Pilgrim Hot Springs Geothermal System Conceptual Model

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Prepared by: The University of Alaska Fairbanks
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SUMMARY: PILGRIM GEOTHERMAL SYSTEM CONCEPTUAL MODEL
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TERMS AND ACRONYMS

ACEP       Alaska Center for Energy and Power
ASTER      Advanced Spaceborne Thermal Emmission and Reflection radiometer
DOE        Department of Energy
NETL       National Energy Technology Lab
EM         Electromagnetic
ETM+       Enhanced Thematic Mapper
FLIR       Forward Looking Infrared Radiometry
IGRF       International Geomagnetic Reference Field
MHG        Maximum horizontal gradient
MT         Magnetotellurics
MW_e       Megawatt Electric
MW_th      Megawatt Thermal
PGS        Pilgrim Geothermal System
PHS        Pilgrim Hot Springs
PSG        Pseudogravity
UAF        University of Alaska Fairbanks
USGS       United States Geological Survey
VNIR       Visible Near Infrared
INTRODUCTION AND HISTORY
The Pilgrim geothermal system (PGS) is located 125 miles south of the Arctic Circle in the southwestern portion of Alaska’s Seward Peninsula, 45 miles north of the city of Nome (Figure 1). The region is the traditional home of the Inupiaq peoples, whose ancestors have called this region their home for thousands of year. The 440 acre property upon which the primary springs are located is listed on the National Register of Historic Places. It was purchased in 2010 by Unaatuq, LLC, a consortium of seven native corporations and regional non-profits who hope to renovate and sustainably develop the site. Also in 2011, the University of Alaska Fairbanks Alaska Center for Energy and Power (ACEP) received funding from the Department of Energy and Alaska Energy Authority to re-examine the PGS and attempt to answer unresolved questions about the amount of available heat, the total surface footprint of the resource, and the precise location of the upflow zone from bedrock.

The PGS is the one of the two most studied and drilled geothermal systems within the Central Alaska Hot Springs Belt (Kolker et al., 2008), which extends from the Seward Peninsula across Central Alaska and into the Yukon Territory. The other is the Chena geothermal system near Fairbanks, which has been producing electrical power from 168 °F (76 °C) fluid since 2006. The PGS was the first subarctic and Alaskan geothermal system to be extensively studied in terms of its geology, geochemistry, geophysics, hydrology, and heat loss in 1979 (Turner and Forbes, 1980). Earlier work, extending as far back as a 1915 chemical analysis (Waring, 1917) and a regional thermal spring chemistry survey (Miller, 1973, Miller et al., 1975) showed that the PGS produced the most saline chloride-rich thermal water of any of the 30+ geothermal systems in the Central Alaska Hot Springs Belt. A series of low-cost, small-scale reconnaissance geophysical surveys were performed at PGS in 1975 (Forbes, et al. 1975, Kirkwood, 1979) as a prelude to the major drilling effort in the summer of 1979.

The first actual drilling occurred in November 1979 with the drilling of the PS-1-H and PS-2-H wells to depths of 160’ and 150’ (48m and 45m) (Kline, et al., 1980). A 1982 drilling program resulted in four more wells: PS-3, PS-4, PS-5, and MI-1 drilled to depths between 260’ and 1001’ (79m and 305m) (Woodward-Clyde, 1983, Kunze and Lofgren, 1983, Economides et al., 1982). Small-scale and short-term interference flow testing was conducted between all six wells and a quantitative conceptual model of the hottest part of the PGS was developed.

After 1982, exploration and development at the PGS languished until a development feasibility study was performed in 2007 following the success of the Chena project (Dilley, 2007). This feasibility assessment was part of a larger Nome Region Energy Assessment funded by the Department of Energy and National Energy Technology Laboratory, which concluded that geothermal energy was a potentially economic option for the region (Sheets et al, 2008).

Based in part from the conclusions of the DOE/NETL study, the Alaska Center for Energy and Power submitted a proposal to the U. S. Dept. of Energy to resume exploration of the PGS to greater depths with a variety of techniques. This proposal was accepted and work began on the ground in 2011 with repair of the leaking wellheads, the installation of 16 Geoprobe temperature probes, and the drilling of two 500 foot (152m) deep temperature-gradient holes. Other work included analysis of satellite-based imagery and use of forward looking infrared radiometry, or FLIR, to map the spatial extent and total heat flow to the surface, and high-resolution airborne magnetic and EM survey. These results have been published as part of UAF’s Phase 1 Final Report (University of Alaska Fairbanks, 2012).
In 2012 an additional 38 Geoprobe holes were installed and three slim holes were drilled to depths of 1000’, 1183’, and 1294’ (305m, 360m, and 394m). The slim holes were intended to drill deep enough to encounter bedrock, which had not been reached during prior drilling efforts, and hopefully demonstrate the presence of temperatures near 293°F (145 °C). Bedrock was reached in the second hole drilled in 2012, at a depth of 1080’ (329m). This was just 80 feet deeper than two previously completed holes at the site, and shallower than had been expected. A maximum temperature of only 194 °F (90 °C) was documented, which was essentially equivalent to temperatures encountered during prior exploration programs, including the shallow subsurface aquifer temperature readings acquired at depths of less than 100’ (30m). In addition, 59 magnetotelluric (MT) soundings were obtained and analyzed.

Preparation of complete conceptual and numerical models for the PGS system are pending the completion of two MS theses, with expected completion dates of August, 2013. This report represents a summary interpretation of the conceptual model for the PGS resource at the request of the Department of Energy, Alaska Energy Authority, and local stakeholders for the purpose of recommending next steps.

**Figure 1.** Index map showing the distribution of recent sediments (tan colored areas) on shaded topographic relief. Red box shows the area outlined in Figure 2a. The location of Pilgrim Hot Springs is shown with star.

