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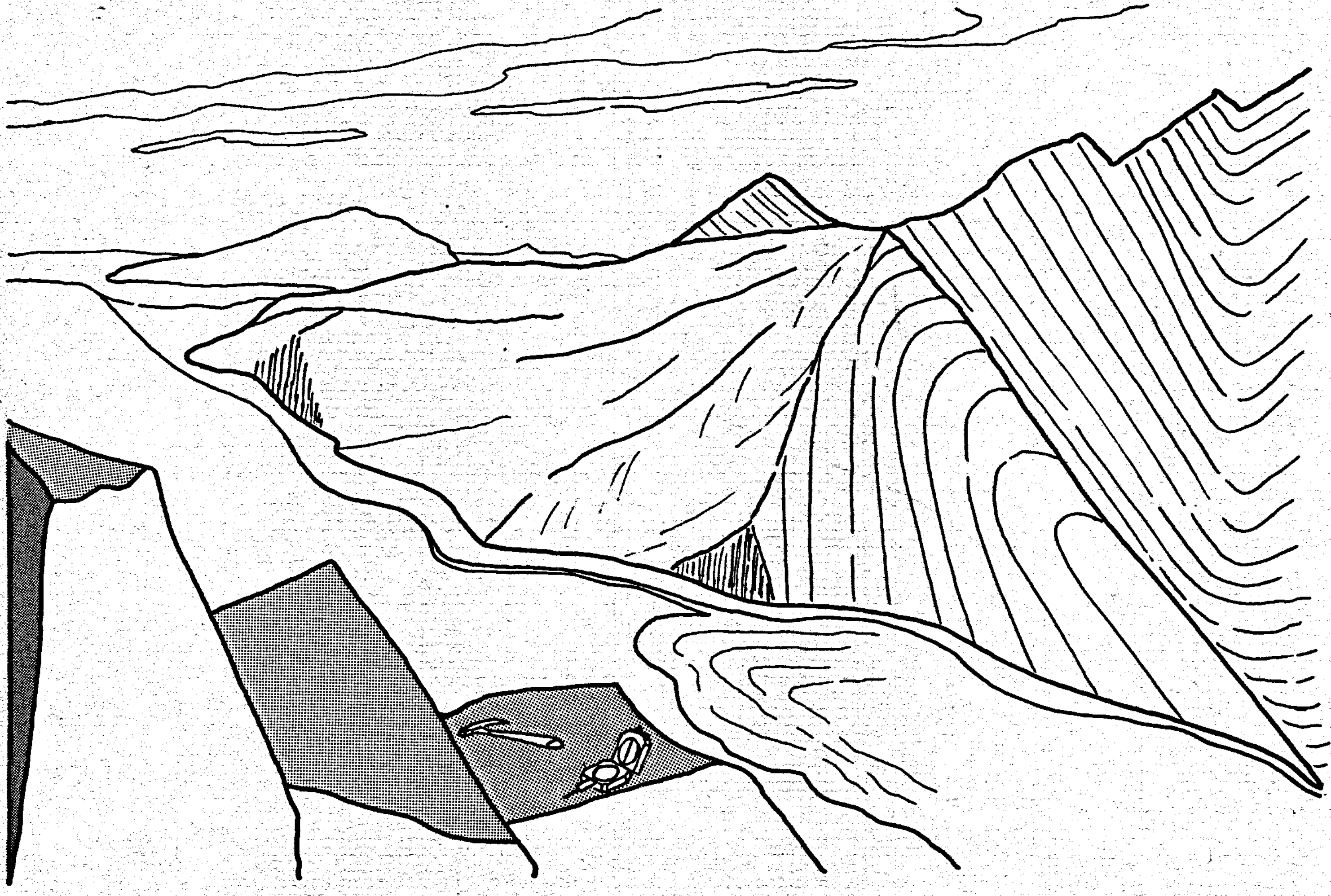

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ASSESSMENT OF THERMAL SPRINGS SITES IN SOUTHERN SOUTHEASTERN ALASKA----PRELIMINARY RESULTS AND EVALUATION

By

Roman J. Motyka, Mary A. Moorman, and
John W. Reeder

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ABSTRACT

Information has been gathered on 13 reported thermal-spring sites, 12 in southern Southeastern Alaska and one in western British Columbia. Five of the reported sites could not be substantiated by DGGS. The eight known thermal spring sites are associated with granitic terrain and, except for Baker Island Hot Springs, occur within or near intensively fractured Cretaceous-age plutons of the Coast Range Batholith.

Thermal-spring surface temperatures range from 21°C (Twin Lakes) to 91.5°C (Bailey Bay). The greatest discharge occurs at Chief Shakes hot springs (450 lpm). Bell Island Hot Springs, which has about a 100-lpm discharge and a 70°C temperature, has had the most development. Two previously unreported thermal-spring sites, Barnes Lake warm springs and Bradfield hot springs, have a low rate of discharge and respective surface temperatures of about 25° and 54°C.

The known thermal springs probably originate from circulation of meteoric waters through deep-seated fracture and fault systems. The chemical constituents of the alkali-sulfate to alkali-chloride thermal waters are probably derived from interaction of the deeply circulating meteoric waters with the granitic wall rocks. Chemical geothermometry suggests subsurface temperatures of 55° to 151°C. If waters are being heated solely by conduction from wall rocks, circulation depths must be about 1.5 to 5 km, assuming geothermal gradients of 30° to 50°C/km.

Variations in temperature, discharge, and chemistry were noted at several thermal springs for which previous records are available. A major decrease in silica and potassium concentrations at Chief Shakes hot springs is suggested by comparing recent analyses of water chemistry to Waring's (1917) original analysis. The rate of discharge at Bell Island Hot Springs may have increased by a factor of two since Waring's visit to the springs.

Subsurface reservoirs associated with thermal springs in southern Southeastern Alaska are of low temperature and are probably limited in extent, compared to geothermal fields now being used elsewhere in the world. Only the Bell Island and Bailey Bay sites now offer any potential for generation of electricity; these sites could also be used for a variety of direct uses such as space heating, wood or lumber processing, and perhaps aquaculture. The other sites have less potential but could be used locally for space heating or agriculture enhancement.

INTRODUCTION

The Alaska Division of Geological and Geophysical Surveys (DGGS) in cooperation with the U.S. Department of Energy, in 1979 embarked on a 3-year program of assessing the geothermal energy resource potential of thermal-spring sites in Alaska. Evaluation of these sites is based on geological and geochemical data, geothermometric models, and geophysical data where available.

More than 100 hot-spring sites have been reported to exist in Alaska. The sites are dispersed throughout the state but significant concentrations and belts of thermal springs occur along the Aleutian Range, in Southeastern Alaska, and across central Alaska. The first comprehensive inventory of the thermal springs of Alaska was published in 1917 by Waring. Since then brief summaries of Alaskan thermal springs have appeared in Waring (1965), Miller (1973), White and Williams (1975), and Muffler (1979). The most recent comprehensive summary of Alaska thermal spring investigations was compiled by Markle (1979).

The scope of the initial assessment of geothermal resources as undertaken by DGGS includes 1) investigating all thermal-spring sites previously reported but not as yet visited or studied by a scientific team; 2) providing up-to-date data and more detailed studies of previously investigated sites; and 3) locating and investigating additional thermal-spring sites not previously reported. The assessment work includes reconnaissance of site geology and hydrology, investigation of thermal-spring characteristics including temperatures and flow rates, and geochemical sampling and analysis of the rocks and thermal waters at the site. This information is used to describe the site, to estimate reservoir temperature, to evaluate the extent of mixing of thermal waters with colder waters, and to determine the geothermal energy potential of the site. These data for all the thermal spring sites of Alaska will be compiled into a geothermal atlas. The information is also being used to select sites with the highest potential for geothermal energy use and to recommend follow-up studies at such sites.

This report presents the preliminary results of investigations on thermal springs in southern Southeastern Alaska, near Petersburg, Wrangell, Ketchikan, and Craig (fig. 1). Three of these sites, which had been reported by Waring (1917) (Vank Island, Unuk, and Saks Cove), could not be substantiated and are presently discounted. South Stikine (Waring, 1917) was reported by a local inhabitant to exist near Mt. Rynda but could not be verified by DGGS. A report of a thermal spring near Virginia Lake was investigated by DGGS; no thermal anomalies were found and the vericity of the report is doubted. Baker Island Hot Springs were verified by the U.S. Forest Service in Craig, but deteriorating weather conditions prevented DGGS field investigation of the site.

Field investigations were performed on seven thermal spring sites during late September and early October, 1979. Four of the sites occur along the Stikine River (Chief Shakes, Twin Lakes, Barnes Lake, Fowler), one along the Bradfield River (Bradfield), and two on and near Bell Island (Bell Island and Bailey Bay). Two of the sites investigated, Barnes Lake and Bradfield, have not been reported in any previous compilation of thermal springs in Alaska. The recent discovery of the Bradfield hot springs in a remote area of the Coast Range Mountains suggests other thermal springs exist but have not yet been detected.

USE OF THE REPORT

The purpose of this open-file report is to provide a ready reference of detailed, current data bearing on the potential uses of geothermal spring areas of southern Southeastern Alaska. The report is organized into several sections so that the reader, depending on his familiarity with geothermal

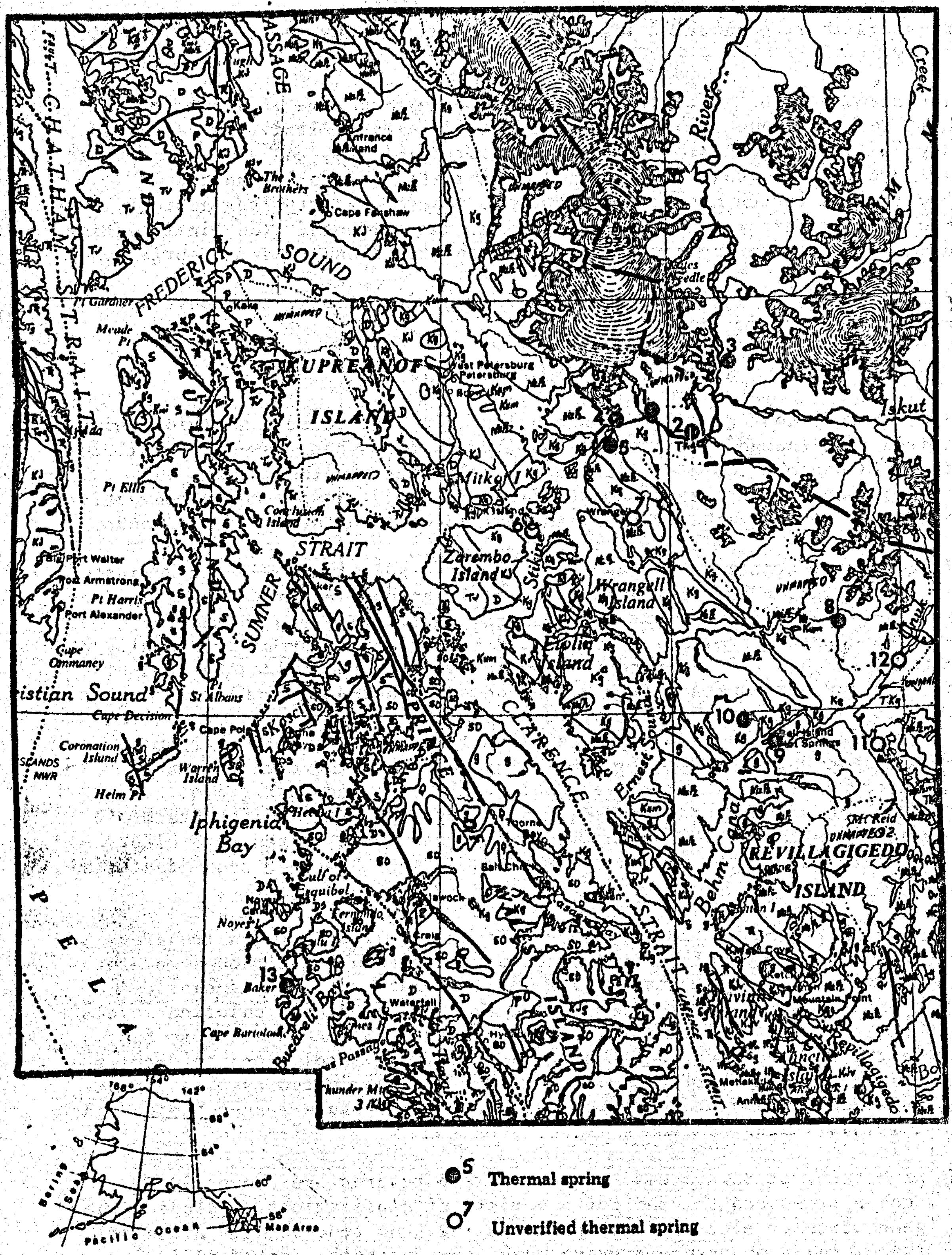

Figure 1. Locations of reported thermal springs, southern Southeastern Alaska. An open circle indicates that existence of spring has not been substantiated and is presently discounted. Location numbers are keyed to text and to the table on page 64. Base map: Beikman, 1975.

resources and their methods of investigation, can use any part of the report separately.

The first section is a brief discussion of hydrothermal convective systems. The next several sections deal with data acquisition and sampling procedures, methods and precision of laboratory analyses, and geothermometric models. Detailed discussions of each site investigated follow next and normally include for each site the location, general description, geology, spring characteristics, reservoir properties, and comments. The final section provides a general summary of southern Southeastern Alaska thermal springs.

For convenience, a list of abbreviations, unit symbols, and conversion factors is presented in appendix A.

HYDROTHERMAL CONVECTION SYSTEMS

Hydrothermal convection systems consist of a heat source, a fluid, and a rock medium with enough vertical permeability to allow hot, low-density fluids to rise while denser, cooler fluids descend elsewhere in the system. In hydrothermal convection systems, most of the heat is transported by convective circulation of fluids rather than by thermal conduction through solid rocks.

Hydrothermal systems are classified into two main types---vapor dominated and hot water---according to the phase controlling heat and mass transfer in the deep thermal reservoir (White and others, 1971). In vapor-dominated systems, the deep reservoirs are controlled predominantly by steam, usually at temperatures of 240°C. Their surface activity is characterized by fumaroles, acid-sulfate springs, and acid-leached ground with hot-spring chloride levels usually below 50 ppm. Such systems are economically attractive for electric power generation because they are generally clean and cheap---they require little more than drilling into the steam reservoir for the development of power. Unfortunately, these systems, occur only under unusual geological conditions. One example of a vapor-dominated system in the United States is the Geysers in northern California, which have been used to generate electricity for several decades. No vapor-dominated systems have yet been identified in Alaska.

Hot-water systems are dominated by circulating liquid, which transfers most of the heat and largely controls subsurface pressures (although moderate amounts of steam may also be present). Their surface activity is characterized by presence of springs discharging thermal water with chloride levels above 50 ppm and with neutral to alkaline pH level. Some hot-water systems boil at depth, and the escaping steam gives rise to fumaroles and acid-sulfate springs, similar to the surficial features of vapor-dominated systems. Hot-water convection systems are divided into three temperature ranges (White and Williams, 1975):

1) High-temperature systems: reservoir temperatures are above 150°C. Such systems can be used for a variety of applications, including the generation of electricity, space heating, and processing. At least three high-temperature systems have thus far been identified in Alaska: Bailey Bay in southern Southeastern Alaska and Geysers Bight

and Hot Springs Cove, both located on Umnak Island (Brook and others, 1979).

2) Intermediate-temperature systems: reservoir temperatures lie between 90°C and 150°C. Such systems can be used for space heating, some industrial processing purposes, and perhaps aquaculture. Future technological advances may eventually make electrical generation from intermediate systems practical. At least 28 intermediate temperature systems have been identified in Alaska; some may prove to be high-temperature hot water systems as more detailed information is obtained.

3) Low-temperature systems: reservoir temperatures lie below 90°C. Such systems can be used for heat and perhaps agriculture purposes (for example, greenhouses), but most likely only in locally favorable circumstances. Over 80 low-temperature systems have so far been identified in Alaska (Turner and others, 1980). Some of these may prove to have higher temperature reservoirs on closer examination.

FIELD TECHNIQUES AND SAMPLING PROCEDURES

A standard pattern of sample and data acquisition was followed at each thermal area visited. At sites with multiple springs, water samples were normally obtained from the thermal spring with the highest temperature and greatest discharge. At some sites more than one thermal spring was sampled. When cold ground-water springs were found in the immediate area, samples were normally taken for silica analysis needed for silica-mixing models.

Water samples were collected, filtered, and treated following procedures described in Presser and Barnes (1974). The samples were always obtained at or as close to the spring source as possible. Filtration was usually through a 0.45-micron filter (although some samples were filtered through a 0.05- or 0.1-micron filter for aluminum and iron determinations). Thermal-water samples obtained for silica analysis were normally diluted in a 1:10 ratio with deionized distilled water to prevent precipitation and polymerization of silica as the sampled cooled. Hydrochloric acid was used for those samples requiring acidification.

Bicarbonate and pH were determined in the field by using methods described by Barnes (1964). Field pH values were determined to the nearest 0.05 pH unit with either a Digi-sense digital pH meter, model 5995-40 or an Orion Research specific ion meter, model 401. The pH meter was normally calibrated against standard buffer solutions before and after each measurement. Readings were normally reproducible to ± 0.05 pH units. Waters were titrated with 0.01639N sulfuric acid for bicarbonate determinations.

Water conductance was measured by using an Electronics Instrument Ltd. Model MC1 MRV portable conductivity measuring set. Conductance is reported in micromhos per centimeter at 25°C with a measurement accuracy of ±3%. The manufacturer's calibration of conductivity cells was verified against standard solutions prior to departure for the field.

