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Assessment of Geothermal Resources of the United States—1975

GEOLOGICAL SURVEY CIRCULAR 726

Prepared in cooperation with the Energy Research and Development Administration

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D. E. White and D. L. Williams, Editors

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CONTENTS

ILLUSTRATIONS

TABLES

Introduction

By D. E. White and D. L. Williams

Over the past 10 years, various individuals and organizations have made estimates of the geothermal resources of the United States (White, 1965; Grossling, 1972; Rex and Howell, 1973; Hickel, 1973; Natl. Petrol. Council, 1973). These estimates have differed by several orders of magnitude. This wide variation has been due in part to lack of or differing assumptions regarding technology and economic conditions, but much of the uncertainty has been caused by an inadequate understanding of the nature and extent of the resources themselves.

Although future technology and economic conditions continue to be difficult to predict, considerable progress has been made during the past few years toward a better understanding of the resources—enough to improve significantly the basis for a comprehensive assessment of the magnitude, distribution, and recoverability of various categories of geothermal resources within the United States. The new assessment presented here is appropriate in light of this improved understanding and is designed to help Government and private industry to evaluate the present and future significance of these resources during their consideration of problems and opportunities for the development of geothermal energy.

Although these new resources estimates rest on a much improved scientific base, they should not be taken as final appraisals that are valid indefinitely into the future. They are limited by the data available to us early in 1975 and will need to be revised at appropriate times as more data and better methods of evaluation become available. A format has been chosen that is amenable to revision.

This assessment of geothermal resources differs from previous resource statements in that it tabulates each identified system or subdivision of the major resource categories of table 1, listing the parameters assumed in calculating volumes, heat contents, and recoverabilities. As technology, economic conditions, and our knowledge of specific systems change, any of the variables may be changed to arrive at updated assessments.

Resource-related terms used in this circular are defined as follows:

Geothermal resource base includes all of the stored heat above 15°C to 10 km depth in all 50 states.

Geothermal resources are defined as stored heat, both identified and undiscovered, that is recoverable using current or near-current technology, regardless of cost. Geothermal resources are further divided into three categories based on cost of recovery:

(1) *Submarginal geothermal resources*, recoverable only at a cost that is more than two times the current price of competitive energy systems;

(2) *Paramarginal geothermal resources*, recoverable at a cost between one and two times the current price of competitive energy; and

(3) *Geothermal reserves*, consisting of those identified resources recoverable at a cost that is competitive now with other commercial energy sources.

Undiscovered resources that are economically recoverable are not differentiated in this report but would be the economic equivalent of reserves.

The first sections of this assessment of geothermal resources include our estimates of the total resource base, subdivided into categories of geothermal systems (table 1). The different categories are related to the fact that, although temperature generally increases with depth below the surface of the Earth, the relative rates of increase differ from place to place, depending on the local subsurface geology and hydrology. Depth below ground surface is an important factor in determining the cost of utilization, so we have arbitrarily limited our calculations of the resource base to two depth ranges: surface to 3 km (roughly the maximum depth yet drilled for geothermal energy) and from 3 to 10 km (roughly the maximum depth yet drilled in search for oil and gas).

The hydrothermal convection systems of categories 1a and 1b(1) (table 1) include all areas that are presently being utilized or explored for generation of electricity; the systems of category 1b(2) are attractive for nonelectrical space and process heating. The relatively high favorability of these systems is due to the fact that convection of water (or steam) transfers heat from the hot deep parts of a system to its near-surface parts. Thus, higher temperatures are attainable in shallower wells at lower costs than in normal temperature-gradient areas where heat is transferred dominantly, if not entirely, by conduction through solid rock. The presence of steam or water as a natural working fluid is an additional major advantage of the convection systems, but the locally available quantity may not be adequate for production on a commercial scale. Systems with assumed maximum temperatures below 90°C are omitted from our tabulations because adequate data are generally lacking, but many are listed by Waring (1965).

The hot igneous systems of category 2 provide challenges for major future utilization. Of all categories, those with molten magma at temperatures above 650°C (and in part as high as ~1,200°C, depending on type of magma) contain the most stored heat per unit of volume or mass; however, the technological problems of utilizing this heat are the most difficult for all categories. These large young magma systems are especially attractive targets for exploration, and their existence accounts for many of the known highly favorable hydrothermal convection systems. The hot dry rocks of category 2b may not always be as hot or as dry at shallow and intermediate depths as generally envisioned because hydrothermal convection in natural fractures may be more common than is indicated from surface evidence.

Category 3 includes the conduction-dominated regions that underlie most of the United States and that constitute by far the largest part of the total resource base. The volumes of rock involved are huge, but average temperatures are low. Convection of natural fluids is a minor factor in many of these conduction-dominated areas. Of special note in this category is the geopressured environment of the gulf coast; the term "fluid resource base" denotes only the fluid part of the total geothermal resource base of the gulf coast.

