. 2H

BB9521B, Biotite

Cumulative % 39Ar released

0 10 20 30 40 50 60 70 80 90 100

Apparent Age Ma

50.1 ± 0.9 Ma
BB9525A, Hornblende

Fig. 2J

Cumulative % 39Ar released

Apparent Age Ma

49.3 ± 0.6 Ma
BB9509, Biotite

Fig. 2K

Apparent Age (Ma) vs. Cumulative % 39Ar released

45.5 ± 0.4 Ma
98% of total $^{39}$Ar released (Fig. 2I). Hornblende from BB9525a is brown, commonly with green hornblende portions and rims. Many of the brown, homogenous grains show no evidence of alteration. These hornblende grains provide an argon release spectra corresponding to a cooling date of 49.3 ± 0.6 Ma, defined by 74% of total $^{39}$Ar released (Fig. 2J). Again, the discrepancy between biotite and hornblende in BB9525a is similar to that region wide, and the hornblende date of 49.3 ± 0.6 Ma is considered closer to the actual emplacement date.

Sample BB9509 is a biotite-quartz high-K basaltic-andesite dike trending N37°E, sampled 100 meters east of Big Pine Creek Rd (Forest Road #555) along Highway 21, west of Lowman (Fig. 9). Although this dike shows alteration effects, such as chloritization and secondary pyrite development, much of the biotite shows only the earliest stages of alteration, and some is essentially pristine. Despite the possible alteration problems, biotite provides an argon release spectra corresponding to a cooling age 45.4 ± 0.4 Ma, defined by 91% of total $^{39}$Ar released (Fig. 2K).

Data concerning extensional timing and directions are sparse in the Boise Basin, and no published dates indicate the timing of strain southwest of the Seafoam area. The dikes analyzed here, however, indicate that extensional strain started no later than 50-51 Ma, and continued with a similar trend for at least 5 million years. This seems to indicate a progression in strain transfer from the Custer graben and Seafoam region to the Boise Basin swarm. While this may reflect regional or local structural influences, it could also
be a function of dike selection, if the analyzed dikes were not the oldest of the swarm. See below for a discussion of strain transfer.

Boise Basin Geochemical Comparisons

The timing and major-element compositions of the two early Boise Basin dikes, BB9525a and BB9521b, indicate they belong to the intermediate subgroup of the quartz monzodiorite suite (Table 2 and Fig. 14). Several plutons of this subgroup are in the Boise Basin swarm area, most notably the Boise Basin and Birch Flat stocks. However, REE data is unavailable and trace-element analyses are very limited for these rocks. Compared to the intermediate Challis compositions, overall REE and trace-element patterns from the dikes are a close match (Fig. 13A,B). One exception to this is a less negative Eu anomaly in the Boise Basin dikes, Eu/Eu* ~ 0.90 for the dikes versus 0.74 - 0.85 for the Challis (c.f.: Norman and Mertzman, 1991). The close chemical and temporal characteristics indicate that the Challis magmatic system operating at about 50 Ma was not localized under the volcanic terrain, but instead was more regional in scope.

BB9509, emplaced 4-5 m.y. after the initial Boise Basin dikes, exhibits major-element characteristics of the mafic subgroup of the quartz monzodiorite suite (Fig. 14). However, unlike the earlier Boise Basin dikes, trace-element trends of BB9509 do not unequivocally correlate with the Challis volcanic rocks. Although comparisons do not preclude correlation, some element concentrations are unusual for the Challis rocks and the associated IMEB dikes (c.f.: Norman and Mertzman, 1991) (Fig. 13A,B). In addition,
Figure 13A: Mafic and intermediate-composition Challis volcanic rocks (Norman and Mertzman, 1991) and Boise Basin dikes, selected trace elements Chondrite C1 normalized.
**Figure 13B**: Mafic and intermediate-composition Challis volcanic rocks (Norman and Mertzman, 1991) and Boise Basin dikes, rare-earth elements Chondrite C1 normalized.
Figure 14: Trend line analysis determined for the quartz monzodiorite suite only.
Figure 14 (cont.): Trend line analysis determined for the quartz monzodiorite suite only. Data for quartz monzodiorite suite from: McIntyre et al., 1982; Norman and Mertzman, 1991; Lewis and Kilsgaard, 1991.
although the REE distribution is generally within Challis limits, some variances exist, especially in Eu concentrations. As REE and trace-element data from other mafic subgroup plutons is scarce, more specific comparisons are difficult.

