Geothermal Alteration of Basaltic Core from the Snake River Plain, Idaho

Christopher Joseph Sant
Utah State University

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GEOTHERMAL ALTERATION OF BASALTIC CORE
FROM THE SNAKE RIVER PLAIN, IDAHO

by

Christopher J. Sant

A thesis submitted in partial fulfillment of the requirements for the degree of
MASTER OF SCIENCE in
Geology

Approved:

______________________            ________________________
John W. Shervais             Thomas E. Lachmar
Major Professor             Committee Member

______________________            ________________________
James P. Evans             Mark R. McLellan
Committee Member             Vice President of Research and
Dean of the School of Graduate Studies

UTAH STATE UNIVERSITY
Logan, Utah
2012
ABSTRACT

Geothermal Alteration of Basaltic Core from the Snake River Plain, Idaho

by

Christopher Joseph Sant, Master of Science

Utah State University, 2012

Major Professor: Dr. John W. Shervais
Department: Geology

The Snake River Plain is located in the southern part of the state of Idaho. The eastern plain, on which this study focuses, is a trail of volcanics from the Yellowstone hotspot. Three exploratory geothermal wells were drilled on the Snake River Plain. This project analyzes basaltic core from the first well at Kimama, north of Burley, Idaho. The objectives of this project are to establish zones of geothermal alteration and analyze the potential for geothermal power production using sub-aquifer resources on the axial volcanic zone of the Snake River Plain. Thirty samples from 1,912 m of core were sampled and analyzed for clay content and composition using X-ray diffraction. Observations from core samples and geophysical logs are also used to establish alteration zones. Mineralogical data, geophysical log data and physical characteristics of the core suggest that the base of the Snake River Plain aquifer at the axial zone is located 960 m below the surface, much deeper than previously suspected. Swelling smectite clay clogs pore spaces and reduces porosity and permeability to create a natural base to the aquifer. Increased temperatures favor the formation of smectite clay and other secondary minerals to the bottom of the hole. Below 960 m the core shows signs of alteration including color
change, formation of clay, and filling of other secondary minerals in vesicles and
fractured zones of the core. The smectite clay observed is Fe-rich clay that is authigenic
in some places. Geothermal power generation may be feasible using a low temperature
hot water geothermal system if thermal fluids can be attained near the bottom of the
Kimama well.

(113 pages)
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The Snake River Plain is located in the southern part of the state of Idaho. The eastern plain, which this study focuses on, is a trail of volcanics from the Yellowstone hotspot. Three exploratory geothermal wells were drilled on the Snake River Plain. This project studies basaltic core from the first well at Kimama, north of Burley, Idaho. The objectives of this project are to establish zones of geothermal alteration and analyze the potential for geothermal power production near the center of the Snake River Plain. Thirty samples from 1,912 m of core were sampled and analyzed. Mineral data, temperature data, and physical characteristics of the core suggest that the base of the Snake River Plain aquifer is located 960 m below the surface, much deeper than previously suspected. Swelling clay clogs pore spaces to create a natural base to the aquifer. Below 1,020 m the core shows signs of alteration including color change, formation of clay, and filling of other secondary minerals in vesicles and fractured zones of the core. Geothermal power generation may be a reasonable option using a hot water low temperature geothermal system if the thermal fluids can be accessed below the base of the Snake River Plain aquifer.
DEDICATION

I have many to thank for their support in assisting me through this project. First, I would like to thank my advisor, John Shervais, my thesis committee, and the many professors from Utah State University and other schools that have mentored and helped me get to the point where I am today. Special regards go out to my thesis committee for their concern and guidance toward me as a student, a scientist, and a researcher. They have taught me valuable lessons that will follow me throughout my life. Apart from my thesis committee I would especially like to thank clay mineralogist Jeff Walker from Vassar College. His insight and assistance with analysis, modeling, and interpretation were an integral part of this research. I thank him for all his valuable assistance, and for becoming a great friend. I would also like to thank my fellow student Katie Potter for allowing me to use her samples and unpublished data. Above all, I would like to thank my wonderful family for enduring my schooling years amidst my almost continual absence and for their love and support through this entire process. My thanks and love go out to my beautiful wife, Elizabeth, and three children, Sebastian, Paige, and Luke. Thanks to my friends, family and fellow students that have been a shoulder to lean on and for being willing to listen, and pretend to be interested, to all my science babble. So many have helped me get to this point that it is impossible to thank them all. To all those who have interacted with me through this process, you know who you are, and I thank you for your support.
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Chris Sant
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INTRODUCTION

OVERVIEW

This project is a small part of the Snake River geothermal drilling project (SRGDP). The SRGDP is a slimhole geothermal exploration drilling project that drilled three deep (up to 2 km) holes on the Snake River Plain (SRP) at Kimama, Kimberly and Mountain Home, Idaho (Figure 1). The primary goal of the SRGDP is to evaluate geothermal energy potential in three different settings of the SRP; the axial volcanic zone, the margins, and in a faulted graben. The Kimama drill site is located 40 kilometers north of Burley, Idaho on the axial volcanic zone of the SRP. The Kimberly drill site is located about ten kilometers southeast of Twin Falls on the margins of the plain. The Mountain Home drill site is located on the northwest corner of the Mountain Home Air Force base in a faulted graben (Figure 1).

Figure 1. Red stars mark the geographic locations of the three drill sites of the snake River Geothermal Drilling Project on the SRP, Idaho. This study focuses only on the Kimama drilling location, shown in black text. Imagery obtained from NASA 10 m DEM data acquired through GeoMapApp (after Shervais et al., 2011).
OBJECTIVES

The objective of this particular study is to establish the nature of and likely zonation of geothermal alteration as sampled in 1,912 m of Kimama core to understand the geothermal history of the axial volcanic zone of the SRP. Clay mineralogy will be used to establish the zones of alteration and assess the potential of geothermal power generation using sub-aquifer resources.

The zone of geothermal alteration can be defined two ways; first, defining the base of the SRP aquifer, and second, understanding the thermal alteration of the SRP. The base of the aquifer and zone of alteration will be defined using X-ray diffraction analysis of clay minerals in the core.

SIGNIFICANCE

This project has great significance to the communities on and around the SRP of Idaho by more correctly estimating the volume of water in the eastern SRP aquifer and estimating the geothermal energy potential on the SRP. Locating the base of the aquifer can help more correctly estimate the volume of water in the SRP aquifer. Also, understanding the thermal history of the SRP can help estimate the geothermal energy potential and future needs of geothermal exploration on the SRP.
BACKGROUND

SETTING

The SRP has potential for geothermal power generation. It houses one of the largest and most productive aquifers in the country and it has high heat flow in excess of 100 mW/m² (Blackwell, 1989).

The SRP is located in southern Idaho and spans almost the entire state from east to west and covers a 40,400 km² area (Lindholm, 1996). Traditionally, the SRP is broken into two parts, eastern SRP and western SRP. For ease of discussing my study area and the eastern SRP I will further divide the eastern SRP into the eastern SRP (ESRP) and central SRP (Figure 2).

Figure 2. Shows the SRP provinces. The western SRP is shown in green, the central province is outlined in yellow, and the eastern in blue. Three prominent locations denoting SRP province divisions, King Hill, Hagerman and the Great Rift are shown in black on the map. Imagery obtained from NASA 10 m DEM data acquired through GeoMapApp.
The division of the western and central SRP, starting from the south, is located along Salmon Falls Creek and transitions to the Snake River as it migrates from the southern part of the plain to its northern edge plain near Hagerman, Idaho, and then along King Hill Creek near King Hill, Idaho (Lindholm, 1996). The division of the central SRP and the eastern SRP is at the Great Rift zone which extends from Craters of the Moon to King's Bowl and the Wapi lava fields (Kuntz et al., 1992).

The CSRP and ESRP are a trail of volcanics created by the Yellowstone hotspot migration (Pierce and Morgan, 1992). The WSRP is a fault-bounded graben. Yellowstone volcanism commenced 17 Ma in present day Nevada and has continued through time to its current location at Yellowstone National Park, Wyoming. From northern Nevada and southern Oregon to Yellowstone National Park there are many volcanic fields and calderas and eruptive centers resulting from the Yellowstone hotspot migration (Pierce and Morgan, 1992) (Figure 3).

The long axis of the plain that trends SW to NE and is a topographic high within the plain. The axis contains the thickest accumulation of basalt, the highest concentration of young volcanic vents, and the smallest accumulation of sediment (Smith, 2004; Mark and Thackray, 2002).
The SRP has little evidence of faults bounding its lateral extent. However, Basin and Range faults trend NW to SE, perpendicular to the plain (Rodgers et al., 1990). Basin and Range northeast-southwest extension in the SRP is manifest by dikes bringing basaltic melt to the surface from the upper mantle (Smith, 2004). This extension can be seen in the Great Rift zone and other proposed volcanic rift zones that parallel basin and range faulting (Prinz, 1970; Kuntz et al., 1992; Hughes et al., 2002).

**Basalts of the SRP**

The majority of the basalts on the ESRP are highly evolved olivine tholeiite pahoehoe flows (Greeley, 1982; Leeman, 1982; Kuntz et al., 1992). The flows on the ESRP range from meters to tens of meters in thickness, and their geochemical signatures
indicate a mantle plume source (Hanan et al., 2008; Welhan et al., 2002). The bulk of the basaltic vents are clustered around the axis of the SRP (Kuntz et al., 1992; Smith, 2004).

**Eruptive Style**

The axis of the plain has the highest concentration of basaltic volcanic vents (Figure 4). The vents on the axis include young fields such as Hell's Half Acre, Cerro Grande, North and South Robbers, and King's Bowl (Kuntz et al., 1992; Smith, 2004).

Figure 4. Map showing the eastern portion of the ESRP, the locations of adjacent mountain ranges, prominent faults, volcanic rift zones, volcanic deposits, general groundwater flow direction, and the location of the Idaho National Laboratory. Figure taken from Kuntz et al. (2002).
Basalts on the SRP have been erupted mainly from low-angle shield volcanoes and some dike intrusions producing many small basaltic eruptions similar in style, but not scale, to the Columbia River flood basalts (Greeley, 1982). Many small eruptions collectively can form thousands of basaltic flow fields on the SRP. Cumulative thicknesses of basalt on the ESRP have measured over one kilometer thick (Lindholm, 1996). With an average flow being seven meters thick, one kilometer cumulative thickness would contain 142 lava flows stacked on one another, assuming no sedimentary interbeds.