The later sections of this assessment consider the parts of the resource base that may be available for recovery and utilization, depending on various physical, economic, and technological constraints. Reliable estimates of recoverability are more difficult to derive than are the estimates of the resource base because of the great uncertainties of future prices, technologies, and governmental and environmental considerations and the natural physical properties of the deep subsurface.

National parks have been given special status in this geothermal assessment. Some national parks and national monuments were created because of their young or active volcanism or because of their spectacular geysers and hot springs, which are the surface evidence of high-temperature convection systems (Yellowstone, Lassen, Crater Lake, Mount Rainier, Hawaii Volcanoes, and Katmai). The parks are included in estimates of the resource base (they are in the United States, and they do have much heat in the ground), but they have been excluded from all considerations of recoverability and utilization. Geothermal exploitation of Wairakei, New Zealand, and Beowawe Geysers, Nevada, has diverted the natural water supply into wells and has destroyed the hot springs and geysers. These hot-water systems clearly cannot sustain both natural activity and exploitation. The vapor-dominated systems (Larderello, Italy, and The Geysers, California) are not so immediately sensitive, but long-range destruction of the natural activity by exploitation is equally certain.

Conversion factors from the metric system of units to English units and explanations of some

Table 1.—Categories of geothermal resource base (heat in the ground at temperatures above 15° C to specified depths and without regard for recoverability)

	Temperature Characteristics	Natural fluid supply
1. Hydrothermal convection systems (relatively high temperatures at shallow depths; heat content estimated only to 3 km depth; see Renner and others, this volume).		
a. Vapor-dominated systems	∿240°C	Available; not always adequate.
b. Hot water systems		
(1) High-temperature systems	>150°C	Available; not always adequate.
(2) Intermediate-temperature systems	150°C to ∿90°C	Available; not always adequate.
(3) Low-temperature systems (not tabulated; many in Waring, 1965)	<90°C	Available; not always adequate.
2. Hot igneous systems (excluding hydrothermal convection systems in (1) above; heat content estimated from 0 to 10 km depth; see Smith and Shaw, this volume).		
a. Assumed part still molten	>650°C	Inadequate.
b. Assumed not molten but very hot ("hot dry rocks")	<650°C	Generally inadequate.
3. Conduction-dominated areas (by heat-flow provinces, utilizing available data on heat flows, radiogenic heat production, and thermal conductivity of rocks; heat contents estimated for 0 and 3 and 3 to 10 km depth; see Diment and others, this volume. This category includes the Gulf Coast geopressured environment with its fluid fraction specially considered by Papadopulos and others, this volume).	15° to ∿300°C	Adequate in parts of sedimentary basins, generally inadequate elsewhere.

terms in this report are shown in table 2.

ACKNOWLEDGMENT

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Table 2.—Metric units used in this volume, conversion factors to other units, and some assumed values for physical parameters.

Length: 1 meter (m) = 3.281 ft; 1 kilometer (km) = 3,281 ft = 0.6214 mi; 1 centimeter (cm) = 0.3937 in. = 6.214×10^{-6} mi.

Area: 1 km^2 = $10^6 m^2$ = 0.3861 mi^2 = 247.1 acres.

Volume: 1 km^3 = 0.239 mi^3 = 10^{12}ℓ; 1 liter (ℓ) = 0.2642 gal; 1 ℓ/min = 5.886×10^{-4} ft^3/sec.

Temperature: °C = 5/9 (°F - 32); 0°C = 235.15°K.

Temperature gradient: 1°C/km = 10^{-3}°C/m rate of increase in temperature with depth: conductive gradient is directly proportional to heat flow and inversely proportional to thermal conductivity of the rocks.

Pressure: 1 bar = 0.9869 atm = 1.020 kg/cm^2 = 14.50 psi = 10^6 dynes/cm^2 = 0.1 meganewtons/m^2. All pressures absolute, with 1.01 bar added to gage pressure at sea level and geothermal areas at low altitudes.

Heat/power: 1 cal = 4.186 joules = 3.9685×10^{-3} BTU = 0.001 kcal = 0.00116 watt h; 1 cal/g = 1.80 BTU/lb. 1 MW (electric)·century = 7.53×10^{14} cal (thermal)/e_c, where e_c is conversion efficiency.

Coal assumed to have a potential heat content of 7.2×10^3 cal/g. A barrel of petroleum (42 gal) assumed to have potential heat of combustion of 1.45×10^9 cal = 5.8×10^6 BTU = 0.223 short tons coal. In this volume, heat contents stated in units of 10^{18} cal, with each unit equivalent in heat content of 690 million barrels of petroleum or 154 million short tons of coal.

Heat flow: 1×10^{-6} cal/cm^2 sec = 4.19×10^{-2} W/m^2 (watts per sq. meter); the world-wide average conductive heat flow is approximately 1.5×10^{-6} cal/cm^2 sec

Thermal conductivity: 1×10^{-3} cal/cm sec °C = 0.418 W/m°K.