**REGIONAL GEOLOGY**

Most of the Seward Peninsula is underlain by Precambrian high grade metamorphic rocks, Cretaceous intrusive rocks, and overlying Paleozoic carbonates (Figure 2) (Amato et al., 2004, Till et al., 2011, Miller et al., 2011). The closest volcanic rocks to the PGS which could provide a direct source of heat for hot springs are the Holocene Lost Jim basaltic lava flows. These flows cover 88 square miles, lie about 30 miles northeast of the PGS and are outside of the Figure 2 map boundary (Till et al., 2011). Because of its distance from the PGS, we consider it unlikely that the Lost Jim lava flows represent a possible magmatic heat source for the PGS. The northern horn of the Seward Peninsula also hosts the world’s largest set of maar craters dated at 21,000 years (Rozell, 2006) but again these are far too distant to suggest a magmatic heat source near the PGS, unless the lava intruded into a fault near PGS and is still cooling.
The PGS lies within an alluvial basin bounded on the south by the Kigluaik Mountains and two prominent hills to the north, Marys Mountain and the Hen and Chickens Mountain (Figure 2a and b). Bedrock exposed in these topographic highs consists of Proterozoic amphibolite to granulite facies metamorphic rocks (Turner et al., 1979; ) The intervening alluvial basin is filled with a fluvial sequence of Quaternary clay, sand, and gravel deposited by the Pilgrim River which meanders through the valley. These sediments are up to 1000 feet (305m) thick and are covered by tundra, thermokarst lakes, and muskeg swamps. These surficial deposits are in turn underlain by discontinuous permafrost. The surface elevation of the basin is only a few tens of feet above sea level.

The dominant structural feature in the PGS region is the east-west trending Kigluaik normal fault located 2.5 - 3 miles south of the hot springs. This fault dips 52° to the north (Amato and Miller, 2004) and separates the Pilgrim Valley graben in the north from the deep crustal rocks exposed at >4000 feet (1219m) in the Kigluaik Mountains to the south (Swanson et al., 1980). The Kigluaik range-front fault is interpreted to be active, based on offsets of glacial moraines (of undefined age) at the base of the mountains by at least 33 feet (10 m) (Swanson et al., 1980), and the presence of sag ponds, offset streams, scarps and other features (Till and others, 2011). The metamorphic rock outcrops of the small Hen and Chickens Mountain are located 2.5 miles north of the hot springs but do not appear to be bounded by faults.

Figure 2a and b. (a) Geologic map of the area surrounding Pilgrim Springs (red star). Map after Till et al. (2010). (b) Topographic map of the area surrounding Pilgrim Valley. The red box in 2a shows the area outlined in 2b.

LOCAL GEOLOGY
There are no local outcrops other than a few feet of floodplain alluvium underlain by clay. Kline et al. (1980) argue that the PGS is located near the western edge of an actively subsiding graben that is being filled by fluvial sediments deposited by the Pilgrim River. The subsurface geology at PGS consists of interbedded unconsolidated to poorly consolidated Quaternary alluvial, fluvial, glaciolacustrine, and brackish lagoon sediments (Figure 3).
Particle sizes range from clay to gravel, with sand, silt, and clay predominating with lesser gravel (Figure 4). The sand is locally indurated with silica cement in the vicinity of most of the deeper wells. The most laterally extensive silt and clay unit is located about 164’ (50 m) above the top of the metamorphic basement.
Metamorphic bedrock is present below a depth of ~1050’ (320 m) and was penetrated by only two wells: PS 12-2 and PS 12-3 (Miller et al., 2013). In these two wells the metamorphic rocks consist of mica schists. The difference in depth to the top of bedrock of ~33’ (10 m) between the two wells (Figure 4) may simply represent an irregular surface or, less likely, a possible fault offset.

The most striking local surface feature at the PGS is a thaw in the permafrost covering an area of ~.58mi² (1.5 km²) that allows anomalous vegetation such as cottonwood trees, alders, grass, and various wildflowers to thrive. Outside of this thermal anomaly, the PGS is surrounded by regional discontinuous permafrost on the order of 328’ (100 m) thick (Osterkamp et al., 1980) (Figure 5).

![Mosaic Image Visual Interpretation](image.png)

Figure 5. Landsurface classes where green and white are unfrozen ground.

The permafrost is an effective aquitard which can form extensive caps and allow development of deeper artesian conditions. The shallow thermal anomaly at the PGS is an opening in the permafrost through which sub-permafrost groundwater can potentially rise to the surface. The artesian head can simply result from the elevation difference between the low elevation Pilgrim Valley and subsurface water flowing downhill from the nearby Kigluaik Mountains.

The local PGS thermal anomaly is not the only hole in the permafrost in the Pilgrim Valley. In 1979 a second permafrost hole was discovered by its vegetation pattern on aerial photographs and confirmed by shallow thermal probing (Turner, D. L., and Forbes, R. B., 1980). This second anomalous area was confirmed with remote sensing methods (University of Alaska Fairbanks, 2012). This suggests that the PGS is not the only thermal anomaly in the area or is part of a larger anomaly.
GEOCHEMISTRY
Geochemistry is the study of the chemical composition of the earth and its rocks and minerals. The science uses the tools of chemistry to explain the mechanisms behind geological systems. The chemistry of the Pilgrim thermal waters from 1915 through 1993 presents a picture of a conventional, low to moderate temperature geothermal fluid (Liss and Motyka, 1994). Analyses of thermal water samples collected in 2010 are in good overall agreement with these earlier results (Benoit, 2013), and one thermal water sample collected from a 150 °F (65 °C) artesian flow from the PS 12-2 well was similar to the PS-3 samples of Liss and Motyka (1994).

The most primitive (saline) thermal water is a near-neutral sodium chloride water with chloride contents of 3350 – 3530 ppm and magnesium contents near 1 ppm (Table 1). Calcium contents of about 500 - 550 ppm are abnormally high. There is no silica or carbonate mineral deposition associated with the thermal springs.

<table>
<thead>
<tr>
<th>pH</th>
<th>T °C</th>
<th>Na</th>
<th>K</th>
<th>Ca</th>
<th>Mg</th>
<th>Li</th>
<th>B</th>
<th>SiO₂</th>
<th>HCO₃</th>
<th>SO₄</th>
<th>Cl</th>
<th>F</th>
</tr>
</thead>
<tbody>
<tr>
<td>7</td>
<td>90</td>
<td>1600</td>
<td>65</td>
<td>520</td>
<td>1</td>
<td>4.0</td>
<td>2</td>
<td>95</td>
<td>25</td>
<td>17</td>
<td>3400</td>
<td>4.7</td>
</tr>
</tbody>
</table>

Static and flowing temperature logs from the artesian wells indicate the various thermal fluids produced from the wells come from a depth range of 60 to 500’ (18m to 152m) and have temperatures from 75 °F to 196 °F (24 °C to 91 °C). The chemistry ranges from dilute groundwater in the deepest fluid entry in well PS-5 to most saline in the shallowest fluid-entry points in wells PS-1 and 2. These samples indicate subsurface mixing in the sediments themselves; there is no indication of mixing within the wellbores.

