THE GEOLOGY, GEOCHEMISTRY, AND HYDROLOGY OF FOUR HOT SPRING AREAS ALONG THE SOUTH FORK PAYETTE RIVER, BETWEEN LOWMAN AND BANKS, BOISE COUNTY, IDAHO.

by

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A study was made of four hot spring areas located along the South Fork Payette River, between Lowman and Banks, Boise County, Idaho. The purpose of this investigation was to determine the detailed geologic, geochemical, and hydrologic setting of the thermal springs.

Much of the study area is underlain by Cretaceous granodiorite. The northeast trending Idaho porphyry belt, a complex of Tertiary dike swarms and granitic stocks, transects the east-central portion of the study area. Northeast and northwest trending major fault zones cut these units and control the course of the South Fork Payette River. Minor structures, as exemplified by the trends of dikes, also exhibit strong northeast and northwest trends consistent with the regional structural pattern suggesting periods of northwest and northeast extension during the Eocene and Oligocene Periods.

The four thermal spring areas are located along major fault zones and were divided into two types. The Goller, Corder, and Pine Flat hot spring areas are associated with Tertiary dike swarms related to the Idaho porphyry belt. Hot spring vent locations are controlled by the dikes having the highest hydraulic conductivity which act as fluid conduits. SiO₂ and Na:K:Ca geothermometry
yielded source temperatures of 71°C which, combined with a measured geothermal gradient of 80°C/km, suggests a 1 km circulation depth.

The Deer Creek hot spring area is distinct geologically and geochemically from the other three areas. Situated in an area lacking dikes, the hot water rises 2 km along the intersection of two major faults from a thermal aquifer at 142°C.

The two types of geothermal systems share several common features. Recharge, with cold meteoric water, occurs along the major fault zones with long (9,000-28,800 years) residence times for waters in the system. Little or no mixing of thermal and non-thermal waters occurs during ascent. Recurrent fault movement has maintained open conduits otherwise plugged by the gradual precipitation of minerals by the rising thermal water.
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INTRODUCTION

Purpose and Scope

A study was undertaken of the geothermal area which occurs along the South Fork Payette River between Lowman and Banks, Idaho. The objectives of this investigation are 1) to determine the detailed petrologic and structural setting of the thermal spring areas; 2) to establish the relationship between the geologic setting and the location of the thermal spring activity; 3) to estimate source temperatures and depths of circulation of the thermal waters; and 4) to interpret the geothermal systems in the area of investigation.

The results of this study should provide a basis for the evaluation of the potential for further development of the geothermal resources in this area. In addition, it is hoped that these results will contribute to a more detailed understanding of geothermal systems within the Idaho batholith.

Location and Access

The thermal springs examined in this study occur in four areas along the South Fork Payette River, between Lowman and Banks, in Boise County, Idaho on land contained within the Boise National Forest (Figure 1). The four thermal spring areas studied are the Pine Flat hot springs, Corder hot springs, Goller hot springs, and the Deer Creek hot springs. A total of 13 km² within these four areas was mapped in detail including portions of the Banks and Garden Valley
Figure 1: Location of the four thermal spring areas.
15 minute Quadrangles and the Pine Flat 7.5 minute Quadrangle. A geologic
road log was made from Lowman to Banks.

Access to the area is by Idaho route 17 which parallels the South Fork
Payette River on its north bank and by Grimes Creek road on the south bank
(Figure 1). The one bridge which crosses the river in the study area connects
these two roads 2 km southeast of Garden Valley. Much of the river front is
privately owned with access by permission of the owners. The remainder of the
study area is accessible only by cross-country travel on foot and by raft.

Method of Investigation

An analysis of the petrology, structure, geochemistry, and hydrology of the
area was undertaken to determine the geologic and hydrologic setting of the
thermal springs.

Using the three previously mentioned topographic base maps, brunton
compass, and tape measure, geologic maps of the four thermal spring areas,
detailed outcrop maps around selected hot springs, and a reconnaissance
geologic map of the study area were prepared. Representative rock samples
were collected and 65 thin sections were made for petrographic study. Whole
rock major element abundances of 28 samples were obtained by x-ray
fluorescence analysis through the kindness of Dr. Peter Hooper. Geometric
analysis of minor structures was carried out using Lambert equal area
stereographic plots of joint, dike, minor fault, and slickenside orientations.

Twenty-two water samples were collected and acidified to pH<3 using
reagent grade HNO₃. Spring temperatures, pH, and discharge rates were
measured in early and late summer to note seasonal variations. Temperatures were measured using a thermometer while pH readings were made using low-ion pH paper. The discharge rates were estimated by noting the time required to fill a container of known volume. The water samples were analyzed for Si, Na, K, Ca, and Mg by atomic absorption spectrophotometry and source temperatures were estimated using the SiO₂ and Na:K:Ca geothermometers (Ellis and Mahon, 1977).

By combining the calculated source temperatures with the known geothermal gradient, the depth of circulation of thermal water was estimated. This information and additional published hydrologic data combined with the results of the geologic survey were used to interpret the geothermal systems in the study area.

Definitions and Spring Numbering

The term thermal spring, as used here, is any spring with a discharge volume in excess of one liter per minute and with water temperature greater than 20°C, pH greater than 7.5, or chemistry that suggests past equilibrium at temperatures above 20°C. A thermal seep is a vent discharging water having these same properties at rates less than 1 liter per minute. A cold spring is any vent discharging water that is less than 20°C and has a pH less than 7.5. The spring numbering system used here groups adjacent springs and seeps together that appear to be related based on geologic setting and areal distribution. The spring complexes are numbered sequentially with a prefix designating the hot spring area, for example: Deer Creek (DCS-1), Goller (GHS-1), Corder (CHS-1), and Pine Flat (PFS-1).
Regional Geology

The thermal springs of interest in this study occur along the South Fork Payette River which drains the west central portion of the Atlanta lobe of the Idaho batholith. The Idaho batholith, as defined by Armstrong (1975), is made up of granitic plutons of Cretaceous age which outcrop in central and northern Idaho and western Montana in two lobes, the northern Bitterroot lobe and the southern Atlanta lobe (Figure 2). A more widely accepted definition (Taubeneck, 1971) includes all of the Mesozoic and Cenozoic plutonic rocks that occur both within the main body of the batholith and as outliers.

The Idaho batholith is bounded to the north and northeast by Precambrian belt supergroup rocks, to the east by the Challis volcanics, to the south by the Snake River Plain volcanics, and to the west by the Columbia River Basalts and Seven Devil's volcanic arc allochthonous terrane (Figure 2). The two lobes of the batholith are separated by a narrow northwest trending belt of Precambrian metamorphic rocks termed the Salmon River Arch by Armstrong (1975). The Idaho batholith is one of a series of granitic batholiths which parallel the western margin of North and South America and was emplaced in response to eastward subduction of the Farallon plate beneath the North American Cordillera (Dickenson, 1979).

The Atlanta lobe can be divided into two major groups of plutonic rocks, Mesozoic and Tertiary. The Mesozoic rocks may be further subdivided into three groups (Kiiilsgaard and Lewis, 1983). The earliest, emplaced 97-73 million years ago, is a batholithic border facies of foliated tonalite grading into
Extrusives younger than Idaho batholith

Challis and related volcanics

Idaho batholith

Mesozoic metasedimentary and metavolcanic rocks associated with the Seven Devils allochthonous terrane

Beltian and Paleozoic sedimentary and metasedimentary rocks

Precambrian high-grade belt and pre-belt rocks

Figure 2: General geology of central and southern Idaho (adapted from Hyndman, 1983).
hornblende-bearing granodiorite. This early group is intruded by massive volumes of biotite granodiorite and muscovite-biotite granodiorite which forms the core of the Atlanta lobe. The biotite granodiorite was intruded between 82-69 million years ago whereas the 2-mica granodiorite ranges from 75-66 million years old. Slightly younger (72-64 million years old) leucocratic granite intrudes all older rocks where it outcrops in the eastern portion of the Atlanta lobe. The older tonalites have characteristics of I-type granitoids whereas the younger granodiorite and granite appear to be of S-type (Kiilsgaard and Lewis, 1983).

The Tertiary anorogenic plutonic rocks of the Atlanta lobe consist of a bimodal suite of 50-45 million year old quartz monzodiorite and 45-42 million year old granite which may be comagmatic and are probably related to intracontinental rifting (Bennett and Knowles, 1983). The Challis volcanics and the numerous Tertiary dikes that intrude the Atlanta lobe are the extrusive and hypabyssal equivalents of the Tertiary plutons. The Tertiary dikes are concentrated in swarms which form a northeast trending igneous belt termed the Idaho porphyry belt by Kiilsgaard and Lewis (1983). The dikes range from basalt to rhyolite and were intruded as an imperfect differentiation sequence commencing with Eocene intermediate intrusives and terminating with rhyolite and basalt dikes in the Oligocene Period (Olson, 1968). The Idaho porphyry belt also contains numerous Eocene stocks and small plutons as well as eruptive centers for the Challis volcanics (Bennett and Knowles, 1983). Bennett (1980) ascribes most of the mineralization in the Idaho batholith to this Eocene-Oligocene igneous event.
The trans-Challis fault system (TCFS) is a major northeast trending structural feature which transects the study area (Figure 16) and continues to the northeast where it aligns with the Great Falls lineament (Bennett, 1986). Movement along this regional tectonic feature has been recurrent from the Precambrian to present. Cretaceous plutonic rocks are cut by these fault systems and the emplacement of the Tertiary Idaho porphyry belt has been guided by them (Kiilsgaard and Lewis, 1983). Anderson (1934) recognized that major Basin and Range faults have broken the Atlanta lobe into many northwest-trending, fault bounded mountain ranges. These northwest trending faults all terminate against TCFS structures (Kiilsgaard and Lewis, 1983).

Previous Investigations

The occurrence of thermal springs along the South Fork Payette River was noted by Stearns, et al (1937) and Waring (1965). Ross (1971) listed seven thermal springs in the study area, including the first published mention of Deer Creek hot springs; temperature, discharge, and usage information were tabulated. Young and Mitchell (1973) published water chemistry and geothermometry data for three springs in the study area. A more thorough review of thermal springs along the South Fork Payette River was presented by Lewis and Young (1980) in which chemistry, geothermometry, and isotope data for several springs in the study area were given, although detailed discussion of the geologic control of the thermal springs was not attempted.

Geologic investigations along the South Fork Payette River have been only reconnaissance in nature until recently. Larson and Schmidt (1958) made a
generalized petrologic survey of the western Atlanta lobe, including the present study area. A detailed study of the geology and tectonic significance of the Idaho porphyry belt is given by Olson (1968). Characteristics of the Tertiary granitic plutons of the Idaho batholith and the mineralization associated with them have been described by Bennett (1980) who includes some information on Tertiary intrusives along the South Fork Payette River. Publication of the Banks 1x2 Quadrangle (Mitchell and Bennett, 1979) and the Challis 1x2 Quadrangle (Fisher, et al., 1983) have provided more recent summaries of the structure and distribution of rock types in the study area. The Symposium on the Geology and Mineralization of the Challis 1x2 Quadrangle (McIntyre, 1983) includes a current synthesis of the petrology, structure, mineralization, and tectonics of the central Atlanta lobe.

This study is the fifth in a series of investigations of the detailed geologic and geochemical setting of thermal springs in the Idaho batholith. Kuhns (1980) and Youngs (1981) studied thermal spring areas in the northern portion of the Bitterroot lobe which provided a framework for this study. Vance (1986) conducted a detailed investigation of three geothermal areas on the southeast margin of the Bitterroot lobe. A similar study by Dingee of three thermal spring areas located along the upper half of the South Fork Payette River is in progress.
GENERAL GEOLOGY OF THE THERMAL SPRING AREAS

The general geology of the study area is shown in Plate 1 (back pocket). Geologic maps of the four thermal spring areas (Figures 3, 4, 5, and 6) show the geologic setting of the hot springs as well as the distribution and numbering of the thermal springs. Geologic outcrop maps (Figures 7, 8, 9, and 10) of selected springs in each of the four thermal spring areas show the detailed geology around the spring vents. A geologic road log (Appendix A) provides additional information on the geology outside of the four thermal spring areas. A more detailed analysis of the petrology and structure of the study area is given in later chapters.