Mass: 1 g = 10^{-3} kg = 10^{-6} metric ton = 2.20×10^{-3} lb = 1.103×10^{-6} short ton.

Volumetric specific heat of pure water at standard temperature and pressure is 1.0 cal/cm^3°C and of average rocks, assumed 0.6 cal/cm^3/°C. Heat in granite magma at 900°C, crystallizing and cooling to ~15°C assumed to release ~300 cal/g or ~7×10^{17} cal/km^3; equivalent heat in molten basalt at 1,100°C is ~375 cal/g.

Hydrothermal Convection Systems

By J. L. Renner, D. E. White, and D. L. Williams

In hydrothermal convection systems, most of the heat is transferred by the convective circulation of water or steam rather than by thermal conduction through solid rocks. Convection occurs in rocks of adequate permeability because of the buoyancy effect of heating and consequent thermal expansion of fluids in a gravity field. The heated fluid tends to rise, and the more dense, cooler fluid tends to descend elsewhere in the system. Convection, by its nature, tends to increase temperatures at higher levels as temperatures at lower levels decrease below those that would otherwise exist.

Worldwide experience gained from geothermal exploration of hydrothermal convection systems indicates that most systems contain liquid water as the dominant pressure-controlling fluid in fractures and pores. Wells drilled into such systems normally deliver at the wellhead a mixture of liquid water and 10 to 30 percent of steam, which forms in the well bore as pressures decrease upward. In a few systems, however, such as Larderello, Italy, and The Geysers, California, wells produce saturated or even superheated steam, typically with no associated liquid. Moreover, in-hole pressures measured in shut-in wells of these systems normally increase only slightly with depth within the reservoir; the increase in pressure is equivalent to that of a column of steam and associated gases and is much less than the pressure gradient in a column of water. Pressures in these relatively rare systems evidently are controlled by vapor rather than by liquid, and thus the systems are called vapor-dominated systems.

VAPOR-DOMINATED SYSTEMS

There is still divided opinion on the origin and fundamental characteristics of vapor-dominated geothermal systems and on why they differ so much in their production characteristics from the more abundant hot-water systems (Truesdell and White, 1973). All successful wells in the Geysers field, the outstanding example of this type of system in the United States, produce saturated or slightly superheated steam containing little or no liquid water and only a small percentage of other gases. Some successful wells initially discharge some water that dries up to pure vapor with time. In-hole temperatures prior to much production tend to be close to 240°C if reservoir depths are greater than about 400 m; initial wellhead pressures are close to 34 bars (James, 1968; Ramey, 1970; White and others, 1971). These characteristics are generally accepted as typical of the deeper "virgin" parts of The Geysers, Larderello, Italy, and Matsukawa, Japan.[1]

The stored heat of the reservoir rocks is probably 85 percent or more of the total heat in the vapor-dominated systems (Truesdell and White, 1973). Production of steam from a reservoir results in a decline in pressures; consequently, water in the pores boils to steam, utilizing heat stored in the reservoir rocks.

Many aspects of vapor-dominated systems are

[1] Other types of vapor-dominated systems exist, such as those near Monte Amiata, Italy (lower in temperature and much higher in gases other than steam; White, 1973, p. 87, 88; Truesdell and White, 1973), and those found in shallow regimes between ground surface and the water table under local topographic highs of hot-water systems. But in this report, the term "vapor dominated" refers to high-temperature low-gas systems such as The Geysers and Larderello.

not well understood, and critical observations within and below the reservoirs either have not been made, or the data have not yet been released by the operating companies. Our interpretations, however, favor steam as the continuous pressure-controlling fluid in the reservoir, but with liquid water being locally available in small pore spaces and on fracture surfaces. Because of surface tension, this water cannot be drained completely by gravity. Below the vapor-dominated reservoir, we envision a deep water table with underlying rocks saturated with water, probably a high-chloride brine (Truesdell and White, 1973). Estimates of reserves and resources of vapor-dominated systems (Nathenson and Muffler, this circular) are based on this model.

Vapor-dominated systems are considered to develop initially from hot-water systems that have a very large supply of heat but a very low rate of recharge of new water. If the heat supply of a developing system becomes great enough to boil off more water than can be replaced by recharge, a vapor-dominated system starts to form. The fraction of discharged fluid that exceeds recharge is supplied from water previously stored in large fractures and pore spaces. Heat, supplied by condensation of rising steam, is conducted outward from the near-surface, nearly impermeable margins of the reservoir and thus accounts for the high conductive heat flows of these systems. The liquid condensate is in excess of the liquid that can be retained by surface tension; the excess drains downward under gravity to the hypothesized deep water table where it is available for recycling along with newly recharged water.