Since 1915 the chemistry of the most saline thermal waters from the hot springs and the PS-1 and 2 wells have apparently been constant. Chloride contents in the medium salinity PS-3, PS-4, and MI-1 wells have increased between 1982 and 2010. The chemistry of the dilute PS-5 water has shown no discernable change.

The noncondensible gas concentrations are dominated by nitrogen and methane with carbon dioxide making up a maximum of 11% to 13.9% of the total gases. The N₂/Ar ratio indicates the nitrogen and argon are derived from dissolved air. This overall gas composition indicates the fluid has recently been in contact with air and is not derived from a high temperature geothermal system, in spite of gas geothermometry values as high as 419 °F to 446 °F (215 °C to 230 °C). This presents the possibility that the noncondensible gas chemistry might have been obtained at shallow depths shortly prior to the sample collection. The helium-3/ helium-4 ratio of 0.9 permits a small mantle component of helium gases. This is possible given the Holocene basaltic volcanic activity elsewhere on the Seward Peninsula. The amount of noncondensible gases as a percentage of the total mass flow was not given by Liss and Motyka (1994).

The stable isotope data indicate that the oxygen shift is small and similar to that of Pilgrim River waters. This suggests that the thermal water is either derived from the Pilgrim River or sourced from the same area as the Pilgrim River. The Pilgrim River drains water from both the north and south sides of the Kigluaik Mountains. Perhaps more detailed isotopic sampling might be able to discriminate between a more local water source on the north side of the mountains versus the Pilgrim River which largely flows from the south side of the mountains. Liss and Motyka (1994) make no mention of tritium dating of the thermal waters.
The geothermometry analyses done to date yield a wide range of most possibilities, from as low as 223 °F (106 °C) (chalcedony geothermometer) to 293 °F (145 °C) (Na-K-Ca geothermometer) (Liss and Motyk, 1994). The gas geothermometers predict unrealistically high temperatures.

**SUBSURFACE TEMPERATURES**

The most costly and most definitive data gathered at Pilgrim to date has come from five “drilling” campaigns. The first subsurface temperature data came from about 70 half-inch pipes hand driven to depths of 16' to 33' (5 m to 10 m) in June 1979 (Osterkamp et al, 1980). These data were used to construct a contour map at a depth of 15' (4.5 m) (Figure 6). This map was used to determine the locations of the PS-1-H and PS-2-H holes (now known as PS-1 and PS-2 and drilled only 270’ (82m) apart) drilled to depths of 160’ and 150’ (48m and 45m) in November 1979. The PS-1 and PS-2 holes were drilled into a 196 °F (91 °C) thermal aquifer with its top at a depth of 40' to 60' (12m and 18m) and produced short-term artesian flow at estimated rates as high as 200 to 400 gpm.

![Figure 6. Contour map of Pilgrim Hot Springs area showing temperature distribution at 4.5 m depth and locations of 1979 and 1982 wells. Contour interval is 10°C (from Woodward-Clyde, 1983).](image-url)
In 1982 four deeper, but still closely spaced, holes were drilled ranging from 260’ to 1001’ (79m to 305m) in depth (Woodward-Clyde, 1982 (Figure 10). These were also sited on the basis of the 4.5 m temperature contour map (Figure 6). The 1982 drilling was the first attempt to encounter the predicted chemical geothermometer temperatures. These holes, along with the 1979 PS-1 and PS-2 wells, were logged in both static and flowing conditions to document the inflow points (Figure 7). The observed inflow points range in depth from as shallow as 60’ to 500’ (18m to 152m) (Figure 7). Unfortunately, all of these four deeper holes (PS-3, PS-4, PS-5 and MI-1) encountered strong negative temperature gradients beneath a shallow thermal aquifer and this result effectively terminated exploration at the PGS for the next 28 years (Figure 7). Only one geochemical sampling effort was made in 1993 during this hiatus.

In 2011 work at Pilgrim resumed with a remote sensing effort to try to better understand the spatial extent of the thermal anomaly. In addition, existing wells were relogged, and the wellheads were refurbished to control artesianal flows (University of Alaska Fairbanks, 2012). Four of the original holes were relogged in a static or “pseudo static” condition (Figure 7). To further define the source of the upflow and hopefully increase the size of the known thermal anomaly, two 500’ (152m) deep, small diameter temperature gradient holes, S-1 and S-9, were drilled as far north of the earlier well cluster as logistically possible (figure 10).
These holes confirmed the thermal upwelling was closer to the original wells (Figure 8).

Also, in 2011 a tracked Geoprobe was utilized to drive pipes to depths as great as 109’ (33m) at 16 sites to better define the shallow thermal anomaly (Figure 10). In 2012 an additional 35 Geoprobe holes were driven to depths as great as 154’ (47m) by using smaller diameter pipes (Figure 10).

Temperature profiles from the 65 holes and wells at Pilgrim Hot Springs show a wide variety of shapes but multiple patterns are readily visible within the overall mass of lines (Figure 9). These have been broken down into various groups of closely spaced holes with similar profiles to define the locations or boundaries of the shallow thermal aquifers, temperature distributions within the aquifers, the depths to the top or hottest points within the thermal aquifers, and minimum temperatures beneath the shallow thermal aquifers (Benoit, 2013; Figures 10, 11, 12, and 13).

**Figure 8. Temperature profiles of all deep boreholes at PGS.**
Figure 9. All Pilgrim temperature profiles (to depth of 160’, which allows presentation of Geoprobe holes along with deeper holes).
Figure 10. Map showing the wells and geoprobe hole locations. The very shallow thermal aquifer, at depths of 10' to 20'(3m to 6m), has been determined based on subsurface temperatures and outlined on the map.
Figure 11. Map showing the shallow thermal aquifers at the PGS with top depths of 50’ to >123’ (15m to >37m). The aquifer has been arbitrarily subdivided into northern and southern units but hydrologically it most likely behaves as a single unit.
Figure 12. Map showing the depth to the top of the shallow thermal aquifer.
Three slim holes, PS 12-1, PS 12-2, and PS 12-3, were drilled with the intent to reach as deep as 2500’ (762m). These deeper wells were drilled to determine the depth to bedrock, to locate the upflow zone and to prove that the higher temperatures predicted by the geothermometer actually existed. The depth to bedrock was determined in two of the slim holes (PS-12-2 and PS-12-3, Figure 4) at shallower...
depths than predicted by old geophysical surveys. The location of the upflow was narrowed to a small area (Figure 13). Unfortunately, these slim holes encountered a maximum temperature of only 194 °F (90°C), suggesting that either the geothermometry is inaccurate, the higher predicted temperatures are present at much greater depths, or the higher predicted temperatures are laterally displaced from beneath the shallow thermal anomaly.