Regional Interpretations

Chronology

South of the Lewis and Clark Line, the northernmost well-studied extensional feature is the Bitterroot metamorphic core complex. Available data does not unequivocally define an oldest limit for the onset of extensional deformation in the Bitterroot core complex. However, the 57 Ma Bear Creek pluton provides a baseline from which subsequent unroofing can be measured. The Bear Creek pluton was emplaced at 10-15 km depth (Toth, 1987), in the east-center of the core complex, north of the original emplacement location of the East Fork dikes. This is the oldest pluton for which depth-of-emplacement data is available, and thus it is not possible to determine pre-Bear Creek depth-of-emplacement trends. However, Toth (1987) suggests that the 54 Ma Paradise pluton was emplaced into the Bear Creek pluton at depths of less than 10 km, implying up to 5 km of unroofing between the emplacements of the two plutons. This infers unroofing rates that are plausible for either tectonic processes or surficial erosion. In addition, using $^{40}$Ar/$^{39}$Ar cooling ages, House (1995) suggests that cooling of the western Bitterroot metamorphic core complex and by inference, extension, started by 55.6 Ma. Given ~55 Ma extension indicated by East Fork dike ages of 55.5 Ma and 53.6
Ma, I suggest that extensional unroofing began before the emplacement of the Paradise pluton, initiating deformation that later included the 110° stretching lineations in the Bitterroot mylonite. Significant extension is evident during the subsequent 6 m.y. Between the 55.8 Ma crystallization (monazite $^{207}\text{Pb}/^{235}\text{U}$) of the mesozonal Skookum Butte stock in the northeastern Idaho-Bitterroot batholith, and the 48.8 Ma crystallization (zircon $^{206}\text{Pb}/^{238}\text{U}$ and $^{207}\text{Pb}/^{235}\text{U}$) of the epizonal Whistling Pig pluton to the southwest, as much as 12-18 km overburden was removed (House, 1995). However, peak tectonic unroofing and subsequent cooling was not recorded in Bitterroot mylonite rocks until 47-43 Ma (Garmezy, 1983; House and Hodges, 1994; Foster, 1995).

South and east of the Idaho-Bitterroot batholith in the Shoup area, dike emplacement demonstrates that extension commenced by ~50.2 Ma at the latest, and possibly as early as 53.6 Ma, an age supported by the 53.4 Ma date from the Seafoam swarm farther southwest. This is about 2-5 m.y. later than initial brittle deformation at the emplacement location of the translocated East Fork swarm, and 20° to the south of the 110° direction indicated in this and other studies. The apparent time difference may be due to any of several reasons. First, temporal differences may be due to migration of strain towards the south, on a smaller scale than has been suggested by previous workers for the Cordillera as a whole. Another possibility is that rheological differences between Idaho batholith host rocks in the north and Precambrian basement and meta-sediments of the Salmon River Arch in the south affected the transfer of stress across the Idaho arc (see
Foster and Fanning, 1997). However, the primary reason for the temporal difference may lie in sampling. Three dike ages do not provide a statistical base, and it is possible that older dikes are emplaced in the Shoup area. Conclusions based on the temporal differences at this point would be speculative.