Our model requires that fluid in excess of that provided by recharging water must be discharged from the system. This feature has important consequences, if true, in that it requires identifiable vent areas. A small vapor-dominated system perhaps could discharge some steam and other gases into surrounding liquid-saturated ground with no conspicuous surface evidence for its existence, but we are skeptical that a large system with high total heat flow and high rate of discharge of steam and other gases can remain concealed without developing the prominent vent areas that characterize all known vapor-dominated systems of this type. The low-temperature, high-gas systems similar to Monte Amiata, Italy (White, 1973, p. 86–87), probably have impermeable cap rocks and little or no surface evidence. Such systems can be considered as thermal natural-gas fields that are high in CO_2 and H_2S, relatively low in temperature, and at least in part characterized by water drive.

Identified systems

The Geysers, California, is the only example of a large vapor-dominated system extensively drilled in the United States (table 3). The extent of the field is not yet known, but the drilling pattern established by more than 100 wells suggests that the commercial limits may have been attained a little northwest of the Sulphur Bank section (about 2 km northwest of the first producing wells at The Geysers). Step-out wells have shown the field to extend at least 3½ km north and 2½ km southwest of the first wells. Drilling is not yet complete to the southeast, but a belt 2 to 5 km wide, 15 km long, and about 70 km² in total area is our present estimate of the extent of the field. Most commercial wells are 1¼ to 2½ km deep, ranging from about 0.2 km in some of the early wells to a present maximum near 3 km. The heat reservoir is assumed to be continuous between 1 and 3 km in depth; thus, its assumed volume is 140 km³. If the average temperature is 240°C, as we assume, then the estimated total heat content is 18.9×10^{18} cal.

The Mud Volcano system in Yellowstone Park was first recognized by its surface characteristics and geochemistry as a probable vapor-dominated system and later confirmed by a single research drill hole (White and others, 1971). The area of surface activity is about 5 km². Resistivity data (Zohdy and others, 1973) suggest that the vapor-dominated part extends to a depth of 1 to 1½ km and is underlain by a better electrical conductor, presumably a deep water table. The vapor-dominated part is assumed to extend from 0.2 to 1.5 km in depth, and its calculated volume is 6.5 km³. If its average temperature is 230°C, then its estimated heat content is $\sim 0.8 \times 10^{18}$ cal.

Outlook for new discoveries

All recognized vapor-dominated systems of the Larderello type are characterized by prominent vent areas with bleached rocks, scanty vegetation, acid-sulfate springs, and no closely associated chloride waters. If these systems do require such vent areas, then few similar unrecognized systems exist for future discovery. The principal possibilities known to us are in Yellowstone National Park and Mount Lassen Volcanic National Park.

Yellowstone Park includes several possible systems other than the Mud Volcano system. The rather young sinter of the Mud Volcano system (White and others, 1971) indicates evolution from a hot-water system soon after the last glacial stage (about 10,000 years ago). This evidence, combined with the resistivity data that suggest a relatively small system saturated with water at depths below about 1½ km, implies a still-evolving system. During the last glacial stage, thick glacial ice and consequent deep meltwater lakes over the thermal areas may have provided high water pressures that resulted in much recharge down the present discharge channels, thereby insuring a water-saturated system. Thus, a vapor-dominated system may become a hot-water system during glaciation. If this is so, then other systems in Yellowstone Park may also have shallow vapor-dominated reservoirs that are still developing.

The thermal activity within the boundaries of Mount Lassen Volcanic National Park has the characteristics of vapor-dominated systems, with chloride waters being completely absent. However, the Morgan Spring group, just outside of the park and about 8 km south of the thermal activity in the park, is a high-temperature chloride-water system that discharges at an altitude of ½ to 1 km below the surface springs in the park. Morgan Springs may be draining the deep chloride part of a large vapor-dominated system within the park.

HOT-WATER SYSTEMS

General characteristics

Hot-water systems (White, 1973) are dominated by circulating liquid, which transfers most of the heat and largely controls subsurface pressures (in contrast to vapor-dominated systems). However, some vapor may be present, generally as bubbles dispersed in the water of the shallow low-pressure parts of these systems.

Most known hot-water systems are characterized by hot springs that discharge at the surface. These springs, through their chemical composition, areal distribution, and associated hydrothermal alteration, have provided very useful evidence on probable subsurface temperatures, volumes, and heat contents. However, springs cannot discharge from convection systems that are capped by impermeable rocks or that exist where the local water table is below the ground surface. Both of these exceptions exist, and many other examples are likely to be discovered.

The temperatures of hot-water systems range from slightly above ambient to about 360°C in the Salton Sea system and the nearby Cerro Prieto system of Mexico. For convenience in this assessment, hot-water convection systems are divided into three temperature ranges: (1) above 150°C (table 4 and figs. 1 and 2); these systems may be considered for generation of electricity; (2) from 90°C to 150°C (table 5 and figs. 2 and 3); these systems are attractive for space and process heating; and (3) below 90°C (not tabulated); these systems are likely to be utilized for heat only in locally favorable circumstances in the United States.