The mapped subsurface temperatures at Pilgrim have completely outlined a very shallow thermal aquifer at depths of 10' to 20' (3m to 6m) (Figure 10) that overlies a larger and deeper shallow thermal aquifer with its top at depths of 50' to >123' (15m to >37m) (Figure 11). Both thermal aquifers are charged by the same upwelling source as they have virtually identical maximum temperatures of 191 °F (88°C) for the very shallow aquifer and 195 °F (90°C) for the shallow thermal aquifer, and these temperatures are found in the same very restricted area. In addition, the chemistry of the thermal spring is identical to the chemistry from the PS-1 and PS-2 wells which produce directly from the shallow thermal aquifer. The shallow thermal aquifer can be subdivided into two aquifers based on differences in depth and the shapes of the temperature profiles with the northern lobe being deeper and thicker. The southern lobe is characterized by very sharp temperature maximums and more rapid cooling below the thermal aquifer. Hydrologically it is most likely that these two subdivisions are somewhat artificial divisions of one aquifer and it indicates the overall water flow in the region is from south to north.

The best temperature data set to define the location of the upwelling zone are the temperature minimums beneath the shallow thermal aquifer (Figure 13). These suggest that the upwelling is located in the general vicinity of the PS-GEO-15 and PS-GEO-12-5 Geoprobe holes. Unfortunately, the deepest available temperature profile at Pilgrim (PS-12-2, Figure 8) has a slightly negative temperature gradient in the uppermost part of the metamorphic basement rocks below a depth of 1150' (350m). This effectively prohibits extrapolation of temperature gradients to greater depths.

Flowing and static logs from the 1979 and 1982 wells (Figure 7) clearly show the depths of what appear to be single fluid entries ranging in depths of 60' to 500' (18m to 152m) and temperatures of 75°F to 196°F (24°C to 91°C). Woodward-Clyde (1982) determined that the artesian wells have increasing artesian head with increasing depth. The location of these wells is also important as this patterns shows higher artesian pressure further from the river.

Repeat static temperature logs in 2011 show that between 1982 and 2011 the shallow thermal aquifer temperature at PS-1 declined from 196°F to 188°F (91°C to 87 °C) and from 176°F to 158°F (80°C to 70°C) at MI-1 (Figure 7). There has also been a substantial change in the shape of the MI-1 profile between the maximum and minimum temperature depths. In contrast, the 2011 repeat log in well PS-3 shows a slight increase in aquifer temperature between 1982 and 2011. These changes suggest that the PGS is dynamically changing. This is supported by changes in water chemistry that suggest that there is interplay between the geothermal fluid and the surrounding cooler groundwater.

No repeat flowing temperature logs have been run to document changes in fluid-entry temperature or the locations of the fluid-entry intervals. However, temperatures of the water measured while chemical sampling wells at the surface indicate that the flowing temperatures in PS-1 declined from 197°F to 174°F (92°C to 79°C) between 1982 and 2010. This is a much larger change than the temperature decreases observed in the downhole measurements noted in the previous paragraph. Temperatures logged at the hot springs also show a large temperature decline with time. As Liss and Motyka (1994) noted “Flowing wells have captured and diverted much of the ascending thermal water away from...
nearby springs, which have dropped dramatically in temperature and discharge.” These changes may have also extended into the subsurface.

**MAGNETOTELLURIC SURVEYS**

In August, 2012 Fugro Electromagnetics obtained 59 MT soundings at the PGS. MT surveying is able to access a larger area along the east-west axis than the Geoprobe probing machine, and the data may help to clarify the margins of the thermal anomaly. Results of the survey are shown at 82’ (25 m) (Figure 14), 656’ and 984’ (200 m and 300 m) (Figure 15), and 1312’ and 1640’ (400 m and 500 m) (Figure 16). The edges of all of the MT maps are dominated by machine contouring artifacts and should be ignored.

At a depth of 82’, approximately the depth of the shallow thermal aquifer as determined by the well data (Figures 10 & 11), the electrical resistivity pattern is fairly circular. The lowest resistivity values of < 3 Ohm meters correspond closely with the mapped outline and margins of the shallow thermal aquifer (Figure 14).

![Magnetotelluric survey map at 82 feet below the ground surface.](Image)

Beyond the margins of the shallow aquifer the resistivity values increase rapidly to > 100 ohm meters in all directions except toward the north or northeast where the change is much more gradual and much smaller in magnitude. This pattern could be reflecting a clay layer overlying and capping the shallow thermal aquifer and/or a change in the salinity of this aquifer. The western and southeastern areas of very high resistivity are interpreted to be due to the presence of permafrost.
At a depth of 656’ the MT resistivity map shows a much smaller and highly constrained low resistivity area centered between the PS 12-1, 12-2, and 12-3 wells (Figure 15). This depth is well below the shallow thermal aquifer and at the top of the thickest and most widespread clay-rich layer (Figures 3, 4). The resistivities mapped at this depth may be more reflective of the coarser material at slightly shallower depths than to the more clay-rich material at slightly greater depths. The location of this small, low resistivity bull’s-eye is just north of the predicted location of the deeper thermal upwelling at this depth. The southwest and southeast margins of the low resistivity anomaly continue to be very sharp while the north and northeast side are more gradational. This can be interpreted as fresh water flowing from underneath the permafrost in course-grained sediments, moving across the geothermal anomaly to the south while being forced around the low resistive area, cooling and mixing with the geothermal fluids.

At a depth of 984’ (300m) (Figure 15) the >50 Ohm meter resistivities have disappeared from the margins of the survey area and only the southeastern margin of the low resistivity area remains characterized by closely spaced contours. A depth of 984’ is about 65’ (20m) above the known metamorphic bedrock and about 82’ (25m) below the bottom of the thickest and most widespread clay-rich layer known from the various drill holes (Figures 3, 4; Miller et al., 2013). The center of the low resistivity anomaly has migrated a few hundred meters northward.