I believe the anomalous N45°E Shoup dike trend is due to local structural interference with the regional stress field. This is supported by the timing and trend of an approximately contemporaneous dike in the Seafoam area trending ~N20°E, and the later Shoup dike trending N18°E. Both of these are consistent with regional strain measurements, indicating 110° extension. These trends suggest that the strain at Shoup was similar to that in the East Fork, and underwent local refraction or rotation.

By about 50.5 Ma, initial intermediate and mafic Challis volcanic rocks were erupted in the northern part of the Challis field 50 km to the south of the Shoup area (McIntyre et al., 1982; Janecke and Snee, 1993) (Fig. 9). Volcanism continued for over 10 m.y., with most of the activity occurring in at least two and possibly a third pulse, between 49 to 45 Ma (McIntyre, 1985; Janecke, 1991; Fisher and Johnson, 1995). Previous workers suggest that extension began as much as ~3 m.y. prior to this volcanism (Moye, 1988). Subsequently, greatest extensional rates in both the trans-Challis fault zone and to the southeast in the Lost River area occurred at approximately 48 - 49 Ma (McIntyre et al., 1982; Janecke, 1991). Coeval maximum trans-Challis fault system extension and volcanism are reflected in the subsidence of the northeast-trending Custer
and Panther Creek grabens and 48.5 Ma eruptions of voluminous intermediate and silicic ash-flows from the Van Horn caldera. Cross-cutting relationships in volcanic strata show that the degree of extensional faulting dropped precipitously just after 48 Ma (Janecke, 1991). This temporal pattern is somewhat older than maximum extension in the Pioneer core complex to the southeast, where unroofing occurred between 48 and 45 Ma at a maximum rate of 1.5 km/Ma. (Wust, 1986; Silverberg, 1988, 1990). Janecke (1992) documents a subsequent, second distinct extensional vector, her ‘episode 2’, originally suggested by Hait (1984), in east central Idaho. \(^{40}\)Ar/\(^{39}\)Ar dates from synextensional tuffs in the Pass Creek fault system show that the second phase of extension, directed 90° from the initial extension and producing N-NW-striking, down-to-the-west normal faults, rapidly replaced the initial extensional vector at 48-46 Ma in east-central Idaho (see Fig. 15).

In the Seafoam dike swarm, the N23°E dike age of 53.4 Ma suggests that 110° extension was active in this region at approximately the same time as in the Shoup area. However, in the Boise Basin swarm to the southwest, dike ages do not fit the temporal pattern established by the other IMEB dikes. While the oldest dike sets were identified as such and sampled based on field determinations and relative age work of Olson (1968), the oldest dikes analyzed in the Boise Basin swarm are 50 - 51 Ma, ~ 2 - 5 m.y. younger than other swarms in the present study. As with the Shoup dikes, the difference in dike timing may be due to one of several reasons. As mentioned above, the discrepancy may
Figure 16: Dates indicate timing of pre-Basin and Range style extension periods, as determined through radiometric methods applied to extensional features. All dates (Ma).

reflect statistical limitations of analyzing only three dikes in the very large Boise Basin dike swarm. However, I assume that these dikes represent the oldest extensional features in the Boise Basin swarm, based on observed field relations and previous work (Olson, 1968). Therefore, the disparity could reflect a delay in the initial transfer of stress through the young, hot Idaho-Atlanta batholith that was emplaced between the Boise Basin and the remainder of the porphyry belt, and that was undergoing significant faulting syn- and post-dike emplacement. Evidence of subsequent stress field rotation supports this ‘delay of transfer’ speculation. Cross-cutting relationships in the Boise Basin swarm show younger, more north- and northwesterly trending dikes, indicating that the stress field rotated to west-southwest extension (episode 2 of Janecke (1992)). However, the youngest dike (trending N37°E) analyzed in the Boise Basin swarm suggests that the initial 110° extensional stress field was still in place as of ~ 45.4 Ma. This is 0.5 - 2.5 m.y. later than the documented stress rotation to the northeast in the Challis volcanic region. Although this is not a large difference, it is possible that rheological effects of the cooling batholith and significant faulting that later would delay rotation of the stress field may have also delayed transfer of initial stress to the area at 50 Ma, causing a gradual migration of strain effects south through this portion of the IMEB.