Direct temperature measurements are made either in surface springs or in wells. The temperatures of springs generally do not exceed the boiling temperature at existing air pressure (100°C at sea level to 93°C for pure water at an altitude of ~2,200 m), although some springs in Yellowstone Park and elsewhere are superheated by 1° to 2°C. At depth in wells, where pressures are much higher, the boiling temperature is also much higher. Wells that tap water initially at temperatures above surface boiling yield a mixture of water and steam ("flash" steam), with proportions depending mainly on the initial water temperature and the pressure in the steam-water separator. For example, water flashed from 300°C to a separator pressure of 4.46 bars (50 lb/in²), near a common operating pressure, yields 33 percent steam; 200°C yields 11 percent, but 150°C (just at boiling for the pressure) yields none (Muffler, 1973, p. 255, fig. 28). Obviously the favorability of a hot-water system for generation of electricity from flashed steam increases rapidly above 150°C. Binary systems may allow utilization of somewhat lower temperatures for generation of electricity.

The waters of these systems range from very low salinity to brines of extreme salinity. The most common range is from 0.1 to 1 percent

(Text resumes on p. 51)

Table 3.—Identified vapor-dominated systems of the United States

	Location		Temperatures °C			
Name	Lati-tude ° ' N	Longi-tude ° ' W	Sur-face 1/	Geochemical 2/ SiO_2	2/ Na-K-Ca	Sub-sur-face 3/
The Geysers, CA	38 48	122 48	101	(not applicable)		∿240
Mt. Lassen Nat'l Park, CA	40 26	121 26	95½	(not applicable)		∿240
Mud Volcano system Yellowstone Nat'l Park, Wyoming	44 37.5	110 26	∿90	(not applicable)		∿230
Totals for 3 systems						

Note: Yellowstone and Mt. Lassen National Parks permanently withdrawn from exploitation.

1/Maximum surface temperature reported from a spring or well.
2/Predicted using geothermometers, assuming last equilibration in the reservoir.
3/Average reservoir temperature based on geothermometry unless otherwise noted in comments.

with probable subsurface temperatures exceeding 200 °C

Reservoir Assumptions				Comments
Subsurface area km^2 4/	Thickness km 5/	Volume km^3 6/	Heat content 10^{18} cal 7/	
70	2.0	140	18.9	Area may range from 50 to 100 km^2; bottom of reservoir may extend below assumed -3 km. >100 well's drilled by early 1975. Present heat production ∿80 times estimated natural heat flow.
∿47	1.0	47	6.3	Likely to be a vapor-dominated system but not confirmed.
5	1.3	6.5	0.8	Reservoir assumed ∿0.2 to 1.5 km thickness underlain by hot-water system indicated by resistivity survey.
∿122		∿194	∿26	

4/From surface manifestations, geophysical data, well records and geologic inference. Assume ∿1.5 km^2 if no data pertinent to size is available.
5/Top assumed at 1.5 km of no data on depth available. Bottom assumed to be-3 km for all systems.
6/Calculated from area and thickness.
7/Calculated as product of assumed volume, volumetric specific heat of 0.6 cal/cm^3°C, and temperature in degrees above mean annual surface temperature (assumed to be 15°C).

Table 4.—Identified hot-water convection systems

Name	Location		Temperatures °C			
	Latitude ° ' N	Longitude ° ' W	Surface 1/	Geochemical 2/ SiO_2	Geochemical 2/ Na-K-Ca	Subsurface 3/
ALASKA						
Geyser Bight	53 13	168 28	100	210	236	210
Hot Springs Cove	53 14	168 21	89	131	154	155
Shakes Springs	56 43	132 02	52	142	175	155
Hot Springs Bay	54 10	165 50	83	152	179	180
ARIZONA						
Power Ranch Wells	33 17.1	111 41.2				180

1/ Maximum surface temperature reported from a spring or fumarole.
2/ Predicted using chemical geothermometers, assuming last equilibration in the reservoir; assumes saturation of SiO_2 with respect to quartz, and no loss of Ca from calcite deposition.
3/ Assumed average reservoir temperature based on data presently available.
4/ From surface manifestations, geophysical data, well records and geologic inference. Assumes 1.5 km^2 if no data pertinent to size is available.
5/ Top assumed at depth of 1.5 km if no data available. Bottom assumed at 3 km depth for all convection systems.
6/ Calculated from assumed area and thickness.
7/ Calculated as product of assumed volume, volumetric specific heat of 0.6 cal/cm^3 °C, and temperature in degrees C above 15°C.

with indicated subsurface temperatures above 150 °C

Reservoir Assumptions				
Subsurface area km² 4/	Thickness km 5/	Volume km³ 6/	Heat content 10^{18} cal 7/	Comments
4	2	8	.9	22 springs and geysers in 3 thermal areas in 2 km long zone, near Okmok Caldera; siliceous sinter deposit.
2	2	4	.3	Hot springs and geysers in area about 1 km² near Okmok caldera.
1.5	1.5	2.25	.2	Several springs discharging ∿380 lpm; chemical data not reliable.
1.5	1.5	2.25	.2	Hot springs and fumaroles on active Akutan volcano.
2.5(?)	1	2.5	.2	No natural springs; two wells ∿1 km apart drilled to 3 km depth with bottom-hole temperatures of 163°C and 184°C; discharge estimated 19,000 l/min. from below 2 km.