Water temperatures and shallow soil temperatures were measured with either an Enviro-labs digital thermistor probe DT-101 or a Brooklyn dial thermometer. All measurements are reported in degrees celsius. The resolution and accuracy of the digital thermometer are 0.05°C and 0.1°C, respectively; for the dial thermometers they are 0.25°C and 0.5°C. Spring-temperature measurements were usually made at the spring orifice or as close to it as possible.

Wherever possible, spring discharge was measured with a Scientific Instruments, Inc. pygmy flowmeter and reported in liters per minute (lpm). Accuracies of these measurements were commonly impaired by shallow channels, low water velocities, and friction in the vane bearings. At some springs the discharge was determined by the time needed to fill a 4-liter bucket. When it was not possible to measure discharge by either of these two methods, an estimate of the discharge was made.

A reconnaissance of site geology was normally made at each thermal area. This entailed examination and sampling of local bedrock, determination of local faults and fracture systems, and examination and sampling of hot-spring deposits (if any).

LABORATORY ANALYSES: METHODS USED AND PRECISION

Whenever possible, laboratory methods of determination used by DGGS were taken from the list of methods of choice for each chemical species as prescribed by Presser and Barnes (1974). Table 1 summarizes the methods used and presents the expected precision of laboratory analyses for 12 constituents commonly found in water as determined by U.S. Geological Survey (USGS) Laboratories. Atomic-absorption analyses were run on a Perkin-Elmer model 603 atomic-absorption spectrophotometer using air-acetylene and nitrous-oxide-acetylene flames under conditions listed in the manufacturer's procedure manual. Colorimetric and turbidimetric absorptions were read on a Beckman DU Vis-UV single-beam spectrophotometer, using a matched set of four 10-mm cells. An Orion model 701 digital pH meter and Orion specific ion and reference electrode were used in the measurements of Cl^-, F^-, and Br^-.

DGGS water analysis procedures are currently undergoing a process of standardization to conform to USGS established methods and quality assurance. The USGS Branch of Water Resources is performing duplicate analyses on selected samples and is serving as a reference lab for the initial stages of the DGGS water analysis program. The final results will be presented in an DGGS open-file report.

Most of the precisions given in table 1 for the various methods of analysis are based on data obtained through multilaboratory analysis of test samples prepared by U.S. Geological Survey laboratories (Skougstad and others, 1979). Where possible, the precision is expressed in terms of a regression equation over a stated range. The precision, expressed in terms of the relative deviation (the ratio of the standard deviation to the mean times 100 percent) for various laboratory determinations is given in Appendix B. (Appendix C lists inductively coupled argon-plasma determinations.)

Table 1. Methods and expected precision: DGGS water-chemistry analyses

Chemical species	Reference	Description of method	Precision[a]
SiO_2	A p. 495	Colorimetric-formation and reduction of cilicomolybdate	S_T = 0.031 x + 0.60 (mg/l) Range: 2.0 to 40 mg/l
Al	A p. 39	AA; chelation with 8-hydroxyquinoline and extraction with MIBK	S_T = 0.073 x + 14.97 (ug/l) Range: 75 to 800 ug/l
Fe	A p. 153	AA; direct aspiration	S_T = 0.056 x + 23.90 (ug/l) Range: 80 to 1,000 ug/l
Ca	A p. 107	AA; $LaCl_3$ added	S_T 0.057 x + 0.343 (mg/l) Range: 0.1 to 60 mg/l
Mg	A p. 177	AA; $LaCl_3$ added	S_T 0.043 x + 0.134 (mg/l) Range: 0.5 to 36 mg/l
Na	A p. 255	AA; direct aspiration	S_T = 0.039 x + 0.448 (mg/l) Range: 3 to 80 mg/l
K	A p. 229	AA; direct aspiration	Appendix B
Li	A p. 171	AA; direct aspiration	S_T = 0.048 x + 4.84 (ug/l) Range: 30 to 500 ug/l
HCO_3	A p. 517, B, C	Electrometric titration in the field of alkalinity using 0.01639N H_2SO_4 and calculation from alkalinity and pH data	± 2.0 ppm (C)
SO_4	D p. 496	Turbidimetric, barium sulfate complex formed using $BaCl_2$	Std dev 9.1% (D)
Cl	A p. 589	Mohr titration using $AgNo_3$	S_T = 0.034 x + 0.33 (mg/l) Range: 1.0 to 210 mg/l
	A p. 331	Colorimetric - ferric thiocyamate	S_T = 0.056 x - 0.002 (mg/l) Range: 0.1 to 10 mg/l
F	A p. 523, E	Direct reading using specific ion electrode, Orion TISABII added. (CDTA adjustor and buffer)	Appendix B

Table 1. Methods and expected precision: DGGS water-chemistry analyses (cont.)

Chemical species	Reference	Description of method	Precision[a]
Br	D p. 291	Colorimetric oxidation of bromide and bromination of phenol red at pH 5 to 5.4	
B	A p. 311	Colorimetric-complexing with carmine	$S_T = 0.463\ x + 20.2$ (ug/l) Range: 30 to 550 ug/l
pH	C	See Barnes, 1964	± 0.05 pH units

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[a] X = Concentration in mg or ug/l. Quoted precisions are from Skougstad and others, eds. (1979) unless otherwise noted.

References

A. Skougstad and others, eds, 1979.
B. Presser and Barnes, 1974.
C. Barnes, 1964.
D. Rand and others, eds., 1975.
E. Instruction manuals for Orion Ionalyzer Specific Ion Electrodes, 1977, by Orion Research Incorporated.

GEOTHERMOMETRY

Chemical geothermometry has become an important tool for estimating reservoir temperatures of hydrothermal systems and has proved very useful in determining the geothermal resource potential of a specific region. Therefore, much of the DGGS assessment work is aimed at providing accurate geothermometry for the individual geothermal sites investigated. Summaries of the various geothermometers that have been applied to geothermal systems can be found in Fournier (1977) and in Ellis and Mahon (1977). The geothermometers are based on temperature-dependent water-rock reactions that control the chemical and isotopic compositions of the thermal waters. The most reliable and most frequently used quantitative geothermometers are related to the silica content and to the sodium, potassium, and calcium content of thermal waters. Recent application of sulfate-water, oxygen- isotope geothermometry to geothermal systems indicates this method may become a third important geothermometer, particularly for higher temperature systems (McKenzie and Truesdell, 1977).

Assumptions inherent in using compositions of thermal waters to estimate subsurface temperatures have been summarized by Fournier (1977):

1. Temperature-dependent reactions involving rock and water fix the amounts of dissolved "indicator" constituents in the water.
2. There is an adequate supply of all the reactants.
3. There is equilibrium in the reservoir or aquifer with respect to the specific indicator reaction.
4. No reequilibration of the indicator constituents occurs after the water leaves the reservoir.
5. No mixing with different waters occurs during movement to the surface.

The attainment of equilibrium in the reservoir depends on a number of factors such as the kinetics of the particular reaction, the temperature of the reservoir, the reactivity of the wallrock, the concentrations of the indicator elements in the water, and the residence time of the water in the reservoir at the particular temperature. Thus, in some situations, equilibrium in the reservoir may be attained for some reactions and not for others.

Whether a water reequilibrates after leaving a reservoir during flow to the surface depends on similar factors: the rate of flow, the path of ascent, the type and reactivity of wallrock traversed, the initial temperature of the reservoir, and the kinetics of the various reactions that may occur. Different reactions may occur in an ascending water at different rates. Therefore, the apparent last temperature of equilibration may be different for different chemical geothermometers.

For this report the silica and Na-K-Ca geothermometers have been used exclusively for estimating subsurface temperatures.

Silica Geothermometry

The silica geothermometer is based on the experimentally derived relationship between silica solubility, temperature, and pressure (Fournier and Rowe, 1966; Fournier, 1973). Dissolved silica found in thermal waters may be supplied by temperature-dependent reactions between the thermal water and either quartz, chalcedony, amorphous silica, or cristobalite. Figure 2, Curve A shows the solubility of quartz as a function of temperature in equilibrium with saturated steam. Curve B shows the amount of silica that would be left in the residual liquid after maximum loss of steam on adiabatic cooling to 100°C and 1 bar pressure. Similar curves can be constructed for the other mineralogic phases of silica.

Fournier (1973) found that above 150°C, quartz controls the silica equilibrium and that the quartz geothermometer generally works best in the range 150-225°C. When initial temperatures are above 225°C, silica is likely to precipitate on ascent to the surface owing to the relatively fast rates of reaction at higher temperatures and the attainment of supersaturation with respect to amorphous silica as the solution cools. For reservoirs below 150°C , Fournier (1973) found that chalcedony and sometimes cristobalite or amorphous silica, rather than quartz, may control the dissolved silica content. However, in granitic rocks Fournier reported that quartz may be the controlling mineral down to temperatures as low as 90°C (Brook and others, 1979). Thus, ambiguities can arise from the application of silica geothermometry in the range 90°-150°C.

Figure 2 or similar curves for the other phases of silica can be used to estimate reservoir tempertures. If the water sample is likely to have cooled mainly adiabatically (by boiling), Curve B, which corrects for the maximum possible steam loss, is used. If the sample cooled mainly by conduction, Curve A is used.

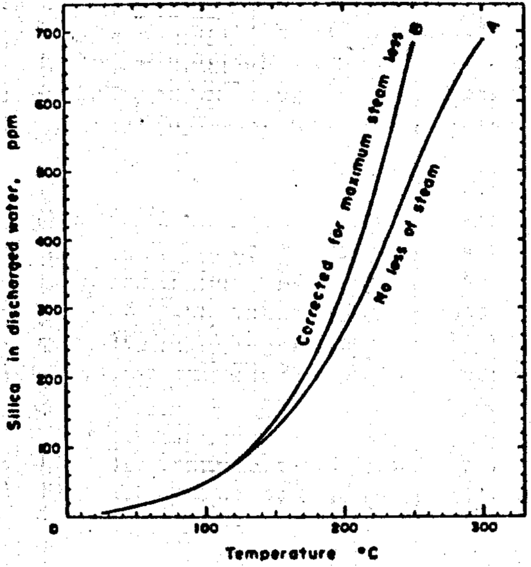

Figure 2. Solubility of quartz as a function of temperature. Curve A shows the solubility in liquid water in equilibrium with saturated steam. Curve B shows the amount of silica that would be left in the residual liquid after maximum loss of steam on adiabatic cooling to 100°C and 1 bar pressure. From Fournier and Rowe (1966).

The following empirical relationships can alternatively be used to estimate reservoir temperatures up to 250°C.

Amorphous silica $$T, {}^\circ C = \frac{731}{4.52 - \log C} - 273.15$$

Beta-cristobalite $$T, {}^\circ C = \frac{781}{4.51 - \log C} - 273.15$$

Alpha-cristobalite $$T, {}^\circ C = \frac{1000}{4.78 - \log C} - 273.15$$

Chalcedony $$T, {}^\circ C = \frac{1032}{4.69 - \log C} - 273.15$$

Quartz conductive $$T, {}^\circ C = \frac{1039}{5.19 - \log C} - 273.15$$

Quartz adiabatic (after steam loss) $$T, {}^\circ C = \frac{1522}{5.75 - \log C} - 273.15$$

"C" is the concentration of SiO_2 in milligrams per kilogram of water.

Brook and others (1979) reviewed the problems associated with interpreting high SiO_2 levels in highly alkaline spring waters (pH >8). In alkaline waters, hydroxide reacts with silicic acid to reduce the proportion of silicic acid to total dissolved silica:

Silicic acid		Hydroxide		Dissociated silicic acid		Water
H_4SiO_4	+	OH^-	=	H_3SiO_4	+	H_2O

The total concentration of dissolved silica measured in the laboratory, however, is H_4SiO_4 plus $H_3SiO^-_4$ and must therefore be reduced by the concentration of $H_3SiO^-_4$ to obtain an accurate estimate of the subsurface reservoir temperature.

To correct for the dissociation of silica in alkaline waters, a simple correction suggested by Brook and others (1979) has been adopted in this report. The concentration of silicic acid is determined at the spring temperature and pH. This concentration, recast as SiO_2, is then used in the appropriate geothermometer. There is considerable disagreement about the value of the first dissociation constant of silicic acid at temperatures above 30°C (Seward, 1974). The values used in this report were taken from Ryzhenko (1967). Corrections for the dissociation of silicic acid are not made unless the correction is 10°C or more.

Silica Mixing Models

One of the assumptions inherent in using geothermometry is that thermal-spring waters are undiluted. However, the waters issuing from many (if not most) thermal springs probably consist of mixtures of deep hot water and shallow cold water. Fournier and Truesdell (1974) described two mixing models that may be applied to springs with large rates of flow and temperatures below boiling. These models are based on the relationship between the enthalpy and silica content of the ascending thermal water, the cold ground water, and the resultant mixed thermal-spring water. In the first model the enthalpy of the hot water plus steam that mixes with and heats the cold water is the same as the initial enthalpy of the deep hot water; the deep hot water may boil below mixing, but all the steam condenses in the cold water. In the second model the enthalpy of the hot water in the zone of mixing is less than the enthalpy of the hot water at depth owing to escape of steam during ascent.

Barrett and Pearl (1978) summarized the additional assumptions, given below, which are implicit in the use of these mixing models:

1. Initial silica content is controlled by the temperature-dependent reactions between the deep thermal water and the various silica phases.
2. Additional silica is neither dissolved nor deposited after mixing.
3. The temperature and silica content of cold springs are similar to of the ground water that mixes with the ascending hot water.

Truesdell and Fournier (1977) devised a simple procedure for applying these models, in which a plot of dissolved silica vs enthalpy is used (fig. 3). For the situation in which no steam is lost before mixing, the silica and heat contents (enthalpies) of the cold and warm spring waters are plotted as two points, A and B. A straight line is drawn through these points to intersect the quartz solubility curve (note that below 100°C the temperature in degrees Celsius is numerically equivalent to cal/g). Point C then gives the original silica content and enthalpy of the deep hot water. The original temperature of the hot-water component is then obtained from steam tables (Keenan and others, 1969). The fraction of hot water in the warm spring is obtained by dividing the distance AB by AC.

For the situation in which the maximum amount of steam is lost from the hot water before mixing, the silica and heat contents of the cold and warm spring waters are plotted as two points, A and D, in figure 3. A straight line is drawn through those points and extended to the enthalpy of the residual liquid water at the assumed temperature of separation and escape of steam, taken here to be 100°C. In this case the residual liquid water before mixing will have an enthalpy of 100 cal/g, point E of figure 3. The original enthalpy of the hot-water component is obtained by moving horizontally across the diagram from point E to the maximum steam loss curve, point F. The original silica content of the hot-water component is given by point G. The fraction of hot water (after steam loss) in the warm spring is obtained by dividing the distance AD by AE. If steam is assumed to escape from water at a temperature above 100°C, the original enthalpy of the hot water will lie at a value along a horizontal line, between the maximum steam loss curve and the quartz solubility curve (no steam loss).

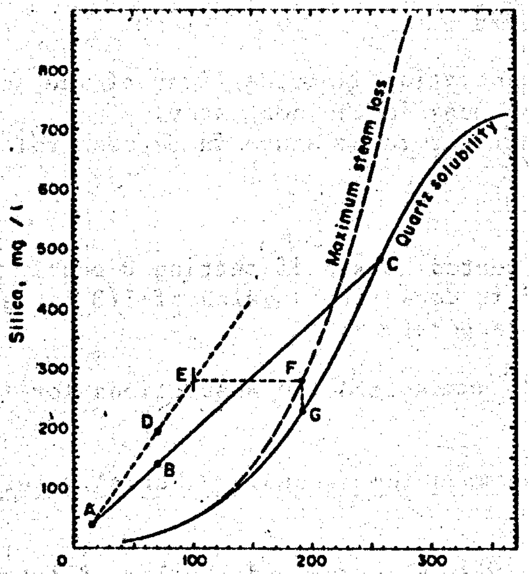

Figure 3. Silica mixing model examples. The graph, which gives dissolved silica vs. enthalpy, is used for determining the temperature of warm spring water derived by mixing a hot-water component with cold water. From Truesdell and Fournier (1977).

Brook and others (1979) pointed out that the problem with any unexplored hydrothermal system is in proving that the water issuing at the surface is indeed mixed. One proof would be a linear trend between measured spring tempertures and chloride concentration (for example, Fournier, 1979). Normal ground-water usually has low chloride concentrations, whereas thermal waters from high-temperature systems usually contain about several hundred milligrams per liter of chloride. A linear trend between the isotopic composition of the water (deuterium or oxygen-18) and dissolved chloride is another proof of mixing (Mariner and Willey, 1976). Unfortunately, very few areas have sufficient springs of different chemical and isotopic composition to prove mixing by such rigorous criteria.

Sodium-Potassium-Calcium Geothermometer

The Na-K-Ca geothermometer is based on an empirical relationship between the molar concentrations of sodium, potassium, and calcium ions and water temperature. Fournier and Truesdell (1973) presented a detailed account of the geochemical theory involved in the development of the Na-K-Ca geothermometer. Temperature is related to water composition, in molality, by the following empirically derived equation:

$$T,\ ^{\circ}C = \frac{1647}{\log \frac{Na}{K} + B \log \frac{\sqrt{Ca}}{Na} + 2.24} - 273.15$$

where:

Na, K, Ca = ionic concentration in moles/liter of the sodium, potassium, and calcium ions in the hot water.
T, °C = estimated subsurface temperature in degrees celsius
B = 1/3 for $T > 100°C$
B = 4/3 for $T < 100°C$

The equation is first tested to see if setting B equal to 4/3 yields a temperature below 100°C; if it does not, a value of 1/3 is used for B to estimate the equilibration temperature.

Barret and Pearl (1978) summarized the assumptions for the use of the Na-K-Ca geothermometer.