**Basalt Facies**

Basalt facies in are important because the different facies can act as enhancements or inhibitors to fluid flow. The basalt facies vary in the vertical and horizontal directions of each flow. Important basalt flow facies include ash, shelly pahoehoe texture, basaltic cinders, basaltic rubble zones, ropey pahoehoe texture, vesiculated zones, and massive basalt.

In the horizontal direction the important facies are shelly pahoehoe texture, ropey pahoehoe texture, and basaltic cinders. Parts of the flow that are close to the vent have different morphologies than the distal parts of the flows. Close to the vent the flows are more platy, and may contain ash layers and cinders from fissure eruptions (Welhan et al., 2002). Distal parts of basalt flows are fed by surface lava channels or enclosed lava tubes just below the surface and produce ropey pahoehoe texturing. Lava tubes range from meters to tens of meters in diameter and are round or sub-round in cross section.
In the vertical direction the important facies include rubble zones, vesiculated zones, and massive zones. The rubble zones are located on the exterior of the flow, like a crust, and are composed of cooling and fracture features from the flow mechanics of the flow. The vesiculated zone is adjacent to the rubble zone and contains sealed bubbles of volcanic gases. The massive zone is in the very center of the flow and contains no gases, cracks or structures of any kind (Figure 5).

Figure 5. Diagram showing different pahoehoe lava flow facies. This diagram shows a cross-sectional view down the long axis of the flow, the flow progressed from left to right. The different facies of a lava flow are, increasing in porosity, the massive interior of the flow, columnar jointing facies with high vesiculation, Type-I Interflow Zone which consists of the fractured, rubbly, shelly crust of the flow, and Type-II Interflow Zone which consists of tension fractures located on the margins of the flow. Diagram from Welhan et al. (2002).

**Basaltic Structures**

Basaltic structures within individual lava flows are important for understanding permeability of flows and possible pathways for groundwater and thermal fluids. The main structures seen in and around basalt flows are tension cracks located adjacent to basaltic vents, tension fractures located along the margins of individual flows, and columnar joints located in the body of the flows themselves.
Tension cracks are large-scale features and have been found in several of the Holocene lava fields. Tension cracks have been identified as cracks that form in young basalt flows, are at least one meter wide, and have no vertical or horizontal motion associated with them (Kuntz et al., 2002; Prinz, 1970). Tension cracks are related to dike injection. They trend parallel to the dike intrusion and are somewhat symmetrical in their distance from the dike. The tension cracks represent dike extensional stresses and are thought to connect to the dikes at a depth range of one-half of a kilometer to one kilometer below the surface (Kuntz et al., 2002). On the surface the tension cracks are spaced one-half to one kilometer away from the dike intrusion and are visible in air photographs at King's Bowl lava field.

Tension fractures of the Wapi and Hell's Half Acre lava fields were studied because they were erupted from low shield volcanoes that represent a modern analogue of the buried SRP basalts (Welhan et al., 2002). Tension fractures are a system or network of cooling fractures that are located along the margins of individual lava flows and sometimes extend the entire length of the flow. Tension fractures are connected with expansion of the interior of a flow due to inflation (Hon et al., 1994; Self et al., 1997; Welhan et al., 2002).

Columnar jointing is another prominent structural feature in basalt flows. Columnar joints are a cooling feature similar to tension fractures. They form perpendicular to the cooling surface of the lava flow and when viewed from the top or bottom of a flow they form a network of columnar polygons that typically are pentagonal.
or hexagonal and range from centimeters to meters in diameter (Long and Wood, 1986; Winter, 2001).

**Sediments on the SRP**

It is important to understand the sedimentary record of the SRP because sediment layers may play a large role in fluid migration. Over the past four million years the SRP has, relative to adjacent regions, subsided about one kilometer (Smith et al., 1994). The result of this subsidence is more space for accumulation within the boundaries of the SRP. On the edges of the SRP in some of the basins, such as the Big Lost Trough, much sediment has accumulated. Sedimentary interbeds between basalts have been measured up to 100 meters in thickness (Smith, 2004).

Along the topographically high axial volcanic zone and other volcanic rift zones on the plain the most common sediments that have accumulated are loess and deposits that accumulated quickly during storm events.

Most of the sediment deposits from the river drainages adjacent to the plain are deposited on the margins and consist of pluvial lakes, laminated lacustrine silts, and clays which have been preserved between basaltic lavas (Bestland et al., 2002). Dry conditions in the Pleistocene also led to deposition of playas, eolian sands and silts, and the formation of paleosols (Smith, 2004). A study of the Big Lost trough revealed that sediments on the ESRP are largely governed by climatic effects. During dry interglacial periods the sediment deposition contains dune, and playa facies, and loess deposits. During the wet glacial periods the sedimentation is largely dominated by lake deposits, fluvial deposits, and gravels.
Fluid Flow in SRP Basalts

Many factors contribute to the porosity and permeability of the SRP, and many of them are the structural and facies elements discussed earlier. These include dikes, fissures, lava tubes, tension cracks, tension fractures, rubble zones, columnar jointing, massive interior of individual flows, and sediment layers. However, these elements may enhance or restrict fluid flow in the plain.

Fluid Flow Enhancements

There are many elements that may contribute to the enhancement of fluid flow in the basalts of the SRP including fissures, lava tubes, tension cracks, tension fractures, rubble zones, and columnar jointing. The axis of the SRP plays a larger than expected role in fluid transport than previously known (Smith, 2004). The composition and primary structures in the basalts on the axis of the plain may influence the porosity and permeability of the rocks. More vents along the axis than at the margins provide more areas for porous shelly pahoehoe, ash, and cinders, increasing overall permeability (Figure 6). More vents on the axis produce more lava flows, increasing the number of lavas stacked on top of each other, thereby increasing the potential thickness of a fluid flow zone. The axis, being a topographic high, collects less sediment that may act as aquitards than the margins of the ESRP.

Large-scale features such as fissures, lava tubes, and tension cracks can be major contributors to fluid flow in the SRP. They may act as open space for water to flow unrestricted in the subsurface. Fissures, lava tubes, and tension cracks can be meters
wide, kilometers long and can transport large amounts of water long distances in a short amount of time.

Figure 6. Diagram showing generalized hydraulic conductivities of a low-angle shield volcano. Conductivity values from high to low are scoria, ash and spatter, proximal flows, distal flows, crater filling flows, and feeder dikes. Figure from Welhan et al. (2002).

Tension fractures, rubble zones, and columnar joints in individual lava flows can also play a large role in fluid flow in basalts. Tension fractures in the Wapi and Hell's Half Acre lava fields are observed along the margins of flows and retain as much as 15% of their original porosity after burial by younger lavas (Welhan et al., 2002). The networks of tension fractures extend sometimes up to several kilometers in length and may connect fracture and rubble zones of multiple flows to create flow paths of high permeability (Welhan et al., 2002). When cooled, rubble zones can retain 50% porosity, providing much space for fluid transport. Overall, rubble zones contribute to the highest permeability and porosity of the ESRP (Figure 7), but tension fractures, columnar
jointing, vesicles, lava tubes, and ash layers also add to the total potential water pathways in the basalt flows (Smith, 2004; Lindholm, 1996). Groundwater takes advantage of all the open spaces, seeking out interconnected pathways for flow (Smith, 2004).

**Figure 7.** (A) Diagram showing how highly permeable interflow zones from individual lava flows can interconnect and provide high fluid flow pathways in the SRP basalts. (B) A cross-sectional view of two stacked pahoehoe lava flows and their relative permeabilities. Diagram from Welhan et al. (2002).

**Fluid Flow Restrictions**

The factors that may restrict groundwater flow in the SRP include the presence of dikes, fissures, massive zones, tension cracks, tension fractures, rubble zones, and sediment layers. Basaltic pahoehoe lava flows contain a massive dense interior with no fractures and little to no vesiculation (Figure 5). Basaltic dikes may also restrict fluid flow along the axial volcanic zone of the plain. The dikes are perpendicular to the groundwater flow direction. Even though the surface concentration of young eruptions are clustered around volcanic rift zones, the dikes are thought to be present throughout the entire plain due to the ever changing location of magmatism (Smith, 2004). Fissures
also play a role in fluid flow inhibition. Open fissures act as conduits for fine-grained sediments to be transported into deep basalts. Once the sediment has reached the bottom of the fissure it can then travel through rubble zones and open spaces in lava flows and lower permeability (Smith, 2004). Sedimentary interbeds collect on the basalts from wind-blown sediment and from deposition of sediment from river valleys adjacent to the margins of the plain during a volcanic hiatus, and may gather in the top rubble zones of flows, clogging up potential fluid pathways.

**Sediment Influence on Fluid Flow**

Hydraulic conductivity values of each sediment facies in the Big Lost trough were measured by Mark and Thackray (2002), and revealed that the non-channel deposits have values ($10^{-1.8}$ to $10^{-4.7}$ cm/s) similar to the massive interiors of basalt flows and restrict groundwater flow. Conversely, channel deposits have similar hydraulic conductivity values ($10^{0.9}$ to $10^{-1.7}$ cm/s) to the basalt rubble zones and act as preferential pathways for groundwater flow. Channel deposits, although they do not cover much geographical area of the SRP, may be of great significance to the groundwater pathway system (Cooke and Shervais, 1999; Mark and Thackray, 2002).

**SRP Aquifer**

The majority of information about the ESRP aquifer comes from studies completed at the Idaho National Laboratory (INL). It is extremely important to understand groundwater characteristics and properties at the INL because contaminants have entered the aquifer. Since 1952 aqueous chemical and radioactive waste has been
disposed of in deep wells at the INL site (Ackerman, 1991). Therefore, in recent years, extensive studies have been conducted to understand the SRP aquifer.