Table 4.—Identified hot-water convection systems with

Name	Location		Temperatures °C			
	Latitude ° ' N	Longitude ° ' W	Surface 1/	Geochemical SiO2 2/	Geochemical Na-K-Ca 2/	Subsurface 3/
CALIFORNIA						
Surprise Valley	41 40	120 12	97	174	159	175
Morgan Springs	40 23	121 31	95	190	229	210
Sulphur Bank mine	39 01	122 39	80	181	157	185
Calistoga	38 34.9	122 34.4		157	155	160
Skagg's H.S.	38 41.6	123 01.5	57	150	153	155
Long Valley	37 40	118 52	94	219	238	220
Red's Meadow	37 37	119 04.5	49	161	130	165
Coso H.S.	36 03	117 47	95	161	238	220
Sespe H.S.	34 35.7	118 59.9	90	133	155	155
Salton Sea	33 12	115 36	101			340
Brawley	33 01	115 31				200
Heber	32 43	115 31.7				190
East Mesa	32 47	115 15				180
Border	32 44	115 07.6				160

indicated subsurface temperatures above 150 °C—Continued

Reservoir Assumptions				
Subsurface area km² 4/	Thickness km 5/	Volume km³ 6/	Heat content 10^{18} cal 7/	Comments
125	2	250	24	7 spring groups, in area of hydrothermal explosion, 1951; minor sinter, 4 wells drilled; maximum reported 160°C, mixing models as high as 225°C.
5	2	10	1.2	25 springs flowing 350 lpm; and considerable sinter; system may be much larger, if connected to Lassen.
2.5	1.5	3.75	.4	Springs discharging into water-filled open pit of large mercury deposit; 4 wells drilled, reported maximum 182°C.
4.5	2	9	.8	4 hot springs and several flowing wells; spring discharge about 30 lpm.
2	1.5	3	.3	3 springs, flowing 57 lpm.
225	2	450	55	Springs and fumaroles in area of about 10 km². Recent caldera; about 10 wells drilled, reported to 181°C, extensive geology and geophysics.
1.5	1.5	2.25	.2	5 springs flowing 38 lpm.
168	2	336	41	1 group of hot springs; weak fumarole areas; geophysics indicates may be a very large system.
1.5	1.5	2.25	.2	4 hot springs flowing 470 lpm.
54	2	108	21	Many low-temperature seeps; 1 group to 101°C, now under Salton Sea; numerous drill holes to 2400 m and temperatures to 360°C in hypersaline brine.
18	1.5	27	3	No surface discharge, reported high temperature based on old oil test; size based on temperature-gradient survey.
50	2	100	11	No surface discharge; much active exploration but no data released; estimated using temperature gradient data and exploration activity.
28	2	56	5.5	No surface discharge; temperature estimated using drilling data, volume from temperature gradient data and drill-hole data.
3	.6	1.8	0.2	No surface discharge; estimated from temperature gradient data and extrapolation of East Mesa geology.

Table 4.—Identified hot-water convection systems with

Name	Location: Latitude ° ' N	Location: Longitude ° ' W	Temperatures °C: Surface 1/	Geochemical: SiO_2 2/	Geochemical: Na-K-Ca 2/	Subsurface 3/
IDAHO						
Big Creek H.S.	45 18.8	114 19.2	93	160	175	175
Sharkey H.S.	45 00.9	113 51.1	52	135	175	175
Weiser area	44 17.9	117 02.9	77	157	142	160
Crane Creek	44 18.3	116 44.7	92	173	166	180
Near Cambridge	44 34.4	116 40.7	26	119	180	180
Wardrop H.S.	43 23.0	114 55.9	66	120	155	155
Murphy H.S.	42 02.2	115 32.4	51	127	160	160
NEVADA						
Baltazor H.S.	41 55.3	118 42.7	80	165	152	170
Pinto H.S.	41 21	118 47	93	162	176	165
Great Boiling (Gerlach) Springs	40 39.7	119 21.7	86	167	205	170
Hot Sulphur Springs	41 28.2	116 09.0	90	167	184	185
Near Wells	41 10.9	114 59.4	61	140	181	180
Sulphur H.S.	40 35.2	115 17.1	93	183	181	190