1) No mixing occurs between the ascending thermal water and shallow ground water

Mixing between the hot thermal water and shallow, dilute, ground water will have little effect on the sodium-potassium ratio but may affect the calculated calcium-sodium ratio because of the square root of calcium term. If the original calcium content of the undiluted thermal water is low, mixing will have little effect on the geothermometer results. If the calcium content of the undiluted thermal water is high (greater than 50 to 100 mg/l), mixing with dilute ground water will cause the subsurface temperature estimate to be too low.

2) Sodium, potassium, and calcium concentrations in the thermal water are controlled by temperature-dependent equilibra with albite, potassium feldspar, and calcium-bearing carbonate minerals.

The sodium, potassium, and calcium ratios are strongly affected by the bedrock mineral suite. Depending on which mineral suite controls the water composition, a wide range in temperature estimates is possible. At similar water temperatures, the sodium-potassium-calcium ratios are widely variable in solutions equilibrated with potassium feldspar and albite, muscovite and albite, alkali-bearing carbonates, or other mineral suites.

For example, waters equilibrated with mineral suites containing potassium feldspar but no albite (sodium-deficient mineral suites) will provide excessive subsurface temperature estimates. On the other hand, waters equilibrated with mineral suites containing albite but no potassium feldspar (potassium-deficient mineral suites) will yield temperature estimates that are too low. Waters in equilibrium with alkali-bearing carbonates (evaporite sequences) generally yield excessive temperature estimates. However, equilibration with zeolites may yield minimum temperature estimates.

3) Little or no reequilibration occurs during ascent

Changes in the sodium-potassium-calcium ratios in thermal waters may be great or negligible, depending on the rate of ascent and the relative reactivity of the rocks and minerals along the flow path. Low-calcium thermal waters generally yield low subsurface temperature estimates because of continued reactions between water and wallrock during ascent (increased aqueous calcium ion concentration). High-calcium-content waters, however, may yield excessive geothermometer temperature estimates because of calcium-carbonate deposition (decreased aqueous calcium ion concentration) during ascent.

Fournier and Potter (1978) reported that the high concentration of magnesium or large magnesium-to-calcium ratios in some waters was interfering with the Na-K-Ca geothermometers. A modification to the Na-K-Ca geothermometer used in this assessment was recently devised by Fournier and Potter to correct for these adverse effects of magnesium. Graphs or empirical formulas are used to determine temperature corrections when waters have Na-K-Ca calculated temperatures above 70°C and values of R less than 50, where $R = Mg/(Mg + Ca + K) \times 100$ in molar equivalents. Waters with values of R greater than 50 are thought to come from relatively cool aquifers about equal to the measured spring temperature, regardless of much higher calculated Na-K-Ca temperatures.

Accuracy of Geothermometry

The accuracy of the geothermometers depends on the accuracy of the laboratory analyses for the various constituents used in the geothermometers. The following example illustrates the possible variations in subsurface temperature estimates resulting from normal laboratory analytical error. The 95 percent confidence limits (two times the standard deviation) can be determined from table 1 and Appendix B. Using data from Bell Island hot springs, the variations in constituents used in the silica and cation geothermometers are:

Temperature:	74°C
SiO_2	108 ± 8 mg/l
Na	176 ± 14 mg/l
K	7.2 ± 1.6 mg/l
Ca	8.3 ± 1.6 mg/l

Applying these ranges of values to the silica and the Na-K-Ca geothermometers gives the following results:

Silica geothermometer temperatures (°C)	Low	Reported	High
Concentration (mg/l)	100	108	116
Adiabadic	133	136	140
Conductive	137	142	146
Chalcedony	110	115	120
Cristobalite	87	91	95
Opal	17	21	25

Cation: Geothermometer temperatures (°C)	Low	Reported	High
Na-K-Ca (4/3)	129	144	158
Na-K-Ca (4/3)	103	117	130

The low and high temperature ranges given above for the cation geothermometers are based on using the respective minimum and maximum values in the 95 percent confidence range for Na and Ca, and the respective maximum and minimum values in the 95 percent confidence range for K. These choices give the widest possible spread in temperature estimates.

DESCRIPTION OF INDIVIDUAL THERMAL SPRING SITES

The location of thermal spring areas in southern Southeastern Alaska are shown in figure 1. The numbers are keyed to the discussion of individual sites and to the table on page 64. The existence of several thermal spring sites that had been previously reported could not be substantiated and are discounted. The thermal areas are described in numerical order with thermal areas in the same general region discussed together.

The following form is used in discussing each thermal spring site investigated:

1. Spring name

2. Location: includes latitude and longitude to the nearest tenth of a minute; the topographic quadrangle map and township, range, section, and one-quarter section in which the site is located.

3. General description: includes distance and directions to the area from the nearest town or other prominent geographic feature; type of surface activity; location and number of springs; area of surface expression; local drainage; topography and terrain; vegetation; type of development if any; land status.

4. Geology: includes discussion of thermal spring host rock; local rock types and contacts; and local and regional faults, fractures, and photo interpretation of lineaments. A geologic map of the area is normally provided. These maps are usually adapted from previously published geologic maps of the area, modified by any findings obtained during DGGS reconnaissance of the area.

5. Spring characteristics: includes temperature measurements; associated deposits; gases; unusual characteristics; water chemistry; water type; other physical properties; and comparisons with earlier studies.

6. Reservoir properties: includes discussion of geothermometry; mixing models where applicable; and estimates of reservoir temperature, volume, and thermal energy content.

   The techniques described by the USGS for estimating reservoir properties have been adopted in this report (Brook and others, 1979; Mariner and others, 1978; Nathenson, 1978). A judgment is made as to the minimum, maximum, and most likely subsurface temperature based on geology and geothermometry, and on geophysics and downhole measurements where available. (Subsurface temperatures derived from silica mixing models are not used in estimating reservoir tempertures in this report unless corraborative evidence for mixing exists, for example, chloride-enthalpy analyses or water oxygen isotope analyses.) A mean reservoir temperature and standard deviation are then calculated following methods described by Nathenson (1978).

   Estimates of reservoir volume are made from available geologic, geophysical, and bore-hole data. Few thermal sites in Alaska, however, have had even cursory geophysical exploration and only one site, Pilgrim Springs (Turner and others, 1980), has undergone exploratory drilling. Except for some shallow seismic studies at Bell Island hot springs (Pyle, 1978), there has been no geophysical exploration of thermal sites in Southeastern Alaska. To obtain a speculative estimate of the thermal-energy content of unexplored hydrothermal systems, a standard mean reservoir volume based on general geological and resources-exploitation constraints was adopted for such sites (Mariner and others, 1978). The standard reservoir is taken to have a circular area of 2 km^2 and a thickness of 1.7 km centered at a depth of about 1.5 km.

The accessible resource base for a hydrothermal system, defined by Brooks and others (1979) as the amount of geothermal energy within a specified volume of rock and at a specified temperature referenced to 15°C, has been adopted for this report. The depth limit of economic production drilling for geothermal resources is currently about 3 km; any system or part of a system below 3 km is therefore not considered.

Thermal energy (Q) of each system is then calculated from the equation:

$$Q = pc\ V\ (T - T_o)$$

where:

pc = volumetric specific heat of rock plus water ($pc = 2.7\ J/cm^3 - {}^\circ C$)
V = mean volume (converted to cm^3)
T = mean temperature (°C)
T_o = reference temperature (15°C)

The volumetric specific heat (pc) is calculated assuming the rock specific heat to be 2.5 $J/cm^3 - {}^\circ C$ and the reservoir porosity to be 15 percent. The method of calculating the standard deviation is given in Nathenson (1978).

7. Comments: this section discusses speculations on the cause of the thermal springs; potential usefulness of the geothermal resource; unusual characteristics; and other miscellaneous items.

8. Tables:
   1) Physical properties and chemical composition of thermal waters, including: sample source, collection date, major element chemical composition, discharge rate of spring samples, temperature, etc.
   2) Geothermometry, including: all cation, all silica, mixing models where applicable.

9. Figures:
   1) Geologic map generalized from available literature.
   2) Photos of site.

Appendix C provides a table of minor and trace-element constituents of the thermal waters as determined by inductively coupled argon-plasma analyses.

1. CHIEF SHAKES HOT SPRINGS

Location

Latitude 56° 43.0' N.; longitude 132° 00.3' W.; Petersburg C-1, 1:63,360 Quadrangle[1] (1953) T. 59 S., R. 85 E.; sec. 34, SE 1/4 of NE 1/4 of Copper River Meridian.

General Description

Chief Shakes hot springs are located north of the Stikine River, about 35 km NNE of the village of Wrangell, Alaska (fig. 1). The springs are accessible by boat via the Stikine River to the Ketili River (a side slough of the Stikine), thence 3 km up a west-flowing tributary located about 4 km from the mouth of the Ketili (fig. 4). During high water, a shallow draft boat can be floated to within 200 m of the springs. The springs are within the Tongass National Forest and are managed by the U.S. Forest Service.

The Stikine Valley is a spectacular glaciated valley flanked by numerous hanging valleys and steep granitic walls that are nearly vertical in places. Glaciers still cover much of the higher elevations. Shakes Glacier, NNW of the springs, descends nearly to river level.

The hot springs lie along the base of a steep glaciated granitic cliff that forms the north side of the Stikine Valley. The prinicipal spring (spring 1) emerges from beneath boulders at the cliff base, about 15 m above average slough-water level. Part of the waters from this spring are channeled through a pipe to a large wooden tub enclosed in an A-frame structure, about 100 m downslope from the spring source. The hot tub and enclosure were constructed by residents of Wrangell, who commonly use the hot springs for recreational purposes.

At least seven additional hot springs are located from 50 to 200 m downslope from the A-frame. These springs emanate from small fissures in the granitic cliff near the base of the cliff and from the alluvium floor. Most of these are minor springs with very low discharge. One of the springs, however (spring 2), has a discharge and temperature comparable to the principal spring. The spring at the lowest elevation in the series lies about 5 m above mean slough water level.

The hot springs waters drain through a gently sloping meadow about 200 m by 300 m, located south of the cliff, and thence into the slough. The meadow is surrounded by willow and tall stands of sitka spruce and hemlock.

Geology

Bedrock in the vicinity of Chief Shakes hot springs has been reported as granodiorite and quartz diorite of Tertiary-Cretaceous age (fig. 4) (Beikman, 1975; Buddington and Chapin, 1929). The bedrock is though to be part of the Coast Range Batholithic complex. The contact between the layered gneises and

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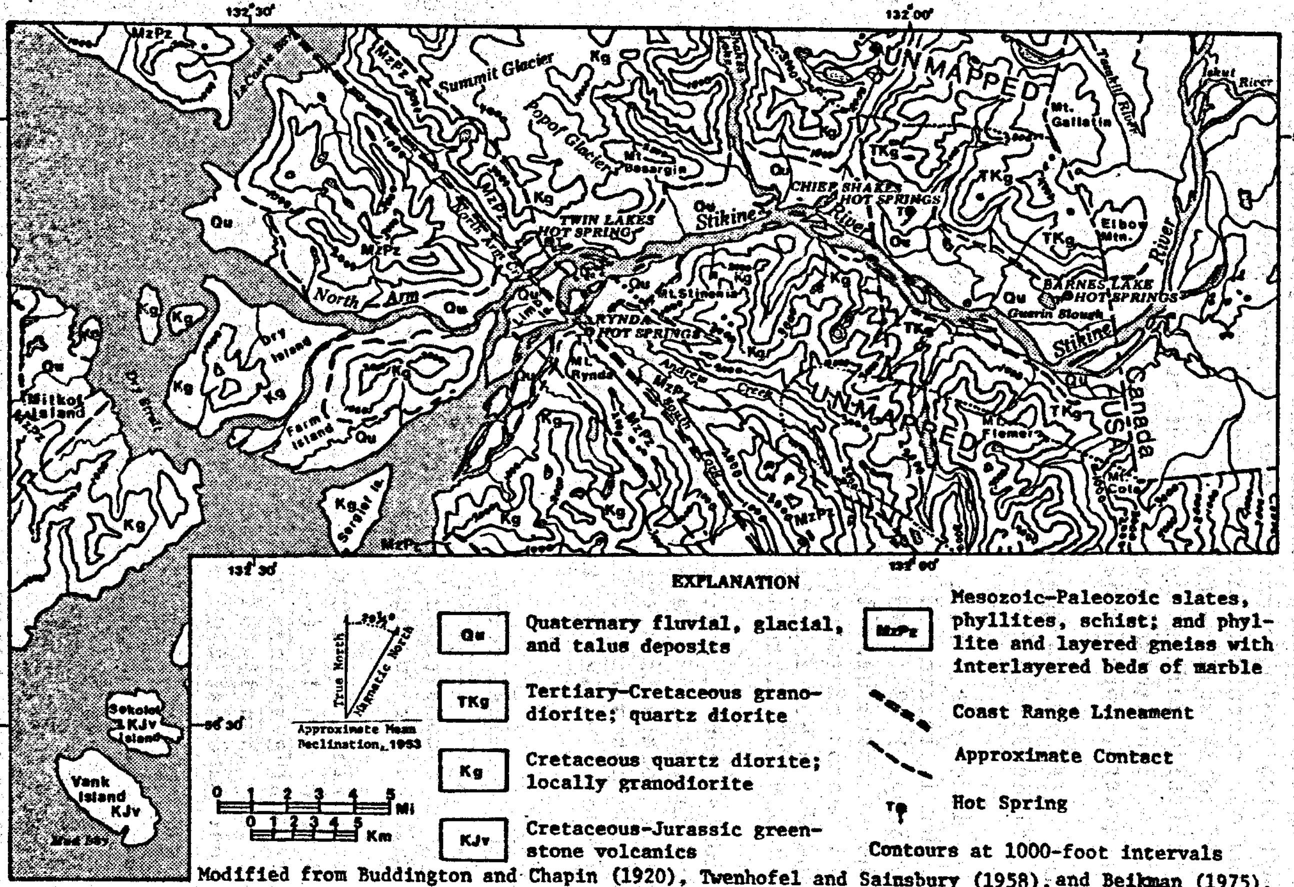
[1] Chief Shakes hot springs actually lie about 3 km due east of the position shown on the USGS topographic maps.

Figure 4. Geologic map of Stikine River area.



Modified from Buddington and Chapin (1920), Twenhofel and Sainsbury (1958), and Beikman (1975).
Base maps: Bradfield Canal (1955) and Petersburg (1960) 1:250,000 Quads

phyllitic schists, which are part of the Wrangell-Revillagigedo belt of regionally metamorphosed rocks, with the west margin of the Coast Range Batholith occurs about 15 km to the west of Chief Shakes.

Buddington and Chapin's mapping was compiled at a 1:500,000 reconnaissance scale. No detailed geologic mapping has yet been done in the Chief Shakes area. During our reconnaissance, bedrock in the immediate vicinity of the springs was found to be a fine- to medium-grained granitic gneiss with distinct to faint foliation defined by alignment of mafic minerals and potassium feldspar crystals. The rocks are composed primarily of plagioclase, potassium feldspar, quartz, hornblende, and biotite.

The base of the cliff at the springs is covered with colluvium; the valley is floored with river alluvium.

The Coast Range lineament, which extends along the western border of the Coast Range Batholith, is located 15 km west of Chief Shakes hot springs (Twenhofel and Sainsbury, 1958). No major linear occurs at the springs itself but the rocks of the area are heavily fractured. Several joint sets are observable on aerial photos of the area. The most prominent sets trend N.-S., E.-W., N. 60 E., and N. 55 W. The cliff face above the springs is approximately parallel to the N. 55 W. trend.

During our reconnaissance, a large fracture with an attitude of N. 57° E., 70° N. was noted in the cliff face several feet above the principal hot spring. Several other joints with this general attitude occur in the immediate vicinity of the hot springs. Another joint set observed at the springs had an attitude of strike N. 48° E., dip 80° N.

Spring Characteristics

Table 2 gives the chemical composition and summarizes physical properties of the two hot springs at Chief Shakes with the greatest discharge. Earlier analyses are included for comparison. The principal spring measured 50.4°C; spring 2 was 45°C. The other springs ranged from 26°C to 50°C.

Discharges from springs 1 and 2 were measured by using a pygmy flowmeter and were determined to be 320 lpm ($\pm$ 10%) and 135 lpm ($\pm$ 20%), respectively. The flow from the remaining springs is very low, the combined discharge estimated to be not more than 15 lpm. The total discharge of thermal springs at Chief Shakes is estimated at 470 lpm.

Waring (1917) reported a discharge of 380 lpm and a temperature of 52°C, apparently for spring 1 at Chief Shakes. More recently Sloan (1976) reported a flow of 200 lpm and a spring temperature of 50.5°C. The quality and accuracy of the earlier flow measurements were not stated.

No odors were detected at any of the Chief Shakes hot springs. Slight gas bubbling was noted at spring 2, probably CO_2. Bright blue-green algomats lined spring channels where temperatures were about 40°C or higher. Whitish carbonate deposits veneered dry parts of the hot spring channels.

Table 2. Chemical composition and physical properties of Chief Shakes Hot Springs. (Previous USGS analyses included for comparison; all chemical analyses in mg/l.)

	DGGS Spring 1	DGGS Spring 2	USGS[a] Spring 1	USGS[b] Spring 1
SiO_2	70	63	70	108
Al	<0.01[c]			1.2
Fe	<0.01	0.017	0.01	
Ca	14	14	15	13
Mg	0.098	0.2	0.1	0.4
Na	85	82	73	87
K	3.0	2.9	3.0	9.3
Li	0.01	0.01	0.02	
HCO_3	49	50	45	43
CO_3				18
SO_4	137	149	140	142
Cl	4.9	4.6	6.3	6.5
F	1.3	1.3		
Br	0.27	0.29		
B	0	0	0.03	
pH, field	7.90	8.05	7.1	
Dissolved solids	338	342	330	409
Hardness	35		38	
Sp conductance	427	415	390	
T, °C	50.4	45	50.5	52
Flow rate, lpm	320	135	238	441
Date sampled	9/18/79	9/18/79	7/7/76	6/19/15

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[a] Water quality analysis file, USGS Central Laboratory, Denver, CO, collected by C.F. Sloan.