Most of the study area is underlain by Cretaceous plutonic rocks of the Idaho batholith (Plate 1). The east-central portion of the study area is transected by the northeast trending Idaho porphyry belt intrusive complex. In the study area, the main body of the Idaho porphyry belt, which is 10 km wide and forms extremely rugged terrain, is made up of a 1 km wide granitic stock and many swarms of dikes that range from basalt to rhyolite. No thermal springs were observed discharging from the Tertiary body although two cold springs issue from alluvium in the eastern portion of this unit (Appendix A).

Major fault zones trending generally northeast and northwest were mapped along the South Fork Payette River (Plate 1). Other major faults in the study area trend north-northwest and northeast.

The Deer Creek hot springs are located west of Garden Valley (Plate 1) at the intersection of northwest and north trending faults (Figure 3). DCS-1 (Figure
7) and DCS-2 (Figure 3) discharge from the alluvium filled linear gully formed by the north trending fault. DCS-3 consists of a series of closely spaced thermal spring vents which issue from prominent, shallow south dipping joints just east of DCS-1 (Figure 3).

The Goller hot springs, located just east of Garden Valley, are situated in an area of higher than average dike density (Plate 1). The dikes are predominantly northwest trending. GHS-1 and GHS-2 discharge from fractures in the Cretaceous granodiorite and from alluvium while GHS-3, -4, and -5 are closely grouped and are located directly across the river from GHS-1 and GHS-2 (see Figure 4). GHS-3 issues from a minor fault zone whereas GHS-4 and GHS-5 discharge from alluvium adjacent to an outcrop of a northwest trending Tertiary dike (Figure 8). A series of cold springs and seeps collectively termed GHS-6 emanates from the dike/granodiorite contact formed by a northwest trending dike which parallels the South Fork Payette River in this area (Figure 4).

The Corder hot spring area is located about 1 km west of the main body of the Idaho porphyry belt (Plate 1). Most of the thermal spring vents are located along the northeast trending Corder fault zone although several others were found further to the west (Figure 5). Dikes are abundant in the Corder hot spring area and trend generally northeast and northwest. All of the thermal springs except CHS-2 were observed to discharge from within or in close proximity to a dike (Figure 5). CHS-2 issues from alluvium adjacent to a major fault zone at the western end of the Corder hot spring area. The detailed geologic outcrop map of the springs and seeps comprising CHS-4 (Figure 9) illustrates how minor structures near the dikes may often control the actual spring vent locations. This is a common feature of all four thermal spring areas.
The Pine Flat hot springs are located immediately east of the main body of the Idaho porphyry belt (Plate 1). A strong northeast trend is evident (Figure 6) for the abundant Tertiary dikes in the Pine Flat area. As in the Goller and Corder hot spring areas, the Pine Flat hot spring vents are closely associated with dikes (Figure 6). In particular, the vents that comprise PFS-3 are almost all located within dikes (Figure 10). The PFS-1 and PFS-2 vents are aligned with dikes but discharge from alluvium (Figure 6). PFS-5 also issues from alluvium adjacent to a dike whereas PFS-4 discharges from alluvium far from any outcrop (Figure 6).
Explanation for Figures 3–6.

Contact, dashed where uncertain

Fault, dashed where uncertain, showing sense of movement

Trend of fault

Trend of dike

Trend of aplite

Thermal spring (>11./minute)

Thermal seep (<11./minute)

Cold spring

Aphyric rhyolite

Aphyric high-K basalt

High-K dacite porphyry

High-K andesite porphyry

Aphyric high-K andesite

Main body of Idaho porphyry belt

Deer Creek biotite granodiorite

Biotite granodiorite
Figure 3: Geologic map of the Deer Creek hot spring area. Box outlines area of Figure 7. Base map: USGS 15-minute Quadrangle, Banks, Idaho.
Figure 4: Geologic map of the Goller hot spring area. Box outlines area of Figure 8. Base map: USGS 15-minute Quadrangle, Garden Valley, Idaho.
Figure 5: Geologic map of the Corder hot spring area. Box outlines area of Figure 9. Base map: USGS 15-minute Quadrangle, Garden Valley, Idaho.
Figure 6: Geologic map of the Pine Flat hot spring area. Box outlines area of Figure 10. Base map: USGS 7½-minute Quadrangle, Pine Flat, Idaho.
Explanation for Figures 7-10

- Alluvium
- Aphyric high-K basalt
- Fault
- Minor fault with slickensides
- High-K andesite porphyry
- Trend of dike
- Trend of pegmatite
- Aphyric high-K andesite
- Trend of aplite
- Pegmatite
- Thermal spring (>1 liter/minute)
- Thermal seep (<1 liter/minute)
- Cretaceous plutonic rocks
Figure 7: Geologic outcrop map of a portion of the Deer Creek hot spring area.
Figure 8: Geologic outcrop map of a portion of the Goller hot spring area.
Figure 9: Geologic outcrop map of a portion of the Corder hot spring area.
PETROLOGY

Cretaceous granitic rocks are by far the most abundant rock type exposed in the study area. Tertiary units consist of dikes and one granitic stock. The general distribution of rock types was discussed in the preceding chapter. Field and petrographic descriptions of each rock unit are presented here as is a discussion of the igneous history of the study area.

Method of Investigation

Observation of macroscopic field relations combined with thin section study and whole rock chemical analysis were used to determine the petrology of the rocks found in the study area. Sixty-five thin sections were prepared and studied to determine mineral compositions and textural relationships of all rock types and to conduct point counts for modal abundance estimates of granitic rocks. Following the methods of Hutchinson (1974) a minimum of 1000 points per sample were counted in order to determine modal mineral abundances (Figure 11-A).

Twenty-eight samples were prepared and analyzed by x-ray fluorescence spectrophotometry for 10 major elements through the kindness of Dr. Peter Hooper. The results are presented in Figures 11-B and 12 and in Appendix B.

A classification scheme for the rocks in the study area has been modified from Kielsgaard and Lewis (1983) which incorporates their terminology wherever possible.
Previous Investigations

Larsen and Schmidt (1958) in a reconnaissance petrologic survey of the western Atlanta lobe identified biotite-rich "Cascade-type granodiorite" west of Garden Valley and "coarse quartz monzonite" to the east. Leonard (1976) described "a porphyritic biotite leucogranodiorite, locally muscovitic" core, north of Garden Valley, which he called the predominant facies of the Idaho batholith suite. The Baker 1x2 Quadrangle (Mitchell and Bennett, 1979) shows quartz diorite west of Garden Valley where Larson and Schmidt's Cascade-type granodiorite was mapped. In 1983 Hyndman summarized existing knowledge of the entire Idaho batholith including petrology, radiometric ages, chemistry, petrogenesis, and structures related to the batholith. In these previous investigations little sampling was done in the present study area.

In 1983 the Challis 1x2 Quadrangle was revised (Fisher, et al.); it contains petrologic descriptions of the rocks along the South Fork Payette River. Fisher et al. (1983), mapped a biotite granodiorite east and west of Garden Valley which grades into a 2-mica granodiorite-granite north and south of Garden Valley. This later unit includes a portion of Larsen and Schmidt's (1958) coarse quartz monzonite and the 2-mica portion of Leonard's (1976) core facies. Kiilsgaard and Lewis (1983) published detailed petrographic analysis of the units shown on the Challis 1x2 Quadrangle.

Knowledge of the Tertiary units outcropping in the study area is due almost entirely to Olson (1968) who described, in detail, the petrology and tectonic significance of the Idaho porphyry belt. Bennett (1980) and Kiilsgaard and
Bennett (1983) briefly describe these rocks, give several radiometric dates, and discuss the role of Tertiary igneous activity in the mineralization of the Idaho batholith.

Results of Petrographic Analysis

Cretaceous Plutonic Rocks

The Cretaceous plutonic rocks are here divided into three types; Deer Creek biotite granodiorite, biotite granodiorite, and muscovite-biotite granodiorite.

Deer Creek Biotite Granodiorite

To the west of Garden Valley, biotite rich granodiorite which is locally tonalitic is exposed (Plate 1). Since the Deer Creek hot springs occur in this area, the rock type is herein named the Deer Creek biotite granodiorite.

This rock is medium grained hypidiomorphic granular with 48% anhedral andesine (An30-37), 29% anhedral quartz, 7% subhedral alkali feldspar, and 10-15% subhedral biotite (Figure 11-A). Accessories include muscovite, allanite, opaques, zircon, apatite, and sphene. Moderate alteration is pervasive and consists of sericitized plagioclase, epidote, and chlorite and opaques after biotite. Hornblende is completely absent.

The alkali feldspar, which occurs as megacrysts up to 20 cm. long, generally exhibits a micropoerthetic texture and is often poikilitic, containing xenocrysts of subhedral andesine and anhedral quartz. The local absence of alkali feldspar
Figure 11: Classification of Cretaceous plutonic rocks based on (A) modal abundances and (B) normative abundances: (−) Deer Creek biotite granodiorite, (+) biotite granodiorite.
megacrysts creates areas of up to one hundred square meters of tonalitic composition. The relatively high percentage of biotite gives the Deer Creek biotite granodiorite a medium gray color in outcrop which distinguishes it from the relatively biotite poor light gray granodiorite to the east of Garden Valley.

The Deer Creek biotite granodiorite is generally massive and unfoliated although some areas (10-100 square meters) of foliated rock do occur. The foliation is due to aligned biotite crystals and grades into unfoliated rock of identical composition. The Deer Creek biotite granodiorite is contained within Larsen and Schmidt's (1958) Cascade-type granodiorite.

Biotite Granodiorite

East of Garden Valley light gray biotite granodiorite occurs over much of the study area (Plate 1). This biotite granodiorite is medium grained hypidiomorphic granular and is composed of 53% anhedral oligoclase (~ An24-30 cores and An17-22 rims), 24% anhedral quartz, 19% subhedral alkali feldspar, and 2-5% subhedral biotite (Figure 11-A). Accessory minerals include muscovite, allanite, apatite, and zircon with secondary sericite, chlorite, opaques and epidote. Hornblende is completely absent. The alkali feldspar is mostly orthoclase and commonly occurs as poikilitic megacrysts containing xenocrysts of subhedral oligoclase and anhedral quartz and biotite. The rock is massive and unfoliated throughout the study area.

The biotite granodiorite exposed in the Pine Flat hot springs area is indistinguishable in hand sample from that occurring to the west. In thin section, however, it is observed to be slightly enriched in quartz at the expense of
plagioclase but it is still considered to be part of the biotite granodiorite unit. This conclusion is supported by comparison of the whole rock chemistry of samples of biotite granodiorite from the Pine Flat and Corder hot spring areas (Appendix B; samples P-14 and C-60). The biotite granodiorite is part of Larsen and Schmidt's (1958) coarse quartz monzonite which, together with the Deer Creek biotite granodiorite, makes up the biotite granodiorite of Kiilsgaard and Lewis (1983).

Muscovite-Biotite Granodiorite

To the north and south of Garden Valley two-mica granodiorite is exposed (Plate 1) which has essentially the same mineralogy as the biotite granodiorite but contains 1-2% primary subhedral muscovite. Contacts with the biotite granodiorite are gradational over tens to hundreds of meters (Kiilsgaard and Lewis, 1983) and are not well exposed in the study area. The two-mica granodiorite correlates with the muscovite-bearing portion of Larsen and Schmidt's (1958) coarse quartz monzonite as well as with Leonard's (1976) core facies, and Kiilsgaard and Lewis' (1983) muscovite-biotite granodiorite.

Aplite and Pegmatite Veins

Aplite veins ranging from 0.5-15 cm. wide occur sporadically east of Garden Valley but are common within the Deer Creek biotite granodiorite. The veins consist of fine grained anhedral quartz and alkali feldspar in equal proportions with 1-3% muscovite and minor andesine (An31). At least two generations of
Aplite veins exist in the Deer Creek area as evidenced by cross-cutting relationships.