Reservoir Assumptions				
Sub-surface area km² 4/	Thickness km 5/	Volume km³ 6/	Heat content 10^{18} cal 7/	Comments
2	1.5	3	.3	15 springs discharging ∿280 lpm and depositing travertine and sinter; mixing model suggests 220°C; few wells.
2	1.5	3	.3	Spring discharging ∿30 lpm; travertine and sinter(?) reported; Na-K-Ca may be inaccurate; mixing temperature 220°C.
35	2	70	6.1	Numerous hot springs and wells; at depth may be connected to Crane Creek. Mixing model indicates possible 228°C.
30	2	60	5.9	Springs discharging ∿200 lpm; extensive sinter, in area of mercury mineralization; Crane Creek and Weiser may be separate in a zone from Midvale, ID to Vale, OR. Mixing model indicates possible 239°C.
1.5	1.5	2.25	.2	Flowing well; Na-K-Ca may be inaccurate.
1.5	1.5	2.25	.2	Numerous springs discharging ∿730 lpm; may be part of a larger system in Camas Prairie; mixing model suggests 160°C.
1.5	1.5	2.25	.2	2 springs discharging ∿260 lpm; mixing model suggests 200°C.
1.5	2	3	.3	Springs discharging 100 lpm; flowing well 90°C, discharging 25 lpm; the area may be large southern extension of Alvord Desert, OR. area.
5	1.5	7.5	.7	Two areas, probably interconnected; 2 springs of eastern area depositing travertine and and discharging 500 lpm; 1 well,western area, flowing 100 lpm. Na-K-Ca may be inaccurate.
10	2.5	25	2.3	2 major groups of springs and 4 others; surface discharge ∿1,000 lpm, calculated total discharge (from heat flow) ∿2040 lpm; well ∿150 m deep, 110°C.
1.5	1.5	2.25	.2	Springs with abundant sulfur.
1.5	1.5	2.25	.2	3 springs discharging 45 lpm; may be part of a more extensive system extending for 4.8 km along the west edge of the Snake Mountains.
4	2.5	10	1.1	Many springs and pools in an area of about .5 km²; abundant sinter.

Table 4.—Identified hot-water convection systems with

Name	Location: Latitude ° ' N	Location: Longitude ° ' W	Temperatures °C: Surface 1/	Geochemical: SiO_2 2/	Geochemical: Na-K-Ca 2/	Subsurface 3/
NEVADA Con.						
Beowawe H.S.	40 34.2	116 34.8		226	242	240
Kyle H.S.	40 24.5	117 52.9	77	161	211	180
Leach H.S.	40 36.2	117 38.7	96	155	176	170
Hot Springs Ranch	40 45.7	117 29.5	85	150	180	180
Jersey Valley H.S.	40 10.7	117 29.4	29	143	182	185
Stillwater area	39 31.3	118 33.1	96	159	140	160
Soda Lake	39 34	118 49	90	165	161	165
Brady H.S.	39 47.2	119 00	98	179		214
Steamboat Springs	39 23.	119 45	96	207	226	210
Wabuska H.S.	39 09.7	119 11	97	145	152	155
Lee H.S.	39 12.6	118 43.4	88	173	162	175
Smith Creek Valley	39 21.4	117 32.8	86	143	157	160

indicated subsurface temperatures above 150°C—Continued

Reservoir Assumptions				
Sub-surface area km^2 4/	Thickness km 5/	Volume km^3 6/	Heat content 10^{18} cal 7/	Comments
21	2	42	5.7	Prior to exploration, about 50 springs and small geysers discharging about 400 lpm from extensive area of sinter deposits; 6 wells drilled up to 600m depth, temperatures to 212°C, 1 deep well but no data available.
1.5	1.5	2.25	.2	Several springs, largest flowing ~20 lpm depositing travertine. Na-K-Ca thermometry may be to high.
4	2.5	10	.9	Several hot springs discharging ~760 lpm; calculated total flow ~900 lpm.
1.5	1.5	2.25	.2	Several springs, largest discharging ~100 lpm and depositing travertine so Na-K-Ca may be inaccurate.
1.5	1.5	2.25	.2	One (3) spring discharging only 20 lpm in area of sinter and travertine; surface temperature low because of low discharge.
10	2.5	25	2.2	No surface springs,but hot wells at least to 115°C; calculated total discharge (from heat flow) ~6,000 lpm.
5	2.5	12.5	1.1	No surface discharge,but small area altered by gases, and 21 km^2 of anomalous heat flow. Shallow wells show 100°C near surface; between 2 recent basaltic eruptive centers.
12	2.5	30	3.6	Several former springs discharged ~200 lpm from small area of sinter; several wells; 214°C reported in 1500 m well; calculated discharge ~2,700 lpm.
6	2.7	16	1.9	About 70 springs discharging ~250 lpm from extensive sinter deposits with ages at least as much as 1 million years, calculated total discharge ~4,300 lpm; more than 20 wells for research, exploration, and spa supply.
1.5	1.5	2.25	.2	Several hot springs of low natural discharge discharge; three wells drilled to maximum of ~670 m, up to 106°C; small area of travertine; area may be larger.
1.5	1.5	2.25	.2	Several springs discharging ~130 lpm from area of sinter.
1.5	1.5	2.25	.2	Several springs, minor travertine.