[b] Waring (1917).

[c] Analyses performed on water samples filtered through 0.05 micron filter.

On the basis of the chemistry given in table 2 the waters are classified as a moderately concentrated sodium-sulfate water. The content of chloride compared to sulfate is notably low.

Reservoir Properties

Table 3 summarizes the application of silica and cation geothermometry to Chief Shakes hot springs. The Na-K-Ca (4/3) geothermometer of 66°C indicates subsurface equilibration took place at temperatures below 100°C. The Na-K-Ca (4/3) temperature is also very similar to the cristobalite geothermometer result of 60°C. Considering the granitic gneiss host rock, however, it is doubtful that cristobalite would be the controlling mineralogic phase for silica equilibration. For granitic rocks, chalcedony or quartz are probably more likely to control silica equilibration when reservoir temperatures are below 150°C (Fournier, 1973; Mariner and others, 1979). In view of the uncertainty of the controlling mineralogy and of the indication of subsurface equilibration below 100°C, the Na-K-Ca (4/3) geothermometer is chosen as representative of the minimum reservoir temperature, the chalcedony as the most likely, and the quartz conductive as the maximum temperature:

	Min	Max	Most likely	Mean	Std Dev
Subsurface T (°C)	66	118	90	91	10

The values chosen are based on DGGS spring-1 water chemistry (table 2). These reservoir estimates supersede earlier estimates appearing in USGS Circular 790 (Muffler, 1979), which were based on Waring's analyses. Although the quality of Waring's chemical analyses is unknown, both the cation and the silica geothermometers based on his reported analyses give much higher reservoir temperatures than the more recent analyses. In addition, Waring reported a spring temperature slightly higher than exists now . This suggests that either reservoir temperatures have actually declined or perhaps mixing of colder water has increased.

The two principal hot springs at Chief Shakes issue from the surface at temperatures below boiling, have large flow rates, and cation geothermometer temperatures considerably above orifice temperatures. These factors suggest mixing of colder waters (Fournier and Truesdell, 1974). A cold spring in the area was found to have a temperature of 8°C and a silica content of 8 ppm. Following the method of Truesdell and Fournier (1977), application of chalcedony mixing models gives the following results:

	Parent hot water		
Mixing model	Max T (°C)	SiO^2 (ppm)	Fraction (%)
1. Maximum steam loss	132	136	46
2. No steam loss	186	275	24

There is no observable steam loss at the ground surface. If steam separates it either goes to heating waters other than that emerging at the springs and model 1 may be applicable, or the steam remains and heats cold waters mixing with the hot water fraction and model 2 may apply.

Temperatures from the chalcedony mixing models must be used with caution. No corroborative evidence exists for mixing. Further, residence time in the granitic rock of cold spring waters sampled may have been relatively short and not representative of waters actually mixing with the hot-water fraction.

No geophysical exploration has been done at Chief Shakes and the extent of the subsurface reservoir is not known. Using the USGS suggested standard reservoir volume and the mean reservoir temperature of 91°C gives the following estimates:

Volume (km^3) = 3.3; std dev = 0.9
Thermal energy (10^{18} J) = 0.67; std dev = 0.22

These values must be viewed as highly speculative estimates.

Comments

The hot springs probably result from deep circulation of meteoric waters along fractures and faults. Fluctuations in spring discharge are indicated by comparing reported flow measurements for spring 1. The difference, however, could be attributed to difference in measuring techniques and the error inherent in measuring water flow in shallow channels.

The springs emerge at the base of a granitic cliff at and near the contact with colluvium and river alluvium. Some of the ascending thermal water may discharge beneath the surface into river alluvium and also perhaps infiltrate the Stikine River and go undetected.

Geothermometry indicates reservoir temperatures are well below that required for generation of electrical power. Chalcedony mixing models, however, suggest reservoir temperatures may be higher than that indicated strictly by geothermometry.

Table 3. Chief Shakes hot springs geothermometry.
(geothermometry based on water chemistry given in table 1;
all temperatures in degrees celsius.)

	DGGS Spring 1	DGGS Spring 2	USGS Spring 1	USGS Spring 1
Surface temperature (measured)	50.4	45	50.5	52
Cation geothermometers				
Na-K-Ca (1/3)	123	116	127	175
Na-K-Ca (4/3)	66	66	63	105
Silica geothermometers				
Adiabatic	117	112	117	136
Conductive	118	113	118	142
Chalcedony	90	84	90	115
Cristobalite	68	62	68	91
Opal	0.1	-4.5	-0.1	21

2. BARNES LAKE (PARADISE) WARM SPRINGS

Location

Latitude 56° 40.8' N., longitude 131° 52.9' W.; Bradfield Canal C-6 1:63,360 Quadrangle (1955) T. 60 S., R. 80 E., sec. 9, SW 1/4 of SW 1/4 of Copper River Meridian.

General Description

Barnes Lake warm springs (also locally known as Paradise warm springs) are located 2.5 km north of the Stikine River and about 2 km due west of the U.S.-Canadian border (fig. 1). These springs had not been reported in any previous compilation of Alaskan thermal springs and were brought to our attention by the U.S. Forest Service in Wrangell and Petersburg, Alaska. The springs are situated 1/2 km ENE of Barnes Lake in a narrow valley cut through bedrock (fig. 4). The springs are accessible by boat up the Stikine River, thence up Guerin Slough, and thence overland northward for about 1.5 km. At very high water, a shallow-draft river boat can be brought onto Barnes Lake via Ketili Creek. The springs are located within the Tongass National Forest and are managed by the U.S. Forest Service.

The region's topography is a result of intense glacier scour. Glaciers still reside on Mt. Gallatin, a 5,100-ft peak directly north of Barnes Lake. A thin veeneer of colluvium and river alluvium floor the tributary valleys surrounding Barnes Lake.

There are two warm springs, one on either side of the stream that flows through the narrow bedrock valley. The springs are difficult to find because of thick underbrush and low temperature of the waters. The first spring emerges in a shallow pool in alluvium near the base of the granitic bedrock knob on the west side of the narrow valley. The pool is about 35 m west of the stream and at an elevation of about 35 m. The second spring lies on the east bank of the stream, about 200 m upstream from the first spring. The spring occurs as a seep in water-saturated muds and is located at the base of a granitic cliff that forms the east wall of the valley. A cleft in this cliff, about 5 m wide, occurs directly above the warm seep.

The area surrounding the springs is thickly vegetated with willows, devil's club, and ferns. Sitka spruce and hemlock also occur in the area.

Geology

Bedrock in the area has been reported as Cretaceous granitic rocks: granodiorites and quartz diorites of the Coast Range Batholith (fig. 4), (Beikman, 1975; Buddington and Chapin, 1929). No detailed geologic mapping has yet been done in the area. Country rocks in the vicinity of the springs examined during our reconnaissance was found to be a medium-grained foliated biotite gneiss containing large nodules of quartz. Foliation attitude as defined by orientation of biotite crystals was strike N. 59° W., dip 48° N. In places, biotite forms up to 40 percent of the rock. Other constituents include quartz, potassium feldspar, and plagioclase.

Major fracture patterns which appear on aerial photos of the Barnes Lake area trend N. 60° E., N. 75° W., N. 40-55° W., and N.-S. One prominent fracture trending N. 60° E. lies just north of the springs. The springs themselves are aligned along N. 60° E. The walls of the cleft above the second spring also have this same general trend and a dip of 66° N.

Spring Characteristics

Table 4 provides a summary of the physical properties and a partial chemical analysis of waters from spring 1. The pool into which this spring issues measures about 3 m in diameter and is 1/2 m deep. Noticeable gas bubbling occurred intermittently at the pool, lasting 1-5 sec with a periodicity of 30-60 sec. Pool temperature was 26°C and water flow out of the pool was estimated at 30 lpm. This flow joins cold springs and eventually drains into the main valley stream.

Flow at the second spring was a trickle from saturated muds. Temperature at 10 cm depth registered 27°C.

On the basis of water chemistry of spring 1, the waters are classified as moderately concentrated sodium-sulfate waters similar to Chief Shakes hot springs.

Reservoir Properties

Table 5 summarizes the application of silica and cation geothermometry to Barnes Lake warm springs. The Na-K-Ca (4/3) geothermometer of 62°C indicates subsurface equilibration took place at temperatures below 100°C. The Na-K-Ca (4/3) temperature is also very similar to the cristobalite geothermometer result of 68°C. In view of the granitic gneiss host rock, however, cristobalite is probably not the controlling mineralogic phase for silica equilibration. For granitic rocks, chalcedony or quartz are more likely to control silica equilibration when reservoir temperatures are below 150°C (Fournier, 1973; Mariner and others, 1979). Because of the uncertainty in the controlling mineralogy and the indication of subsurface equilibration below 100°C, the Na-K-Ca (4/3) geothermometry is chosen as representative of the minimum reservoir temperature, the chalcedony as the most likely, and the quartz conductive as the maximum:

	Min	Max	Most likely	Mean	Std Dev
Subsurface T (°C)	62	118	90	90	10

The low flow rates of the springs suggest that thermal waters cooled conductively on ascent. Some mixing of colder water at shallow depths probably also occurs.

No geophysical exploration has been done at Barnes Lake and the extent of the subsurface reservoir is not known. Using the standard mean reservoir volume and the mean reservoir temperature of 90°C gives the following estimates:

Volume (km^3) = 3.3; std dev = 0.9
Thermal energy (10^{18} J) = 0.67; std dev = 0.22

These values must be viewed as highly speculative estimates.

Table 4. Chemical composition and physical properties of Barnes Lake warm springs waters.[a] (All chemical analyses in mgl.)

	Spring 1
SiO_2	71
Al	0.06
Fe	0.17
Ca	8.9
Mg	0.44
Na	72
K	3.5
Li	0.02
HCO_3	n d[b]
CO_3	n d
SO_4	110
Cl	13
F	2.3
Br	n d
B	0.05
Ph, field	6.95
Dissolved solids	249
Hardness	24
Sp conductance	455
T, °C	26
Flow rate lpm	30
Date sampled	9/25/79

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[a] Analyses performed by Chemical and Geological Laboratories of Alaska.
[b] Not determined.

Table 5. Barnes Lake (Paradise) warm springs geothermometry, spring 1. (geothermometry based on water chemistry given in table 4; all temperatures are in degrees Celsius.)

Surface temperature (measured)	26
Cation geothermometers	
Na-K-Ca (1/3)	112
Na-K-Ca (4/3)	62
Silica geothermometers	
Adiabatic	118
Conductive	119
Chalcedony	90
Crystobalite	68
Opal	0.76

Comments

These warm springs probably result from deep circulation of meteoric waters along fractures and faults in the granitic host rock. Some of the ascending thermal water may discharge beneath the surface and go undetected. Geothermometry indicates reservoir temperatures are well below that required for generation of electrical power. These springs are also very remote and have low temperatures and low flow rates, making them impractical for most geothermal applications.

3. FOWLER HOT SPRINGS (CANADA)

Location:

Latitude 56°50' N., longitude 131°45' W.; Bradfield Canal 1:250,000 Quadrangle (1955), Copper River Meridian.

General Description:

Fowler Hot Springs are located in Canada about 8 km from the U.S.-Canadian border (fig. 1). The springs occur less than 1/2 km east of the Stikine River at the base of Warm Spring Mountain and directly opposite the terminus of the Great Glacier (fig. 5). These springs were visited by DGGS while assessing other hot springs that occur along the Alaskan part of the Stikine River. Fowler Hot Springs were included in this study because of their possible relationship to the hot springs on the lower Stikine and because of their potential bearing on interpreting the nature and cause of hydrothermal activity along the Stikine River. No other hot springs are known to exist farther upriver past the Fowler site.

Fowler Hot Springs can be reached by boat up the Stikine River about 25 km past the US-Canadian border. A shallow-draft boat or canoe can then be taken up a small tributary creek whose mouth lies nearly opposite a homestead cabin on the west bank of the Stikine River. This creek drains a series of beaver ponds into which the hot springs either drain or issue. Wright (cited in Waring, 1917, p. 25) reported the existence of as many as 18 individual springs issuing at scalding temperatures from fissures in bedrock at the base of a granitic wall. Most of these springs have since been submerged by beaver ponds. At least four low-discharge hot springs still remain above water level. Local travelers commonly use the tepid waters of the beaver ponds for bathing.

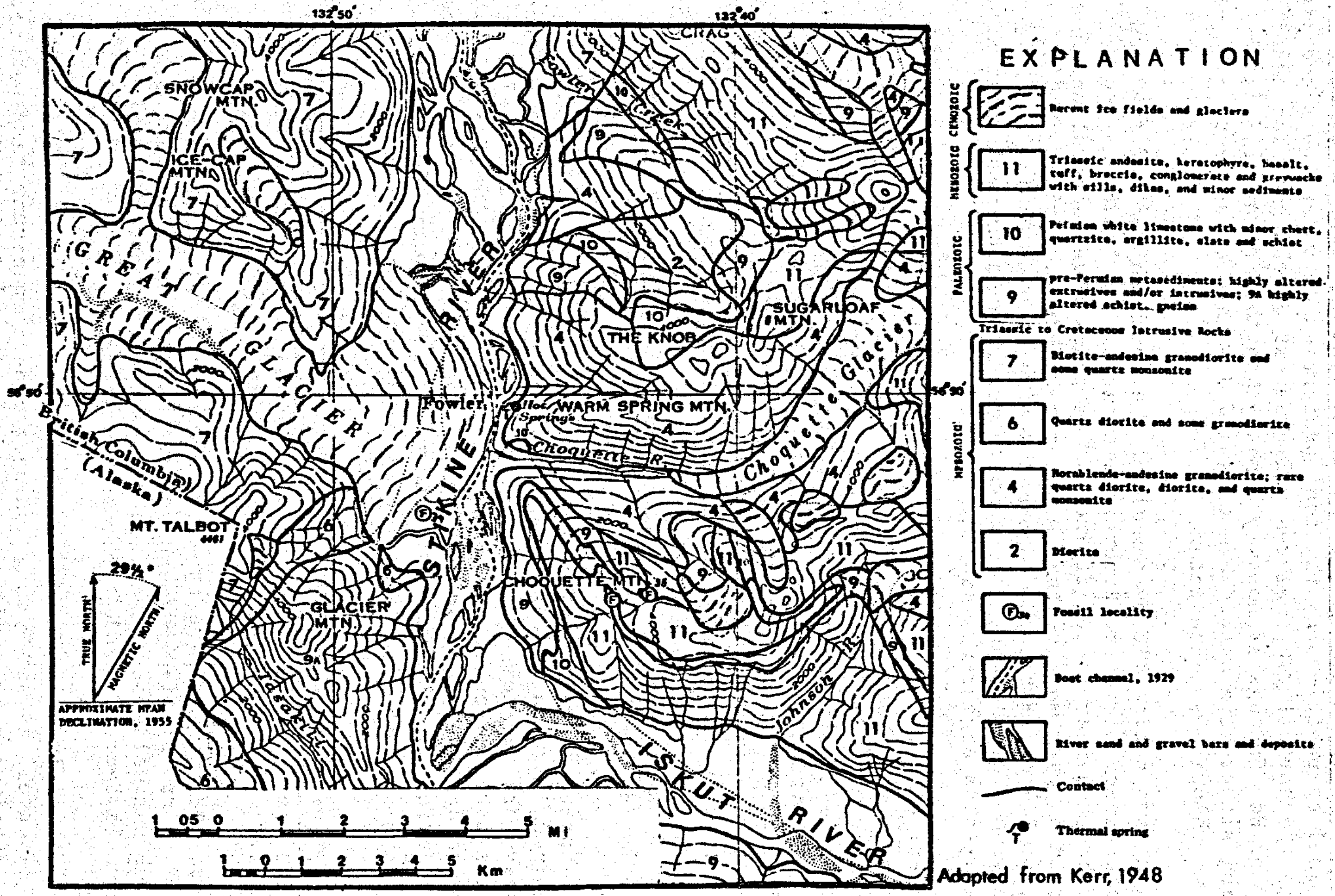
As along its lower course, the Stikine River at Fowler Hot Springs flows through a spectacular glaciated valley. Higher elevations are still glacier clad and the Great Glacier descends almost to river level. Valley slopes are steep but heavily forested with Sitka spruce. Tree line ranges from 600 m (2,000 ft) to 900 m (3,000 ft) in elevation.

Geology:

Bedrock in the vicinity of the hot springs was reported as hornblende andesine granodiorite of Mesozoic age (fig. 5) (Kerr, 1948). The intrusive is part of the Coast Range Batholitic complex. Contacts with Permian and Pre-Permian limestones and schists occur several kilometers north and south of the spring location.

Rocks examined at the springs site were found to be medium- to small-grained biotite hornblende granodiorites with numerous xenoliths present. In places, mafics are 30-40 percent of the rock.

At least two joint systems occur near the springs site. Their attitudes are strike N. 36° W., dip 62° N. and strike N. 16 E., dip 88° S. The cliff face above the springs trends N. 18° W.



Adopted from Kerr, 1948

Figure 5. Geologic map of Fowler Hot Springs area.

Spring Characteristics:

The four visible springs are dispersed along the base of the granitic valley wall over a linear zone 200 m wide. The springs issue from fissures just above pond water level. Spring temperatures range from 50°C to 60°C and the combined flow of the four springs is estimated at 40-50 lpm. Table 6 gives the chemical composition and physical properties of waters obtained from the spring with the hottest temperatures and greatest discharge. The chloride content of Fowler Hot Springs is notably higher than that at other Stikine River hot-spring sites. The waters are classified as a moderately concentrated sodium chloride-sulfate water.