Integrating IMEB extension with regional Eocene tectonics

Following the cessation of Mesozoic WSW-ENE shortening, the onset of the initial, NW-SE-directed Cenozoic extension was not synchronous over the North
American Cordillera. Nor was the extent and direction of extension always similar between regions. The question arises as to the temporal relationship of the Idaho portion of the magmatic arc, where initial extensional timing and direction is defined in the IMEB, and regions immediately to the north and south.

The southern Omineca Belt, consisting of a complex assemblage of Tertiary core complexes, fault zones, and intrusions, terminates in northern Idaho at a regional scale northwest-trending fault system, the Lewis and Clark line (LCL) (Fig. 16). The timing and relative motion on major faults of the LCL is the subject of debate. Sheriff et al. (1984) suggest the LCL is an intercratonic transform fault system, involving a total of 40 km of right-lateral slip that compensates for observed crustal thickness variations across the line. Other workers describe primarily normal movement on the Hope fault, a major fault just north of and subparallel to much of the LCL (Yin, 1991; Fillipone and Yin, 1994; Fillipone et al., 1995). In either case, early Eocene extension was active north and south of the LCL at slightly different times. Following abrupt, widespread cessation of thrusting north of the LCL, extension commenced variably between 58 Ma in the Valkyr shear zone of the Valhalla Complex and ~52 Ma in the Okanagan complex, southern British Columbia (Parrish et al., 1988), and 54 - 52 Ma on the Newport fault zone in northern Idaho (Harms and Price, 1992).

Although this extension is broadly contemporaneous with that identified by this study in the IMEB, the new precision provided herein indicates that extension north of
the Lewis and Clark line began up to 3 m.y. prior to that in the East Fork swarm. In addition, extension halted in many portions of the Omineca belt significantly before that to the south, slowing or ceasing altogether in the Valhalla by 51 Ma and in the Okanagan complex, northern Monashee Mountains, and the Newport fault zone by 45 Ma (Sevigny et al., 1988; Harms and Price, 1992). This contrasts with active, non-Basin and Range extension continuing near the IMEB until at least 43 Ma, and possibly as late as 33 Ma in the Pioneer core complex, the only suggested Oligocene extension north of the Snake River Plain (Silverberg, 1990). There is however some doubt as to the validity of the Pioneer core complex dates (Janecke, 1991). The identified temporal trend consists of initial extension north of the LCL followed soon by extension in the Idaho platelet, which also continued later (Figure 16).

South of the Snake River Plain, however, determining the origin of extensional features becomes more difficult, as extreme Miocene Basin and Range deformation has overprinted and obscured most early Tertiary structures. Most pre-Miocene (pre-Basin and Range) extensional features, in particular metamorphic core complexes, appear to be from latest Eocene and Oligocene times (Coney, 1980, 1987). However, several workers identify early and mid-Eocene extensional features. For instance, Thorman and Snee (1988) describe ~55 Ma extension in the Wood Hills of northeastern Nevada. Lee (1993) and Mueller and Snoke (1993) document ~ 50 Ma initial detachment and cooling related to normal faulting events in the northern Nevada Snake Range and Windemere Hills,
respectively. In addition, Dallmeyer et al. (1986) describe a ~ 45 to 20 Ma cooling event associated with mylonitized Mesozoic igneous and metamorphic rocks in the Ruby Mountains-East Humboldt Range, Nevada, but the period from 30 to 20 Ma may represent initial Basin and Range-style extension. These cooling dates indicate that initial extension south of the Snake River Plain was approximately contemporaneous with that in central Idaho (see Janecke, 1994). However, in some areas south of the Snake River Plain, early to mid-Tertiary extension continued substantially later than in regions farther north, possibly continuing to and merging with late Oligocene and Miocene Basin and Range-style deformation. North of the Snake River Plain, however, the two extensional phases may have been separated by as much as 20 m.y. or more (Coney, 1987).