At depths of 1312’ and 1640’ (400m and 500m) the resistivity maps (Figure 16) are within the metamorphic bedrock. At these depths, the lowest resistivites continue to occur beneath the shallow thermal aquifer, but have moved several hundred meters to the south. There remains a pronounced northeasterly trend to the overall resistivity anomaly.
Figure 15. Magnetotulleric survey map at 656 and 984 feet (200m and 300m) below ground.
Figure 16. Magnetotulleric survey map at a depth of 1312 and 1640 feet below the ground.
AIRBORNE ELECTROMAGNETIC AND MAGNETIC SURVEY

In collaboration with the USGS, an airborne electromagnetic (EM) and magnetic survey was carried out over Pilgrim Hot Springs using the Fugro RESOLVE system during Fall 2011. Frequency-domain airborne EM data was recorded at 6 frequencies (400 Hz, 1800 Hz, 3300 Hz, 8200 Hz, 40,000 Hz, and 140,000 Hz). In-phase and quadrature components of the coplanar coil-pair at a given frequency were used to calculate apparent resistivity maps using a pseudo-layer, half-space model. In this model higher frequencies are sensitive to shallow depths whereas lower frequencies are sensitive to greater depths of investigation.

Interpretation of apparent resistivity and differential resistivity maps (Figure 17) shows low resistivities around Pilgrim Springs. This conductive region extends tens of meters below the surface with the most conductive regions extending to the north and northeast. EM data are sensitive to saline geothermal fluids while higher temperatures in this region likely give rise to more conductive sediments. More moderate resistivities characterize the regions surrounding rivers and streams. These moderate to low resistive areas are likely due to variations in clay content of the sediments.

The high resistivities (> 1000 ohm-m) associated with the mountain ranges reflect the bedrock that comprises these ranges. An equally resistive region exists between the range front of the Kigluaik Mountains and the dense stream channels surrounding Pilgrim Springs. Although subtle topography exists in this region, it is north of the Kigluaik range front. This region of high resistivity is likely indicative of regions of resistive permafrost at depth. This interpretation agrees with remote sensing based permafrost mapping in the area.

An east-west trending, low resistivity (100-200 ohm-m) trend exists on both the apparent resistivity and the differential resistivity maps at all frequencies and depths, respectively. This linear trend follows the base of the Kigluaik Mountains and is preliminarily interpreted to indicate a range-front fault. Fault zones can be conductive when they are comprised of rocks that are fractured and may have hosted fluid flow and subsequent mineralization.

Several conductive anomalies appear in the data at greater depths that are more subdued or missing at shallow depths. These include the conductive regions to the southeast of Pilgrim Springs, the area east of Pilgrim Springs at the eastern edge of the map, the region immediately north of Mary’s Mountain, and a narrow conductive conduit that appears at depth between Pilgrim Springs and the conductive region north of Mary’s Mountain.
Figure 17. Apparent resistivity at 140 KHz (left) and 400 Hz (right) overlayed on topography. Waterways and stream channels are shown in blue and faults are shown in red. Wells are shown with black dots. Apparent resistivity maps show regions of low resistivity (high conductivity) around Pilgrim Springs. At 140 KHz, the areas near rivers and streams are characterized by moderate resistivities (50-300 ohm-m) whereas the Hen and Chickens Mountain, Marys Mountain, and the Kigluaik Mountains are characterized by high resistivities (> 1000 ohm-m). At 400 Hz, the mountainous areas are less resistive. This is likely due to the lack of sensitivity of the data at low frequency as opposed to the mountains getting more conductive at depth. However, more conductive regions southeast of Pilgrim Springs appear in the map that are not seen at higher frequencies.

Airborne magnetic data was acquired using a Cesium vapor magnetometer sampled at a rate of 10 Hz, with a sensitivity of 0.01 nT. Fugro performed basic processing of the magnetic data that included removal of an international geomagnetic reference field (IGRF) and light-line leveling. The USGS performed additional processing applying a variety of derivative and filtering methods to aid in interpretation by helping to delineate structures and to constrain their geometry (Figure 18).

The various filtering methods applied reveal a number of interesting features spanning the shallow to mid-crustal levels. The longest wavelength features are revealed by the pseudogravity (PSG) transformation, that shows a broad high extending southeastwards from Mary’s Mountain to Pilgrim Springs. Similar highs within the survey area are seen further to the southeast over the flanks of parts of the Kigluaik Mountains. A prominent low is observed over the Kigluaik Mountains due south of the springs and northeast-trending elongate low bounds the springs to the southeast. This low is flanked by sharp gradients at its margins and is sub-parallel to the trend of the pseudogravity low described above.
Figure 18. Differential pseudogravity map with magnetic lineations interpreted from maximum horizontal gradients. Pseudogravity highs appear as reds and pinks, gravity lows as blues and purples. Location of Pilgrim Hot Springs is indicated by a red star.

Maximum horizontal gradients (MHG) of the PSG reveal much more detail and can be used to locate sharp contrasts in magnetic properties that occur, for example, at faults or contacts. These contrasts are overlain as magnetic lineations on the PGS map in Figure 18.

Regionally, the MHG can be used to define structural domains. A series of northeast-trending structures is clearly observed southeast of Pilgrim Hot Springs. In contrast, a dominant northwest-trending fabric characterizes the northeastern portion of the survey area between the hot springs and Hen and Chickens Mountain. This trend is similar to that seen far north and south of the study area and may reflect deep basement structures. The area south of the hot springs, however, is dominantly characterized by east-west-trending range-front-parallel structures. These are likely late Cenozoic features associated with north-south extension that formed the basin. A similar east-west trend extends into the area immediately over the springs.

Regionally, the hot springs are characterized by a magnetic high, but this is punctuated by several east-west trending magnetic lows, the most prominent occurring directly over the hot springs. These lows may result from the demagnetization of magnetic material along range-front parallel faults that dissect the basin.
A set of northeast-striking, narrow magnetic highs, located between the springs and Marys Mountain, have a signature consistent with mafic dikes. Furthermore, their trends are similar to the trends of Tertiary dikes that outcrop in the Kigluaik Mountains.

**GRAVITY SURVEYS**

Gravity data were obtained in the vicinity of the PGS in 1979, 1980, (Kienle and Lockhart, 1980, Lockhart, 1981) and by the U. S. Geological Survey in 2010 (Figure 19).