Gas bubbling was noticeable over much of the southern part of the pond, indicating the presence of several submerged springs. At least one zone of upwelling was observed near the cliff face. The ponds cover several acres and surface temperatures over much of this area were 25°-26°C. Wright in 1905 (Waring, 1917, p. 25) estimated the combined spring flow at Fowler at about 3,000 lpm. Judging from the warm pond temperatures, the submerged springs are still discharging at a high rate and temperature.

Reservoir Properties:

Table 7 summarizes the application of silica and cation geothermometry to Fowler Hot Springs. The Na-K-Ca (4/3) geothermometer temperature of 85°C indicates subsurface equilibration took place at temperatures below 100°C. The Na-K-Ca (4/3) temperature is very close to the chalcedony temperature of 86°C. Chalcedony then probably controls silica equilibration and is chosen as representative of the minimum and most likely reservoir temperature. The quartz conductive geothermometer temperature is taken as the maximum:

	Min	Max	Most Likely	Mean	Std Dev
Subsurface T (°C)	86	115	86	96	7

Fowler Hot Spring issue from the surface at temperatures below boiling and have a combined large rate of flow and a cation geothermometer temperature considerably above orifice temperature. These factors suggest the mixing of colder subsurface waters (Fournier and Truesdell, 1974). With a silica content of 10 ppm and a temperature of 10°C as representative of the subsurface cold waters and following the method of Truesdell and Fournier (1977), application of chalcedony mixing models gives the following results:

	Parent hot water		
Mixing model	Max T (°C)	SiO_2 ppm	Fraction (%)
1. Maximum steam loss	114	108	56
2. No steam loss	144	156	38

Temperatures derived from the chalcedony mixing models must be treated with caution. No corroborative evidence exists for mixing. Further, the silica content of the cold water fraction may be underestimated. A higher silica content would decrease both temperature estimates.

No geophysical exploration has yet been done at Fowler and the extent of the subsurface reservoir is not known. Using the mean reservoir temperature of 96°C and the standard mean reservoir volume yields the following estimates:

Table 6. Chemical composition and physical properties of Fowler Hot Springs.[a]
(All chemical analyses in mg/1.)

SiO_2	66
Al	<0.05
Fe	0.03
Ca	52
Mg	0.25
Na	204
K	9.8
Li	0.10
HCO_3	29
CO_3	- -
SO_4	230
Cl	251
F	1.6
Br	0.95
B	0.1
ph, Field	7.95
Dissolved solids	830
Hardness	131
Sp Conductance	1247
T, °C	60
Flow rate lpm	15-20
Date sampled	9/24/79

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[a] Analysis performed on waters obtained from spring having hottest temperature and largest discharge.

Table 7. Fowler Hot Spring geothermometry.
(All temperatures are in degrees celsius.)

Surface temperature (measured)	60
Cation geothermometers	
Na-K-Ca (1/3)	139
Na-K-Ca (4/3)	85
Silica geothermometers	
Adiabatic	114
Conductive	115
Chalcedony	86
Cristobalite	65
Opal	-2.5

Volume (km^3) = 3.3; std dev = 0.9
Thermal energy = (10^{18} J) = 0.72; std dev = 0.22

These values must be viewed as highly speculative estimates.

Comments

The hot springs probably result from circulation of meteoric waters along deep-seated fractures. Geothermometry indicates reservoir temperatures are well below that required for generation of electrical power. The high flow rate and relatively high surface temperatures of the hot spring waters, however, make them suitable for space heating and perhaps for heating of greenhouses.

4. TWIN LAKES (WEST SHAKES) WARM SPRINGS

Location

Latitude 56° 42.0' N., longitude 132° 16.8' W.; Petersburg C-1 1:63,360 Quadrangle T. 60 S., R. 84 E., S of NW.

General Description

Twin Lakes warm springs are located 25 km NW of Wrangell, Alaska and 16 km west of Chief Shakes hot springs (fig. 1). The springs are situated 1/2 km north of the Stikine River at the NW end of a lake labeled Figure Eight Lake on USGS topographic maps and locally known as Twin Lakes (fig. 4). These springs are believed to be the springs Waring (1917) reported as West Shakes Hot Springs.[1] The Twin Lakes warm springs are located in the Tongass National Forest and come under the management of the U.S. Forest Service.

A short Forest Service trail leads from the north bank of the Stikine River to a narrows between the two lakes and affords access to the springs. Alternately, at high water, a shallow-draft boat can be brought up a tributary slough almost directly to the springs.

There are two springs. Both lie close to the NW lake shore at the base of a granitic cliff that forms the northern boundary of the Stikine Valley. Spring 1 emerges from under boulders at the bottom of an avalanche chute; spring 2 emerges from alluvium, near the base of the cliff. Twin Lakes water level fluctuates with Stikine River stage, the latter acting as a dam on tributaries. At high water the springs may become submerged.

The Stikine River south of Twin Lakes is multichannelled, about 1.5 km wide and heavily laden with silt during runoff months. The river flows through a spectacular glaciated valley flanked by steep walls and hanging valleys. Glaciers presently cap the mountains immediately north of Twin Lakes and descend to as low as 600 m (2,000 ft) elevation. Tree line in the vicinity ranges from 600 m (2,000 ft) to 900 m (3,000 ft) in elevation.

Local flora consists of willows, grasses, and shrubs interspersed with stands of Sitka spruce and hemlock.

Geology

The springs are located near the contact between Cretaceous granitic rocks (believed to be part of the Coast Range Batholith) and Mesozoic and Paleozoic schists and layered gneisses (fig. 4) (Beikman, 1975; Buddington and Chapin, 1929). Bedrock in the immediate vicinity of the springs is a foliated, medium-grained hornblende-biotite quartz diorite with well-defined gneissoid structure.

The Coast Range lineament lies immediately west of the springs (Twenhofel and Sainsbury, 1958). Predominant fracture patterns as determined from aerial

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[1] Waring did not visit these springs and relied on information obtained from local inhabitants in judging their location.

photos of the area trend N. 55° W. and N. 75°-80° W. The valley wall here trends E.-W.

A 3- to 4-m wide cleft occurs in the granitic wall about 10 m above the warm springs. The cleft trends N. 50° W. and has nearly vertical walls. Attitude of other joints near the springs measured strike N. 70° W., dip 60° S. and strike N. 48° E., dip 85° S.

The base of the granitic wall is covered with colluvium and alluvium.

Spring Characteristics

The temperature of spring 1 measured 21°C; spring 2 was 18°C. Some fluctuation in water temperature was noted at spring 1. During a previous visit waters were noticeably warmer to the touch. A local resident measured a winter season spring temperature of 26°C. Cold surface water flowing down the cliff face was noted above the springs and some mixing of these colder waters probably occurs during summer months.

Table 8 gives the chemical composition and summarizes physical properties of spring 1, the warmer of the two springs. Samples were obtained at the point where the spring issues from the boulders. The combined discharge of the two springs measured with a pygmy flowmeter was 270 lpm ± 10 percent.

No odors were detected at the spring site nor were there noticeable hydrothermal deposits of any kind.

Reservoir Properties

Table 9 summarizes the application of silica and cation geothermometry to Twin Lakes warm spring 1. The Na-K-Ca (4/3) geothermometer temperature of 49°C indicates equilibration took place well below 100°C. The Na-K-Ca (4/3) temperature is also similar to the chalcedony temperature of 41°C, indicating that chalcedony probably controls silica equilibration. The chalcedony temperature is chosen as minimum reservoir temperature, Na-K-Ca (4/3) as most likely, and quartz conductive as maximum:

	Min	Max	Most Likely	Mean	Std Dev
Subsurface T (°C)	41	73	49	54	7

Some mixing with colder subsurface waters probably occurs. Silica concentrations and spring temperatures are much too low, however, to allow accurate application of silica-mixing models.

No geophysical exploration has been done at Twin Lakes and the extent of the subsurface reservoir is not known. Using the mean reservoir temperature of 54°C and the standard mean reservoir volume gives the following estimates:

Volume (km^3) = 3.3; std dev = 0.9
Thermal energy (10^{18} J) = 0.35; std dev = 0.14.

These values must be viewed as highly speculative estimates.

Table 8. Chemical composition and physical properties of Twin Lakes (West Shakes) warm springs 1. (All chemical analyses in mg/l.)

SiO_2	26
Al	<0.05
Fe	0.027
Ca	5.4
Mg	0.3
Na	26
K	1.3
Li	<0.01
HCO_3	30
CO_3	- -
SO_4	32
Cl	60
F	0.28
Br	0.13
B	0.1
ph, field	6.95
Dissolved solids	112
Hardness	15
Sp Conductance	150
T, °C	21
Flow rate lpm	270±10%[a]
Date sampled	9/21/79

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[a] Combined flow rate for springs 1 and 2.

Table 9. Twin Lakes (West Shakes) warm springs geothermometry. (All temperatures are in degrees celsius.)

Surface temperature (measured)	21
Cation geothermometers	
Na-K-Ca (1/3)	128
Na-K-Ca (4/3)	49
Silica geothermometers	
Adiabatic	78
Conductive	73
Chalcedony	41
Cristobalite	24
Opal	-38

Comments

These warm springs probably result from deep circulation of meteoric waters along fractures and faults in the granitic host rock. Some of the ascending thermal water may discharge beneath the surface and go undetected. Geothermometry indicates reservoir temperatures are well below that required for generation of electrical power. These springs are also very remote and have low temperatures and low flow rates, making them impractical for most geothermal applications.

5. MT. RYNDA (SOUTH STIKINE), REPORTED WARM SPRING

Location

(Approximate) latitude 56° 39.5'N., longitude 132° 16' W.; Petersburg C-1 1:63,360 Quadrangle (1953) map.

General Description

Waring (1917) reported a thermal spring on the south bank of the Stikine River that he called South Stikine hot springs. Waring did not visit the site but relied on information obtained from prospectors and trappers in estimating the location. A local inhabitant informed us that during his winter explorations at the base of Mt. Rynda (fig. 4), he came across a snow free area surrounding a warm seep. The location of this zone approximately fits the location of Waring's South Stikine hot springs. The site described by the informant could not be found during an intensive search for the spring made in September 1979. The thick fall undergrowth, however, could have easily obscured the location.

The area lies 22 km NNE of Wrangell, Alaska near the mouth of Andrew Creek, which can be reached by boat up the Stikine River (figs. 1 and 4). The seep was reported to be located on a low ridge south of Andrew Creek at an estimated elevation of 30 m (100 ft). The area lies within Tongass National Forest and comes under the management of the U.S. Forest Service.

The slopes are heavily forested with spruce and thick underbrush.

Geology

The Mt. Rynda site lies within a belt of Mesozoic and Paleozoic slates, schists, and phyllites with interlayered beds of marble that occur along the west flank of the Coast Range Batholith (fig. 4) (Buddington and Chapin, 1929; Beikman, 1975). The contact with Cretaceous quartz diorites lies about 1 km east of the site. The Coast Range lineament follows Andrew Creek, just east of the site. The area is highly fractured. Inspection of aerial photos show two dominant trends: N. 54° W. and N. 12° W. The region has been intensely glaciated, which has accentuated the regional structural grain. The valley is floored with alluvium and colluvium; the lower mountain slopes are covered with a thin layer of soil.

Spring Characteristics:

The spring is reported to occur as a warm seep in the alluvial cover. No other information is available.

Reservoir Properties:

No information is available.

Comments:

The spring apparently has a low surface temperature and a very low discharge, making it impractical for most geothermal applications. Subsurface conditions, however, are unknown. An overflight of the area in winter could help in locating the spring.

6. VANK ISLAND, REPORTED HOT SPRING, UNSUBSTANTIATED

Location

(Approximate) latitude 56° 27' N., 132° 36' W.; Petersburg 1:250,000 Quadrangle (1960); T. 62 S., R. 82 E., Copper River Meridian.

Comments

On the basis of a second-source account, Waring (1917) reported a possible hot-spring site on the south end of Vank Island (fig. 1). The existence of these springs could not be substantiated by DGGS. Local residents who had lived extensively on the south end of the island were interviewed. None of them had any knowledge of thermal springs on Vank Island. A cursory search of the southern shoreline by DGGS did not turn up any signs of thermal springs. These springs probably either do not exist or have dried up since Waring's earlier report.

7. VIRGINIA LAKE, REPORTED WARM SPRING (UNSUBSTANTIATED)

Location

(Approximate) latitude 56° 28.5' N., longitude 132° 09.5' W.; Petersburg 1:250,000 Quadrangle (1960); T. 62 S., R. 85 E., Copper River Meridian.

Comments

A long-time resident of Wrangell, Alaska reported the existence of winter warm-water flow near the south-central shore of Virginia Lake, located several km east of Wrangell, Alaska (fig. 1). An intensive search of the reported location by DGGS failed to disclose any thermal anomalies. The existence of the warm flow could not be substantiated by other area residents, including bush pilots and U.S. Forest Service personnel familiar with Virginia Lake.

8. BRADFIELD HOT SPRINGS

Location

Latitude 56° 13.9' N., longitude 131° 16.2' W.; Bradfield Canal A-4 1:63,360 Quadrangle (1955); T. 65 S., R. 91 E., Copper River Meridian.

General Description

Bradfield hot springs are located on East Fork Bradfield River about 15 km east of Bradfield Canal (fig. 1). The springs lie on a steep slope about 300 m SE of the river near the confluence of East Fork Bradfield River with a major northern tributary to the river (fig. 6). The springs were discovered during a recent clear-cutting operation on the SE slope of the valley and were brought to our attention by U.S. Forest Service personnel in Wrangell, Alaska. The springs can be reached via a logging road that begins at the main logging camp at the head of Bradfield Canal and continues past the springs site. The springs are within the Tongass National Forest and come under the management of the USFS.

There are two groups of springs. The first group is composed of three springs issuing from fissures in the granitic bedrock about 20 m above road level. The second group consists of one spring issuing from a bedrock fissure and several adjacent small seeps situated about 45 m above road level. The seeps occur in the thin veneer of soil covering the slopes.

The East Fork Bradfield River valley is a glaciated valley, heavily forested on its lower slopes and flanked by glacially scoured bedrock domes at high elevations. Small glaciers still cap the mountains north and south of the spring site. Tree line ranges from 600 m (2,000 ft) to 900 m (3,000 ft).

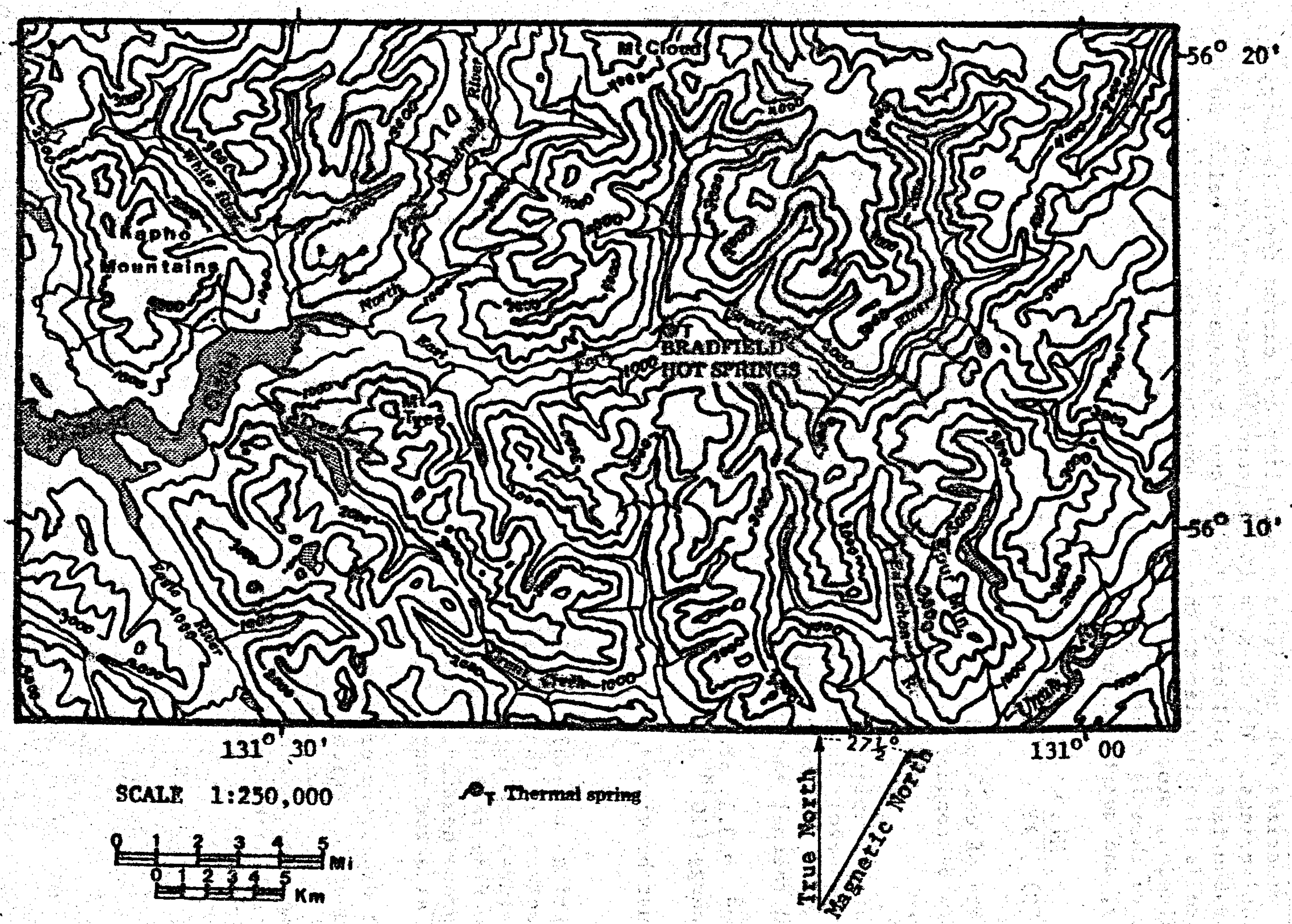
Geology

No geologic mapping of the East Fork Bradfield River area has been done. Bedrock examined at the springs site was found to be a medium- to coarse-grained, hornblende-biotite granodiorite with gneissoid structure. The bedrock contains many conspicuous, dark-gray fine-grained xenoliths, some over 1/2 m long. The host rock is probably part of the Coast Range Batholith, which occurs both north and south of the Bradfield area.