Although some workers dispute the significance of these temporal variations (Janecke, 1994), the more precise dates presented herein indicate a very rapid, though discernible, strain migration north to south across the LCL, with a temporal convergence of the main extensional periods continuing into northern Nevada (Fig. 16). In addition, although it is apparent that there was only a slight delay in the onset of initial early Tertiary extension from north to south, the pre-Basin and Range extensional phase was substantially longer in the northern Great Basin than in regions north of the Snake River Plain.
Summary and Conclusions

1. In each area studied, the earliest dikes predate previously dated extensional features. In most cases, the new dates demonstrate that the onset of extensional deformation pre-dated significant normal faulting and volcanism by several million years.

2. Episode 1 of extension and dike intrusion began at about 56-55 Ma in the eastern Bitterroot metamorphic core complex, 54-53 Ma in the central trans-Challis fault system, and possibly as late as 50-51 Ma at the southwestern end of the IMEB, in the Boise Basin dike swarm.

3. The data indicate that the onset of NW-SE extension was delayed up to 3 million years in the Boise Basin area. This was followed by a delay of a similar amount of time in rotation of the stress field at the onset of episode 2. These delays may have been caused by strain accommodation due to rheological effects of the still hot Idaho-Atlanta batholith and significant syn-dike normal faulting between the Boise Basin swarm and the remainder of the IMEB.

4. All but the latest IMEB dikes belong to the quartz monzodiorite compositional unit, which includes the Paradise pluton and early phase Challis intermediate volcanic rocks. The affiliation of the latest IMEB dike is indeterminate, although it clearly does not belong to the pink granite compositional unit.
5. On a northern Cordilleran scale, the more precise dates presented herein illustrate a very rapid but discernible southward progression of strain through time from the Omineca belt of southern British Columbia south across the Lewis and Clark Line and into central Idaho. Central Idaho and northern Nevada appear to have a synchronous initiation of extensional deformation, between 55 and 50 Ma.
References


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Appendix A
Outcrop locations and rough sample descriptions.

**EF9505** Dike trending N60°E @ 45°46.92'N, 113°46.76'W, in Schultz Saddle, at the head of Meadow Creek in the East Fork Bitterroot River drainage, southeast of Sula, MT
ROCK: Clinopyroxene-hornblende granodiorite
THIN SECTION: Phaneritic, numerous brown hornblende grains with green hornblende and chlorite rims, and a fair number of chlorite grains in contact with titaniferous phases such as sphene, replacing biotite. Plagioclase crystals, with quartz and corroded K-feldspar in interstices.
  Dated hornblende: Brown euhedral crystals, rimmed with green hornblende and chlorite. Many grains are fractured.

**EF9514a** - Dike trends N45°E @ 45°56.25'N, 113°44.14'W in the East Fork Bitterroot River drainage, east of Sula, MT.
ROCK: Porphyritic biotite rhyodacite
THIN SECTION: Biotite and corroded feldspar phenocrysts in a fine grained matrix. Assorted chloritized minerals.
  Dated biotite: Primarily euhedral. Some alteration is present, seen in opaque contamination and broad zones of uneven extinction in grains perpendicular to cleavage. In addition, some grains show green and red zones, possibly the result of alteration. Grains are mechanically intact.

**EF9513** - Dike trends N44°E @ 45°56.16'N, 113°44.46'W, in the East Fork Bitterroot River drainage, east of Sula, MT.
ROCK: Slightly porphyritic biotite rhyodacite
THIN SECTION: Numerous small-medium biotite crystals, corroded plagioclase in v. fine grained matrix. Occasional granophyre and anhedral quartz grains filling interstices.
  Dated Biotite: Generally unaltered, euhedral crystals. Some grains show inhomogenous coloration and/or opaque and rutile contamination.