**Figure 19.** Gravity data for PGS region showing high values for the reds and pink color and low values for the blue and purple colors. Gray circles show 1979 and 1980 data and white circles show 2010 data USGS.
The older data lack precise elevation control and most sample points are far enough apart that the data are regional or reconnaissance in nature. The 2010 USGS data were collected along lines spaced .3mi to .6mi (0.5 to 1 km) apart with station spacing of 328’ to 984’ (100m to 300m).

The gravity data show a pronounced low centered about 2.5 mi (4 km) SW of the Pilgrim thermal spring and higher gravity associated with the Kigluaik Mountains and Marys Mountain. Most likely this represents a deepening of the top of the denser metamorphic bedrock. In the vicinity of the Pilgrim wells the gravity data are not adequate to resolve any induration of the Quaternary sedimentary material as is indicated on Figure 4. This would require a high precision closely spaced network of stations and filtered processing to detect. Kienle and Lockhard (1980) concluded on the basis of this gravity pattern that the PGS is located at the northeastern corner of a subsided basement block but this should be viewed as a highly speculative interpretation.

REMOTE SENSING SURVEYS
A variety of airborne and satellite remote sensing data have been used for mapping of surface geothermal features and to identify targets for field activities both inside and outside of the known geothermal area. In addition, surface heat losses have been estimated from the remote sensing data, and have been used to support resource assessment of the PGS.

Multitemporal and multispectral satellite (ASTER, Landsat, WorldView-2), optical, and thermal data were used to map surface geothermal phenomena. These phenomena include; seasonally-dependent thermal anomalies related to hot springs and thermal waters (ASTER nighttime thermal data, Landsat); anomalous areas of snow-free conditions (ASTER) (Figure 20); unseasonal vegetation growth (WorldView-2) related to near-surface ground heating effects; and anomalous ice-free conditions on the Pilgrim River (Figure 21) suspected to be associated with secondary geothermal fluid outflow north-east of the known hot springs area (Haselwimmer, et al., 2011). Recent analysis of multitemporal/multipolarized, very high resolution satellite radar imagery (TerraSAR-X) has outlined the extent of heated ground at the known hot springs site through mapping of unfrozen soils and perturbed snow conditions as well as variable river ice conditions (i.e. thickness) potentially related to secondary geothermal outflow into the river. These geothermal features and land cover mapped from satellite imagery are consistent with the extents of the thermal aquifer identified from subsurface temperature measurements and airborne geophysical data.
Figure 20. Time series of ASTER visible to near-infrared (top) and thermal (bottom) data from Pilgrim Hot springs showing snow-free areas and vegetation growth anomalies associated with geothermally-heated ground.

Figure 21. A larger subset of ASTER winter-time false color composite image with 15m spatial resolution / pixel size. Two prominent snow-free areas are shown with red arrows. Left arrow points to the area near the PHS. Right arrow shows a persistent snow free area (dark) seen in the Pilgrim River, northeast of the PHS.
Two airborne remote sensing surveys were carried out using a fixed wing plane over Pilgrim Hot Springs during Fall 2010 and Spring 2011. These surveys acquired very high spatial resolution optical (~10 cm pixels) and thermal imagery (~1 m pixels). This imagery was used to map land cover (Figure 5) and provided indications of the extent of the thawed area. The imagery was also used to identify surface geothermal features including hot springs/seeps, thermal pools, and areas of heated ground (as manifested as areas of anomalous snow melt) (Figure 22).

Surface temperatures of hot springs and thermal pools mapped from the thermal imagery were used to estimate the hot spring heat flux and outflow rate using a thermal budget model (Haselwimmer, et al., 2013). Using this approach, the convective heat flux and outflow rate of the hot springs has been conservatively estimated at ~ 4.7–6.7 MWth and ~ 976–1400 l/min respectively. These estimates are 2–3 times higher than field measurements. The higher estimate using the airborne thermal data probably reflects the synoptic coverage of the airborne thermal data (i.e., the airborne imagery was able to record the heat flux and flow rate from all hot springs/seeps and thermal waters, not just the limited number measured on the ground).

We have also estimated the convective/conductive heat flux associated with areas of heated ground by using shallow ground temperature measurements (47in (120 cm) depth) to calibrate a map of snow-melt intensity derived from multitemporal aerial and satellite imagery (Haselwimmer et al., 2012) (Figure 23). Using this approach the total ground heat flux has been estimated at 2.39 MWth.

All together, these surface manifestations of geothermal heat loss observable by remote sensing methods encompass ~7-9 MWth of heat energy.
Figure 22. Aerial imagery for the study area at Pilgrim Hot Springs; (A) optical data acquired with a Nikon D700 camera showing locations and temperature of hot springs mapped from the thermal imagery (red dots) that form two broad groups (dashed outlines); (B) thermal imagery acquired with a FLIR A325 camera on 09/10/2010; (C) thermal imagery acquired on 04/28/2011. The thermal data outlines the location of hot springs, seeps, thermal pools, and thermal streams draining the geothermal area. Heated ground is manifested as areas of anomalous snow melt in the winter-time thermal image (C). The sites of field hot spring flow measurements are shown as black dots on (B) (From Haselwimmer et al., 2013)

Figure 23. Map of shallow ground heat flux derived from mapping changing extents of snow-free areas in remote sensing data coupled with limited in-situ ground calibration measurements (From Haselwimmer et al., 2012)
NUMERICAL SIMULATION OF THE GEOTHERMAL SYSTEM

Numerical simulation of the geothermal system has resulted in a suite of models that describe potential thermal liquid distribution and flow patterns. The models simulate groundwater flow with buoyancy effects and generates their own distribution pattern based on measured lithography from the latest wells (Figure 3 and 4) and a constant ground water flow across the domain flowing from the south to the north into the Pilgrim River (Figure 24). These models show that a single source of high temperature water can describe the observed temperatures in the basin. Two scenarios to simulate the system show that the source of geothermal water needs to have an additional pressure head of 33-39 ft (10-12m) and a temperature of 203–248°F (95-120°C). The model with a heat source of 248°F showed the best results in matching the borehole temperatures. The total heat released from the surface of this simulation model is 26 MW\textsubscript{th} (thermal energy).