Prominent fractures are apparent on aerial photos of the springs area. These trend N. 13° E., N. 38° E., and N. 77° E. One prominent fracture with a trend of N. 65° E. occurs on the ridge directly above the springs. The attitude of joints located at the springs measured strike N. 43° E., dip 77° S.

Spring Characteristics

Water flow from the springs is quite low, with the combined discharge estimated at less than 40 lpm. Spring temperatures ranged from 46° to 57°C. Dark-green algomats line the upper channels of the hot springs. No odors or gas bubbling were noticeable at any of the springs. Table 10 gives the physical properties and chemical composition of thermal waters obtained from the springs having the highest temperature and greatest discharge. The waters are classified as a moderately concentrated sodium-sulfate water.



Contours at 1000 foot intervals

Base map: Bradfield Canal (1955) 1:250,000 Quad

Approximate Mean Declination, 1955

Figure 6. Location map, Bradfield Hot Springs.

Reservoir Properties

Table 11 summarizes the application of silica and cation geothermometry to Bradfield hot springs. The Na-K-Ca (4/3) geothermometer temperature of 92°C suggests subsurface equilibration took place at temperatures below 100°C. This temperature is also similar to the chalcedony temperature of 102°C, indicating chalcedony probably controls silica equilibration. The Na-K-Ca (4/3) temperature then is chosen as minimum reservoir temperature, chalcedony the most likely, and quartz conductive as the maximum:

	Min	Max	Most Likely	Mean	Std Dev
Subsurface T (°C)	92	130	102	108	8

The low rate of combined flow of the hot springs suggests that the waters cool principally by conduction during ascent. Some mixing of colder waters may also occur.

No geophysical exploration has been done at Bradfield and the extent of the subsurface reservoir is not known. Using the standard mean reservoir volume and the mean reservoir temperature of 108°C gives the following estimates of stored thermal energy:

Volume (km^3) = 3.3; std dev = 0.9
Thermal energy (10^{18} J) = 0.83; std dev = 0.9

These values must be viewed as highly speculative estimates.

Comments

The hot springs probably result from circulation of meteoric waters along deep-seated fractures in the country bedrock. Geothermometry indicates reservoir temperatures are well below that required for generation of electrical power. The low flow rate of the springs limits their use for other applications such as space heating. Local inhabitants expressed interest in developing the site for recreational and bathing purposes.

The recent discovery of these hot springs suggests that other hot springs exist in unfrequented regions of the Coast Range Mountains and have not as yet been detected. Vapor plumes emanating from such sites would be most prominent during winter months. Aerial reconnaissance of Coast Range drainages during winter could help locate additional hot-spring sites.

Table 10. Chemical composition and physical properties of Bradfield hot springs.[a]
(All chemical analyses in mg/l.)

SiO_2	87
Al	<0.01)[b]
Fe	0.02)[b]
Ca	13
Mg	0.04
Na	118
K	5.7
Li	0.04
HCO_3	42
CO_3	- -
SO_4	235
Cl	30
F	2.5
Br	22
B	0.1
ph, Field	8.15
Dissolved solids	512
Hardness	33
Sp Conductance	885
T, °C	57
Flow rate lpm	14
Date sampled	10/2/79

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[a] Analysis performed on spring with highest temperature and greatest discharge.

[b] Analysis performed on acidified 0.05-micron filtrate

Table 11. Bradfield hot spring geothermometry.
(All temperatures are in degrees celsius.)

Surface Temperature (measured)	57
Cation Geothermometers	
Na-K-Ca (1/3)	142
Na-K-Ca (4/3)	92
Silica Geothermometers	
Adiabatic	126
Conductive	130
Chalcedony	102
Cristobalite	79
Opal	10

9. BELL ISLAND HOT SPRINGS

Location

Latitude 55° 56.0' N., longitude 131° 33.4' W.; Ketchikan (D-5) 1:63,360 Quadrangle (1952); T. 68 S., R. 90 E., sec. 31, SE 1/4 of NW 1/4 of Copper River Meridian.

General Description

Bell Island Hot Springs are located near the southwest end of Bell Island, about 65 km north of Ketchikan (fig. 1). The principal springs are situated on the north bank of a small creek that drains into a narrow cove (fig. 7). The springs are about 400 m from and 5 m above high-tide level.

Waring (1917) reported that the waters issue from a narrow fissure about 15 m long in biotite granite. The springs are encased in five concrete basins. A sixth spring emerges at the base of the valley wall, about 120 m downstream from the main group.

The surrounding terrain has been intensely glaciated, accentuating NE-SW regional structures grain. The valley in which the springs occur is narrow with steep slopes rising to 1500 ft in elevation. The slopes are densely forested with spruce and have numerous exposures of bedrock. The island is surrounded by highly scenic, long, narrow passages flanked by steep, heavily forested slopes.

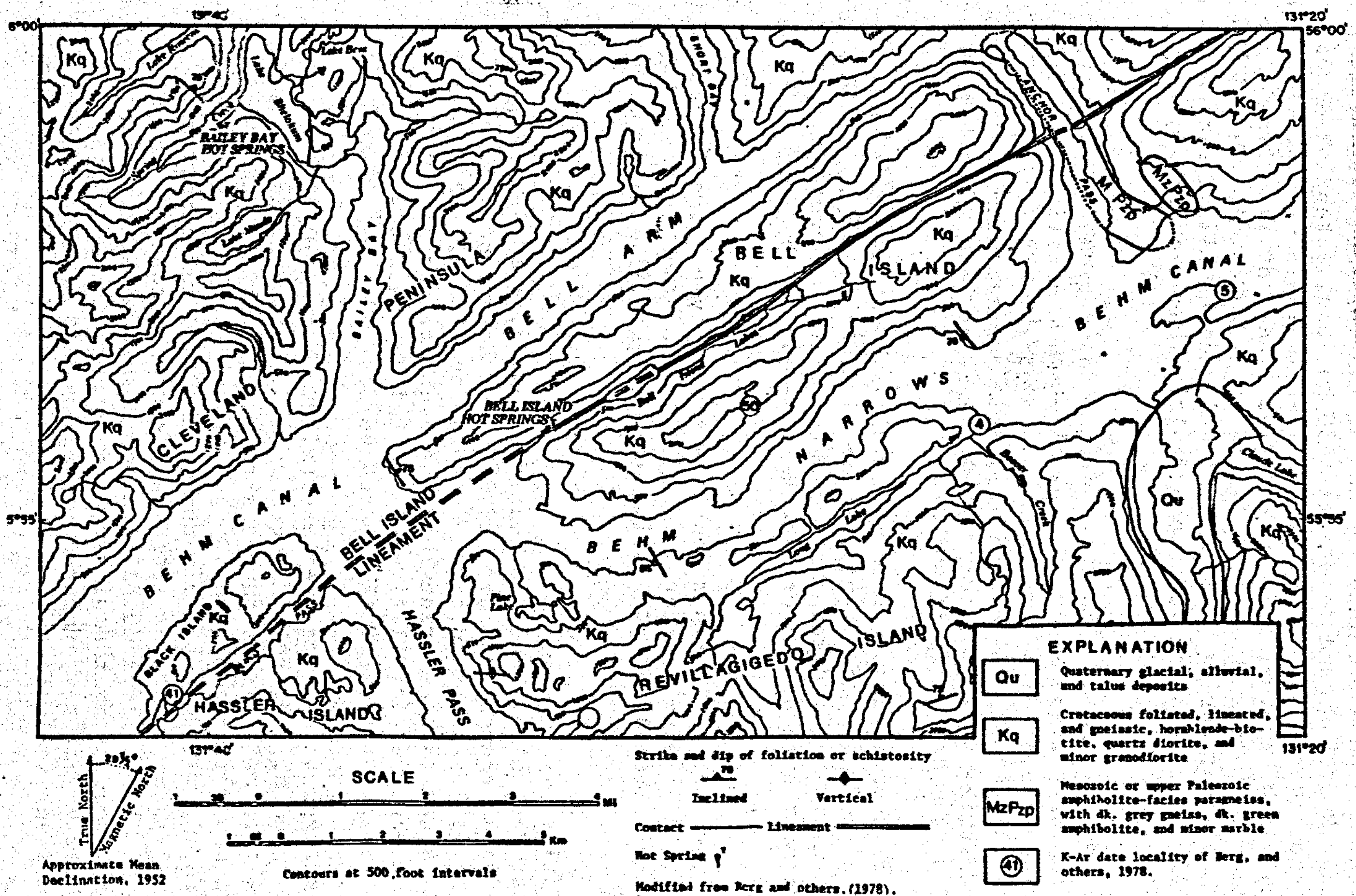
The springs are privately owned and serve as a major attraction for the Bell Island Fishing Resort. The resort uses the hot-springs waters to heat the main lodge, 14 other buildings and cabins, and a large swimming pool. The springs have been used in a similar fashion periodically since 1902. The resort includes 11.05 acres surrounding the springs and is presently owned by Don Peterson. Surrounding lands are within the Tongass National Forest and are managed by the U.S. Forest Service.

Bell Island can be reached by either boat or float plane from Ketchikan. The resort maintains a dock and float plane ramp and a 1/2-mile-long boardwalk that connects the dock to the main resort area and hot springs. The resort is closed to the public from October 1 to late spring.

Geology

Predominate rock types in the Bell Island are lineated, gneissic hornblende-biotite quartz diorites and minor granodiorites that are probably part of the Coast Range Batholith (fig. 7) (Berg and others, 1978). Radiometric K-Ar ages in the western part of the pluton indicate emplacement occurred about 72 to 84 m.y. ago. The intrusive rocks are cut by numerous dikes and veins of light-gray-weathering, quartz-feldspar-(biotite-garnet) pegmatite. The pluton also contains abundant ellipsoidal dark inclusions that commonly are parallel to aligned hornblende crystals. Foliation attitude at the SE. end of Bell Island was reported as strike N. 35° W., dip 75° N.

Bedrock examined near the spring site is a faintly foliated, medium-grained gneissic hornblende quartz diorite. Hornblende occurs in sub-



Modified from Berg and others.(1978).
Base map: Ketchikan D-5 & D-6 (1955) 1:63,360 Quads

Figure 7. Geologic map of Bell Island and Bailey Bay Hot Springs area.

hedral crystals 2-3 mm long, which are parallel to foliation. Pegmatitic dikes cutting outcrops vary in width from a few centimeters to 30 cm.

The hot springs lie on a prominent lineament that bisects Bell Island. The lineament trends N. 58° E. and is at least 30 km long, extending across Anchor Pass to the northeast and through Blind Pass to the southwest (fig. 7). Clearly visible on aerial photos, this distinctive lineament is defined by a steep, narrow valley that runs the entire length of Bell Island. The lineament may be fault related; Waring (1917) found evidence of slickensiding in rocks near the springs, suggesting that some faulting has occurred in the vicinity.

Inspection of aerial photos shows the surrounding region to be highly fractured with a dominant trend N. 60° E., parallel to the Bell Island lineament. Other prominent joint sets visible on the photos trend N. 75° E., N. 40° E., N-S, and S. 30°-45° E. A joint system examined near the hot springs has an attitude of strike N. 58° E., dip 77° N. Another less prominent joint set trends N. 33° W. and has a dip of 60° N.

Aphanitic to fine-grained quartz porphyry dikes have invaded a number of the N. 60° E.-trending fractures. Ranging from 20-30 cm in width, the dikes cut the pegmatitic dikes. The dikes are probably related to porphyritic granite and quartz monzonite stocks of Miocene age reported by Berg and others (1978).

The valley floor near the springs consists primarily of granular alluvial and glacial deposits overlying bedrock (Pyle, 1978). on the basis of shallow refractive seismic profiles, Pyle estimated sediment fill at the head of the cove at 12-15 m.

Spring Characteristics

Water temperatures in the enclosed concrete basins on 10/5/79 ranged from 74.1°C for the southeasternmost basin to 67.4°C for the northeasternmost (table 12). Waring (1917) reported a temperature range of 52°-72°C for the five basins, which suggests the springs have become hotter since 1915. Noticeable bubbling occurred in all five basins and the waters smelled slightly sulfurous. The sixth spring measured 46°C.

Spring temperatures were remeasured several times from 10/5/79 to 10/10/79 and were found to fluctuate as much as 1°C. Comparisons to measurements made in recent years by other investigators (Ogle, 1976; Baker and others, 1977) suggest spring temperatures can vary as much as 3°C during summer. A temperature log was kept briefly during 1978 and indicated spring temperatures during winter may be several degrees higher than during summer. This was corroborated by the 1979-80 Bell Island Resort caretaker, who claimed that spring temperatures during the winter were as much as 10°C higher than during summer. These higher temperatures perhaps result from reduced winter infiltration of cold surface waters.

The total rate of discharge from the concreted springs is estimated at 100 lpm. Waters from the concreted springs are directly piped to the various buildings and pools and no provision exists for measuring flow directly at the source. The estimate is based on measurements of overflow and discharge at the individual cabins and swimming pool; the estimate neglects minor leaks and evaporation from the pool surface. These estimates compare well with other recent estimates of flow (Baker and others, 1977; Sloan, 1976a-c). The measured flow rate however, differs substantially from Waring's visual estimate of 35 lpm in 1915 and suggests that a significant increase in spring discharge has occurred. The estimated flow of the sixth spring is 2 lpm.

Table 13 gives the chemical composition and physical properties of waters obtained from basin 5. Previous analyses indicate that chemical composition of the waters has remained relatively stable. The waters are classified as a sodium chloride-sulphate water.

Reservoir Properties

Table 14 summarizes the application of silica and cation geothermometry to Bell Island Hot Springs using the DGGS water-chemistry analysis. Previous geochemical analyses give similar results. Cation geothermometry indicates subsurface equilibration took place at temperatures above 100°C and that therefore the Na-K-Ca (1/3) temperature applies. This temperature is also very similar to the quartz conductive geothermometer temperature, indicating quartz probably controls subsurface silica equilibration. In estimating reservoir temperature, the chalcedony temperature is taken as minimum, quartz conductive as most likely, and Na-K-Ca (1/3) as maximum:

	Min	Max	Most Likely	Mean	Std Dev
Subsurface T (°C)	115	144	142	134	7

The hot springs at Bell Island issue from the surface at temperatures below boiling, have a combined flow rate of over 100 lpm, and a cation geothermometer temperature considerably above orifice temperatures. These factors suggest the mixing of colder waters (Fournier and Truesdell, 1974). The apparent increase in spring temperatures during winter also suggests some mixing of colder waters. A cold spring in the area was found to have a temperature of 10°C and a silica content of 6 ppm. Following the method of Truesdell and Fournier (1977), application of quartz mixing models gives the following results:

	Parent hot water		
Mixing model	Max T (°C)	SiO_2 (ppm)	Fraction (%)
1. Maximum steam loss	150	125	73
2. No steam loss	217	330	31

There is no observable steam loss at the ground surface. If steam separates it either goes to heating water other than that emerging at the springs and model 1 may be applicable, or it remains and heats cold waters mixing with the hot-water fraction and model 2 may apply. The maximum temperature of 217°C from model 2 seems high for a hydrothermal system circulating in deep fractures in the granitic bedrock. The temperature predicted by model 1, however, is quite close to that indicated by geothermometry.

Table 12. Bell Island Hot Spring temperatures, 10/5/79.
(Numbers correspond to concrete basins going NE to SW.)

	1	2	3	4	5
Temperature (°C)	67.4	72.0	71.8	71.3	74.1

Table 13. Chemical composition and physical properties of Bell Island Hot Springs.
(Previous analyses included for comparison;
all chemical analyses in mg/l.)

	DGGS	USGS[a]	USGS[b]	USGS[c]	USGS[d]
SiO_2	108	110	125	113	105
Al	<0.01[e]				1.2
Fe	0.009[e]	0.01			
Ca	8.3	10	10	9.5	4.6
Mg	0.01	0.1	0	0.0	1.0
Na	176	200	197	200	201
K	7.2	7.2	6.6	7.2	
Li	0.15	0.15			
HCO_3	58	60	53	30	37
CO_3	1.5	14		15	13
SO_4	118	130	79	130	129
Cl	197	200	189	186	188
F	4.6				4.4
Br	0.14				
B	0.6	0.46			
pH, field	8.50	8.4	8.95	8.7	
Dissolved Solids	650	702			674
Hardness	21	25			
Sp Conductance	950	920			
T, °C	74.1	74	72	71	72
Flow rate lpm	100[f]	100[f]			35 (est.)
Date sampled	10/5/79	7/6/76	?	?	6/6/15

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a. Water-quality analysis file, USGS central laboratory, Denver, CO, collected by C.E. Sloan.
b. Mariner and others (1978).
c. USGS, lab 4683, 1958.
d. Waring (1917).
e. Analyses performed on samples filtered through 0.05-micron filter.
f. Combined flow, all springs.

Table 14. Bell Island Hot Springs geothermometry.
(All temperatures are in degrees celsius;
geothermometry is based on DGGS water-chemistry given in table 14.)