Irregular pods and veins of extremely coarse pegmatite occur throughout the Deer Creek biotite granodiorite with individual crystals up to 15 cm. in length. The pegmatites consist of equal proportions of quartz and alkali feldspar with minor muscovite occurring as books up to 1 cm. in width.

The aplite and pegmatite veins generally have sharp contacts and cross-cut one another suggesting that they formed together as residual components of the granodiorite intrusion.

Tertiary Igneous Rocks

The Tertiary igneous rocks exposed in the study area are almost all part of the Idaho porphyry belt plutonic and related hypabyssal units. These include a granitic stock, and dikes ranging from basalt to rhyolite. A younger basalt dike outcrops in the western portion of the main body of the Idaho porphyry belt (Appendix A).

Idaho Porphyry Belt

The northeast trending Idaho porphyry belt transects the study area, separating the Pine Flat hot springs area from those to the west (Plate 1). As no thermal springs issue from the main body of the Idaho porphyry belt in the study area it is not a major focus of this study. The reconnaissance petrographic work completed here on the main body of the Idaho porphyry belt suggests a granitic
stock intruded and surrounded by numerous basic to silicic, aphyric and porphyritic dikes. Only one rhyolite dike was found outside of the main body of the Idaho porphyry belt, in the western end of the Pine Flat hot springs area (Figure 6); it is not considered in detail here.

The vast majority of the dikes mapped in the four thermal spring areas are basic to intermediate and were emplaced as part of the Idaho porphyry belt igneous event (Olson, 1968). These dikes include aphyric and porphyritic high-K andesite, high-K dacite porphyry, and aphyric high-K basalt (Figure 12).

Aphyric High-K Andesite

By far the most common type of dike exposed in the study area is aphyric high-K andesite (Figures 4 and 5) which occurs in all four thermal spring areas and is the only type found in the Deer Creek hot springs area.

These dikes are medium gray to black and range from 1 cm. to 2 m. wide with an average of 1 m. They are composed of 55-70% subhedral andesine (An35-46), 5-30% mafic minerals (predominantly hornblende with lesser biotite and rare clinopyroxene), 5-20% anhedral alkali feldspar, and 0-10% anhedral quartz. Alteration of the dikes is common and consists of sericitized plagioclase and chlorite, hematite and black opaque oxides after the mafic minerals.

High-K Andesite Porphyry

Medium to dark gray porphyritic dikes ranging from 0.3-1.5 m. thick occur sporadically east of Garden Valley (Figures 4 and 5). The matrix of these dikes
Figure 12: Classification of Tertiary dikes: (o) aphyric high-K andesite, (+) high-K andesite porphyry, (c) high-K dacite porphyry, (v) aphyric high-K basalt, (o) aphyric basalt, (A) aphyric rhyolite (classification after Peccerillo and Taylor, 1976).
is identical to the aphyric high-K andesite and the chemistry of the two types of dikes is very similar (Figure 12 and Appendix B). The high-K andesite porphyry dikes contain 2-15% phenocrysts of subhedral to euhedral hornblende and lesser biotite. Alteration minerals include sericite, chlorite, hematite, black opaque oxides, and epidote.

High-K Dacite Porphyry

Several widely scattered (Figures 5 and 6) small groups of distinctive high-K dacite porphyry dikes were found on both sides of the main body of the Idaho porphyry belt. These dikes range from 2 to at least 20 m. wide and are traceable along strike for up to 1 km., further than any other type of dike. These dikes are easily recognized by their 5 mm. long, white euhedral andesine (An₃₇) phenocrysts which, together with the much less abundant subhedral biotite and hornblende phenocrysts, make up to 30% of the rock. The phenocrysts are set in a light gray groundmass composed of 50-55% subhedral andesine (An₃₂), 35% anhedral alkali feldspar, 5-10% anhedral quartz, and 5% subhedral mafic minerals. Alteration of these dikes is similar to the others; sericite, chlorite, hematite, and black opaque oxides are common.

Aphyric High-K Basalt

Two swarms of aphyric high-K basalt dikes were identified from XRF data (Figures 5 and 6 and Appendix B). However, distinction between these dikes
and aphyric high-K andesites was not possible in the field and, hence, the
proportion of each type is uncertain.

The two swarms consist of northeast trending subparallel anastomosing dark
gray to black dikes that range from 0.5-2 m. thick (Figures 9 and 10). They are
composed of 60-65% subhedral andesine (An33-43), 20-35% subhedral mafic
minerals (mostly hornblende with lesser biotite and clinopyroxene), 5-10%
anhedral alkali feldspar, and trace anhedral quartz grains. Alteration is
ubiquitous and consists of sericite, chlorite, hematite, and black opaque oxides.

Aphyric Basalt

One medium grained black 1 m. wide aphyric basalt dike intrudes rhyolite
dikes in the western portion of the main body of the Idaho porphyry belt (Plate
1). Based on whole rock major element abundances (Appendix B), this dike is
correlative with the Weiser Embayment flows of the Columbia River Basalt
group (Dr. Peter Hooper, personal communication) and is the youngest igneous
rock found in the study area.

Distribution of Rock Types

The Cretaceous plutonic rocks include the Deer Creek biotite granodiorite
which outcrops in the western portion of the study area and is bounded to the
east by the Boise Ridge fault (Plate 1). The central and eastern portions of the
study area are underlain by biotite granodiorite which grades into 2-mica
granodiorite to the north and south of Garden Valley.
The Tertiary dikes occurring outside of the main body of the Idaho porphyry belt are not evenly distributed along the South Fork Payette River. Instead, there is a pronounced increase in dike density near the thermal spring areas (Plate 1). The aphyric and porphyritic high-K andesite dikes showed the widest range of exposure whereas the high-K dacite porphyry and aphyric high-K basalt dikes were limited to the Corder and Pine Flat hot spring areas and the far eastern end of the study area (Appendix A and Figures 5 and 6).

Relative Timing of Emplacement of Igneous Rocks

A lack of exposure of contacts between the three Cretaceous plutonic units and the paucity of cross-cutting relationships between the various dike types prohibits a detailed discussion of relative ages of the igneous rocks found in the study area. However, several published radiometric dates, rare cross-cutting dikes, and generally accepted ideas on the petrogenetic and structural history of the Atlanta lobe allow a partial reconstruction of igneous events.

Kiilsgaard and Lewis (1983) group the Cretaceous rocks in the study area into biotite granodiorite (herein subdivided into the Deer Creek biotite granodiorite and biotite granodiorite) dated at 82-69 million years old and a 75-66 million year old core forming 2-mica granodiorite-granite noting a general increase in age from the core unit to the margins of the batholith. This implies that the Deer Creek biotite granodiorite may be slightly older than the biotite granodiorite east of Garden Valley and that the 2-mica granodiorite is even younger.
Faulting related to the trans-Challis fault system (page 9) which transects the Cretaceous intrusives (Plate 1) guided the emplacement of the Idaho porphyry belt (Kiliksgaard and Lewis, 1983) the main phase of which was emplaced during the early to middle Eocene (Olson, 1968). The Idaho porphyry belt is the intrusive equivalent of the Challis volcanics which covered most of Idaho during the Eocene and are still present east of the Atlanta lobe (Figure 2). A biotite from the granitic stock outcropping within the Idaho porphyry belt in the study area yielded a potassium-argon age of 49.2 million years (Bennett, 1980).

According to Olson (1968) the Tertiary dikes were intruded as an imperfect differentiation sequence commencing with intermediate volcanism in the Eocene and becoming more silicic with time. The sequence terminated in the Oligocene with both basic and rhyolite dikes. Siems and Jones (1977) observed the same petrogenetic trend but note that several reversals have occurred. A rhyolite dike from the Little Falls molybdenum prospect in the western portion of the main body of the Idaho porphyry belt (Appendix A) gave a fission track age of 29.2 million years (Bennett, 1980).

With the available data few unequivical age relationships can be established for the Tertiary rocks. It appears that the andesite and dacite dikes are probably middle to late Eocene whereas the rhyolites are Oligocene in age. No definite age can be assigned to the high-K basaltic dikes which may range from Eocene to Miocene (Olson, 1968). The Weiser Embayment-CRB correlative basalt dike which intrudes the rhyolite at the Little Falls prospect (Appendix A) is of Miocene age.
Both reconnaissance and detailed field mapping revealed major and minor structures including major fault zones that control the course of the South Fork Payette River, other large scale faults, minor faults and fault zones, pervasive jointing, and dikes. Descriptions of these features are given here whereas geometric analysis of the minor structures is included in a later section of this chapter.

The terms fault, fault zone, and joint follow the usage of Hobbs et al. (1976). A fault is a planar discontinuity across which movement parallel to the structure has occurred on a macroscopic scale. A fault zone is a tabular region containing many parallel or anastomosing faults. A joint is a discontinuity across which displacement, if present, is microscopic in scale.

Major Structures

Major Fault Zones

Geologic mapping revealed a system of major fault zones oriented generally N50E and N50W (Plate 1). Northeast trending fault zones as long as 6.5 km., such as those in the Corder and Pine Flat hot spring areas, are separated by northwest trending structures of similar length. Along most of its length in the study area, the South Fork Payette River was observed to flow within these major fault zones.
These fault zones are probably deep seated features of the batholith and are very similar to shear systems identified by Kuhns (1980) and Youngs (1981) in the Bitterroot lobe and to shear zones described by Ginther (1981) along the North Fork Payette River. Young (1985) notes that most thermal springs in the Idaho batholith are found along large fault structures which control the courses of the major rivers.

In the study area, the fault zones range from approximately 50 to 150 m wide and are steeply dipping. Movement along these structures, as indicated by slickensides (Appendix C) and offset dikes (Figure 13), includes both normal dip-slip and strike-slip components. Deformation is entirely brittle in nature with dikes, veins, and crystals offset across discrete fractures. No evidence of ductile movement was seen in the study area. In addition to aligned drainages, these fault zones are expressed as zones of highly fractured rock, rare areas of bleached rock, slumps, and vegetation changes.

The following description incorporates field observations made throughout the study area and in particular from the exposure of the northeast trending fault zone in the Corder hot springs area where the South Fork Payette River leaves the structure in a very sharp bend (Plate 1). This bend creates the best cross-sectional exposure of a major fault zone in the study area. To facilitate description of the fault zone it is here subdivided into four zones A, B, C, D with zone A at the margin and zone D at the center of the fault zone.

An increasing degree of deformation is observed from the margins of the fault zones to their centers. At the margins the joint density increases from 0.5-3 joints/meter to 20-40 joints/meter over a distance of 5-10 meters. This increase in fracturing in zone A is not evenly distributed but instead is concentrated along
Figure 13: Photograph of a minor fault zone offsetting a Tertiary dike in Zone B of the Corder fault zone.
narrow bands of intense fracturing separated by wider bands of less fractured rock. Zone A is generally about 10 m wide.

The fracturing increases inward over a span of 5 m to 30-45/m in zone B and definite minor fault zones are developed along the zones of intense fracturing (Figure 13). Movement along any one of these minor fault zones appears to be less than a few centimeters except in rare cases with observed displacements of up to 4 m. Outside of these minor fault zones is less fractured (10-25/meter) rock in which curved fractures become more common. Zone B is usually about 40 m wide.

Within zone B is zone C, a 15 m wide zone of highly fractured (50-100/meter) rock which exhibits moderate alteration. The minor fault zones become more numerous (about 1/meter) and the intervening rock is finely broken by sets of curved fractures (Figure 14).

Zone D, a 5 m wide zone of bleached and altered nonresistant fault gouge forms the center of the major fault zone in the Corder hot springs area. No distinct joint/fracture patterns are present as the granodiorite is completely fractured and crumbles in the one outcrop where it was seen.

Other Major Faults

In addition to the major fault zone system which controls the course of the South Fork Payette River other large scale faults in the study area are present which can be divided into northeast and north-northwest trending groups.

The previously described N50E trending fault zone which cuts the Corder hot springs area is the only segment of the major fault zone system which can be
Figure 14: Photograph of fracturing in granodiorite in Zone C of the Corder fault zone.
traced away from the river from both ends (Plate 1). This structure is herein
named the Corder fault zone. It dies out quickly to the northeast but continues
southwest out of the study area into the Quartzburg area (Kiilsgaard and
Bennett, 1983) and is the western most mapped fault belonging to the trans-
Challis fault system. A subparallel splay fault off the Corder fault zone,
evidenced by aligned saddles displaying vegetation changes and rare
exposures of altered granodiorite, trends S35W out of the study area.