The SRP aquifer is one of the most productive aquifers in the United States (Lindholm, 1996). The ESRP aquifer is defined as the saturated zone in the SRP where fluid flow is rapid. Recharge for the aquifer comes from four main sources including percolation of irrigation water, streamflow, groundwater from adjacent mountain drainage basins, and precipitation. The main drainage basins that add water to the aquifer come from the mountain drainage systems north of the plain, including the Big Lost River, Little Lost River, Birch Creek, Camas Creek, and Mud Lake with the Big Lost River being the largest contributor to the aquifer (Orr, 1997; McLing, 1994). The waters of the aquifer are discharged mainly at two locations; King Hill and Hagerman, Idaho. An average annual discharge at Thousand Springs near Hagerman, Idaho is approximately $5.3 \times 10^5$ m$^3$ of water (Wood and Low, 1986). The main river drainages, mentioned above, drain into the aquifer rather than into the Snake River because the topographically high axial volcanic zone separates the drainages from the Snake River on the southern edge of the plain.

The SRP aquifer is mainly in basalt (Welhan et. al., 2002), and has transmissivity values that range from 10,220 m$^2$/day to 70,600 m$^2$/day at the INL, and a hydraulic gradient of 1.9 m/km with an average flow velocity of three meters per day.

The depth to water has a large range, and at the INL has been measured anywhere from 60 to 270 meters below the land surface (Ackerman, 1991; Knobel et al., 1992). The thickness of the aquifer is debated. Some suggest that the base of the aquifer is marked by
a sedimentary bed that overlies a 1.8 Ma basalt unit (Anderson and Bowers, 1995; Mann, 1986). Whereas others conclude that the depth of the aquifer is not known (Ackerman, 1991). Others suggest that the base of the aquifer can be correlated with the temperature inflection in wells as well as alteration characteristics (Morse and McCurry, 2002). Morse and McCurry (2002) suggest that authigenic mineralization is clogging pore space and restricting flow of aquifer waters, and that the mineralization correlates with the temperature inflection in the well.

Water temperatures of the aquifer are 6° C at the northeast end of the ESRP and show an increase in temperature to 16° C at the discharge zone at Hagerman, thought to be warmed by residual heat from magmatic intrusions in the rocks of the SRP (Smith, 2004). From top to bottom, the temperature of aquifer waters is nearly isothermal. Once the aquifer is penetrated, the temperature steadily rises, marking the base of the aquifer (Smith, 2004). The base of the aquifer at the INL has been measured in well WO-2 at 550 meters based on the temperature inflections (Morse and McCurry, 2002). The temperature inflection represents the transition from permeable rocks with rapidly flowing groundwater above, to impermeable rocks in which very little water movement occurs (Smith, 2004). In some wells at the INL site the temperature inflection also marks the boundary of hydrothermal alteration and mineralization of basalts (Morse and McCurry, 2002).

**Previous Work**

Contaminants from INL have found their way into the SRP aquifer, and study of their transport in the subsurface is a high priority. A significant study was completed at
the INL in 2002 about the local geology and its contribution to contaminant flow. This study contained geologic studies in the ESRP that included its sedimentary systems, volcanologic characteristics, groundwater chemistry, and bioremediation. Some of the influential studies that pertain to this work are summarized below.

Morse and McCurry (2002) investigated five wells at the INL site to examine if there is correlation between basaltic alteration and authigenic mineralization with depth of burial, age of the basalts, temperature inflection in the well, and/or interaction with upwelling of hot geothermal fluids. Through petrologic and geochemical investigations they found that there was no correlation between alteration or secondary mineralization with depth of burial or age of the basalts. What they did notice was that there was a strong correlation between the temperature inflection and basalt alteration depth in four of the five wells, interpreted as warm waters from the warmer lower part of the aquifer coming into contact with the upper cooler, faster flowing water of the aquifer.

Anderson and Bowers (1995) studied the SRP geology and its aquifer in relation to contaminant transport at the INL site. They examined rocks from many wells drilled at the INL, including the deepest drill hole on the ESRP, INEL-1, a 3,159 m deep well. They also examined rocks at Test Area North (TAN), part of the INL. Anderson and Bowers (1995) state that the effective base of the aquifer is located 304 m below the land surface and is marked by a sediment layer that is composed of sand, silt, and clay overlying a 1.6 Ma altered basalt. At a different well at TAN the base of the aquifer is located at 270 m below the surface, and is over 183 m thick. The depth to water at the INL site varies, but it ranges from tens of meters to over two hundred meters below the
surface. Geohydrologic conditions change considerably at different locations within the INL site. The aquifer at INL ranges in thickness from 183 m to 365 m. In some places wells can yield up to thousands of gallons per minute (Ackerman, 1991) and that the deeper the rocks are buried the less permeable they become (Mann, 1986).

Smith (2004) reviewed of all the major tectonics, volcanics, sedimentation, and geophysics, and related their effects on the groundwater system in the ESRP. This included summaries of the history of the ESRP, its style of volcanism, subsidence, crustal extension, heat flow, and topography. This is the first paper that gives an overall summary of the literature on the ESRP and how it impacts the aquifer. Ideas of note in this review are how all the different basalt facies contribute to the overall porosity and permeability of the aquifer system, and how the temperature changes from the northeastern end to the discharge area of the ESRP near Thousand Springs. This summary is important to this particular study because there are many variables that have an influence on fluid migration in the SRP, and it is crucial that all variables are considered. Smith (2004) also summarized the temperature measurements in deep wells drilled at INL, and their associated base of the aquifer (Figure 8). The average depth of the base of the aquifer drilled at INL from Figure 8 is 365 m below the surface, with the deepest base located at 550 m depth in well ANL-1. The average thickness of the aquifer at INL is 205 m, with the thickest aquifer being 364 m, also in well ANL-1. Figure 9 provides the same data, but in a different format that is easier to view. Figure 9 shows the depth and thickness of the aquifer at INL.
Figure 8 - Temperature gradient profiles of deep wells drilled at INL. Figure from Smith (2004).

Figure 9 - Graph showing the base of the aquifer levels and aquifer thicknesses of the wells drilled at INL. Blue bars indicate aquifer thickness. Red bars indicate the base of the aquifer of wells drilled at INL. Data obtained from Smith (2004).
HAWAIIAN GEOTHERMAL STUDIES

The ESRP's most closely related geologic relative is the Hawaiian volcanic system. The geologic conditions of the Hawaiian islands is very similar to the basalts on the ESRP. In the early 1990s Ingebritsen and Scholl (1993) summarized data from previously drilled geothermal exploration wells drilled on the island of Hawaii. This data is valuable to this work because it can serve as a model for what can be expected to be found in the Kimama well.

Temperature Data

In a summary of geophysical characteristics of hydrothermal systems on Hawaii, Kauahikaua (1993) compared six different well temperature profiles against each other. Four out of six wells had an inflection point in the temperature data in which the geothermal gradient suddenly changed. Well Pu'u Wa'awa'a stayed at a constant 20°C throughout the depth of the well. The temperature in SOH-4 was first measured at 120°C at about 900 m depth and steadily rose, but no major inflection point was observed. Up until the inflection point, each well measured almost a constant temperature. Four out of the six wells measured groundwater temperature above the inflection point between 20°C and 30°C. One well (Ashida-1) measured about 40°C. The wells with their respective temperatures and depths of the inflection points can be seen in Figure 10.

Alteration

Ingebritsen and Scholl (1993) summarized the geohydrology of Kilauea volcano, Hawaii. They revealed that permeability of basalts is affected by three factors, including the morphology of the lava flows, the frequency and quantity of dikes, and the degree of
hydrothermal alteration. Ingebritsen and Scholl (1993) also stated that hydrothermal alteration acts to reduce the permeability of the rocks by partially filling the vesicles through secondary mineralization.

Ingebritsen and Scholl (1993) listed the permeability of dike-free Hawaiian basalts, and concluded that they have comparable permeability values to karst limestone, well-sorted sands, and gravels reported by Freeze and Cherry (1979). Permeability reduction by hydrothermal alteration has been documented in the Cascade Range (Blackwell and Baker, 1988), Hawaii (Thomas, 1987; Ingebritsen and Scholl, 1993) and in Iceland (Tomasson and Smarason, 1985). Stone and Fan (1978) found chlorite and montmorillonite in some of the Hawaiian basalts that filled vesicles and fractures.

Figure 10 - Well temperature data from the island of Hawai'i. Figure taken from Kauahikaua (1993).
CLAY MINERALOGY

Minerals in volcanic rocks that are exposed to weathering and alteration processes commonly weather into different types of clays. The typical minerals found in volcanic rocks are quartz, feldspars, pyroxenes, amphiboles, olivine, and micas. All of these minerals weather to clay with the exception of quartz (Wilson, 1987). It can be expected to find clay minerals at Kimama, similar to what was found at the INL site (Morse and McCurry, 2002).

Clay minerals are silicate minerals and members of the phyllosilicates or "sheet silicate" group. The structure of clays can be broken up into two main atomic units, the tetrahedron and the octahedron.

The tetrahedron is a three-sided pyramid composed of four oxygen atoms that occupy the corners or tips of the pyramid with a single cation, typically silicon, in the very center of the pyramid. This structure is very strong. The bond strength of a silica tetrahedron is intermediate between an ionic and a covalent bond. The silica tetrahedra share pyramid points to form sheets and the sheets are arranged so that the bases are all oriented on the same plane and the tips all point in the same direction (Grim, 1968).

The octahedral unit can be described as almost a merging of two silica tetrahedra. If two tetrahedra were to join with the bases against each other it would form the exterior structure of the octahedron. The octahedron is an eight-sided bi-pyramid composed of six oxygen atoms and a single cation in the center that is usually aluminum, iron or magnesium. These structures also merge to form sheets.
Each clay mineral is composed of the two basic atomic layers already discussed (tetrahedral and octahedral), and it is the stacking order of these layers and cationic substitution (the atom in the center of the tetrahedron or octahedron) that distinguishes one type of clay from another.