Surface temperature (measured)	74
Cation geothermometers	
Na-K-Ca (1/3)	144
Na-K-Ca (4/3)	117
Silica geothermometers	
Adiabatic	137
Conductive	142
Chalcedony	115
Cristobalite	91
Opal	21

Temperatures from the quartz mixing models must be used with caution. No corroborative evidence yet exists for mixing. Information is insufficient to determine spring temperature vs chloride-concentration trends, and no information is available on the isotopic composition of either the thermal or the cold waters. Further, the silica content and temperature of the cold spring water used in the analysis may not be representative of waters actually mixing with thermal waters.

Geophysical exploration at Bell Island has been limited to shallow seismic refraction investigations; the extent of a deep subsurface reservoir is not known. Using the standard mean reservoir volume and the mean reservoir temperature of 134°C gives the following estimates of stored thermal energy:

Volume (km^3) = 3.3; std dev = 0.9
Thermal energy = (10^{18} J) 1.06; std dev = 0.31

These values must be viewed as highly speculative estimates.

Comments

The hot springs probably result from deep circulation of meteoric waters along fractures and possibly faults associated with the Bell Island lineament. The caretaker at Bell Island resort reported that other hot springs exist and lie along the lineament northeast of the resort springs and southwest in Blind Passage. Despite an intensive field search the existence of these springs could not be verified. An aerial reconnaissance during winter might aid in locating vapor plumes from these reported springs.

Some of the ascending thermal water may be undetected, discharging beneath the surface into the sediments overlying bedrock SW of the spring site. Some hot springs may also be submerged beneath the sea or under lakes that lie along the lineament.

Geothermometry indicates the reservoir temperatures are below the minimum value required for generation of electrical power. If subsurface mixing of colder waters is occurring, however, mixing models suggest reservoir temperatures may exceed 150°C, which could be enough to generate electricity.

The surface resource is presently being used for heating but much of the water is being discharged into the sea at temperatures of 40-50°C. Additional cascaded uses for the resource are possible, perhaps for greenhouses or aquaculture. The present owner of Bell Island Hot Springs Resort has expressed interest in the increased use of the Bell Island geothermal resources.

10. BAILEY BAY HOT SPRINGS

Location

Latitude 55° 59.0' N., longitude 131° 39.8' W.; Ketchikan D-5 1:63,360 Quadrangle (1952); T. 68 S., sec. 9, SW 1/4 of SW 1/4 of Copper River Meridian.

General Description

The Bailey Bay Hot Springs are located 80 km north of Ketchikan, off Behm Canal and near Bailey Bay (fig. 1). The springs are 0.5 km up from the mouth of Spring Creek, which drains into Lake Shelokum (fig. 7). Lake Shelokum in turn drains via a steep cascading bedrock channel into the head of Bailey Bay. The springs are accessible by boat to the head of Bailey Bay, thence over an unmaintained Forest Service trail for 5 km. Alternately, a float plane can be taken to Lake Shelokum. The springs are within the Tongass National Forest and are managed by the U.S. Forest Service.

Ten principal springs and a number of seeps and smaller springs are dispersed over a 12,000-m^2 area on the steep northwest-facing slope of Spring Creek valley (fig. 8). The springs lie between 112 m (40 ft) to 45 m (150 ft) above Spring Creek and issue from fissures in the granitic bedrock or occur as seeps and pools in the alluvial cover.

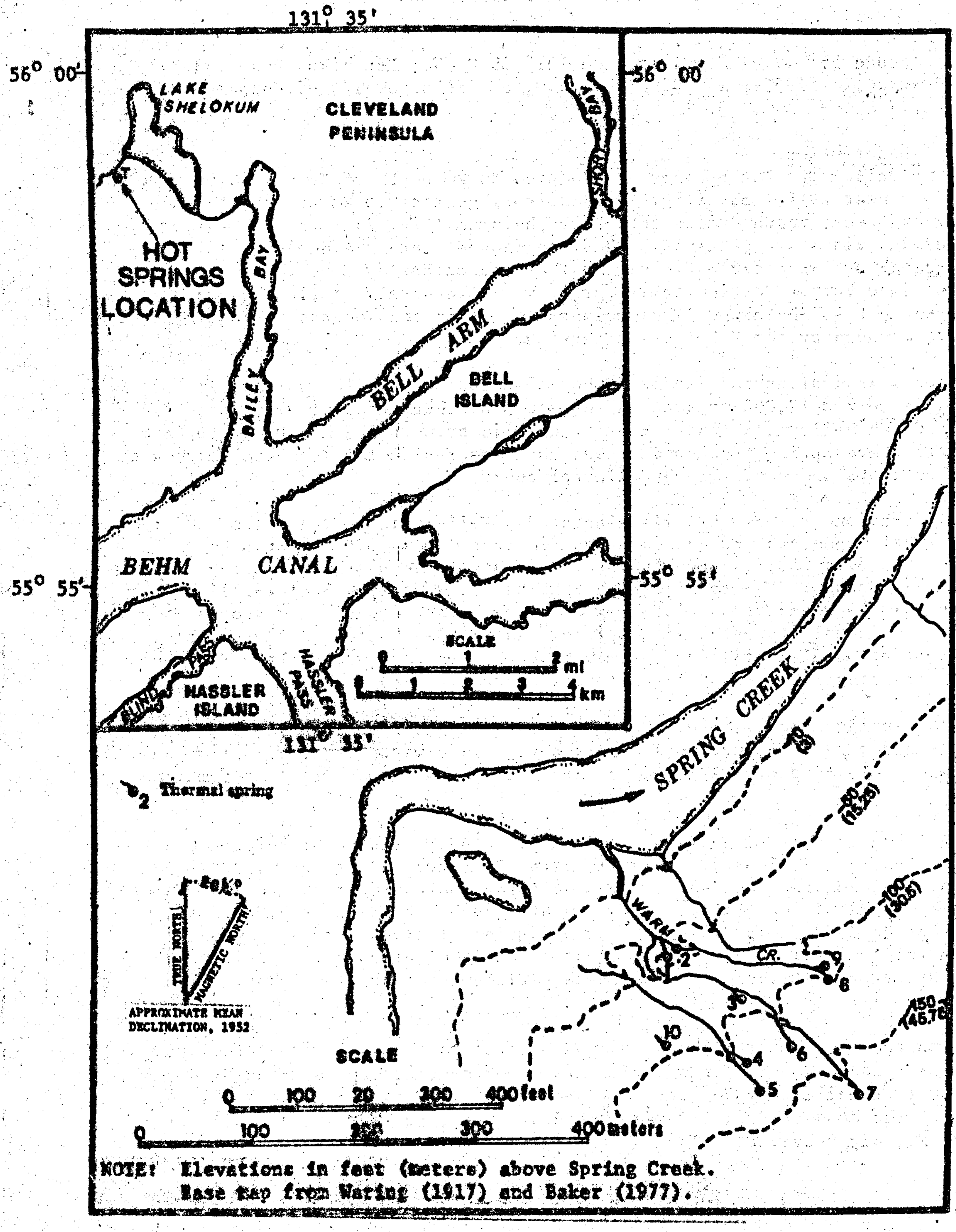
The region has been heavily glaciated. Valley slopes range from 30° to 45°. Local topography ranges in elevation from 106 m (350 ft) at Lake Shelokum to over 730 m (2,400 ft) at the crest of the northwest-facing slope above the springs. The area is highly scenic with steep, glacier-polished, granitic walls remeniscent of Yosemite valley. The soil-covered slopes and valley are densely forested with spruce. The valley bottom is flat as a result of alluvial in-filling and may have once been an arm of Lake Shelokum (Baker and others, 1977). The valley is susceptible to flooding.

The springs were more extensively used during the early part of this century and were once developed as a resort with a wagon trail to Bailey Bay. The resort closed during WWII and a three-sided shelter is all that remains.

Geology

Bedrock in the Bailey Bay area is a lineated, gneissic hornblende-biotite quartz diorite and minor granodiorite (fig. 7) (Berg and others, 1978). The pluton also contains abundant ellipsoidal dark inclusions that are parallel to aligned hornblende crystals. Foliation attitude near Bailey Bay Hot Springs is strike N. 35 W., dip 75° N. Dikes and veins of light-gray weathering quartz-feldspar-(biotite-garnet) pegmatite cut the intrusive. K-Ar ages in the western part of the pluton indicate emplacement took place about 72 to 84 m.y. ago.

Inspection of aerial photos shows the area to be highly fractured with a dominant northeasterly trend. The most prominent fractures trend N. 60° E. Fractures at Bailey Bay Hot Springs have trends ranging from E to N. 60° E. Those at the head of Bailey Bay measured strike N. 60° E., dip 78° N.; strike N. 50° E., dip near-vertical; and strike N-S, dip near-vertical.



NOTE: Elevations in feet (meters) above Spring Creek.
Base map from Waring (1917) and Baker (1977).

Figure 8. Bailey Bay Hot Springs.

A series of aphanitic to fine-grained, quartz porphyry dikes have invaded many of the fractures that trend N. 50° E. to N. 60° E. Ranging 20 to 40 cm in thickness, the dikes are probably related to porphyritic granite stocks and quartz monzonite stocks of Miocene age reported by Berg and others (1978).

Spring Characteristics

Ten principal springs issue from fissures in the granitic bedrock and from pools formed in alluvium covering the steep valley slope (fig. 8). Nine of these springs were reported previously (Waring, 1917; Baker and others, 1977); the tenth spring was found during DGGS investigations. This newly reported spring emerges in a gravel-bottomed pool, located about 50 m SW of Spring 3, and has a temperature of 71°C. The other springs range from 56.0° to 91.5°C (table 15).

Compared with previous measurements, most of the hot springs at Bailey Bay appear to have fluctuated at least a few degrees. These variances in measured temperatures could reflect slight changes in the mixing fraction of colder waters. They could also be due to differences in measuring technique and measurement location. Hot spring water temperatures can decrease several degrees just a few centimeters away from the spring sources. All ADGGS measurements were made by inserting the sensitive portion of the thermal probe either directly into the spring orifice or as close to it as reasonably possible. Probe location was always adjusted to obtain a maximum temperature reading. The 20° to 25°C increases in temperature at springs 4 and 5 since 1914, however (table 15), are much too large to attribute to differences in measurement techniques. These changes are more likely due to a real increase in the temperatures of waters arriving at the surface.

Where possible, flow rates were measured by DGGS using a pygmy flowmeter; otherwise, flow rates were estimated (table 15). Spring discharges ranged from 1 to 70 lpm, with the largest discharges occurring at springs 1, 3, and 5. The combined October 1979 flow for all springs is estimated at 300 lpm ± 15 percent. Given the limited degree of accuracy of flow measurements and estimates, the overall flow rates appears to have remained relatively constant. Flow at three of the individual springs, however, appears to have changed appreciably, with springs 8 and 9 decreasing and spring 7 increasing since 1914.

One of the three sources of spring 3 issues in a spout or jet of hot water which measured about 17 cm in height on October 12, 1979. This small jet of water is probably due to thermal artesian pressure. Previous observations indicate jet height and temperature fluctuates with time (table 16).

Analyses of water chemistry were performed on thermal waters obtained from springs 1, 3 and 5 (table 17). The springs are nearly identical in chemistry and other properties. Some moderate changes appear to have occurred since 1915 but the accuracy of Waring's earlier analysis is not known.

Waring (1917) reported considerable bubbling of gas at Bailey Bay Hot Springs but no bubbling was noticeable in October 1979. Carbonate deposits

Table 15. Temperatures and flow rates, Bailey Bay Hot Springs.

Spring	Temperature °C			Flow, (lpm)		
	10/79[a]	6/77[b]	7/14[c]	10/79[a]	6/77[b]	7/14[c]
1	84.0	84	82	70	75	87
2	83.0	-	81	20	30	34
3a	87.0	85	86		55	53
3b	84.5	84		60	20	
3c	84.5	82			15	
4	86.5	84	63	10	10	15
5	91.5	90	66	50	55	57
6	72.5	72	67	1	4	4
7a	81.0	75	76		4	11
7b	77.5	77		40	15	
7c	56.0	45			4	
8	89.5	89	83	35	65	42
9	88.0	83	88		40	15
10	71.0			10		
			Total[d]:	300 ± 15%	398	318

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a. DGGS, this report.
b. Baker and others, 1977, estimated.
c. Waring, 1917
d. An overall total flow of 274 lpm for springs 1-9 was reported by Sloan, 1976.

Table 16. Bailey Bay Hot Springs, spring-3 water jet height and temperature

Date	Height (cm)	Temperature (°C)	References
1905	38	95	Wright and Wright, 1908
1915	30	86	Waring, 1917
1976	15	90	Ogle, 1976
1977	28	85	Baker and others, 1977
1979	17	87	DGGS, this report

Table 17. Chemical composition and physical properties of Bailey Bay Hot Springs. (Previous USGS analyses included for comparison; all chemical analyses in mg/l.)

	DGGS	DGGS	DGGS	USGS[a]	USGS[b]	USGS[c]
	Spring 1	Spring 3	Spring 5	Spring 3?		Spring 3
SiO_2	139	140	138	150	160	142
Al	<0.01[d]					2.2
Fe	<0.01[d]	<0.01	<0.01	0.01		1.5
Ca	1.3	1.3	1.3	2.1	2.0	13
Mg	0.0	0.0	0.0	2.2	0.0	2.1
Na	101	102	102	89	78	54
K	4.5	4.5	4.4	4.3	4.2	
Li	0.06	0.06	0.06	0.07		
HCO_3	138	132	129	145	118	27
CO_3	4.5	8.0	6.0	0		53
SO_4	41	38	39	43	44	32
Cl	51	51	49	53	47	11
F	3.2	3.2	3.1			
Br	0.0	0.0	0.0			
B	0.4	0.6	0.6	0.33		
pH, field	8.80	8.85	8.85	8.2	7.2	
Dissolved Solids	414	414	407	415	453	
Hardness	3.3	3.3	3.3	14		
Sp conductance	484	482	479	444(lab)		
T °C	84.5	87.5	91.5	90.0	88.	86
Discharge lpm	70	60	50			53
Date Sampled	10/12/79	10/12/79	10/12/79	7/7/76		6/17/15

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[a] Water-quality-analysis file, lab 216065, USGS Central Laboratory, Denver, CO, collected by C.F. Sloan.

[b] Mariner and others, 1978.

[c] Waring, 1917.

[d] Analyses performed on acidified 0.05-micron filtrate.

occur at and near springs 1-5. Many of the spring channels are lined with brilliant and colorful orange-red bacterial mats that change to blue-green algomats where temperatures drop below about 60°C.

The multiplicity of springs at Bailey Bay indicate that single conduits near the surface are too narrow to accommodate the volume of ascending hot water. Channel permeability may have been reduced by deposition of silica and other hydrothermal minerals on the conduit walls as the thermal waters cool on ascent. Some of the thermal water must be discharging beneath the surface, because much of the soil cover in the vicinity is 15°-20°C at only a few cm depth. The spring waters must also go to heating some of the local bedrock; the surface of an area measuring several square meters on a moss-covered rock face directly above spring 5 registered 42°C.

Reservoir Properties:

Table 18 summarizes the application of silica and cation geothermometry to Bailey Bay Hot Springs. The Na-K-Ca (4/3) geothermometer temperature of 142°C indicates subsurface equilibration took place at temperatures well above 100°C and that therefore the Na-K-Ca (1/3) geothermometer temperature of 153°C is applicable. This temperature is very similar to the quartz-conductive temperature of 157°C. The pH levels measured by DGGS for three springs at Bailey Bay Hot Springs are nearly identical and notably higher than that reported by previous investigators. Applying a pH correction to these springs using the DGGS pH values results in a quartz-conductive temperature of 142°C. This temperature is taken as the minimum reservoir temperature, the uncorrected quartz conductive temperature is taken as the maximum, and the Na-K-Ca (1/3) as the most likely:

	Min	Max	Most Likely	Mean	Std Dev
Subsurface T (°C)	142	157	153	151	3.4

The principal hot springs at Bailey Bay issue from the surface at temperatures below boiling and have a large combined flow rate and cation geothermometer temperatures considerably above orifice temperatures. These factors suggest the mixing of colder waters (Fournier and Truesdell, 1974). A cold spring near Bell Island Hot Springs, located in similar terrain several km south of Bailey Bay Hot Springs, was found to have a temperature of 8°C and a silica content of 8 ppm.

Using these parameters to characterize the cold water fraction and following the method of Truesdell and Fournier (1977), application of quartz mixing models gives the following results:

	Parent hot water		
Mixing model	Max T (°C)	SiO^2 (ppm)	Fraction (%)
1. Maximum steam loss	154	135	89
2. No steam loss	225	370	36

Temperatures from the quartz mixing models must be used with caution. No corroborative evidence for mixing such as from chloride-temperature analysis or water oxygen isotope analyses is available. Further, residence time in the granitic rock of cold springs waters which were sampled may have been

Table 18. Bailey Bay Hot Springs geothermometry. (Based on spring 3 water chemistry, DGGS analyses; all temperatures are in degrees celsius.)

Surface temperature (measured)	87
Cation geothermometers	
Na-K-Ca (1/3)	153
Na-K-Ca (4/3)	142
Silica geothermometers	
Adiabatic	149
Conductive	157
Conductive, pH corrected	142
Chalcedony	133
Cristobalite	107
Opal	35

relatively short and not representative of waters actually mixing with the hot water fraction.

Although there is no observable steam loss at the ground surface, steam may be separating below the surface and heating waters other than that emerging at the springs sampled. The large area of heated ground and the similiarity of temperature estimated by the steam loss model to temperatures estimated by geothermometry suggests this may be the case. The high temperature predicted by model 2 seems unlikely for waters circulating in fractures in granitic bedrock.

No geophysical exploration has been done at Bailey and the extent of the subsurface reservoir is not known. Using the standard reservoir volume and the mean reservoir temperature of 150°C gives the following estimates:

Volume (km^3) = 3.3, std dev = 0.9
Thermal energy (10^{18} J) = 1.20; std dev = 0.35

These values must be viewed as highly speculative estimates.