A northeast trending fault mapped by Mitchell and Bennett (1979) cuts the
Deer Creek biotite granodiorite 2 3/4 km. east of Deer Creek hot springs but is
concealed by alluvium and vegetation where it crosses the Payette River.

The north-northwest trending faults are of two types, large regional structures
and faults of relatively local extent. The Boise Ridge fault and the Deadwood
fault zone are both major, north trending high angle faults with their east sides
downdropped (Plate 1). The Boise Ridge fault uplifted the Deer Creek biotite
granodiorite during the middle to late Miocene thereby ponding the South and
Middle Forks of the Payette River in Garden Valley (Kiilsgaard and Lewis,
1983). The Deadwood fault zone trends south down the Deadwood River
canyon but dies out as it reaches the South Fork Payette River east of the Pine
Flat hot springs.

A N32W,68E trending fault which extends for at least 1.5 km forms a perfectly
straight valley south of the Payette River and appears to cut directly through the
Deer Creek hot springs area (Figure 3). A large exposure of slickensides south
of the river indicates left lateral oblique movement whereas small outcrops of
highly fractured rock along the linear gulley containing the hot springs DCS-1
and 2 suggests continuation of this fault north of the river.
Two relatively small north-northwest trending faults were identified. The first, which cuts an aplite vein and is closely associated with a cold seep, trends N23W, 72E and forms a fault gouge filled gully 30 m long about 3/4 km east of Deer Creek hot springs (Figure 3). The second, which also forms a grus and breccia filled gully 50 m long, trends N1E, 74W and is located 1/4 km west of the Pine Flat hot springs (Figure 6).

The straight stretches of many creeks and canyons throughout the study area are undoubtably controlled either by faults or master joints. However, hard field evidence for even the major structures discussed is sometimes absent and topographic expression is often the only evidence found along segments of the faults.

Minor Structures

Minor Fault Zones

Numerous minor fault zones were observed in the Cretaceous plutonic rocks in all of the hot spring areas. These minor structures are recognized by tabular breccia zones, slickensides, and offset of dikes and veins (Figure 13). Most of these minor fault zones occur within the major fault zones (page 40) and were rarely observed more than 200 m from a known major structure. The minor fault zones range from 1 to (rarely) 10 cm wide and can be traced for up to 30 m. They are made up of numerous closely spaced, subparallel, anastomosing fractures each of which shows little movement but together may account for up to 4 m of offset. Secondary calcite fracture filling was observed in several minor
fault zones. Rarely, a minor fault zone can be seen to enter a Tertiary dike at a high angle, bend so as to be parallel to the strike of the dike, and then dissipate into numerous curved joints (Figure 9).

Minor Faults

In the Deer Creek and Goller hot spring areas normal dip-slip and oblique strike-slip movement is evidenced by slickenside formation on joint surfaces covered by secondary white mica. These minor faults differ from the minor fault zones in that movement is clearly along discrete joints which are always part of moderate to shallow dipping major joint sets. The amount of movement, when possible to determine, is always less than 1 m.

Joints

Joints are the most pervasive structural features in the study area. An analysis of the joint density in all rock types both within and away from the thermal spring areas was undertaken. As the porosity and hydraulic conductivity (page 72) of plutonic rocks is due almost entirely to jointing (Freeze and Cherry, 1979) a detailed knowledge of joint density should identify the most favorable geothermal fluid conductors.

The density of jointing in all four thermal spring areas for virtually all rock types is higher near (<10 m) the hot spring vents than at a distance (>50 m) from them (Table 1). As discussed previously (page 40), jointing increases with proximity to the major fault zones. The increase in joint density observed in the
Table 1: Comparison of joint density in various rock types near (<10 m) and far (>50 m) from thermal spring vents.

<table>
<thead>
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<th>LOCATION</th>
<th>ROCK TYPE</th>
<th>JOINT DENSITY (JOINTS/METER)</th>
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</tr>
<tr>
<td></td>
<td></td>
<td>aplit rare</td>
</tr>
<tr>
<td></td>
<td></td>
<td>pegmatite common</td>
</tr>
<tr>
<td>away from springs</td>
<td>granodiorite</td>
<td>0.2-3</td>
</tr>
<tr>
<td></td>
<td></td>
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sampled thermal and non-thermal springs in the study area.
rocks near the hot spring vents is, therefore, an expected result of the thermal springs occurring along these major structures. The granodiorite is consistently lower in joint density than adjacent dikes. The dikes exhibit a wide range of joint density within individual dikes and among different dike types. Overall, there is a general decrease in joint density in the order: aphyric high-K andesite, aphyric high-K basalt, high-K andesite porphyry, and high-K dacite porphyry. It is notable that the vast majority of hot springs issue from or are in very close proximity to aphyric high-K andesite and aphyric high-K basalt dikes.

Dikes and Veins

The dikes and veins are included as structural features because they are considered to have been emplaced preferentially into joints or fractures formed or opened in response to the existing stress regimes. Their orientations, therefore, may provide clues concerning the orientation of the stress regime present at the time of their emplacement.

Geometric Analysis of Minor Structures

Geometric analyses were carried out of joint, dike, vein, minor fault and fault zone, and slickenside orientations for each of the four thermal spring areas in order to find similarities and differences in structural patterns within the study area and to relate the orientation of Tertiary intrusive units to the structures observed. The results of this analysis are presented in Appendix C. The Lambert equal area stereographic plots of joints in the Cretaceous plutonic
rocks were contoured at an interval of 1% of the total data in 1% of the area of the stereonet. The terms maxima and peak as used here refer to orientations having relatively large numbers of measured data points.

Deer Creek Hot Springs

Joint orientations in the Deer Creek biotite granodiorite yield five maxima of 3-4% with the three broadest peaks oriented N28E,62E, N19W,76E, and N62W,84N (Appendix C-1). Two maxima, representing relatively shallow dipping joint sets trend N66E,24S and N1E,28W.

Most aplite and pegmatite veins show very similar northeast trends (Appendix C-2) which correlate with no major joint set in the granodiorite. A small number of veins are oriented north-northwest and north-northeast, parallel to N19W,76E and N28E,62E joint sets. The similarity between aplite and pegmatite vein orientations would be expected for genetically related features.

The plot of minor faults (Appendix C-3A) shows a tight cluster, oriented N10W,25W, which are part of the N1E,28W joint set and which also parallel the general vein orientation. The few minor fault zones found trend north-northwest and north-northeast, subparallel to joint sets in the granodiorite.

Nearly all of the slickensides were found on the N10W, 25W trending minor faults and suggest normal movement along these surfaces in the direction 230, N80W (Appendix C-3B).
Goller Hot Springs

The contoured plot of joint orientations in the granodiorite at Goller hot springs (Appendix C-4) reveals a strong (4-5%) N50W, 82E trend with two other northwest trending 3-4% maxima oriented N23W, 83E and N70W, 72E. A north-northeast trend averaging N19E, 66E is given by two closely spaced maxima. The only shallow dipping joint set in this area is oriented N88E, 34S. This joint pattern is similar to that of the Deer Creek area (Appendix C-1) except for a pronounced N1E, 28W joint set which is absent in the Goller hot spring area. Jointing in the dikes is characterized by a general scatter of orientations (Appendix C-5A) although a N45W, 70E trend parallel to the most pronounced joint set in the granodiorite is evident.

Orientations of the two types of dikes found in the Goller hot springs area, aphyric and porphyritic high-K andesite, exhibit a very strong N45W, 77E trend (Appendix C-5B). This is nearly identical to the N50W, 82E joint set in the granodiorite suggesting that, in the Goller hot springs area, this joint set was preferentially intruded by Tertiary magma.

The minor fault zones group into two orientation maxima, N16E, 73E and N22W, 62E (Appendix C-6A), which parallel two of the major joint sets. A peak trending N62W, 26S corresponds to white mica coated minor faults which are part of the N88E, 34S joint set.

The few slickensides observed (Figure 8) were all found on the minor fault surfaces and trend 80, N78W and 80, S71E (Appendix C-6B) indicating oblique, near strike-slip motion.
Corder Hot Springs

Joint orientations in the granodiorite at Corder hot springs are dominated by a strong (4-5%) maxima corresponding to N58E,88S with two lesser maxima (3-4%) trending N15W,87W and N34W,86W (Appendix C-7). Two poorly defined (2-3%) maxima trend N6E,86E and N24E,89E. The three strongest maxima are similar to joint sets in the Deer Creek and Goller hot spring areas (Appendices C-1 and C-4) but a pronounced northeast trending joint set is conspicuously absent in the Corder hot spring area. Jointing in the dikes, as in the Goller hot springs area, is not in distinct sets but, rather, is more random in distribution (Appendix C-8A) with only a north-northwest trend evident.

The dikes of the Corder hot spring area show two major trends (Appendix C-8B), northeast and northwest, both of which are steeply dipping. The aphyric and porphyritic high-K andesites trend predominantly to the northeast averaging N46E,70S which correlates to no major joint set but is parallel to the Corder fault zone (Plate 1). A smaller number of high-K andesites, most of which are west of the Corder fault zone (Figure 5), have an average trend of N25W,82W, parallel to the two north-northwest oriented joint sets. A less distinct group of N83W,45S trending andesite dikes appears to have intruded joints which are part of the major south dipping joint set. The two high-K dacite porphyry dikes found (Figure 5) trend N30W,75E, subparallel to the N34W,86W joint set. The swarm of aphyric high-K basalt dikes identified by XRF analysis (Appendix B; sample C-37) trends N50E, dips steeply and is parallel with the margin of the Corder fault zone where it is located (Figure 5).
A wide variety of orientations of both minor fault zones and slickensides was noted in the Corder hot spring area (Appendices C-9A and C-9B). Minor fault zone orientation maxima include N10W,85W and N42W,68E which parallel major joint sets and N78W,60S and correlates with no other structural feature. A fourth minor fault zone trend, N20E with steep east and west dips, is subparallel to the Corder fault zone. The range of slickenside orientations reflects the variety of minor fault zone orientations which display both strike-slip and dip-slip characteristics.

Pine Flat Hot Springs

Joint orientation maxima in the granodiorite at the Pine Flat hot springs area include a strong (4-5%) N33E,85E trend, two subparallel trends N41W,85E and N58W,75E, and the only moderately dipping joint set, N89W,37S (Appendix C-10). This joint pattern shares northwest and northeast trending and east-west trending moderately south dipping joint sets in common with the other three thermal spring areas and is especially similar to the Goller hot springs area (Appendix C-4). Joints in the various dikes (Appendix C-11A) show the random distribution typical throughout the study area.

A strong northeast trend and much weaker northwest trend was observed for the dikes in this area, all of which dip steeply (Appendix C-11B). The aphyric and porphyritic high-K andesites are predominantly northeast trending and are the only dikes in the Pine Flat area found trending northwest. The two high-K dacite porphyry dikes found trend N20E,81W. The swarm of high-K basalt dikes (Appendix B; samples P-2 and P-4) mapped at the site of highest thermal spring
activity (Figure 10) trend N39E,80E. The strong northeast trend of all dike types suggests they intruded into the dominant N33E,85E joint set of the granodiorite with relatively few dikes emplaced into the northwest trending joint sets.

Summary of Results

Most of the minor faults, minor fault zones, and dikes can be correlated with joint sets present in the Cretaceous plutonic rocks. Formation of the minor faults is due to movement occurring along joint surfaces generally having shallow dips (Appendix C-3A). Minor fault zones correlate less well with joint sets than do other minor structures (Appendix C-9A) which reflects the association of major fault zones and minor fault zones (page 40). The dike orientations best show the control of minor structures by the jointing in the plutonic rocks. A synoptic plot of all dike orientations measured in the study area (Figure 15) shows prominent northeast and northwest trends which can be correlated with joint sets in the plutonic rocks of all four thermal spring areas (Appendices C-1, C-4, C-7, C-10).

The general northeast and northwest trend of the minor structures is similar to the orientations of the major fault zones which control the course of the South Fork Payette River (Young, 1985).