There are many different types of clays but only smectite and illite will be reviewed because preliminary X-ray diffraction analysis of the Kimama core samples suggests the presence of smectites, and it is common for smectite to alter to illite. The stacking order for smectite and illite is the same and is composed of two tetrahedral layers, with an octahedral layer in between, known as T-O-T (Grim, 1968). The tetrahedral layers are bonded to the octahedral layer by the points or tips of the tetrahedrons and the bases are exposed, available to bond to other clay particles or molecules (Figure 11).

The main difference between smectite and illite, simply put, is its expandability. Smectite is expandable and illite is not. The reason for this is simple. Both smectite and illite contain an overall negative charge. The negative charge is satisfied by ions or molecules that bond to the tetrahedral sheets. The T-O-T layer charge of illite is more negative than that of smectite. A slight negative charge (smectite) attracts hydrous molecules such as water or organics that have positive polar charges. A more negative charge (illite) attracts ions with a purely positive charge that can bond tightly to the tetrahedral sheet, such as potassium. The size of the ion or molecule that binds to the tetrahedral sheet governs how far the clay layers expand from each other. A molecule of H₂O or CH₄ that bonds to smectite takes up much more space than a K⁺ ion that bonds to
illite, and causes the clay layers to expand from each other and take up more volume (Figure 11).
A subdivision of some clay minerals is described as dioctahedral or trioctahedral units. Cation substitution can occur in the octahedral layer of the clay particle. The “normal” cation in the octahedral sheet is Al\(^{3+}\). The ratio of Al\(^{3+}\) to other ions in the octahedral layer dictates whether the clay is dioctahedral or trioctahedral. A dioctahedral clay is when up to two-thirds of the Al\(^{3+}\) cations are substituted for a different cation of similar size. A trioctahedral clay is when three-thirds, or a full substitution, has occurred leaving no remaining Al\(^{3+}\) in the octahedral layer.

Clays were thoroughly studied in the late 1960s and 1970s by many researchers due to significance of clay transformations to petroleum exploration. The transition from smectite to illite is very important to the petroleum industry for two reasons. First, smectites contain water and organic molecules, potentially hydrocarbons, between individual clay layers and can give space (when altered to illite) to transport hydrocarbons to more porous reservoirs (Powers, 1967; Wilson, 1987). Second, the transition from smectite to illite occurs around 100° C, in the middle of the oil window (Burst, 1969). Just as the smectite to illite transition is important to the petroleum industry, it can also provide the same type of information about temperature and fluid migration in a geothermal system.

All minerals, except quartz, dissolve or decompose into a variety of clay minerals when exposed to weathering. Feldspars and micas weather into kaolinite and smectites. Ferromagnesian minerals such as olivine, pyroxenes, and amphiboles weather into trioctahedral smectites, vermiculite, and chlorite based clays (Wilson, 1987).
The reaction of dioctahedral smectite to illite in shale is thought to take less than one million years at temperatures between 75°-200° C (Johnston, 1983). Smectite is unstable above about 100° C and begins to react or transform to illite. There are two mechanisms by which this process is thought to occur; a dissolution and re-precipitation from only clay minerals (Boles and Franks, 1979; Nadeau et al., 1984; Ahn and Peacor, 1986), and an ionic substitution from sources other than the clays themselves (Hower et al., 1976; Eberl, 1984). Before transformation to illite begins, the smectite clays have a slight negative charge on the tetrahedral layers, thus, attracting polar molecules such as water and organic molecules. In the transformation of smectite to illite it is understood that that Al\(^{3+}\) substitutes for Si\(^{4+}\) in the tetrahedral layers resulting in a more negative charge for the layer overall. As a result, the water molecules are replaced by an ion with greater bond strength, almost always K\(^+\), and sometimes Ca\(^{2+}\), or Na\(^+\) (Figure 12). The single ion sits in the hexagonal-shaped hole in the tetrahedral layer, and satisfies the negative charge. Since there is no negative charge attracting polar molecules to bind between the clay minerals, the new mineral, illite, is un-expandable (Johnston, 1983; Grim, 1968; Moore and Reynolds, 1989).

**X-ray Diffraction Interpretation**

X-ray diffraction is the best method to identify individual clay species. Interpretation of X-ray diffraction of clays takes practice. Different types of clays have slight differences in the diffraction signal, and it can be difficult to distinguish one from another. However, with proper preparation techniques different species of clays can be differentiated. X-ray diffraction is very precise and accurate in the higher 2θ angles.
However, the best manifestations of clays occur in the low angles, below 20° 2Θ. The {001}, or basal reflection, is the most intense peak for clay X-ray diffraction, making it the most important for identification purposes.

Figure 12. The tetrahedral site as viewed from one side of the sheet, with the silica tips pointing up and out of the page. White circles represent oxygen atoms. The silica atoms are unseen, lying under the central circle in each triangle-shaped tetrahedron. Dashed circle in the middle of the tetrahedral structure is the site where the polar molecule or other ion binds to satisfy the negative charge of the sheet. The hexagonal structure interconnects with other tetrahedra to form a sheet that extends indefinitely in the lateral direction, indicated by gray tetrahedra.
Smectite, a commonly found clay, has a characteristic X-ray diffraction signal. An air-dried sample in a relatively humid location has a peak at $6.0^\circ \theta$ (15 Å), a peak at $5.2^\circ \theta$ (16.9 Å) after ethylene glycol treatment, and a peak at $10.0^\circ \theta$ after heat treatment (Moore and Reynolds, 1989). Illite has a peak at $10.1^\circ \theta$ that remains in the same location through all three treatments (Moore and Reynolds, 1989).

Peaks can be masked or suppressed by ionic influence or presence of other minerals. For example, Fe suppresses clay peaks in the $10^\circ$ to $15^\circ \theta$ region that are very small to begin with. Also, the presence of a mineral, such as pyroxene, in the low angles can have higher intensities than the clay present so that it has the effect of masking the clay to make it seem absent in the sample.

Peak shape can change due to an effect called defect broadening. Defect broadening is when clay unit cells, or layers, do not line up perfectly on the z-axis (Ergun, 1970). Defect broadening can be an offset, a rotation, or non-parallel units stacked on each other. The effect of defect broadening is a less intense broader peak. Defects in the clay minerals come from non-favorable crystallization conditions.

**Previous Work**

Hower et al. (1976) outlined a proposed mechanism for burial metamorphism of shale off of the Gulf Coast. This study is significant to basalts because during low-grade metamorphism and hydrothermal alteration of basalts the rocks are exposed to the same temperatures required for alteration of shales. During this study they investigated the transition of smectite to illite and suggested a mechanism for the chemical reaction. The suggested reaction is:


\[
\text{smectite} + \text{Al}^{3+} + \text{K}^+ \rightarrow \text{illite} + \text{Si}^{4+}
\]

Hover et al. (1976) suggest that this reaction is a process in which Al\(^{3+}\) replaces Si\(^{4+}\). This creates a negative layer charge and releases the water between the individual smectite layers. The final step is that K\(^{+}\) replaces the water or other molecule that was housed between the two layers. Throughout the process the tetrahedral layers stay more-or-less intact as opposed to completely dissolving and re-precipitating. The K\(^{+}\) is derived from potassium feldspar in the shale. The Al\(^{3+}\) comes from a source other than the neighboring clays. Hover et al. (1976) state that the reaction begins at around 100° C and provides a mixed-layer clay structure, and that at higher temperatures of 200° C all the smectite has changed to illite resulting in pure illite.

Boles and Franks (1979) also studied the transformation mechanism of smectite to illite in sandstones of southwest Texas. They propose that the reaction from smectite to illite proceeds as follows:

\[
\text{smectite} + \text{K}^+ \rightarrow \text{illite} + \text{Si}^{4+}
\]

This reaction uses only elements from the clay structures themselves, no outside sources. However, during the process of the reaction some smectite is dissolved in order to create the necessary ions (mainly Al\(^{3+}\)) to form illite. Also, this mechanism produces a lot of Si\(^{4+}\) as a by-product which may act as a cement in the sandstones they investigated. This mechanism could explain the loss of volume of illites as compared to smectites in addition to the collapse of the structure due to water being driven off.

Johnston (1983) outlined the mixed-layer proportions at given temperature ranges of the smectite to illite reaction. Illite starts to appear at 40°-50° C. Above 90° C smectite
becomes unstable and begins to react to form illite. The allevardite phase (50-65% illite) occurs from 90° to 150° C. Transformation to the kalkberg phase (75-80% illite) occurs from 100° to 175° C. Pure illite occurs above 200° C.

Ergun (1970) and Moore and Reynolds (1989) revealed that defect broadening changes the shape in the basal x-ray diffraction pattern. Defect broadening, or the defect-free distance, is a measure of how perfectly the clay unit cells are aligned with adjacent unit cells, or layers. The defects in the clay layers can be a rotation or an offset of the unit cells along the z-axis. Defect broadening is described using the variable N. The variable N is a distance measurement of the number of unit cells, or clay layers. The lowest value of N is 1, which implies complete disorder in the stacking of clay layers. The highest N is about 50, which implies 50 unit cells stacked in perfect order. The average smectite N is around 10 unit cells thick. The higher the value of N, the thicker the crystal. Lower values of N suggest more defects in the crystal. The effects of N change the shape of the diffraction peak. Low values of N broaden the peak, and high values of N heighten the intensity and narrow the breadth of the peak.

Garcia-Romero et al. (2005) studied clay minerals in sediments derived from basaltic parent material in Spain. They observed the presence of smectite, chlorite, mixed-layer clays, and some micas. The results of this study revealed that the clays present were formed from detrital, hydrothermal, and neoformed origins. Garcia-Romero et al. (2005) concluded that the factors controlling the composition of the clays includes the source and chemical makeup of the parent material, the depositional environment, and
early diagenetic processes. They also concluded that smectite, calcite and zeolites may originate from a hydrothermal origin.

Mizota and Faure (1998) studied clays of volcanic ash deposits from composite volcanoes in Japan. They sampled six ash deposits from three different volcanoes. X-ray data suggest that the bulk of the material is smectite with minor amounts of kaolinite and micas present. Mizota and Faure (1998) concluded that the smectites found in the ash deposits were formed by hydrothermal processes.
METHODS

SAMPLE COLLECTION

Core samples from the Kimama drill hole were collected using deep slimhole diamond coring. Drilling of the Kimama drillhole by DOSECC (Drilling, Observation, and Sampling of Earth's Continental Crust) began in September 2010 and was completed January 2011. The final depth of the drillhole was 1,912 m (Shervais et al., submitted). HQ core (6.35 cm diameter) was drilled to a depth of 1,184 m, and NQ (4.76 cm diameter) was collected to TD.