**EF9301** - Dike trends N40°E, @ 45°52'N, 113°47'W, in Meadow Creek of the East Fork Bitterroot River, MT dike swarm. Dike dips ~70° NW.
ROCK: Clinopyroxene-hornblende basaltic andesite
THIN SECTION: Substantial chloritized minerals in v. fine grained groundmass, with large hornblende and small clinopyroxene phenocrysts.
  Dated hornblende: Large (1mm-1cm) brown euhedral laths, all fractured somewhat but unaltered.
Shoup (2)

SH01 - Dike trends N18E @ 45°22.98’N, 114°16.23’W, 1/2 mile west of Shoup, ID.
ROCK: Hornblende andesite
THIN SECTION: Numerous plagioclase phenocrysts and indistinct chloritized masses, surrounded by small subhedral brown hornblende lathes, in a fine grained matrix. No biotite, although much of the chlorite may be altered biotite.

  Hornblende: This rock may contain as much as 15% hornblende, with most of it as fine grained subhedral crystals. Much of the hornblende is greenish-brown, but does not appear to be chloritized. Some grains are contaminated with opaque material.

SH02a - Dike trends N44E, 70° SE dip, @ 45° 21.63’N, 114°17.66’W, southwest of Shoup, Idaho.
ROCK: Biotite-hornblende dacite porphyry
THIN SECTION: Prominent biotite, green and brown hornblende, plagioclase and quartz phenocrysts in a v. fine grained matrix.

  Dated biotite: Mostly unaltered. However, some individual grains show pleochroic inhomogeneity that may indicate slight alteration. In addition, some grains have rutile, sphene, and/or opaque contamination.

  Dated hornblende: Brown and green, mostly euhedral and unaltered. Some hornblende grains have significant embedded opaque minerals. Several instances of green hornblende with a brown hornblende (high temperature phase) rim; also, brown hornblende with a very thin green rim.

Seafoam Dike Swarm (2)

SF9517 - Dike trends N23E @ 44° 31.21’N, 115° 06.70’W, ~2 miles west of the SeaFoam ranger station, NW of Stanley, Idaho.
ROCK: Hornblende rhyodacite
THIN SECTION: Heavily corroded feldspar phenocrysts, fractured and chloritized hornblende, in a fine grained matrix. No biotite. The rock in general is quite altered.

  Dated hornblende: Euhedral and fractured.
SF9518b - Dike trends N15E @ 44° 33.73'N, 115° 04.18'W, northwest of Stanley, ID.
ROCK: Greenish gray hornblende dacite.
THIN SECTION:
   Dated hornblende: Large brown euhedral lathes. Some grains are substantially fractured and shattered, but there are also many complete grains. Some show slight chloritic alteration; also, some have green hornblende crystallized in layers or sections at the exterior of the grain.

Boise Basin, S. Fork Payette River

BB9509a - Dike trends N37E @ 44° 04.23'N, 115° 45.21'W on the South Fork of the Payette River, west of Lowman, ID, west of the Deadwood River.
ROCK: Biotite quartz latite
THIN SECTION: Large biotite phenocrysts in a chlorite/feldspar matrix. Everything except the biotite shows significant alteration. 1-5mm pyrite grains are present.
   Dated biotite: Euhedral and subhedral grains. Many grains show indications of earliest alteration, such as chlorite along grain boundaries, although most are large enough to provide substantial unaltered areas.

BB9521b - Dike trends N45E @ 44°04.36'N, 115°40.15'W on the South Fork of the Payette River, west of the Deadwood River.
ROCK: Gray hornblende-biotite latite
THIN SECTION: Numerous large plagioclase phenocrysts, biotite and hornblende. Many completely altered grains are present, in particular the feldspars, which are often totally sericitized.
   Dated biotite: Selected grains are usable for dating.
   Dated hornblende: Much of the hornblende is quite altered, although there are some grains that are in usable shape. Some of the hornblende has green rims.
BB9525a - Dike trends N18E @ 44° 03.75’N, 115° 41.90’W west of Loman, ID, on the South Fork of the Payette River.