Stress Analysis

Determination of the principal orientations of stress responsible for observed structures requires knowledge of the orientation of the structures, the direction
Figure 15: Synoptic equal-area plot of poles to dike orientations.
and sense of initial movement of the structures, and the angle between $\sigma_1$, the maximum compressive stress direction, and the structures (Hobbs, et al, 1976). All of these data are rarely known and so reliable principal stress orientations are not commonly obtained directly from field data.

The use of joints, the most pervasive structures in the study area, for determination of the principal stress orientations is rarely reliable. Anderson (1951) and Aydin and Reches (1982) have used multiple sets (including conjugate sets) of fractures to determine the orientations of principal stress applied to a rock mass. However, it must be possible to show that the multiple sets of fractures are contemporaneous for their methods to be accurately applied. In the study area, it was not possible to clearly relate multiple sets of fractures to one deformation event.

Ramsay (1967), Hobbs, et al (1976), and Davis (1985) discuss the effects of rheology and previous structures on fracture formation in rocks subjected to an applied stress. In anisotropic rock, faults and joints can be oriented at any angle between $0^\circ$-90$^\circ$ from $\sigma_1$, they need not contain $\sigma_2$, and displacement need not occur parallel to the $\sigma_1-\sigma_3$ plane (Hobbs, et al, 1976). Anisotropy may be due to rheological differences caused by multiple lithologies or pre-existing linear or planar fabrics in the rock. In addition, previously formed joints and pre-existing faults may be reactivated producing changing stress orientations in an area through time. The presence of aplite and pegmatite veins, at least five types of dikes, as well as primary jointing in all of the rocks and faults active since the Precambrian strongly implies that anisotropic conditions have existed in the rocks of the study area since they were emplaced.
and that no certain principal stress orientations can be determined from the field data.

The general northeast and northwest trends of the dikes in the study area (Figure 15 and Plate 1), suggests that extensional stresses approximately normal to these two orientations may have existed during portions of the Eocene and Oligocene Periods.

Comparison with Regional Structures

The study area is dominated by northeast and northwest trending major and minor structures with less pronounced north-south and east-west trends. This general pattern is consistent with mapped regional structures (Figure 16) which are dominantly northeast and northwest trending.

The trans-Challis fault system (TCFS) is a major crustal discontinuity expressed by numerous northeast trending normal faults, grabens, eruptive centers for the Challis volcanics, and many mineral deposits and has been active from the Precambrian to present (Bennett, 1986). The TCFS continues into western Montana where it aligns with the Great Falls lineament of O'Neill and Lopez (1985). The Corder fault zone (Plate 1) is the western most mapped TCFS structure and separates the study area into two halves. The eastern half, situated within the TCFS, is dominated by a northeast structural grain, whereas in the western half northwest trending structures predominate.

The Idaho porphyry belt and related dikes in the Pine Flat hot springs area are evidence of the control of the TCFS on emplacement of Tertiary igneous rocks. Major northwest trending Basin and Range faults directly to the south
Figure 16: Selected structural features in Idaho and western Montana (adapted from Bennett, 1986). Box outlines the study area.
and east of the study area (Figure 16) all terminate against the TCFS structures (Kiilsgaard and Lewis, 1983). However, Bennett and Knowles (1983) note that ancestral northwest trending basement faults may have cut the rocks west of the TCFS but most traces of them were destroyed by later faulting and erosion related to uplift of the Atlanta lobe. These hypothesized ancestral faults do not appear on any published geologic map. The northwest trend of the structures in the Deer Creek and Goller hot spring areas may be related, in part, to these ancestral structures. The trends of the Boise Ridge fault and the Deadwood fault zone differ from the general regional structural pattern. However, the sense of movement along these faults in the study area (Plate 1) and the timing of movement along them (Kiilsgaard and Lewis, 1983) suggests that they may be related to Basin and Range faulting.

The TCFS marks the boundary between Eocene extensional features to the north and Eocene extensional features overprinted by younger (Miocene) Basin and Range structures to the south (Bennett, 1986). Although no certain principal stress directions could be obtained from the field data the observed structural patterns are compatible with this proposed stress theory.

Discussion

The major fault zone system which controls the course of the South Fork Payette River is one of several deep-seated fault systems which control the major drainages of the Atlanta lobe (Young, 1985). These fault zones may have been active before batholith emplacement and the northeast and northwest trending segments may be related to early TCFS faults and the ancestral
northwest trending faults noted by Bennett and Knowles (1983). Recurrent movement along faults appears to be necessary to re-open fluid conduits closed off by mineral precipitation in many geothermal systems (Shirley, et al. 1978). Many northwest and northeast trending faults in the Atlanta lobe are still active today (Kiilsgaard and Lewis, 1983) and recent movement along these faults may be rejuvenating existing thermal spring systems and initiating the development of others.

The majority of thermal springs in the Idaho batholith occur along the major drainages which are fault controlled (Mitchell, et al, 1980; and Young, 1985). These geothermal systems use these major structures as conduits for deep circulation of groundwater (Kuhns, 1980; Youngs, 1981; and Young, 1985). In the study area, all thermal springs are located along the major fault zones which control the course of the South Fork Payette River. The high hydraulic conductivity (page 72) produced by intense fracturing of rock in the fault zones has permitted rapid movement of substantial quantities of water through the geothermal systems operating in the study area.
A chemical survey of a geothermal area is an essential part of an overall evaluation. Chemical analysis of thermal spring waters provides a rapid means of estimating source temperature, degree of water-rock equilibrium, solute content, and degree of mixing of rising thermal waters. Source temperature is of prime importance in evaluating the economic potential of a geothermal area.

Method of Investigation

Twenty-two 1 liter samples were collected in August of 1985 from 20 hot springs, one warm spring, and the South Fork Payette River. Sampling was done late in the summer to minimize surface dilution effects. Temperature, pH, and discharge rates were measured at the time of collection. The temperature was measured using a thermometer placed as close to the spring vent as possible while the pH was obtained using low-ion pH paper. The temperature and pH measurements are believed to be accurate to ±0.5°C and ±0.3, respectively. Discharge rates were measured by noting the time required to fill a container of known volume. The samples were acidified to a pH<3 using reagent grade HNO₃ to minimize loss of solutes by oxidation and/or precipitation and by adsorption to the surfaces of the polyethylene sample containers (Brown, et al, 1970).

Water samples were analyzed for SiO₂, Na, K, Ca, and Mg by atomic absorption spectrophotometry. Errors in the water chemistry data could be due
to sample contamination, instrument error, inaccurate standard solutions, and random error. Analysis of the HNO₃ used to acidify the samples (Table 2) and monitoring of baseline drift of the chart recorder allowed correction for sample contamination and baseline drift. Variations in machine response to a constant cation concentration were measured and are generally less than 1-2%. No value was estimated for random error and no correction was applied for possible error due to inaccurate standard solutions.

Source temperatures for the thermal spring waters were estimated using the SiO₂ and Na:K:Ca chemical geothermometers (Ellis and Mahon, 1977). Relatively large errors in the water chemistry can be tolerated with little significant change in estimated source temperatures because the cation concentrations are used in logarithmic functions in the geothermometer formulas (pages 64, 65, and 67).

General Characteristics of the Thermal Waters

Thermal spring water along the South Fork Payette River is characterized by low concentration of total dissolved solids (<600 mg/l) and is of a sodium bicarbonate type (Lewis and Young, 1980). Discharge from thermal springs ranges from seepage to 120 liters/minute at 33°C to 80°C (Table 2). Temperature, pH, and discharge measurements made in May and August revealed no variation over this period. The hot spring water is slightly alkaline with pH ranging from 8.1 to 9.3 whereas nonthermal waters have a pH<7.0. This distinction provides a useful field tool for identification of cooled thermal versus nonthermal water.
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Table 2: Results of water chemistry analysis and geothermometry of sampled thermal and non-thermal springs in the study area. All temperatures are given in degrees Celsius.
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</table>
Results of Chemical Analysis

The water chemistry data obtained from the 22 field samples is presented in Table 2. Two types of hot spring water can be identified. First, that of the Deer Creek hot springs area which is characterized by relatively high SiO₂, Na, K, and Ca. Second, the thermal waters of the other three areas to the east which are relatively low in SiO₂, Na, K, and Ca. Waters from all four areas have very low Mg concentration, generally <0.1 ppm which is close to the detection limit of the analytical system. Within any given area there is great variation in SiO₂ content but very constant Na, K, Ca, and Mg concentrations for hot spring waters.

The warm spring, GHS-6 (Figure 4), with a temperature of 17°C and pH of 6.8, yields anomalous values. Chemical analyses of Danskin Creek hot springs, located 10 km east of Crouch, and Warm Springs, located 4 km north of Crouch, are given although these thermal springs are not located within the four thermal spring areas studied. Water from the South Fork Payette River is chemically distinct from the thermal waters sampled (Table 2). This difference should aid in the identification of thermal waters that have mixed with near surface non-thermal water.

Chemical Geothermometers

The application of quantitative geothermometers in evaluation of a geothermal area is indispensable in economic feasibility studies. In addition,
the estimation of subsurface source temperature has important implications in hydrogeothermal modelling.

Quantitative chemical geothermometers are based on the effect of temperature on the equilibrium concentration of reactive solutes in geothermal fluids (Ellis and Mahon, 1977). Mineral solubility and ion exchange are the two primary types of reactions for which quantitative geothermometers have been developed. Basic to all chemical geothermometers are the following assumptions (Fournier, et al. 1974): 1) Temperature dependent reactions involving water and rock occur at depth 2) All constituents involved in the temperature dependent reaction are present in excess quantities 3) Water-rock equilibration occurs at the reservoir temperature 4) No re-equilibration of the "indicator" constituents occurs at lower temperatures as the water rises from the reservoir to the surface and 5) Either no mixing of the hot water from the reservoir and the cold shallow groundwater occurs or evaluation of the results of mixing is possible.

If the petrology of the geothermal area and the residence time of the thermal waters is known the first three assumptions can be verified. The fourth is not easily determined and can be highly variable for different geothermal systems. The fifth assumption can be validated if the chemistry of the near surface groundwater is known.

Attainment of equilibrium between water and rock both in the reservoir and during ascent may be dependent upon such factors as the kinetics of a particular reaction, temperature of the reservoir, reactivity of the wallrock, concentration of the indicator elements in the water, residence time of water in the reservoir, rate of ascent, and volume of ascending water (Fournier, 1977).
Thus, equilibrium may be attained for some reactions but not for others both in the reservoir and on ascent. Therefore, different chemical geothermometers may yield widely varying source temperatures for the same hot spring. Because of this, the assumptions and limitations of each geothermometer must be considered before a given temperature is considered valid.

Mineral Solubility Geothermometers

The only widely used and accepted quantitative geothermometer based on mineral solubility is the silica geothermometer (Ellis and Mahon, 1977). It is based on the solubilities of quartz, chalcedony, cristobalite, and amorphous silica in water which are temperature dependent reactions.

Equations relating the concentration of SiO₂ (concentrations in mg/l or ppm) to the subsurface last temperature of equilibrium (between 0°C-250°C) for various pure silica minerals are (Fournier, 1977):

1. Amorphous silica
   \[ T(°C) = \left[ \frac{731}{4.52 - \log C} \right] - 273.15 \]

2. Beta-cristobalite
   \[ T(°C) = \left[ \frac{781}{4.51 - \log C} \right] - 273.15 \]

3. Alpha-cristobalite
   \[ T(°C) = \left[ \frac{1000}{4.78 - \log C} \right] - 273.15 \]

4. Chalcedony
   \[ T(°C) = \left[ \frac{1032}{4.69 - \log C} \right] - 273.15 \]

5. Quartz (conductive)
   \[ T(°C) = \left[ \frac{1309}{5.19 - \log C} \right] - 273.15 \]
(6) Quartz (adiabatic) \[ T(\text{OC}) = \frac{1522}{(5.75 - \log C)} - 273.15 \]

Amorphous silica, beta and alpha cristobalite, and chalcedony are not present in abundance in the rocks of the study area and, therefore, the temperature estimates from the geothermometers based on their control of silica solubility are not considered valid. Quartz is an abundant mineral species in the rocks of the study area (Figure 11) and the quartz geothermometers are, possibly, reliable means of estimating source temperatures.