Samples were collected from the core by three methods. The first method was cutting the core using a rock saw. The samples removed approximately one quarter of the core cylinder volume about 10 cm long at each sample location. Samples were taken every 45 meters, for a total of 42 samples throughout the depth of the borehole. The second method of sample collection consisted of sampling each massive flow interior by taking mini core samples from the Kimama core. Katie Potter collected over 500 samples. However, only 10 of those samples were used in this study, and are between the depth interval from 917 to 1,038 meters. Only 10 of the samples were used between 917 to 1,038 meters because preliminary analysis showed that between the previously mentioned interval was the zone in which clays first started to appear in basalt vesicles. The third method of sample collection consisted of picking clay samples out of filled vesicles using dental tools and tweezers. Eighteen samples were collected using the dental tool method. Seventy samples were examined for this study.
SAMPLE PREPARATION

Samples were crushed with a rock hammer, a personal sized "chipmunk" crusher, hand crushing in an agate mortar and pestle, and a "shatter-box." The samples were crushed on a hard surface using a rock hammer to create pieces small enough to enter the chipmunk crusher, the shatter-box, or the mortar and pestle. Samples were crushed into pebble-sized pieces using the ceramic plate chipmunk crusher. These pieces were then ground into finer pieces using an agate mortar and pestle to separate out clays by gravity settling. After many samples were prepared to silt-size particles, I started to use the shatter-box instead of the chipmunk crusher and mortar and pestle because the samples deep in the hole were very hard to crush by hand. The shatter-box employs a tungsten-carbide-coated dish with a puck inside. Samples are loaded into the dish and clamped into the box-shaped apparatus and it is shaken until the material inside the dish is crushed into rock flour. The advantages of this method are its ease of use, speed, and its ability to easily crush very hard rocks. These crushed samples are then prepared into two different states for two different X-ray diffraction analyses.

The first X-ray diffraction method used examines whole-rock powders. The rock powder is back loaded into a disk-shaped depression about 2.5 cm in diameter and 0.4 cm deep so that the down-facing side, the scanning surface, is completely flat. The whole rock is pressed by hand into the depression and it produces a mostly-random oriented powder. This preparation method was used to scan for bulk whole-rock analyses.

Clay separates were prepared for detailed clay analyses. This method is a gravity settled clay separate. This process begins by mixing the rock powder into a solution of
de-ionized water and sodium-hexametaphosphate, a clay dispersant, in a plastic graduated cylinder for settling. The graduated cylinder was left standing, undisturbed and covered. The solution was left to stand for 30 minutes to separate the clays from the rest of the minerals in the bulk rock. After the specified time had lapsed based on Stokes' law of settling, the top three cm of the solution were withdrawn and deposited onto a glass slide using an eyedropper. The slide was dried in an oven at 75°C. Samples prepared at Vassar College were prepared using the filter membrane transfer method onto a glass slide.

After settling onto the glass slide has been accomplished the sample goes through a three-step treatment process, with an X-ray diffraction run after each step. Step one consists of being exposed to room temperature atmosphere. Step two consists of exposing the samples to ethylene glycol vapor at 60°C for 24 hours. Step three consists of heating the sample to 500°C for 2 hours.

SAMPLE ANALYSIS

X-ray diffraction analyses at Utah State University have been carried out using a Philips PANalytical X'Pert X-Ray Diffractometer (XRD). The software programs associated with the XRD are the PANalytical X'pert Data Collector, version 2.0b, published February 27, 2003, and the PANalytical X'Pert HighScore, version 2.2.0, published January 02, 2006. The Data Collector is used to collect the raw data from the XRD, and HighScore is used to refine and interpret the data from the XRD readings. All scans were run using CuKα radiation at 45kV and 40mA.

Analysis of four samples were performed at Vassar College using a Seimens D-5000 theta: 2-theta diffractometer at 40kV and 30mA with a 0.5° slit size. Samples were
run with a 0.02 step size at 1° 2Θ per minute from 0° to 35° 2Θ. The four duplicate samples run at Vassar College by my colleague Jeff Walker are from 1,234, 1,396, 1,676, and 1,798 m depths.

Twenty-two whole-rock, randomly oriented powders were analyzed in a 15-minute scan. The scanning parameters were a line scan at 2° 2Θ/min with 1° slit size and a 20 mm window from 3 to 33°. Clay separates were run once at room temperature with no chemical treatments applied. All 22 samples analyzed, with their corresponding samples depths, are shown in Table 1.

Clay separate scans were analyzed using a 1 or 2-hour scan depending on the treatment. The scanning parameters for each treatment were a line scan at 0.25° 2Θ/min with 0.5° slit size and a 10 mm window. Each clay separate sample was run after each stage of a three-stage treatment process. The first stage is exposure to room temperature atmosphere and it is run from 3° to 33° 2Θ. The second stage of the treatment subjected the samples to ethylene glycol vapors at 60°C for 24 hours and then run in the XRD from 3° to 18° 2Θ. Due the volatility of ethylene glycol, after about 1 hour the sample begins to change thickness which could cause inaccurate results. For this reason, glycolated samples were only run for 1 hour, from 3° to 18° 2Θ. Stage three consisted of heating the samples to 500°C for 2 hours, then cooling and running them from 3° to 33° 2Θ. All nine samples analyzed in the detailed clay study are shown in Table 2.
Table 1 - Whole-rocks powder diffraction samples analyzed with their names and corresponding depths.

<table>
<thead>
<tr>
<th>Sample Name</th>
<th>Depth (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>KMA-508</td>
<td>155</td>
</tr>
<tr>
<td>KMA-1000</td>
<td>305</td>
</tr>
<tr>
<td>KMA-1502</td>
<td>458</td>
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<tr>
<td>KMA-2002</td>
<td>610</td>
</tr>
<tr>
<td>KMA-2502</td>
<td>763</td>
</tr>
<tr>
<td>KMA-3000</td>
<td>914</td>
</tr>
<tr>
<td>KA1B-3009</td>
<td>917</td>
</tr>
<tr>
<td>KA1B-3061</td>
<td>933</td>
</tr>
<tr>
<td>KA1B-3153</td>
<td>961</td>
</tr>
<tr>
<td>KA1B-3160</td>
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<td>1005</td>
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<tr>
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<tr>
<td>KMA-5000</td>
<td>1524</td>
</tr>
<tr>
<td>KMA-5500</td>
<td>1676</td>
</tr>
<tr>
<td>KMA-6000</td>
<td>1829</td>
</tr>
</tbody>
</table>

All clay mineral modeling was completed using the NEWMOD software, 1985 version. NEWMOD is a computer program for the calculation of one-dimensional diffraction patterns of mixed-layered clays.
Table 2- Detailed clay analysis samples with their name, samples depth, and whether they were collected from vesicle fillings or from crushed whole-rock samples.

<table>
<thead>
<tr>
<th>Sample Name</th>
<th>Depth (m)</th>
<th>Collection Source</th>
</tr>
</thead>
<tbody>
<tr>
<td>KMA-3419</td>
<td>1042</td>
<td>Vesicle Fill</td>
</tr>
<tr>
<td>KMA-3557</td>
<td>1084</td>
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<td>KMA-4050</td>
<td>1234</td>
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<td>KMA-4300</td>
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</tr>
<tr>
<td>KMA-4580</td>
<td>1396</td>
<td>Vesicle Fill</td>
</tr>
<tr>
<td>KMA-4825</td>
<td>1471</td>
<td>Vesicle Fill</td>
</tr>
<tr>
<td>KMA-5500</td>
<td>1676</td>
<td>Whole-Rock</td>
</tr>
<tr>
<td>KMA-5900</td>
<td>1798</td>
<td>Whole-Rock</td>
</tr>
<tr>
<td>KMA-6000</td>
<td>1829</td>
<td>Whole-Rock</td>
</tr>
</tbody>
</table>
RESULTS

INTRODUCTION

I discuss three types of data in this study: the physical character of the core, temperature data, and mineralogical data. The physical data consist of a description of the Kimama core. Temperature data consist of a review of the temperature log data from the Kimama borehole. The mineralogical data review the results of the X-ray diffraction analyses.

PHYSICAL DESCRIPTION OF KIMAMA CORE

The final depth of the Kimama drill hole is 1,912 m below the surface. The drill hole yielded 2,118 m of core due to 128 m of duplicate core being drilled between 206 m and 334 m below the surface (Figure 13). Out of the 1,912 m drilled, 1,803 m is basalt, and 109 m is sediment (Potter et al., 2011). The basalt varies from fresh and unaltered to a state of alteration that it is almost unrecognizable as basalt at the bottom of the borehole. Flows range from meters to tens of meters in thickness.

The Kimama core contains 47 basalt flow groups and 557 flow units (Potter et al., 2011). The Kimama basalts are typically a light grey porphyritic diktytaxitic vesicular olivine tholeiitic basalt with phenocrysts of plagioclase and olivine in a fine-grained matrix of plagioclase, pyroxene, and glass. Phenocryst content ranges from about 10-20%, and vesicle content ranges from about 10-50%. The Kimama basalts are very similar to the basalts seen at the INL, described by Morse and McCurry (2002).
Figure 13 - Kimama drilling progress as a cartoon schematic. Note the duplicate cored section from 206 to 334 m. Diagram is not to scale. Diagram obtained from the DOSECC website (www.dosecc.org).