Comments:

Bailey Bay Hot Springs have both the highest surface temperature and highest estimated reservoir temperature of any hot spring site investigated in southern Southeastern Alaska. Surface temperatures and flow rate are sufficient for a variety of applications requiring thermal energy. The estimated reservoir temperature is above that presently required for generation of electrical power.

Bailey Bay Hot Springs probably result from circulation of meteoric waters along deep-seated fractures. Fluctuations in discharge and temperature at individual springs apparently have occurred, as indicated by comparing reported flow and temperature measurements (table 15).

11. SAKS, REPORTED HOT SPRINGS, UNSUBSTANTIATED

Location

(Approximate) latitude 55° 52' N., longitude 131° 05' W.; Ketchikan 1:250,000 Quadrangle (1955); T. 69 S., R. 94 E.

Comments

Relying on information obtained from USFS in 1915, Waring (1917) reported the existence of hot springs located 8 km (5 mi) SE of Saks Cove off Behm Canal (fig. 1). The USFS representative described the springs as similar to Bell Island Hot Springs with "water issuing in part at the rock shore and in part directly from a fissure in granite 200 ft back from the shore" (Waring, 1917, p. 23).

DGGS interviewed several persons familar with the Saks Cove region, including representatives from the USFS and the Alaska Department of Fish and Game, local bush pilots, and local trappers and explorers. Two of the persons interviewed had searched extensively for the springs but could not find any sign of them. None of the others had any knowledge of the springs other than what had been reported by Waring (1917). The springs either probably did not exist at the time of Waring's report and Waring was misinformed or perhaps the springs have dried up since 1915.

12. UNUK, REPORTED THERMAL SPRINGS, UNSUBSTANTIATED

Location

(Approximate) latitude 56°08' N., longitude 131°00' W.; Bradfield Canal 1:250,000 Quadrangle (1955); T. 66 S., R. 93 E.

Comments

Waring (1917) recorded a vague report of thermal springs located about 9 km (6 mi) above the mouth of the Unuk River, which drains into Burroughs Bay (fig. 1). The springs were believed to have a small flow and to be neither very hot nor notably mineralized. Several individuals familiar with the Unuk River were interviewed by DGGS. None had any knowledge of thermal springs along the Unuk River. Carbonate springs located near the international border reported by Waring (1917) as Boundary Springs were reported to be now underwater as a result of a shift in river channel. The thermal springs may have either met a similar fate or perhaps never actually existed.

13. BAKER ISLAND HOT SPRINGS

Location

(Approximate) latitude 55° 17.5'; longitude 133° 40.5'; Craig 1:250,000 Quadrangle (1957), R. 76 E., T. 75 S. of Copper River Meridian.

Comments

A hot-springs site was reported by local fishermen to be located on the west coast of Baker Island (fig. 1). The existence of these springs was recently substantiated by the USFS office in Craig, Alaska. The Forest Service reported that a cluster of springs issue from a steep granitic cliff, several meters above tidewater at the head of Veta Bay (fig. 9). The USFS plans to build a trail to the spring from Port San Antonio on the east side of the island.

Deteriorating weather conditions prevented field investigation of this site during fall, 1979. The site will probably be visited during the 1980 DGGS field season.

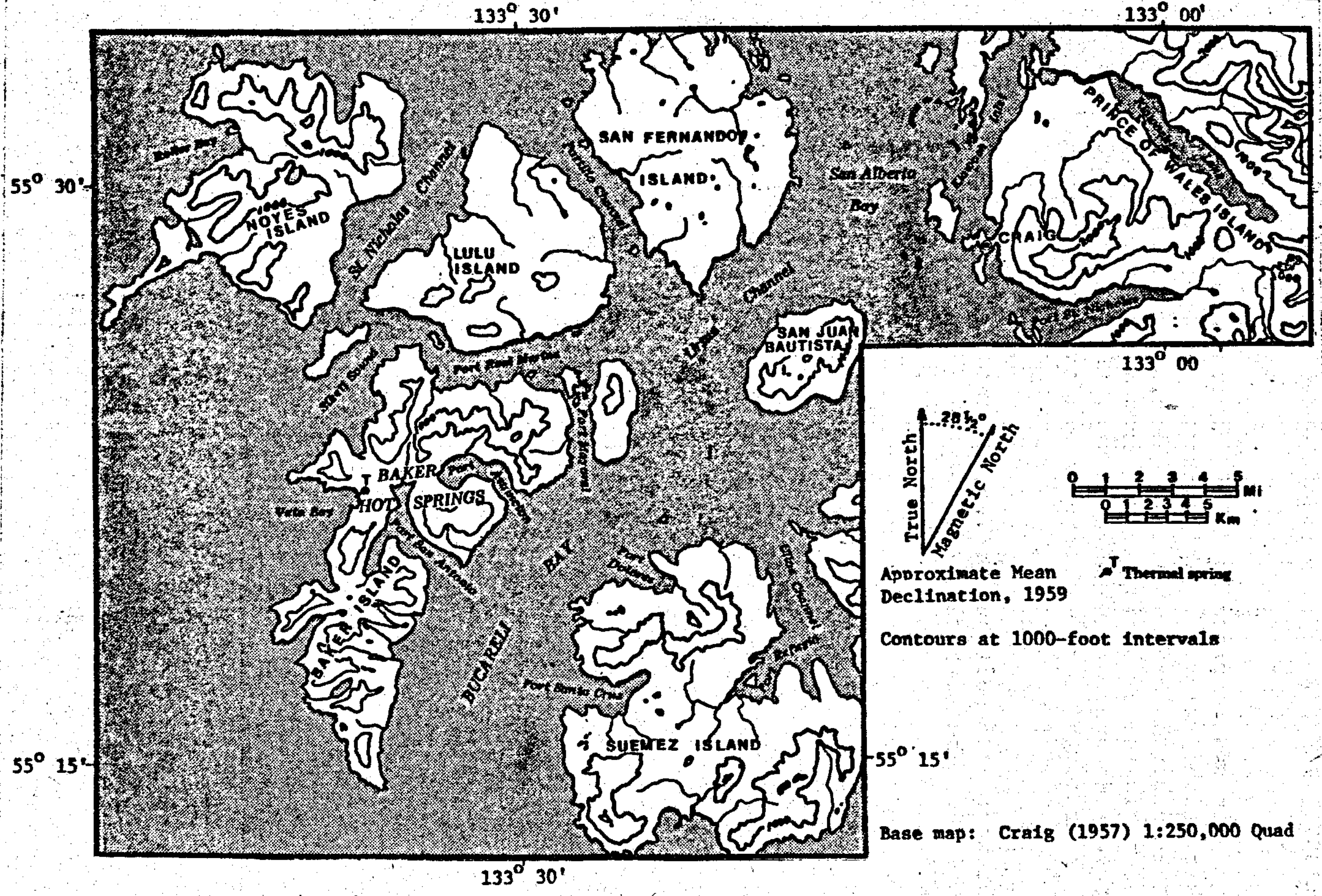

Figure 9. Location map, Baker Island Hot Springs.

SUMMARY, SOUTHERN SOUTHEASTERN ALASKA THERMAL SPRINGS

Locations of the thermal spring sites in southern Southeastern Alaska are shown in figure 1. Table 19 provides a brief summary of each site. Three sites reported in Waring (1917) (Vank Island, Saks Cove, and Unuk River) could not be verified and are presently discounted. These springs may have existed at one time and perhaps have dried up since Waring's account.

The Bailey Bay Hot Springs site (No. 10) has the highest surface water temperature (71° to 91.5°C) and the highest estimated subsurface temperature (151°C) in southern Southeastern Alaska. The largest flow rate (450 lpm) was measured at Chief Shakes Hot Springs (site 1), located in Shakes Slough north of the Stikine River. Bell Island hot springs (site 9) has had the most development. The five principal hot springs there, which have an average temperature of 70°C and a combined flow rate of over 100 lpm, are used by a fishing resort to heat 14 buildings and an olympic-size swimming pool. The Bell Island Hot Springs occur on a prominent southwest-northeast trending, 30 km long lineament that bisects the island.

Two thermal spring sites that had not been previously reported in the literature were investigated during the September-October field excursion. The first site, Barnes Lake warm springs (No. 2), is located about 4 km north of the Stikine River near the Canadian border and consists of a warm pool measuring 25°C and a warm seep measuring 26°C. The second site, Bradfield Canal hot springs (No. 8), occurs on the south bank of the East Fork Bradfield River near the confluence of the two main tributaries to the river. The site consists of several small thermal springs issuing from fissures in the granite bedrock at temperatures ranging from 53°-57°C. Total flow is estimated at 20 lpm.

Except for Bradfield hot springs, all the presently known thermal springs in southern Southeastern Alaska occur either near a major river or near tidewater. This correlation is probably partly a reflection of accessibility and occurrence along travelled routes. The discovery of Bradfield hot springs in a remote area of the Coast Range mountains suggests other such sites exist but have not yet been found. Vapor plumes from thermal springs are most visible during cold weather. A late-winter aerial reconnaissance might disclose the locations of additional thermal spring sites in southern Southeastern Alaska.

Except for Baker Island hot springs, all known thermal springs in southern Southeastern Alaska occur on or near the western flank of the Coast Range Batholith and probably originate from deep circulation of meteoric waters along fractures and faults in granitic host rock.

The chemical compositions of the southern Southeastern Alaskan thermal springs are roughly similar and fall under the category of alkali-chloride to alkali-sulfate waters. The constituents of the thermal waters were probably derived from deep-seated interaction of hot water with rock. Previous chemical analyses are available for three of the thermal-spring sites: Chief Shakes, Bell Island, and Bailey Bay. Concentrations of most of the chemical constituents appear to have undergone at least minor variations. The overall chemistry of the thermal waters from Bell Island and Bailey Bay appear to have remained substantially the same. Two significant changes in chemistry, how-

Table 19. Southern Southeastern Alaska thermal springs sites

Site	No. of springs	Total Flow (lpm)	Spring temperature(°C)	Estimated reservoir temperature (°C)	Remarks
Chief Shakes	7	450	50	91	Springs located in Chief Shakes Slough N. of Stikine River. Springs emerge from fissures at base of granitic cliff.
Barnes Lake	2	30	26	90	Springs located on either side of creek draining into Barnes Lake.
Fowler (Canada)	5	Unknown	59	96	Located in Canada on S. side of Stikine River, about 16 km from Alaska border. Springs emerge from fissures at base of granitic cliff. Beaver-dam impoundment has submerged several springs.
Twin Lakes	2	270	21	54	Springs located on N. shore of Twin Lakes at base of granitic cliff near fracture zone. Mixing of cold waters probable. Springs are probably the W. Shakes Springs mentioned in Waring (1917).
Mt. Rynda	Reported by local inhabitants as warm seep near base of Mt. Rynda on south bank of Stikine River. This spring is probably the South Stikine Spring mentioned in Waring (1917).				
Vank Island	Reported by Waring (1917) to be located on south end of Vank Island but unsubstantiated. Local inhabitants familiar with the island know of no thermal springs on the island.				
Virginia Lake	Reported by local inhabitants of Wrangell, Alaska. Search of designated location did not turn up any thermal anomalies.				
Bradfield Canal	7	20	57	108	Located about 150 m S. of E. Fork Bradfield River. Springs issue from several small fissures in granitic bedrock on hillside.
Bell Island	5 major 1 minor	100	74	134	Located on lineament which bisects Bell Island. Terrain is granitic. Springs used to heat resort.
Bailey	10 major	280	91.5	151	Located on hillside S. of Spring Creek. Springs issue from several fissures in granitic terrain.
Saks Cove	Reported by Waring (1917) but unsubstantiated. Extensive searches by local residents could not locate any sign of springs.				
Unuk	Reported by Waring (1917) but unsubstantiated. Local trappers using river have no knowledge of any spring in the general area indicated by Waring.				
Baker Island	U.S. Forest Service from Craig, Alaska verified existence of hot springs on west side of Baker Island. Several springs were reported issuing from granite cliff several meters above tide water.				

ever, appear to have occurred at Chief Shakes. Waring's 1915 analysis is notably higher in silica and potassium than either the USGS 1976 analysis or the DGGS 1979 analysis. The higher concentrations of these two constituents in 1915 suggest subsurface waters had been equilibrating at substantially higher temperatures than at present. One possible explanation for this change is that the deeper part of the conduit system may have become sealed perhaps as a result of seismic activity or from deposition of hydrothermal minerals.

Fluctuations in hot spring temperatures of 1°-2°C was noted over a several-day period at Bell Island. Winter temperatures of these thermal water can apparently increase by 5°-10°C above summer temperatures.

Most of the Bailey Bay Hot Springs were hotter in October 1979 than in either June 1977 or June 1914. Two of the spring temperatures increased substantially since 1914; spring 4 increased by 23°C and spring 5 by 25°C.

Discharge rates appear to have changed at all three sites for which previous records are available. The accuracy of the earlier measurements is not known. Measurements and estimates of flow rates are often subject to large errors and it is often difficult to judge if any significant changes in discharge have actually occurred. The measured flow rate in September 1979 and the reported flow rate of June 1915 (Waring, 1917) for spring 1 at Chief Shakes are both notably higher than the rate reported for July 1976 (table 5). The overall discharge from springs at Bailey Bay appear to have remained about the same since 1915, although some readjustment of flow volume seems to have occurred among the several available conduits. Waring's visual estimates of discharge from the Bell Island Hot Springs are less than half of the more recent minimum estimates of flow rate, which indicates that either flow has substantially increased or Waring's estimates were wrong.

Chemical geothermometers suggest subsurface temperatures in the range of 50° to 110°C for thermal springs along the Stikine and East Bradfield Rivers. If the thermal-spring waters derive their heat solely from deep circulation, they must reach depths of 1 to 3.5 km, assuming geothermal gradients of 30° to 50°C/km. The estimated subsurface temperatures for Bailey Bay and Bell Hot Springs are in the range at 135°-151°C. Circulation depth for these waters would range from 3 to 5 km.

Interest has been expressed in several possible uses of the thermal springs in southern Southeastern Alaska, including aquaculture and electrical generation. Aside from Bell Island Hot Springs, however, the thermal waters of southern Southeastern Alaska remain unused except for occasional recreational purposes. Thus far, our investigations indicate that only the Bailey Bay and Bell Island Hot Springs sites offer any potential for future large-scale development.

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APPENDIX A: ABBREVIATIONS, UNIT SYMBOLS, AND CONVERSION FACTORS

1. Abbreviations

ADGGS - Alaska Division of Geological and Geophysical Surveys
ICAP - Inductively coupled argon plasma
n.d. - not determined
std dev - standard deviation
USFS - United States Forest Service
USGS - United States Geological Survey

2. Unit symbols

C° - degrees Celsius
cm - centimeter
J - joule
kg - kilogram
km - kilometer
l - liter
lpm - liter per minute
m - meter
mg - milligram
u - microgram
u/l - microgram per liter
mm - millimeter
m.y. - million years
ppm - parts per million

3. Conversion factors

°C - 5/9 °Fahrenheit - 32
°C/km - 5/9 °F-32 /0.621 mile
cm - 0.394 inches
gm - 0.035 ounce
joule - 0.239 calorie (cal)
- 9.480 x $10^{\circ -4}$ British thermal unit (Btu)
10^{18} J - 10^{15} Btu = 1 quad
kg - 2.205 pounds
km - 0.621 mile
l - 0.264 gallon
lpm - 0.264 gallons per minute
m - 3.281 feet
mm - 0.039 inch

APPENDIX B. PRECISION OF WATER ANALYSES
(from Skougstad and others 1979).

Constituent	Number of labs	Mean (mg/1)	Relative deviation (percent)
SiO_2	12	5.87	9
	19	36.5	3
Al	4	.0750	23
	3	.433	9
Fe	17	.100	31
Ca	17	12.6	7
	23	110	8
Mg	23	1.98	9
	20	22.0	5
	17	35.6	17
Na	26	3.44	9
	19	43.7	4
	23	78.8	4
K	15	0.8	14
	32	5.2	11
Li	10	.054	9
	18	.484	5
Cl	10	1.56	26
	9	194	4
F	13	0.62	16
	6	1.1	26
	14	3.8	11
B	9	.072	78
	5	.522	14

APPENDIX C. INDUCTIVELY COUPLED ARGON PLASMA DETERMINATIONS OF MINOR AND TRACE ELEMENTS.

(Analyses performed by Chemical & Geological Laboratories of Alaska)
Blank indicates sample was below detection limit.

Element Detection Limit (mg/l)	Ag	As	Au	Ba	Bi	Cd	Co	Cr	Cu	Hg	Mo	Mn	Ni
	0.05	0.05	0.05	0.05	0.05	0.01	0.05	0.05	0.05	0.05	0.05	0.05	0.05
Location													
Southern S.E. Alaska													
Bailey H.S. #2													
Bailey H.S. #3					0.08								
Bailey H.S. #5													
Barnes Lake													
Bell Island													
Bradfield Canal													
Chief Shakes H.S. #1													
Chief Shakes H.S. #2					0.19								
Fowler													
Twin Lakes													

Element Detection Limit (mg/l)	P	Pb	Pt	Sb	Se	Sn	Sr	Ti	W	V	Zn	Zr
	0.01	0.05	0.05	0.05	0.05	0.05	0.05	0.05	0.05	0.05	0.05	0.05
Location												
Southern S.E. Alaska												
Bailey H.S. #2						0.18	0.07					
Bailey H.S. #3				0.06			0.07				0.14	
Bailey H.S. #5							0.05					
Barnes Lake	0.07						0.11					
Bell Island				0.10			0.40					
Bradfield Canal				0.07			0.46					
Chief Shakes H.S. #1	0.10						0.19					
Chief Shakes H.S. #2	0.08			0.08			0.22					
Fowler				0.08			0.62					
Twin Lakes	0.08			0.09			0.06					