Experimental work by Rimstidt and Barnes (1977) showed that 10% of the initial silica in ascending waters will precipitate out in 8 hours at 250°C but at 100°C 13 days are required. For this reason, Fournier (1977) suggests an upper limit of 225°C for reliable application of the quartz geothermometers. As the estimated source temperatures for the study area are all less than about 150°C, potential loss of silica due to precipitation would occur very slowly and will be neglected. Fournier also suggests a lower limit for application of the quartz geothermometers of 150°C whereas Arnorsson (1975) believes 180°C is the lowermost limit of reliable quartz control on silica solubility.

The quartz adiabatic geothermometer assumes cooling occurs only by separation of a steam phase at depth from ascending waters. Source temperatures must be in excess of 100°C for adiabatic cooling to be possible. As with the quartz conductive method, the quartz adiabatic cooling geothermometer is reliable only over the range 150-225°C.
An advantage of the silica geothermometers is that dissolved silica is usually not influenced by common ion effects, formation of complexes, loss of volatile components, pressure or salinity effects and some form of silica is almost always available in sufficient abundance (Fournier, 1977). Restrictions on the ranges of appropriate temperatures and the large errors caused by dilution and evaporative loss are serious limitations.

**Ion Exchange Geothermometers**

Equilibrium constants for certain ion exchange reactions are temperature dependent allowing quantitative geothermometers to be developed which correlate the ratios of dissolved constituents to changing temperatures.

The Na|K quantitative geothermometer has been devised based on the exchange reaction:

\[ K^+ + \text{Na-feldspar} = \text{Na}^+ + \text{K-feldspar} \]

Both types of feldspars are found in the study area in abundance (Figure 11). Unlike the silica geothermometers, Na|K ratios are unaffected by evaporative concentrations (steam loss) and the rate of re-equilibrium during ascent is slower than for silica. This causes the Na|K ratios to yield generally higher source temperatures than silica geothermometry. Ellis (1979) found, however, that the Na|K ratio was a reliable measure of subsurface equilibrium temperatures only over the range 180-350°C. Below 180°C the Na|K ratio gives erroneously low temperatures (Fournier and Truesdell, 1973) due to the
effect of Ca on Na-K exchange. Due to these temperature range limitations and
the presence of Ca in the geothermal water samples, application of the Na|K
geothermometer is not considered valid for thermal waters in this study area.

Because of the limitations of the Na|K geothermometer, Fournier and
Truesdell (1973) proposed the empirical Na:K:Ca quantitative geothermometer:

\[
T(\degree C) = \frac{1647}{[\log(Na/K) + \log(Ca^{-5}/Na) + 2.24]} - 273.15
\]

where:

\[
B = \frac{4}{3} \text{ if } Ca^{-5}/Na > 1 \text{ and } T_{4/3} < 100\degree C
\]

\[
B = \frac{1}{3} \text{ if } Ca^{-5}/Na < 1 \text{ and } T_{1/3} > 100\degree C
\]

with concentrations in molal.

The Na:K:Ca geothermometer is applicable to higher Ca waters and lower
temperatures than the Na:K ratio. Due to the square root of the concentration of
Ca in the empirical formula, the Na:K:Ca geothermometer is subject to error
from dilution unless the original calcium concentration is small relative to
sodium.

Paces (1975) proposed a correction factor for thermal waters when the
Na:K:Ca estimated temperature is less than 75\degree C and the partial pressure of
CO₂ in the aquifer is above 10^{-4} atmospheres. These criteria are not met in the
study area hence this correction has not been applied.

Fournier and Potter (1979) proposed a Mg correction to the Na:K:Ca
geothermometer for waters with estimated source temperatures greater than
70\degree C and having 5>R>50 where \( R = [Mg/(Mg + Ca + K)] \times 100 \) with cation
concentrations expressed in equivalents. This correction factor:
\[ T_{mg} = 10.66 - 4.7415R + 325.87(\log R)^2 - \frac{[1.032 \times 10^5(\log R)^2]}{T} - \frac{[1.968 \times 10^7(\log R)^2]}{T^2} + \frac{[1.605 \times 10^7(\log R)^3]}{T^2} \]

when positive, is subtracted from the Na:K:Ca estimated source temperature. This correction factor was applied to DCS-2, GHS-3, and GHS-6 Na:K:Ca geothermometry.

Results of Chemical Geothermometry

Results of the application of the quartz conductive, quartz adiabatic, and Na:K:Ca geothermometers are presented in Table 2. Although, as discussed previously, only the Na:K:Ca geothermometer satisfies all of the criteria for valid application, the other geothermometer results are given so they can be compared to previous work in this area and in similar geothermal areas.

Calculated source temperatures are consistently higher for the Deer Creek thermal springs area than for the other three areas for all geothermometers. Both of the silica geothermometers show wider ranges of estimated source temperatures than Na:K:Ca geothermometry for the Goller, Corder, and Pine Flat hot springs. This reflects the variation in SiO₂ content of the thermal waters in these three areas due to non-equilibrium conditions resulting from relatively low subsurface source temperatures.

As none of the quartz conductive or quartz adiabatic estimated source temperatures are within the 150-225°C range, these figures may not be reliable. However, the Deer Creek hot springs results are close to the minimum
acceptable temperature and, therefore, quartz equilibrium may have been reached. The fact that the quartz temperatures are almost identical to the Na:K:Ca source temperatures supports this view.

The Na:K:Ca geothermometer is the only method discussed which satisfies all of the required assumptions and the results are, therefore, considered to be the most accurate. There is close agreement among Na:K:Ca source temperatures for all hot springs sampled from within a given area. As with the silica geothermometry, the Deer Creek hot springs area had a significantly higher average source temperature (141.9°C) than did Goller hot springs (72.0°C), Corder hot springs (73.7°C), or Pine Flat hot springs (68.6°C).

The cold spring, GHS-6, yielded an estimated source temperature of 6.3°C which, along with its anomalous chemistry, may indicate mixing with near surface non-thermal water whereas the warm spring, PFS-4, with a relatively high pH appears to have re-equilibrated with wallrock during ascent along a longer pathway (Fournier and Truesdell, 1973).

Comparison with Previous Work

A reconnaissance geochemical survey by Lewis and Young (1980) of thermal spring areas in the Payette River basin included analyses of nine thermal springs in the present study area. Reservoir temperatures for these nine springs were estimated using the quartz-conductive, Na:K:Ca, and sulfate-water isotope geothermometers. Their results are in close agreement with the source temperatures estimated in this study (Table 2). Their quartz-conductive source temperatures exhibit the same variability and cover a similar range as
those obtained here. Agreement between the two sets of Na:K:Ca source temperatures was excellent for all four thermal spring areas and the two sulfate-water isotope temperatures agree very closely with the results of Na:K:Ca geothermometry.

Discussion

Temperatures calculated from the Na:K:Ca geothermometer are considered to be the best estimate of subsurface temperatures. The Na:K:Ca geothermometer is known to be susceptible to dilution effects. A wide range of Na:K:Ca temperatures would be expected if thermal waters from a common source had mixed with near surface non-thermal water. However, in each of the four thermal spring areas there is good agreement among all hot spring Na:K:Ca values. This suggests that all thermal springs in a particular area are fed by a common source and that little mixing with non-thermal waters has occurred or that all hot springs in an area have had equal mixing at depth.

The observed wide variation in silica content of springs in each area is thought to be due to non-equilibrium conditions with respect to silica at depth in the thermal aquifer. Equilibrium was not reached for chalcedony, alpha and beta cristobalite, and amorphous silica due to a lack of abundant available supply of these minerals whereas insufficiently high subsurface temperatures inhibited quartz equilibrium except, possibly, in the Deer Creek hot springs area.

The source temperatures estimated for the Deer Creek hot springs are all above 100°C, the temperature at which adiabatic cooling becomes possible.
Although the quartz adiabatic cooling geothermometer is not valid for
determination of source temperature, adiabatic cooling of the ascending
thermal water may be an important factor in the Deer Creek hydrogeothermal
system.

The striking similarity of the water chemistry and the average Na:K:Ca source
temperatures of Goller hot springs and Corder hot springs (and to a lesser
extent Pine Flat hot springs) as well as their geographic proximity strongly
suggest that they may share a common subsurface thermal aquifer and may be
part of a large scale geothermal system. The Deer Creek hot springs have
distinctive water chemistry and source temperatures which implies they are
derived from a separate thermal aquifer.
HYDROLOGY

The interpretation of the hydrogeothermal systems in the study area is one of the primary goals of this project. Synthesis of geologic and geochemical data discussed previously combined with hydrologic data presented in this chapter is used to describe the predicted flow of groundwater through this area.

Hydraulic Conductivity

The hydraulic conductivity (K) of a rock unit is a constant of proportionality which describes the rate of flow of a particular fluid through the rock unit. It is a function of both the intrinsic permeability (k) of the rock mass and the density (ρg) and viscosity (u) of the fluid (Freeze and Cherry, 1979):

\[ K = k \rho g / u \]

For a given fluid, such as cold meteoric water, the hydraulic conductivity is dependent solely upon k. The intrinsic permeability is a function of the size, shape, and number of pore spaces in a rock unit and the degree to which they are interconnected.

In plutonic igneous rocks the intercrystalline voids are minute and not interconnected which results in extremely small primary permeabilities. Intrinsic permeability in these rock types is due almost entirely to fractures produced by thermal contraction and tectonic stress (Fetter, 1980). Thus, the fracture density is a good estimate of the relative hydraulic conductivity of the various igneous rock units.
As discussed previously (page 46) the granodiorite is consistently lower in fracture density than adjacent dikes (Table 1). The dikes will, therefore, have a higher relative hydraulic conductivity and will allow freer flow of water than the granodiorite. Among the various dike types the relative hydraulic conductivity decreases in the order: aphyric high-K andesite, aphyric high-K basalt, high-K andesite porphyry, and high-K dacite porphyry. As would be predicted, the vast majority of hot springs issue from the two rock types with the highest value of K, that is, the aphyric high-K andesite and aphyric high-K basalt dikes.

Fetter (1980) notes that joints in igneous rocks are generally closed at depths below about 100-300 m. due to lithostatic and tectonic stress. Because of this, hydraulic conductivity becomes extremely small below these depths, essentially stopping groundwater flow. However, granitic rocks remain brittle to depths of at least several kilometers and the dense fracturing which occurs within fault zones may allow deep circulation of groundwater (Freeze and Cherry, 1979). The major fault zones present in the study area are linear zones of high hydraulic conductivity and are deep seated features which provide conduits for circulation of large volumes of groundwater to depths of several kilometers.

The presence of secondary calcite noted in the fractures of several dikes associated with hot springs (page 43) suggests that gradual solution at depth and precipitation of minerals near the surface is occurring. This process is common in hot spring systems (Shirley, et al, 1978) and tends to enlarge original flow conduits and seal off secondary flow routes in newly formed geothermal systems.

In summary, the hydraulic conductivity of the rock units varies with the fracture density. The major fault zones are the only conduits for deep circulation
of large volumes of groundwater to depths of several kilometers. Based on the observed fracture density differences it is probable that deep within the major fault zones, the dikes have equal or higher fracture densities than the granodiorite and, therefore, are probable sites of rapid fluid movement. In the upper several hundred meters, joints open producing the observed fracture densities. This results in the preferential development of thermal spring vents in the aphyric high-K andesite and aphyric high-K basalt dikes. Enhancement of original fluid conduits has occurred by the gradual solution and precipitation of minerals by the rising thermal water.