Authigenic mineralization in unaltered Kimama basalts consists of calcite and a mineral deposited near flow boundaries, thought to be an SiO$_2$ mineral, perhaps chalcedony. The first appearance of calcite in the core is around 100 m below ground level (BGL) and, based on visual observations, becomes more abundant with depth. Calcite crystals are observed lining and filling vesicles. The calcite crystals are euhedral, translucent dog-tooth and rhombohedral crystal forms. The SiO$_2$ mineralization is opaque, white in color, hard, and is first seen in vesicles and fractures around 840 m BGL.
Most vesicles above 950 m BGL are clear of sedimentation and authigenic mineralization (Figure 14). The vesicles in this depth range that are filled and lined are near flow boundaries. Rubble zones and vesicles above 1,100 m are mostly free of sediment, except for some calcite fill in rubble zones (Figure 16). The shallowest depth at which all vesicles are filled is 1,042 m. The greatest depth at which open vesicles are observed is at 1,833 m below the surface.

Figure 14 - Core from 877 m BGL showing the lack of sedimentation in vesicles, and rubble zone located in the top row of the box.

The first time authigenic clays appear as vesicle fillings is at 962 m BGL. The clay that is seen in the vesicle fillings is minor amounts of a green clay (Figure 15).
Below 1,100 m BGL vesicles alternate between being open, lined with authigenic minerals, and filled with authigenic minerals. Vesicles have been seen to be lined with calcite, clay or zeolites. Completely filled vesicles are filled with clay, a white silica-based mineral, or filled with a combination of clay, calcite and the white siliceous mineral (Figures 14 and 16).

Figure 15 - Core from 962 to 965 m BGL. The first appearance of clays can be seen as green vesicle fillings in this photograph in the upper left portion of the box at 962 m BGL.
Figure 16 - Core from approximately 1,060 m BGL, showing vesicles filled with clays, a white siliceous mineral and calcite.

At approximately 1,100 m BGL other minerals such as clays and zeolites begin to appear. Vesicles from 1,243 m to 1,432 m are filled with the white silica-based mineral, and occasionally thereafter until T.D. (Figure 17).
The color of the core changes with depth. The majority of the core above 1,020 m BGL is a fresh, unaltered grey color with a few zones that are oxidized to red. A sudden color change occurs from grey to green at about 1,020 m depth (Figures 18 and 19).
Figure 18 - Basalts from 1,009 m to 1,014 m depth. Basalts are still grey with open vesicles and no authigenic mineralization is present.

Figure 19 - Basalts from 1,025 m to 1,028 m depth have a greenish tint as well as clay deposition and other secondary minerals filling vesicles and fractures.
TEMPERATURE DATA

Temperature logs were run in the Kimama hole by three organizations. Two logs were run by DOSECC as well as bottom hole temperature during drilling, one by the Operational Support Group (OSG) of ICDP, and one by Southern Methodist University (SMU). Three of the logs show a sharp temperature inflection around 960 m BGL.

Drilling of the Kimama well was completed in January 2011. During drilling DOSECC measured bottom hole temperature periodically. In May of 2011, approximately five months after completion, SMU ran their temperature log (Blackwell, 2012). In June 2011 DOSECC and OSG ran their temperature log in the Kimama hole (Nielson et al., 2012). The most accurate log from each organization was used for temperature inflection measurements. The logs from SMU, OSG, and DOSECC Down were used to measure the temperature inflections in the well. The DOSECC Up and DOSECC Drilling log temperature inflections were not used to measure temperature because the DOSECC Down log is more accurate. The temperature inflections for the three chosen logs are measured at 960, 945 and 969 m BGL for the SMU, DOSECC Down, and OSG logs, respectively. SMU temperature log data can be seen in Figure 20. DOSECC and OSG's data can be seen in Figure 21. The temperature inflections mark the depth at which there is a change in the geothermal gradient of the SRP. The geothermal gradient for each of the three chosen geophysical logs has been calculated. Above the temperature inflection the geothermal gradients range from 4.13° to 5.00° C/km, with an mean of 4.51° C/km. Below the temperature inflection the geothermal gradient ranges from 75.3° to 82.1° C/km, with an mean of 78.5° C/km. Calculated values are in Table 3.
Table 3 - Calculated geothermal gradients for the Kimama drillhole based on temperature logs runs by DOSECC, OSG, and SMU. Gradients calculated in degrees Celsius per kilometer. Inflection depth is mentioned alongside the gradients.

<table>
<thead>
<tr>
<th></th>
<th>Above inflection gradient (°C/km)</th>
<th>Below inflection gradient (°C/km)</th>
<th>Inflection depth (m BGL)</th>
</tr>
</thead>
<tbody>
<tr>
<td>DOSECC</td>
<td>4.13</td>
<td>82.1</td>
<td>945</td>
</tr>
<tr>
<td>OSG</td>
<td>4.42</td>
<td>75.3</td>
<td>969</td>
</tr>
<tr>
<td>SMU</td>
<td>5.00</td>
<td>78.0</td>
<td>960</td>
</tr>
</tbody>
</table>

Figure 20 - Kimama temperature log data from SMU. The left most column is measured temperature, the middle column is the calculated geothermal gradient, and the right-most column is counts per second of the instrument. Temperature inflection is marked by blue arrow at 960 m BGL, and the geothermal gradient below 960 m BGL is marked with a red trendline at approximately 78°C/km. The crosses in the left-most column are actual temperature measurements made during drilling by DOSECC.
Figure 21 - Kimama temperature log of all logs run at Kimama with the exception of SMU which is shown in the previous figure. Blue line corresponds to the DOSECC log on the way up and out of the hole. Red line corresponds to the DOSECC log while being lowered down the hole. Purple line corresponds to bottom hole measurements taken during drilling. Green line corresponds to the OSG log. Inflection points of the DOSECC Down data (red) and OSG data (green) are shown by arrows in their matching colors along with their corresponding depth values.
MINERALOGICAL DATA

Whole-Rock Powders

Since this study focuses on clay analysis, and because the best clay peaks show up below 20° 2Θ, we focus on the data below 20° 2Θ in the detailed X-ray diffraction data. Whole-rock randomly oriented powders do not provide the best intensities for clay studies. The best intensities for detailed clay analysis are obtained from oriented samples, which will be presented later.

Twenty-two whole-rock powders were analyzed using X-ray diffraction. The samples were analyzed at various depths (see Table 1) below the ground surface. Prominent consistent peaks below 20° 2Θ in all of the 22 diffraction patterns occur at approximately 13.5°, 17°, and 19° 2Θ. All nine samples above 963 m have no clay peak present. The shallowest occurrence of clay in the samples is in sample 963 m. Between 963 m and 1,038 m the samples that show a clay peak are 963, 969, 972, and 1,038 m. Samples between 963 m and 1,038 m that do not have a clay peak are 970, 995, and 1,005 m. All seven samples below 1,038 m have a low-angle clay peak present below 10° 2Θ. All of the whole-rock analysis data can be seen in the Appendix.

Clay Separate Samples

Nine clay separate samples were analyzed using X-ray diffraction. Random powder mounts and oriented clay mounts were analyzed at both Utah State University (USU) campus and at Vassar College. The results for all nine samples are presented in the following section. Samples 1234, 1396, 1676, and 1798 were analyzed at USU and Vassar College. Clay separate samples were taken starting at 1,042 m BGL because that
is the approximate location where clays become abundant enough to collect a sufficient amount of clay for X-ray analysis.

Sample 1042 m is from a non-authigenic sedimentary interbed (Figures 22 and 23). The diffractogram has two main clay peaks. One peak is at 11.89 Å in the air-dried line and the other is at 16.14 Å in the glycolated line, but no clay peak is present in the heated line (Figure 24). This sample is not a classic smectite example, but does contain smectite derived from surface sediments. This sample also contains quartz at 3.33 Å. The peak at 8.95 Å may be pigeonite, a member of the clinopyroxene group.

Figure 22 - Sample location for sample 1042 m marked by red arrow. Sample taken from clay filling at tip of arrow.
Figure 23 - Close-up photo of the sample location of sample 1042 m. Sample taken from the brown clay.

Figure 24 - X-ray diffractogram from 1042 m depth. Clay peak present in the air-dried line at 11.89 Å. Clay peak present at 16.14 Å in the glycolated line. No clay peak present in the heated line. This clay is interpreted to be dioctahedral smectite.

Sample 1084 m is from broken pieces of the core in a vesicular section of a basalt flow (Figures 25 and 26). The diffractogram is a classic example of an authigenic
smectite (Figure 27). It has an air-dried peak at 12.0 Å, a strong glycolated peak at 16.9 Å, and a heated peak at 9.94 Å. The suppressed peaks above 10° 2Θ suggest that this is an Fe-rich smectite.

Figure 25 - Sample location for sample 1084 m marked by red arrow. Sample taken from broken-up basalts.

Figure 26 - Close-up photo of the sample location for sample 1084 m. Sample taken from the broken pieces of basalt.
Figure 27 - X-ray diffractogram from 1084 m depth. Clay peak present at 12.0 Å in the air-dried line. Clay peak present at 16.9 Å in the glycolated line. Clay peak present at 9.94 Å in the heated line. All three clay peaks indicate smectite clay. This clay is interpreted to be trioctahedral smectite.

Sample 1234 m is from fracture fill several feet above a sedimentary bed (Figures 28 and 29). This sample was run at USU and at Vassar College using both a random powder mount and an oriented sample. The oriented sample run at USU shows no clay peak present in the heated line (Figure 30). The Vassar College oriented sample shows two strong smectite peaks at 15 Å in the air-dried line, and 16.9 Å in the glycolated line (Figure 31). A random powder mount was also run for this sample to search for the [060] peak, which showed that this sample is dioctahedral. Full-width-half-max (FWHM) was
also measured to detect defect broadening, and measured at 1.32° 2Ө. This sample contains smectite, but it may not be authigenic due to the lack of a heated peak.

Figure 28 - Sample location for sample 1234 m marked by red arrow. Sample taken from clay filling a fracture at indicated location.

Figure 29 - Close-up photo of sample location for sample 1234 m. Sample taken from the fracture filling green tinted clay.
Figure 30- X-ray diffractogram from 1234 m depth. Clay peak present at 11.7 Å in the air-dried line, and an extremely suppressed peak at 17.0 Å in the glycolated line. No clay peak is present in the heated line. This clay is interpreted to be dioctahedral smectite.