![Figure 24. Simulated temperature distribution along a cross section of the model domain from south to north.](image)

This model was also used to test production of thermal water in a single production well near the up flow point from the bedrock. Using a production rate of 2000 GPM the well produced 46 MW\textsubscript{th} at a temperature of 180°F (82°C). Alternate scenarios include a second production well to reduce the cold water pressure, which produced 48 MW\textsubscript{th}. A scenario that incorporated reinjection of water produced 50 MW\textsubscript{th}. This last scenario indicates also that the pressure in the sediment layers does not decline when the produced water is injected.

CONCEPTUAL MODEL

The first primitive model of the PGS recognized the shallow thermal aquifer, estimated its size at .4 - .6mi\textsuperscript{2} (1 – 1.5 km\textsuperscript{2}) and its thickness at ~164ft (50m) based on electrical resistivity methods, and calculated an accessible resource base of 500 MW\textsubscript{th} and a beneficial resource base of 30 MW\textsubscript{th} referenced to 0 °C (Turner, et al, 1980).
In 1983 Woodward-Clyde presented the first simple conceptual model of the geothermal system showing a vertical fracture system flowing 1.6 – 2.0 cubic feet/second of 212°F (100°C) water into a shallow aquifer system. No geology or geophysics was superimposed on this model. Economides et al. (1982) expanded on this model on the basis of deeper temperature gradients to predict a 302°F (150°C) reservoir at a depth of around 5000’ (1524m) and noted the thermal water in the shallow thermal aquifer was flowing south toward the original six wells. Woodward-Clyde (1983) also prepared a detailed heat and water balance for the PGS to calculate an accessible resource base for the modeled part of the system at about 24 MWth based on a thermal influx of 19 to 24 MWth and a stored thermal reserve in the shallow aquifer system of 1 x 1015 Joules used over a 20 year period. No attempt is made in this document to either verify or challenge these 1983 numbers and/or calculations, as they appear fairly consistent with our findings to date.

In late 2012 a slightly revised conceptual model was developed by the University of Alaska Fairbanks to calculate the heat loss from the PGS utilizing forward looking infrared radiometer (FLIR) data. Convective flows at the surface of .43 to 0.9 cubic feet/second were obtained. It is important to recognize that these flow rates do not include flows through the shallow thermal aquifer.

An updated and expanded conceptual model based on the 2011 and 2012 drilling, geology, and geophysics consists of a “local” recharge area for the geothermal fluid based on the stable isotope data. This local area may either be the northern side of the Kigluaik Mountains or from the Pilgrim River which also drains much of the southern side of the Kigluaik Mountains. Whether this water flows down somewhere along the Kigluaik range-front fault, or down in some other location makes no difference in exploiting the overall system. The meteoric water must descend deep enough to reach at least the minimum measured temperature of 203°F (95°C). If a background bedrock thermal gradient of 77°F/.6mi (25°C/km) is assumed the water needs to reach a minimum depth of ~2.5mi (4 km) to heat to 212°F (100°C). Higher fluid temperatures require either higher depths and/or a higher background gradient. This means that the thermal fluid must spend nearly all of its subsurface time within metamorphic bedrock and/or in crystalline igneous rocks acquiring its chemistry prior to its rapid rise up toward the surface. No tritium analyses have been obtained to assess the length of time the Pilgrim thermal water has been out of atmospheric circulation.

The most obvious recognized structural feature that might allow waters to descend to and rise several kilometers is the Kigluaik Mountains normal range-front fault. However, just because this feature is most obvious does not mean that it is most likely, especially for the fluid to travel in both directions. For example, in the Basin and Range province, where several extensively drilled and studied geothermal systems are documented along range-front faults, no author has yet shown that the recharge to the geothermal system utilizes the fault as a downward channel. However, in a cold climate recharge from the surface to the deeper groundwater system can only happen when a sufficiently large conduit exists for water to pass by the permafrost layer. Smaller conduits will freeze as the snow melt enters the frozen zone.

The surface trace of the Kigluaik range-front fault is located 2.5 miles south of the Pilgrim thermal spring. If it actually dips 52° or more to the north to great depth then is passes far beneath the thermal springs and is unlikely to be the upward conduit. If it dips at 20 or 30°, then it passes beneath the thermal springs at a relatively shallow depth but thermal water would need to enter the fault miles to the north of the thermal springs to rise up and discharge. This seems quite unlikely. Parallel range-front faults, such as have been postulated at Dixie Valley Nevada (Blackwell et al., 2000), buried beneath the
Pilgrim Valley might be properly located to channel water to the thermal spring but there is not yet any evidence for the presence of these faults, which would have to be optimally oriented and critically stressed for fracture failure. Miller et al. (2013) show north-south trending inferred and buried faults passing a short distance to the west of the thermal springs. Even if these exist, it is unclear as to why they might host permeability or serve as deep upwelling conduits.

The deepest temperature profile at the PGS, PS 12-2 which extends 230’ (70m) into the metamorphic bedrock, has a slightly negative temperature gradient in the bedrock. This is indicative of convective and perhaps lateral flow of 194°F (90°C) thermal water near the top of the bedrock. This eliminates any possibility of credibly extrapolating the temperature gradient to greater depths.

Other, as yet unrecognized or unhypothesized permeability channels may very well exist. The most obvious or simplest of these possibilities might be that the thermal water actually rises from the southwest where the bedrock is deeper based on the gravity data. The thermal water might flow within a gravel unit or permeable layer just above the top of the metamorphic rocks. However thermal waters at that depth would strongly affect permafrost above this conduit. Thermal conduction in any sediment would thaw permafrost if 194 degree Fahrenheit (90°C) water exists within a depth of 6561’ (2000m). Other flow mechanisms may very well exist, but without additional evidence such as a deep seismic survey for the existence of conduit in the bedrock they are at best speculative.

Above the top of the metamorphic bedrock at depths of ~1050m (320m) the thermal fluid flow pattern is much better defined by the existing drilling. Some type of vertical or near-vertical channel allows the thermal fluid to rise to the surface through a sequence of unconsolidated Quaternary fluvial material. If there is any elongation or dip to this channel it has not yet been recognized. It is possible the access limitations for drill hole locations might allow some NW-SE elongation which could then be hypothesized as evidence for a fault. The temperature profile from the PS12-2 well shows identical temperatures of 194°F at depths of 126’ and 1148’ indicating that the geothermal fluid loses no temperature as it rises from the top of the bedrock to the shallow thermal aquifer. Therefore, we speculate that the hottest and most saline fluid samples collected from the thermal springs and the shallow thermal aquifer have probably not been diluted by any shallow groundwater within the unconsolidated Quaternary fluvial material. This indicates pressures are higher within the plumbing hosting the thermal flow than in the surrounding Quaternary material. This allows production of thermal fluid from the deeper groundwater system by pumping the closest well to the upwelling point.