Source of Water

The discussion of the hydraulic conductivity suggests that recharge of the geothermal systems is by downward percolation of water along the major fault zones. Source water for geothermal systems may include meteoric, magmatic, juvenile, and water released during metamorphic reactions (Ellis and Mahon, 1977). Young (1985) presents isotope data for the hot springs DCS-1, GHS-1, CHS-2, and PFS-1 which suggests that the source water for these springs is meteoric in origin (Table 3). During circulation of thermal water, deuterium proportions remain essentially unchanged due to the small amounts of hydrogen in rock minerals with which the water reacts. $^{18}O$ content in thermal water is generally enriched due to reactions with host rock minerals containing abundant $^{18}O$. The enrichment of $^{18}O$ values relative to deuterium values for the sampled thermal springs (Table 3) implies that the discharging hot water is not derived from current precipitation. The local meteoric water is enriched in
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</table>

Table 3: Isotope analyses and radiometric ages from sampled thermal springs and local meteoric water (Scott Mountain cold spring), (from Young, 1985).
deuterium compared with the thermal waters (Table 3). Young (1985) concluded that this is due to recharge of the geothermal system having occurred when the local climate was cooler and deuterium levels in local meteoric water were lower.

Tritium ($^3$H) and $^{14}$C radioactive isotope studies confirm a long residence time for thermal waters in the study area (Table 3). Tritium is a radioactive isotope of hydrogen which is incorporated in rain and snow and can be used to measure how long a particular water has been isolated from the atmosphere. Young (1985) obtained a residence time of at least 80 years, near or exceeding the maximum tritium-age dating limit, for hot water from GHS-1 (Table 3). Groundwater dissolves CO$_2$ from the air trapped in soil and application of the $^{14}$C radiometric age dating method to DCS-1 and GHS-1 spring waters yielded residence times of 28,800 and 9,000 years, respectively. Thus, it seems certain that recharge occurs by deep circulation of meteoric water along major fault zones with long residence times for the water in the geothermal system.

**Geothermal Gradient**

Geothermal gradients were measured by Brott, et al. (1976) from two mining exploration drill holes near the study area. The holes are located 3 km southwest and 5 km southeast of the Corder hot spring area and yielded geothermal gradients of 80°C/km and 36°C/km, respectively. The higher value is from Cretaceous biotite granodiorite whereas the lower value was measured from within the main body of the Idaho porphyry belt. As all of the thermal spring areas in this study area are underlain by Cretaceous granodiorite, the
higher (80°C/km) geothermal gradient, measured from this rock type, is used to estimate the depth of circulation of thermal waters. A geothermal gradient was not measured for the Deer Creek biotite granodiorite but the mineralogical differences (Figure 11) between it and the biotite granodiorite to the east are considered too slight to have a significant effect on the geothermal gradient in the Deer Creek hot spring area.

### Depth of Circulation

Using the average Na:K:Ca estimated source temperatures and the measured geothermal gradient from the Cretaceous biotite granodiorite, the depth of circulation for waters in each of the four thermal spring areas was estimated. A circulation depth of 2 km was obtained for the Deer Creek hot springs whereas the Goller, Corder, and Pine Flat hot spring waters all circulate to a depth of 1 km. These depths of circulation are similar to estimates made for other thermal spring systems in the Idaho batholith (Youngs, 1981; and Vance, 1986).

### Sources of Heat

The main source of heat in the earth is radioactive decay of potassium, uranium, and thorium (Brott, et al., 1976). This produces an imbalance between the hot core/mantle and the cool surface. Equilibrium is maintained by heat flow to the surface by three basic mechanisms: conduction, mass movement, and fluid circulation. Heat flow is measured in heat flow units (HFU) which quantify
the amount of heat flowing out of the earth's surface in a given time. The average heat flow for the world is 1.5 HFU and values in excess of 3.0 HFU are usually found only in geothermal areas (Brott, et al., 1976).

Surface heat flow has been calculated from the field measurements of the geothermal gradient and thermal conductivity of the rock units in two drill holes near the study area. Heat flow values of 4.3 HFU and 3.0 HFU were obtained from sites 3 km and 5 km from the Corder hot spring area, respectively (Brott, et al., 1976). The fact that areas of high heat flow occur several kilometers from the nearest hot springs suggests the presence of a thermal anomaly greater in areal extent than the surface manifestations.

Sources of the above normal heat flow in the study area may include conduction from the core/mantle, radioactive decay and conduction from the Idaho batholith granitic rocks, residual heat from a young shallow igneous intrusion, and frictional heat resulting from fault movement. No attempt to quantify the amounts of heat resulting from each of these four sources has yet been attempted. Brott et al. (1976) attributed a contribution of at least 0.5 HFU to the total heat flow in the study area to radioactive decay of the batholithic rocks. The presence of a near surface intrusive body young enough (3-5 million years) to supply significant heat (Rybach and Muffler, 1981) seems unlikely as the latest significant intrusive event in the study area ended in the Oligocene Period (page 36). The recurrent movement along the trans-Challis fault system (page 57) suggests that mechanical heat from faulting may supply some heat to the area although it is probably not a significant portion of the total heat flow. Therefore, it seems likely that conduction from the core/mantle and from the
batholithic rocks is the main source of heat driving the geothermal systems in the study area.

Driving Forces

Flow in thermal spring systems is produced by two forces acting on the water, hydraulic head and, what White (1968) termed, thermoartesian head. Hydraulic head is the force acting on the water due to the altitude difference between the spring vent and the recharge area. Recharge in the study area occurs upriver from the spring areas along the major fault zones. The circulation system created by hydraulic head differences is enhanced by density differences between the cold recharge water and the hot discharge water. When cold meteoric recharge water sinks down along the major fault zones it is heated at depth whereupon it becomes less dense and tends to rise. This phenomenon is thermoartesian head and is more pronounced with higher subsurface source temperatures. The density difference between recharge water (about 10°C) and Deer Creek source water (142°C) amounts to 7.4% whereas for the Goller, Corder, and Pine Flat source waters (71°C) it is 2.2%.

Comparison of the Four Geothermal Areas

Based on the hydrogeologic data, two main types of geothermal areas have been identified. The first includes the Goller, Corder, and Pine Flat hot spring areas. The second type is exemplified by the Deer Creek hot spring area.
The Goller, Corder, and Pine Flat hot spring areas are very similar in terms of geologic setting and geochemistry. All three areas occur along the major fault zones in the biotite granodiorite. The three thermal spring areas are sites of large numbers of Tertiary dikes (Plate 1) which appear to control the location of thermal springs (Figures 4, 5, and 6). The chemistry of the thermal waters and geothermometry from each of the areas is nearly identical (Table 2). The average Na:K:Ca source temperatures for each of the three areas are all near 71°C, corresponding to a depth of circulation of 1 km.

The many similarities between these three thermal spring areas and their geographic proximity (especially the Goller and Corder hot spring areas) suggests they may be part of a larger scale geothermal system. However, it is unclear whether the Pine Flat hot springs are fed by the same thermal aquifer as the Goller and Corder hot spring areas or two smaller geothermal systems exist. If two, or even three, smaller distinct systems are present, then the same depth of circulation with the same equilibrium concentrations of elements in solution at the same source temperature must be present. Only a prohibitively expensive hydrologic drilling program could potentially identify the presence of one large geothermal system versus two or three smaller ones.

The Pine Flat hot spring area is separated from the others by the main body of the Idaho porphyry belt (Plate 1). No thermal springs were observed discharging from this Tertiary body and a distinctly lower geothermal gradient was measured within it (page 76). It appears that the Tertiary intrusives in the study area have a much lower heat flow than the surrounding Cretaceous plutonic rocks, in contrast to the thermal characteristics of the Idaho batholith in general (Bennett, 1980). Most Tertiary intrusives in the Idaho batholith are
granites which are enriched in U, Th, and K giving them a high heat flow (Bennett, 1980). The Idaho porphyry belt dike rocks are, however, intermediate in average composition and are relatively depleted in these elements (Olson, 1968). This produces the lower heat flow values measured and accounts, at least in part, for the absence for thermal springs noted in the Idaho porphyry belt (Appendix A).

The Deer Creek hot spring area is distinct from the three areas to the east in its geologic setting and its geochemistry. The hot springs in the Deer Creek area discharge from Deer Creek biotite granodiorite in which Tertiary dikes are almost completely absent (Figure 3). The intersection of a major fault zone and a second large fault appears to control the location of thermal spring vents. The thermal water from these springs is enriched in SiO₂, Na, K, and Ca relative to the other three areas (Table 2). The average estimated source temperature (142°C) is double that of the other areas and suggests circulation to a depth of 2 km. With a source temperature in excess of 100°C, ascending thermal water may have cooled by separation of a steam phase. Adiabatic cooling from 142°C will have no significant effect on fluid composition (Fournier and Rowe, 1966) and, hence, it is not possible to determine if separation of a steam phase has actually occurred in the Deer Creek thermal spring system. The water chemistry and geothermometry of Warm Springs (Table 2) is very similar to that of the Deer Creek hot springs suggesting a possible common thermal aquifer or a similar type of geothermal system.

The Boise Ridge fault (Plate 1) separates the two types of geothermal systems in the study area. This large north-northwest trending normal fault
divides the area into two parts in terms of geology, geochemistry and hydrology
and appears to act as a hydrologic barrier.

Hydrogeothermal Systems

Analysis of the geologic, geochemical, and hydrologic data suggests the
presence of two types of geothermal systems in the study area. The first type is
represented by the Goller, Corder, and Pine Flat hot spring areas. The second
type of geothermal system addresses the different hydrogeologic setting of the
Deer Creek hot spring area.

Goller, Corder, and Pine Flat Hydrogeothermal System

Recharge of the geothermal system supplying the Goller, Corder, and Pine
Flat hot springs occurs along the major fault zones which control the course of
the South Fork Payette River. Cold meteoric water sinks along these zones of
relatively high hydraulic conductivity, upriver from the thermal spring areas. The
water circulates to a depth of 1 km where it is heated to about 71°C. At this
temperature equilibrium is reached with respect to the feldspar minerals but not
with respect to quartz. When the laterally moving water encounters the Tertiary
dikes it ascends along these high hydraulic conductivity units. This ascent is
driven by both hydraulic and thermoartesian head gradients. As discharge
temperatures are only 10-20°C less than estimated source temperatures (Table
2), rapid movement of thermal water probably occurs with only moderate heat
loss by conduction to the wallrock.
The similar chemistry of all spring waters and the areal extent of the thermal spring areas suggests that little or no mixing of thermal and non-thermal water has occurred. The estimated depth of circulation for these areas is 1 km but hot springs issue, in the Corder hot spring area, along a 1.5 km stretch of river. It seems more likely that little or no mixing has occurred than that equal mixing along numerous widely spaced flow conduits is taking place simultaneously. This conclusion is supported by the very low Mg concentrations in thermal spring waters compared with non-thermal groundwater (Table 2; Payette River sample).

Within several hundred meters of the surface, joints open and hydraulic conductivity increases for all rocks but the aphyric high-K andesite and aphyric high-K basalt dikes remain the dominant geothermal fluid conduits. Very close to the surface, minor faults, minor fault zones, and prominent joints may control the exact location of the thermal spring vents but the hot springs are always in close proximity to a Tertiary dike.

Mineral solution at depth and gradual precipitation of minerals during ascent has occurred. This process tends to enlarge primary conduits and seal off secondary routes in young geothermal systems, thereby enhancing the original flow volume and temperature. This decreases the potential for contamination due to mixing with non-thermal water (GHS-6) or for re-equilibration at lower temperatures due to cooling during a longer, slower ascent (PFS-4). With time, however, near surface precipitation of minerals may completely seal off a thermal water conduit. Recent fault movement has undoubtably served to re-open spring vents and initiate the establishment of new thermal spring conduits. By this process, the hot spring areas may retain the general recharge-discharge
circulation patterns with only minor changes in the actual upflow conduit locations. The many spring vents in each of the three areas have constant year long flow volumes, temperatures and chemistry implying the presence of mature geothermal systems (Ellis and Mahon, 1977).