Figure 31- Diffractogram of sample 1234 m. Red line represents air-dried, and the green line represents the glycolated line. The most intense peak in the air-dried line is located at 6.0° (15 Å). Most intense peak in the glycolated line is located at about 5.2° (16.9 Å). Figure courtesy of Jeff Walker, Vassar College.
Sample 1311 m is from a massive section of a basalt flow that was fractured and then sealed with mineralization (Figures 32 and 33). This sample contains a clay peak at 12.3 Å in the air-dried line, and two very strong clay peaks at 16.8 Å in the glycolated line, and 9.8 Å in the heated line (Figure 34). These data identify smectite clay in the matrix of the whole-rock as opposed to only in vesicle fillings.

Figure 32 - Sample location for sample 1311 m marked by red arrow. Sample taken from massive part of a basalt flow which contains secondary mineralization.

Figure 33 - Close-up photo of the sample location for sample 1311 m. Sample taken from a crushed portion of the basalt core.
Figure 34- X-ray diffractogram from 1311 m depth. Clay peaks at 12.3 in the air-dried line, 16.8 in the glycolated line, and 9.8 Å in the heated line indicate smectite clay. This clay is interpreted to be trioctahedral smectite.

Sample 1396 m is from teal colored vesicle fillings of a vesicular basalt flow (Figures 35 and 36). This sample was run both at USU and Vassar College using both a random powder mount and oriented samples on a glass slide. The USU data show a clay peak in the heated line at 9.96 Å (Figure 37). This sample also shows highly suppressed peaks in the angles above 7° 2Θ suggesting an Fe-rich sample. The Vassar college data show intense peaks at 15 Å in the air-dried line and 16.9 Å in the glycolated line (Figure 38). A random powder mount was run to search for the [060] peak, and the analysis showed that this sample is trioctahedral. Full-width, half-max was measured to detect...
defect broadening, and was measured at 0.83° 2Θ. This sample is a nontronite, an Fe-rich member of the smectite clay group.

Figure 35 - Sample location for sample 1396 m marked by red arrow. Sample taken from green/teal clay in fracture marked by arrow tip.

Figure 36 - Close-up photo of the sample location for sample 1396 m. Sample taken from the green clay fillings vesicles.
Figure 37 - X-ray diffractogram from 1396 m depth. Clay peak present at 14.29 Å in the air-dried line. Clay peak present at 13.63 Å in the glycolated line. Clay peak present at 9.96 Å in the heated line. This clay is interpreted to be trioctahedral smectite.

Figure 38- Sample from 1396 m depth. Red line is air-dried line and green represents glycolated line. Two major peaks are at 6.0° (15 Å) in the air-dried line, and a peak at 5.2° (16.9 Å) in the glycolated line. Figure courtesy of Jeff Walker, Vassar College.
Sample 1471 m is from vesicle fillings (Figures 39 and 40). This clay is the same brown hue as sample 1042 m that was sampled from a sedimentary interbed. The data show much the same results as in sample 1042 m. There is a clay peak present in the air-dried line at 12.3 Å, and at 16.9 Å in the glycolated line (Figure 41). This sample contains smectite. This sample could be the same material as sample 1042 m, a non-authigenic smectite clay. There is a chance that some drilling mud could have solidified in the vesicles here.

Figure 39 - Sample location for sample 1471 m marked by red arrow. Sample was taken from vesicle fillings.
Figure 40 - Close-up photo of the sample location for 1471 m. Sample taken from vesicles fillings in basalt core.

Figure 41- X-ray diffractogram from 1471 m depth. Clay peak present at 12.3 Å in the air-dried line. Clay peak present at 16.9 Å in the glycolated line. No clay peak present in the heated line. This clay is interpreted to be dioctahedral smectite.
Sample 1676 m is from a massive portion of a basalt flow (Figures 42 and 43). This sample was run at USU and Vassar College using only oriented samples. This sample contains no heated clay peak, but two excellent clay peaks at 15 Å in the air-dried line, and at 16.9 Å in the glycolated line (Figures 44 and 45). This is much like samples 1042, 1234, and 1471 discussed earlier that do not have a heated clay peak. The X-ray data suggests that this sample contains smectite. The full-width, half-max ratio was measured in this sample at 0.83° 2Θ. There are also strong peaks at 28° 2Θ that are most likely anorthite, the Ca end-member of the plagioclase family.

Figure 42 - Sample location for sample 1676 m marked my red arrow. Sample taken from a massive section of basalt.
Figure 43 - Close-up photo of the sample location for sample 1676 m. Sample taken from the massive basalt.

Figure 44 - X-ray diffractogram from 1676 m depth. Clay peak present at 14.11 Å in the air-dried line. Clay peak present at 16.72 Å in the glycolated line. This clay is interpreted to be dioctahedral smectite.
Figure 45- X-ray diffractogram from 1676 m depth. Air-dried line is red, and glycolated line is green. A clay peak is present in the air-dried line at about 6.0° (15 Å). A clay peak is present in the glycolated line at about 5.2° (16.9 Å). figure courtesy of Jeff Walker, Vassar College.

Sample 1798 m is from an altered massive section of a basalt flow (Figures 46 and 47). This sample was run at USU and Vassar College using both a random powder mount and oriented samples on a glass slide. The Vassar College data shows an air-dried peak at 15 Å, but neither data set show a heated clay peak (Figures 48 and 49). The Vassar College data show a glycolated peak at 16.9 Å (Figures 49). This sample is much like the previous sample, and does contain smectite clay. The air-dried peak in the Vassar College sample was only run to about 16° 2Θ because of a power outage. A random powder mount was run on this sample to search for the [060] peak. Analysis suggests that
the sample consists of dioctahedral smectite. The full-width half-max was ratio measured at 0.73° 2Θ.

Figure 46 - Sample location for 1798 m marked by red arrow. Sample taken from massive portion of a basalt flow.

Figure 47 - Close-up photo of the sample location for sample 1798 m. Sample taken from the altered portion of the basalt core shown.
Figure 48 - X-ray diffractogram from 1798 m depth. Clay peak present at 12.68 Å in the air-dried line. Clay peaks present at 16.72 and 13.53 Å in the glycolated line. This clay is interpreted to be dioctahedral smectite.

Figure 49 - X-ray diffractogram from 1798 m depth. Note: air-dried line is green, and glycolated line is red (opposite of other diffractograms from Vassar College). A clay peak is present in the air dried line at about 6.0° (15 Å). A clay peak is present in the glycolated line at about 5.2° (16.9 Å). Figure courtesy of Jeff Walker, Vassar College.
Sample 1829 m is from an altered massive section of basalt (Figures 50 and 51). This sample shows an air-dried peak at 13.0 Å, a glycolated peak at 16.8 Å, and a small and almost non-existent heated peak at 10.1 Å (Figure 52). It is possible that the two previous samples are the same material, but the heated peak is non-existent or extremely suppressed like this particular sample. This sample contains smectite, as most other samples analyzed. This sample also contains plagioclase at 28° 2θ.

Figure 50 - Sample location for sample 1829 m marked by red arrow. Sample taken from altered massive basalt.
Figure 51 - Close-up photo of the sample location for sample 1829 m. Sample taken from the altered portion of the basalt shown.

Figure 52 - X-ray diffractogram from 1829 m depth. Clay peaks present at 13.0 Å in the air-dried line, 16.8 Å in the glycolated line, and a suppressed, almost non-existent, 10.1 Å peak in the heated line indicate smectite. This clay is interpreted to be dioctahedral smectite.
Clay Modeling

Clay mineralogist Jeff Walker modeled smectite clay minerals with different defect-free distances using the NEWMOD software with the objective to determine the degree or quality of crystallinity of the smectite clay. He modeled one type of clay with two slight variations. Both clays modeled are a glycolated Fe-rich dioctahedral smectite. The only difference between the two models is the amount of defect broadening, or the value of N.

The first sample modeled, with low values of N, was modeled with a mean N of 1 (no perfect staking) and high N of 5 (Figure 53). The second sample modeled uses higher values of N, with a mean value of 2.5, and a high value of 12 (Figure 54).

The low N model has a less intense and more broad peak with a FWHM value calculated as 1.5° 2Θ. The sample modeled with low values of N look very similar to sample 1234 m, suggesting that sample 1234 m contains an Fe-rich dioctahedral smectite. The high N model has a high narrow peak with a FWHM ratio calculated as 0.7° 2Θ. The sample modeled with high valued of N looks very similar to sample 1798 m, suggesting that sample 1798 m is a Fe-rich dioctahedral smectite with a larger defect-free distance than sample 1234 m.

The implications of defect broadening are very apparent in the FWHM measurements. A decrease from a mean of 2.5N to a mean of 1N more than doubles the peak width at half the max height from 0.7° to 1.5° 2Θ. Thus, it can be seen that clays with a high degree of crystallinity (high N) are much easier to identify using X-ray diffraction because the diffraction peaks are more narrow and intense as opposed to being
short and broad with a potential to be masked by other X-ray signals. Measuring the FWHM ratios and also using a visual peak width comparison, this modeling technique can help determine the crystallinity quality, and thereby give some indication of what ideal conditions were like at the time of formation.

Figure 53 - Dioctahedral Fe-rich smectite model created in NEWMOD showing the influence of defect broadening. Low N values shorten and broaden the [001] peak. The full-width half-max ratio measured at $1.5^\circ 2\Theta$.
Figure 54 - Dioctahedral Fe-rich smectite model created using NEWMOD showing the effect of defect broadening. Higher values of N narrow the peak and heighten the [001] peak. The full-width half-max ration is measured at 0.7° 2Θ.
DISCUSSION

BASE OF AQUIFER

Three main observations highlight the base of the aquifer on the axial volcanic zone of the SRP. The three observations come from the physical characteristics of the core, temperature log data, and mineralogical data.

The first signs of alteration occur around 960 m depth when clays first begin to appear as vesicle linings. The color of the core remains a light grey color, then a sudden color change occurs at 1,020 m depth. The color changes from light grey to green due to alteration, the vesicles begin to be filled with clay and other minerals, including hydrothermal minerals such as zeolites.

Second, the temperature log data exhibits a major change in the geothermal gradient in the Kimama borehole at 960 m depth. The geothermal gradient changes from 4.5 to 78.5 °C/km indicating that the fast flowing aquifer waters are not present below 960 m BGL.