The Quaternary material surrounding the rising geothermal fluid is capable of sustaining artesian flow from the shallow thermal aquifer and deeper intervals. Woodward Clyde (1983) performed a series of two, six hour flow and interference tests on the wells drilled in 1979 and 1982 and concluded “the differences in artesian heads are affected largely by the depth of the well” and that the direction of groundwater flow could not be determined. There was a general trend of the deeper artesian zones having progressively lower transmissivity (gpd/ft) and lower hydraulic conductivity (gpd/ft²). Thus the thermal pattern of a hot shallow thermal aquifer overlying a temperature minimum and then positive temperature gradients beneath the minimum is superimposed on an artesian basin. The high positive and negative temperature gradients show that in the natural state there was basically no vertical movement of fluid, either up or down, other than the rising thermal water in the narrow upwelling conduit. The 2011 static temperature log of the MI-1 well does show the development of some localized vertical flow within or near this wellbore between 1982 and 2011 (Figure 7). Beneath the shallow thermal aquifer there is no evidence in the temperature logs of gravel layers hosting more
convective flow than clay-rich layers. If there is an active and large scale lateral flow of cooler groundwater beneath the shallow thermal aquifer it has remarkably little impact on the static temperature profiles. Identifying any groundwater discharge from this artesian flow system below the shallow thermal aquifer in a swampy environment such as surrounds the PGS has not been attempted and would be a very difficult task.

The most laterally extensive thermal flow system documented at the PGS is the shallow thermal aquifer with its top as close as 50 or 60 feet below the surface. This is the primary geothermal discharge zone of geothermal fluid within the PGS. This fluid ultimately must discharge into the shallow groundwater system and the Pilgrim River. Above the shallow thermal aquifer a thinner and less extensive very shallow thermal aquifer also exists but it is a relatively minor feature. The very shallow thermal aquifer is also charged by the same thermal upwelling as the shallow thermal aquifer and at the same initial temperature. This very shallow thermal aquifer presumably directly supplies fluid to the thermal springs.

A second thermal anomaly has been documented 2.5mi (4 km) northeast of Pilgrim Hot Springs. However, its inaccessibility for drilling and the absence of any thermal fluid on the surface for sampling has prohibited the past gathering of any significant subsurface exploratory data.

**DISCUSSION: FUTURE POSSIBLE ACTIVITIES**

Extensive geological, geochemical, geophysical studies and drilling have been conducted at PGS since the 1970s. The primary drivers of these repeated exploration efforts have been the relatively high geochemically-predicted subsurface temperatures of up to 293°F (145°C), along with the relatively good accessibility. These exploration efforts have resulted in the definition of a modest-sized shallow thermal anomaly that is fed by 194°F (90°C) thermal water rising up from metamorphic bedrock at a depth of 1050’ (320m). This conduit appears to be quite restricted in size. The 2012 drilling failed to prove the presence of temperatures >194°F. Unfortunately the negative temperature gradient in the bottom 200’ of the hottest and deepest of the 2012 holes does not allow predictions to be made as to where or how deep the geochemically-predicted temperatures might be located within the metamorphic bedrock. This raises the question of what are the possible next steps in further assessing and developing the Pilgrim resource. Past exploration efforts have been hampered by limited access to much of the surface and there is no reason to expect future access issues will be reduced.

Geoscientists can always provide some rational for additional geological mapping or improved geophysical studies such as more comprehensive gravity surveying, more aerially extensive electrical resistivity surveying or surface mapping. Surface mapping of fracture patterns in the exposed bedrock combined with a reflection seismic survey would be very useful in locating faults and fractures in the bedrock and identifying deep permeability patterns. This information would be able to guide future drilling programs that attempt to find the source of PGS. These types of efforts are long term in that any new or improved resource models will still need to be tested by future drilling.

**Option1:** One possibility is to continue in the near term with deeper drilling into the bedrock in the continued hope of finding the higher predicted temperatures. This would be a relatively high priced and risky action as there currently is no geological model predicting the locations of higher temperatures at the PGS. If the higher temperatures do not closely underlie the known thermal anomaly then significant new road building is likely required.
Option 2: One other drilling possibility with higher odds of success and minimal road building costs is to drill into the upflow zone to test how much fluid the most potent and permeable part of the known geothermal system is capable of producing. This course of action is anticipated to initially produce ~90 °C fluid. There is a chance that over time hotter fluid will be drawn up from depth during a large flow rate test. There is also the chance that cooler fluid from shallower depths can also be drawn into the well over time.

Option 3: One other possibility is to simply cease fieldwork on the project with or without preparing a more extensive document than this on the project results to date.

Our recommendation is to proceed with Option 2, should funding be identified to support an additional drilling program. The high cost of energy in the region appears to make even a modest project such as a 2MW, binary plant economic to develop. This is substantiated by a recent (May, 2013) agreement between the land owner, Unaatuq LLC, and Potelco, Inc (a subsidiary of Quanta Services), to develop the resource and export power to Nome should the resource be shown capable of sustaining a minimum production level of 2MW_e.

Alternatively, should this drilling program be unable to confirm the 2MW_e threshold established for export power, there are still numerous opportunities for direct use of the resource which would benefit from the proposed Option 2 exploration program. The site has significant historical significance to the residents of the region, including the peoples of Mary’s Igloo whose traditional lands surround the springs and considerable thought has been put into sustainable development of the site to support economic opportunities throughout the region. In fact, the land owner, Unaatuq, LLC, was created as a partnership between seven regional organizations including: Bering Straits Native Corporation, Sitnasuak Native Corporation, Norton Sound Economic Development Corporation, Mary’s Igloo Native Corporation, Teller Native Corporation, White Mountain Native Corporation, and Kawerak, LLC.

Finally, there is a significant graphite prospect less than 10 miles from the springs that has been identified and extensively explored over the past 10 years. Should this prospect be developed, mining activities could serve as another load to support resource development - something that is of significant interest to the mining company leading this effort.

In summary, we strongly believe the Option 2 program would almost certainly lead to development of the Pilgrim Hot Springs resource. The actual results of this program would strongly influence the exact nature and course of this development, particularly related to any future private sector investment.
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