Deer Creek Hydrogeothermal System

The Deer Creek hot springs share some characteristics in common with the three thermal spring areas to the east although several significant differences are evident. As with the three thermal spring areas to the east, recharge to the Deer Creek hot spring system is by downward movement of cold meteoric water along major fault zones. The water descends to a depth of 2 km where it is heated to about 142°C, reaching equilibrium with respect to feldspars and quartz. The absence of dikes in the Deer Creek area has resulted in a more homogeneous hydrologic setting. The intersection of two large faults has created a narrow zone of very high hydraulic conductivity. The laterally migrating hot water, after a long residence time in the system, rises rapidly along this highly fractured conduit. Hot spring vents occur only over a limited areal extent and equal mixing of all thermal water with non-thermal water at depth could have occurred. However, the very low Mg concentrations of thermal waters suggests that little or no mixing has taken place. Cooling of the thermal waters by separation of a steam phase during ascent may have occurred. The actual spring vents are located along the north trending fault and a shallow south dipping prominent joint set which intersects this fault near the surface.
As with the three geothermal areas to the east, recurrent movement along faults in the Deer Creek area has probably rejuvenated flow conduits plugged by precipitation of minerals from rising thermal waters.

Comparison With Other Geothermal Systems

The four thermal spring areas in the study area show many characteristics in common with other geothermal systems in the Idaho batholith. It seems that variations in rock fracture densities within a geothermal area control the recharge and discharge locations. Kuhns (1980), Youngs (1981), and Vance (1986) all suggest that recharge occurs along major fault/shear zones. Kuhns (1980) thought that andesite dikes in the Lochsa geothermal area acted as impermeable boundaries forcing hot water to the surface but no discussion of the relative fracture densities of the rock types was given. Youngs (1981) found that rhyodacite dikes had the highest fracture density of any rock type in the Running Springs geothermal area and that they acted as fluid conductors, in agreement with the model presented here. Weakly fractured quartz latite porphyry dikes which cross cut a zone of intensely fractured country rock force thermal water to ascend in the Horse Creek geothermal system (Vance, 1986).

Youngs (1980) noted one thermal spring occurrence at the intersection of two faults. Vance (1986), working mainly in metamorphic rocks on the southeastern margin of the Bitterroot lobe, also found that thermal spring areas were located where a fracture zone imbricates and dissipates and where two faults intersect. These settings are similar to that observed in the Deer Creek hot spring area.
The aforementioned authors' estimated geothermal gradients (40-51°C/km) are lower than the value (80°C/km) measured immediately south of the present study area. Their estimated source temperatures (40-200°C) are slightly less to significantly greater than the values (71-142°C) estimated here. This results in estimated depths of circulation (1-4 km), similar or slightly deeper than those (1-2 km) obtained here.

Young (1985), in a summary paper on thermal springs in the Idaho batholith, notes the association of hot springs with large regional fault systems that control the major drainages. An investigation of the role of Tertiary dikes or other geologic features controlling the location of hot springs along these fault zones was not attempted. The hot springs in the study area appear to be typical of the hot water dominated geothermal systems throughout the Idaho batholith.

Economic Potential of the Study Area

A goal of this study was to determine the potential for future development of the geothermal resources in the study area. As reservoir temperatures are all below 150°C, production of electricity in conventional steam plants, which require temperatures in excess of 200°C (Grose, 1971), is not possible. Binary cycle power plants are currently producing electricity from 175°C water (Nelson and Lacy, 1986) and testing with lower temperature waters continues. Based on measured spring discharge volume and the relatively small volume of highly permeable thermal aquifers, sufficient quantities of thermal water are not available for sustained binary power generation (Nelson and Lacy, 1986).
Residents of the study area use the hot spring water for bathing, space heating of homes and greenhouses, and for medicinal purposes. The few wells drilled for hot water in the study area have been shallow (<75 m) and constant volumes (up to 240 liters/minute) of water at temperatures similar to nearby springs have been obtained. It appears that sufficient hot water is present in this sparsely populated area to accommodate increased development of this resource for most direct use applications. In addition, the scenic setting of the hot spring areas and their proximity to a major population center (Boise) suggests that careful development of these areas for tourism and recreation may ultimately yield the greatest economic returns from this resource.
CONCLUSIONS

This study was designed to determine the geologic, geochemical, and hydrologic setting of the thermal springs occurring along the South Fork Payette River, between Lowman and Banks, Idaho.

The study area is underlain by Cretaceous granitic rocks of the Idaho batholith which were intruded by dikes ranging from basalt to rhyolite during the middle Tertiary (Olson, 1968). Large northeast and northwest trending major fault zones, many active since the Precambrian, control the course of the South Fork Payette River (Kiilsgaard and Lewis, 1983). Minor structures, particularly the dikes, are often joint controlled and exhibit general northeast and northwest trends. This pattern is consistent with structures of the Atlanta lobe (Bennett, 1986).

Two types of geothermal systems were identified in the study area. The Goller, Corder, and Pine Flat hot spring areas compose the first type whereas the Deer Creek hot spring area exemplifies the second.

Recharge to the Goller, Corder, and Pine Flat hot springs occurs by downward percolation of cold meteoric water along the major fault zones. A geothermal gradient of $80^\circ C/km$ and source temperatures of $71^\circ C$ estimated from chemical geothermometry suggests a depth of circulation of 1 km for thermal waters. The heated water rises rapidly with little or no mixing along the Tertiary dikes having the highest hydraulic conductivities. Flow along these dikes has been enhanced by solution and precipitation of minerals. The actual
spring vent locations may be controlled at the surface by minor structures, although a close association with dikes is always found.

The Deer Creek hot springs are distinct geologically and geochemically from the other three areas. Recharge to the Deer Creek system occurs along the major fault zones. However, estimated source temperatures of 142°C suggest a circulation depth of 2 km, twice that of the other spring areas. Isotopic work yields residence times of 28,800 years for waters of the Deer Creek area compared with 9,000 years for the other three areas. The absence of dikes in the Deer Creek area has apparently resulted in the control of hot spring location by cross cutting major faults. The intersection of two faults provides a zone of high hydraulic conductivity allowing rapid ascent of thermal waters with little or no mixing.

The association of thermal springs with major structures in the Atlanta lobe seems clear. More detailed mapping and geochemical analyses from other thermal spring complexes in the Atlanta lobe would aid in the understanding of the control of hot spring locations along these major structures. Geothermal gradient measurements are needed in the entire Idaho batholith to better understand the association between thermal anomalies and hot spring locations. If funding became available, a drilling program could be undertaken to better delineate the volume and directions of fluid flow in geothermal areas and the relation of adjacent thermal spring areas to a possible common thermal aquifer.
REFERENCES CITED


Bodvarsson, G., 1974, Dikes as fluid conductors in the extraction of geothermal energy: Geothermal Energy Mag., v. 2, n. 4, p. 42-50.


APPENDIX A:

GEOLOGIC ROAD LOG FROM LOWMAN TO BANKS, IDAHO.

The road log station locations are shown on Plate 1. The traverse is from Lowman to Banks along Idaho route 17.
APPENDIX A: GEOLOGIC ROAD LOG FROM LOWMAN TO BANKS, IDAHO

Station #

1. Two 2 m wide aphyric high-K andesite dikes intrude Cretaceous biotite granodiorite (Kbg) and trend N48W, 64S and N5W, 71W.

2. A 10 m wide aphyric high-K andesite dike intrudes Kbg and trends N22E, 77W.

3. Three 1 m wide aphyric high-K andesite dikes intrude Kbg and trend approximately N30E.

4. Eight closely spaced, subparallel, aphyric high-K andesite dikes ranging from 0.5-10 m wide intrude Kbg and trend N23E, 78E. A single 20 m wide high-K dacite porphyry dike parallels and forms the southeast border of this dike swarm. A minor fault trends N59W, 66S immediately to the north of the dike swarm outcrop.

5. A 0.5 m wide aphyric high-K andesite dike intrudes Kbg and trends N71W, 76N.

6. Six aphyric high-K andesite dikes ranging from 0.5-2 m wide intrude Kbg and trend N66W, 65S.

7. A 25 m thick high-K dacite porphyry dike trends N55E, 69N. Four aphyric high-K andesite dikes ranging from 10 cm to 10 m thick trend from N40W, 55W to N70W, 69S. All dikes intrude Kbg.

8. A 2 m wide aphyric high-K andesite dike intrudes Kbg and trends N12E, 79E.

9. An approximately 50 m wide swarm of cross cutting, subparallel, aphyric high-K andesite and high-K dacite porphyry dikes with a general trend of
Station #

N41E, 86E intrude Kbg.

10. A 2 m wide aphyric high-K andesite dike intrudes Kbg and trends N25E, 56W.

Refer to Figure 6 for the geology between Stations 10 and 11.

11. Gradational contact between Cretaceous biotite granodiorite (Kbg) and the main body of the Tertiary Idaho porphyry belt (Tpb).

12. Cold spring (15°C, pH=7.3) discharging 1.5 liters/minute from alluvium.

13. Cold spring (16°C, pH=7.0) discharging 1 liter/minute from alluvium.

14. Little Falls Molybdenum prospect. A 1 m wide aphyric basalt dike intrudes rhyolite dikes and trends N32E, 80E, immediately east of the main adit, was correlated to the Weiser Embayment flows of the Columbia River Basalt group by XRF analysis (Appendix B).

15. Highly weathered contact between the main body of the Tertiary Idaho porphyry belt (Tpb) and Cretaceous biotite granodiorite (Kbg).

16. Danskin Creek hot spring (Table 2) located in Kbg at the intersection of a northwest trending gully, which is possibly fault controlled, and the Payette River.

17. A 2 m wide aphyric high-K andesite dike intrudes Kbg and trends N16E, 66W.

Refer to Figure 5 for the geology between Stations 17 and 18.

18. A 1 m wide aphyric high-K andesite dike intrudes Kbg and trends N35E, 50W.

Refer to Figure 4 for the geology between Stations 18 and 19.

19. Exposure of Cretaceous muscovite-biotite granodiorite (Kmbg). View to
Station #

the west of the Boise Ridge fault scarp.

20. Exposure of Deer Creek biotite granodiorite (Kdcg).

Refer to Figure 3 for the geology between Stations 20 and 21.

APPENDIX B:
RESULTS OF X-RAY FLUORESCENCE WHOLE ROCK CHEMICAL ANALYSIS

Compositions are given in weight percent. Key to sample labels: D= Deer Creek hot spring area; G= Goller hot spring area; C= Corder hot spring area; P= Pine Flat hot spring area; R= Road log. Analyses obtained through the kindness of Dr. Peter R. Hooper.
### Appendix B: Results of whole rock x-ray fluorescence chemical analysis.

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APPENDIX C:
LAMBERT EQUAL-AREA STEREOGRAPHIC PLOTS OF STRUCTURAL DATA

The orientations of minor structures are plotted on Lambert equal-area stereonets. Minor structure orientations plotted include joints, dikes, minor faults, minor fault zones, and slickensides. Contour diagrams of joints in the Cretaceous plutonic rocks are contoured on the interval of 1% of the total data in 1% of the area of the stereonet.
Appendix C-1: Poles to joints in Deer Creek biotite granodiorite in the Deer Creek hot spring area (A) and contour diagram of joints (B).
Appendix C-2: Poles to aplite veins (A) and pegmatite veins (B) in the Deer Creek hot springs area.
Appendix C-3: Poles to minor faults (A) and trend and plunge of slickensides (B) in the Deer Creek hot spring area.
Appendix C-4: Poles to joints in the biotite granodiorite in the Goller hot spring area (A) and contour diagram of joints (B).
Appendix C-5: Poles to joints in dikes (A) and dike orientations (B) in the Goller hot spring area. (+) aphyric high-K andesite; (x) high-K andesite porphyry.
Appendix C-6: Poles to minor fault zones (A) and trend and plunge of slickensides (B) in the Goller hot spring area.
Appendix C-7: Poles to joints in biotite granodiorite in the Corder hot spring area (A) and contour diagram of joints (B).
Appendix C-8: Poles to joints in dikes (A) and dike orientations (B) in the Corder hot spring area. (+) aphyric high-K andesite; (x) high-K andesite porphyry; () high-K dacite porphyry.
Appendix C-9: Poles to minor fault zones (A) and trend and plunge of slickensides (B) in the Corder hot spring area.
Appendix C-10: Poles to joints in biotite granodiorite in the Pine Flat hot spring area (A) and contour diagram of joints (B).
Appendix C-11: Poles to joints in dikes (A) and dike orientations (B) in the Pine Flat hot spring area. (+) aphyric and porphyritic high-K andesite; (x) high-K basalt; ( ) high-K dacite porphyry.