Third, the mineralogical data suggest that smectite clay appears in the basalts just below 960 m BGL. The presence of clay in the basalts causes pore spaces to be clogged and the smectite clay provides an explanation for slower moving thermal waters below 960 m BGL that have been noticed at other locations on the SRP (Smith, 2004).

Two of the three observations are seen at about 960 m BGL, including the temperature logs indicate a sharp major change in the geothermal gradient and X-ray diffraction patterns show the appearance of clays. The physical characteristics of the core show a gradational change starting at 960 m BGL from a fresh light grey basalt with no
clay to an altered darker green basalt with clay filled vesicles. All three observations point to clogging of the basalt pore spaces to create a natural boundary between the fast moving fresh water and the slower moving thermal waters. Morse and McCurry (2002) and Smith (2004) found similar results on the SRP at varying depths.

I suggest, based on these data that the base of the aquifer on the axial volcanic zone of the SRP at Kimama is 960 m below the surface. Compared with estimated depths of 500 m or less at the INL (Smith, 2004; Morse and McCurry, 2002) on the margins of the SRP, the thickness of the aquifer on the axial zone at Kimama is much greater than previously suspected, almost double.

ALTERATION ZONES

Using observations from core samples, temperature log data, and mineralogical data one major boundary of alteration is present in the volcanic axis of the SRP at Kimama. The major boundary between unaltered basalts and altered basalts begins at 960 m BGL. It is at this depth which the temperature inflection is located, when smectite clay begins to clog the pore spaces in the rock, and secondary mineralization begins to occur. Above 960 m BGL the basalts show no significant signs of alteration. Between 960 m BGL and 1,020 m BGL the core shows a gradational change from fresh to altered basalts that is manifest by an increasing abundance of clay lining and filling the vesicles and pore spaces. Below 1,020 m the basalts show signs of alteration such as a color change, vesicle filling, and other secondary mineralization.
THERMAL ALTERATION

The chemical analyses in this study confirm the presence of smectite clay in the Kimama borehole between 960 and 1829 m BGL. The presence of smectite is expected because it has been found in basalts of the SRP (Morse and McCurry, 2002) and in the basalts of Hawaii (Tomasson and Smarson, 1985) which have a very similar geochemical signature to the basalts of the SRP. Smectites from Kimama are both dioctahedral and trioctahedral.

Dioctahedral clays are commonly found in sedimentary rocks and form from the weathering of K-feldspars. Dioctahedral smectites, with increasing temperature, convert to illite. The included models suggest that two of the samples (1234 and 1798 m) that do not have a heated smectite peak are dioctahedral smectite. Therefore, it can be assumed that all samples that contain smectite peaks in the air-dried and glycolated lines, but do not have a heated peak at 10 Å are dioctahedral smectites. The samples 1042, 1234, 1471, 1676, 1798, and 1829 m may be interpreted as dioctahedral smectites. Samples 1042 and 1234 m were samples a short distance from sedimentary interbeds so it is expected that they are dioctahedral.

Trioctahedral clays are derived from the weathering of mafic minerals such as pyroxene, olivine, and basaltic glass. With increasing temperature and pressure trioctahedral clays convert to chlorite instead of illite. X-ray diffraction data reveal that sample 1396 m is a trioctahedral smectite. Sample 1396 m has a heated clay peak, whereas, the dioctahedral samples do not. Therefore, we assume that all the samples with
a 10 Å peak in the heated line are trioctahedral. The samples 1084, 1311, and 1396 m may be interpreted as trioctahedral smectites.

Since six of the nine the samples analyzed are interpreted as dioctahedral smectites and because smectites should become unstable at generally the same conditions whether they are dioctahedral or trioctahedral, it is sufficient to use the model by Johnston (1983) as a temperature of formation guide. Johnston (1983) suggests that above 90° C smectite begins to become unstable and converts to illite. No mixed-layer clays of smectite/illite or smectite/chlorite were observed. Therefore, the temperatures since the formation of the present smectite clays present in the basalt has remained below 90° C. The time period since formation of the smectites is unknown.

Defect broadening of dioctahedral clays was modeled and has higher values of N, the peak width decreases and the intensity of the peak increases. This observation is measured using the FWHM method. Four FWHM measurements were made among the nine clay samples analyzed. The samples with FWHM measurements are 1234, 1396, 1676, and 1798 m, and have values of 1.32°, 0.83°, 0.83°, and 0.73°, respectively. As the FWHM values decrease with depth, the values of N increase with depth. Based on the FWHM measurements it can be observed that the clay crystals become less defective, or better crystallized with depth. Better established crystallization with depth implies that the conditions for smectite crystallization become more favorable with depth.

It is expected that the geothermal gradient will remain constant with depth at the calculated 78.5° C/km. If the geothermal gradient stays constant until 2300 m depth the temperature would reach 120° C. At temperatures of 120° C conditions would promote a
layered smectite/illite mineral with 75-80% illite (Johnston, 1983). It is also expected that the unstable smectites would form into a mixed-layer clay.

GEOTHERMAL POWER GENERATION

Under the geohydrologic conditions on the axial zone of the SRP, the most likely geothermal system is a hot water geothermal system. A hot water geothermal system ranges in water temperatures from 50° to 150° C (Gupta and Roy, 2007). The temperatures measured in the Kimama well were less than 60° C. This system could classify as a low temperature hot water geothermal system.

After all the observations and geochemical analyses are considered an assessment of geothermal power generation potential can be made. The alteration boundary at 960 m depth marked by the temperature inflection, geothermal alteration, and smectite clay could act as a great cap or seal for a hot water geothermal system. The swelling smectite clays clog porosity and permeability creating a natural seal for rising sub-aquifer thermal waters.

This geohydrologic condition in the axial zone is a resource that is worth further exploration by drilling exploration wells that are deeper than 2 km, different geothermometry analyses, and additional geophysical techniques including additional temperature measurements, or more wells drilled on the axis of the ESRP further to the NE in younger, presumably hotter volcanic fields. At the present time, it could be debated whether geothermal resources are economically reasonable in this area. Well depths of 2 km or greater should yield temperatures above 100° C and the formation of illite which enhances fluid flow of thermal waters. The same caliber resources may be found closer to
the surface at the margins of the SRP where the aquifer is thinner, thereby decreasing initial cost of geothermal production.
SUMMARY

After study and analysis of the alteration conditions along the axial volcanic zone of the SRP it has been confirmed that the base of the aquifer is located at 960 m BGL. The base of the aquifer is marked by the temperature inflection in the well, a color change in the rocks, appearance of smectite clays in vesicles, fractures, and the matrix of the basalts, and the appearance of secondary authigenic mineralization.

The alteration boundary is marked at 960 m depth, where all basalts above are unaltered. Basalts below 960 m BGL show signs of geothermal alteration, such as a color change, appearance of clays, and other secondary mineralization.

Smectite clay was observed in the Kimama drillhole. No transition to illite or chlorite was observed. Highly defective Fe-rich smectites were observed, in both the dioctahedral and trioctahedral varieties. It is expected that at greater depths and higher temperatures the smectite would become unstable and form into illite and/or chlorite. It is concluded that the smectite clays and other hydrothermal alteration mineralization are clogging pore spaces in the basalts at depth to create a natural base of the aquifer. The smectite clay is sourced from weathered material from the surface (dioctahedral clays) and from weathered basalt minerals such as pyroxenes and olivine (trioctahedral clays). Of all the clays analyzed, dioctahedral clays are twice as abundant as trioctahedral clays.

Producible geothermal power generation from sub-aquifer resources is a low temperature resource. This system could be classified as a low temperature, hot water geothermal system. It could provide low temperature geothermal energy to residents and businesses on the SRP. However, the depth required to reach such conditions is over 1
km. Economical geothermal prospects may be more abundant along the margins of the plain where the aquifer is thinner. Further studies should be conducted elsewhere on the ESRP to establish a more correct idea of the available geothermal resources.
REFERENCES


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APPENDIX
X-RAY DIFFRACTION PATTERNS
Figure A1 - X-ray diffractogram from 155 m depth. No clay peak present.

Figure A2 - X-ray diffractogram from 305 m depth. No clay present.
Figure A3 - X-ray diffractogram from 458 m depth. No clay peak present.

Figure A4 - X-ray diffractogram from 610 m depth. No clay peak present.
Figure A5 - X-ray diffractogram from 763 m depth. No clay peak present.

Figure A6 - X-ray diffractogram from 914 m depth. No clay peak present.
Figure A7 - X-ray diffractogram from 917 m depth. No clay peak present.

Figure A8 - X-ray diffractogram from 933 m depth. No clay peak present.
Figure A9 - X-ray diffractogram from 961 m depth. No clay peak present.

Figure A10 - X-ray diffractogram from 963 m depth. A small, but noticeable, clay peak is present at about 6° 2Θ.
Figure A11 - X-ray diffractogram from 969 m depth. A small, but noticeable clay peak is present at about $6^\circ \ 2\Theta$.

Figure A12 - X-ray diffractogram from 970 m depth. No obvious clay peak present, but perhaps a minor amount of clay is present at about $6^\circ \ 2\Theta$. 
Figure A13 - X-ray diffractogram from 972 m depth. A subtle clay peak is present at about 6° 2θ.

Figure A14 - X-ray diffractogram from 995 m depth. No clay peak present.
Figure A15 - X-ray diffractogram from 1005 m depth. No clay peak present.

Figure A16 - X-ray diffractogram from 1038 m depth. Clay peak present at about 6° 2θ.
Figure A17 - X-ray diffractogram from 1068 m depth. Clay peak present at about 6° 2Θ.

Figure A18 - X-ray diffractogram from 1221 m depth. Sharp clay peak present at about 6° 2Θ.
Figure A19 - X-ray diffractogram from 1372 m depth. Sharp clay peak present at about 6° 2θ.

Figure A20 - X-ray diffractogram from 1524 m depth. Clay peak present at about 6° 2θ.
Figure A21 - X-ray diffractogram from 1676 m depth. Clay peak present at about 6° 2θ.

Figure A22 - X-ray diffractogram from 1829 m depth. Clay peak present at about 6° 2θ.