ROCK: Gray biotite hornblende latite porphyry

THIN SECTION: Hornblende, biotite, plag and quartz in a v. fine grained feldspar/quartz matrix. A fair amount of alteration is present on most minerals.

  Biotite: Much of it has gone to chlorite, and almost all grains are pocked or contain opaque minerals to a certain extent. In addition, rutile needles embedded in otherwise clean biotite grains indicate some expulsion of titanium, possibly during the early stages of chloritic alteration.

  Dated hornblende: Brown hornblende commonly has green portions and rims, although it is mostly green hornblende, and not chlorite. Brown portions of the large grains show no alteration effects.
Appendix B:

Argon release spectra and Isotope correlations
SH01, Hornblende

$\frac{^{36}Ar}{^{40}Ar}$ vs. $\frac{^{39}Ar}{^{40}Ar}$

<--- Air
SH02A, Hornblende

Apparent Age Ma

Cumulative % 39Ar released
SH02A, Hornblende

36Ar/40Ar vs. 39Ar/40Ar

Calculated age: 53.6 ± 0.5 Ma
SH02A, Biotite

36Ar / 40Ar

0.0035

0.0030

0.0025

0.0020

0.0015

0.0010

0.0005

0.0000

0.00 0.02 0.08 0.10 0.12 0.18 0.20 0.22 0.24 0.26

39Ar/40Ar

--- Air
Calculated age: 49.8 ± 0.4 Ma
BB9509, Biotite

36Ar / 40Ar

--- Air
BB9521B, Hornblende

Cumulative % $^{39}$Ar released

Apparent Age Ma

51.0 ± 0.8 Ma
BB9521B, Biotite

36Ar / 40Ar

39Ar/40Ar

<--- Air
BB9525A, Biotite

--- Air
Appendix C:

Major, trace, and rare earth element compositions, and CIPW norms.
## Major Element Analyses (Weight %)

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<th>EF9505</th>
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<th>SH02A</th>
<th>SF9517</th>
<th>SF9518b</th>
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Appendix D: Suggestions for future research

1. Isotope geochronology research on individual dike swarms, especially in swarms that contain various dike trends, will provide a better database on temporal changes in crustal stress regimes. This work should be supported by detailed mapping of swarms. In addition, a statistical trend analysis would provide insights into regional vector changes with time.

2. Dikes are mapped in the West Fork of the Bitterroot River. They are possibly emplaced into both the footwall and the hanging wall of the core complex. Geochronologic data from these dikes would provide a more precise estimate of deformational timing and stress trends on the southern portion of the Bitterroot mylonite zone.

3. Areas north and south of the Lewis and Clark line seem to have been extending at slightly different times. This seems to indicate that the LCL might have been acting as an intracontinental transform, as per Sheriff et al. (1984). Are there extensional structures near the Idaho portion of the LCL, particularly on the south side, that might be studied to further constrain differential extension across the line?
4. The relationship between pre-50 Ma magmatism and extensional tectonism is enigmatic in much of the northern Cordillera. This is especially true in central Idaho, where relatively little geochronologic and geochemical data is available. Some workers suggest that thermal weakening precipitated by magmatism initiated extension. Others suggest that extensional unroofing resulted in significant pressure relief melting. It seems likely that both processes operated in the Eocene magmatic arc. To what degree was each process active? Separate from the main-phase Idaho batholith and the Challis volcanic rocks, large scale geochemical and textural studies of the earliest Tertiary plutonic suite, the quartz monzodiorite, have not been done. This data would provide regional insights into magmatic origin and depth of pluton emplacement.