Table 4.—Identified hot-water convection systems with

Name	Location: Latitude ° ' N	Location: Longitude ° ' W	Temperatures °C: Surface 1/	Temperatures °C: Geochemical SiO_2 2/	Temperatures °C: Geochemical Na-K-Ca 2/	Temperatures °C: Subsurface 3/
NEW MEXICO						
Valles Caldera	35 43	106 32	87			240
Lightning Dock area	32 08.5	108 50	99	156	169	170
OREGON						
Mickey H.S.	42 40.5	118 20.7	73	180	207	210
Alvord H.S.	42 32.6	118 31.6	76	148	199	200
Hot Lake	42 20.1	118 36.0	96	165	176	180
Vale H.S.	43 59.4	117 14.1	73	153	158	160
Neal H.S.	44 01.4	117 27.6	87	173	181	180
Lakeview	42 12.0	120 21.6	96	157	143	160
Crumps Spring	42 15.0	119 53.0	78	173	144	180
Weberg H.S.	44 00	119 38.8	46	125	170	170

Reservoir Assumptions				
Subsurface area km² 4/	Thickness km 5/	Volume km³ 6/	Heat content 10^{18} cal 7/	Comments
65	2	130	18	Pleistocene caldera with 1 group acid-sulfate springs (Sulphur Springs) and very extensive hydrothermal alteration; more than 6 geothermal wells drilled, but no detailed data available; suspected as having small vapor-dominated cap underlain by high-chloride hot-water system with temperatures over 240°C.
1.5	1.5	2.25	.2	No surface springs; shallow water wells at boiling. The area may be much more extensive. Drill hole 3 km to north showed 121°C at 2 km depth. Better estimate may be avg T = 130°C, area 4 km², thickness 2 km, heat content .5 x 10^{18} cal.
6	2	12	1.4	Several springs discharging ∿100 lpm and depositing sinter; surface manifestations over 0.1 km².
3	1.5	4.5	.5	Several springs in area of .5 km² discharging ∿500 lpm. If Hot Lake, Mickey, and Alvord H.S. are one large system with temperature as at Mickey, the heat content would be 30 x 10^{18} cal; three separate systems is preferred model.
6	2	12	1.2	Thermal springs and 1 very large pool (lake) discharging surface manifestations over 0.1 km². Small spring N. of Hot Lake, 98°C.
50	2	100	8.7	Hot springs discharging ∿75 lpm; large area indicated.
2	2	4	.4	1 spring discharging ∿90 lpm.
8	2	16	1.4	About 16 springs including Hunter's and Barry Ranch discharging ∿2,500 lpm in an area of ∿5 km²; several wells at Hunter's for heating spa.
4	2	8	.8	Spring and well (121°C at 505 m) that has erupted as a geyser; discharging 0 to 50 lpm; in small area of sinter.
1.5	1.5	2.25	.2	Hot spring discharging 40 lpm.

Table 4.—Identified hot-water convection systems with

Name	Location		Temperatures °C			
	Latitude ° ' N	Longitude ° ' W	Surface 1/	Geochemical SiO2 2/	Geochemical Na-K-Ca 2/	Subsurface 3/
UTAH						
Roosevelt (McKean) H.S.	38 30	112 50	88	213	283	230
Cove Fort-Sulphurdale	38 36	112 33	--			200
Thermo H.S.	38 11	113 12.2	90	144	200	200
WASHINGTON						
Baker H.S.	48 45.9	121 40.2	42	151	162	165
Gamma H.S.	48 10	121 02	60	161	220	165
Kennedy H.S.	48 07	121 11.7	43	155	199	160
Longmire H.S.	46 45.1	121 48.7	21	169	168	170
Summit Creek (Soda)	46 42.2	121 29.0	13	169	161	170
WYOMING						
Yellowstone National Park	44 36	110 30	~96	250	270	250
Totals (63 systems)						

Reservoir Assumptions				
Sub-surface area km² 4/	Thickness km 5/	Volume km³ 6/	Heat content 10^{18} cal. 7/	Comments
4	2	8	1.0	Hot springs decreasing from 88°C (1908) to 55°C (1957), then ceased discharging from SiO_2 sealing; extensive siliceous sinter; area and volume may be much larger.
15	1.5	22.5	2.5	No springs but active gas seeps; altered areas mined for sulfur; no reliable chemical data; possibly a vapor-dominated system.
1.5	1.5	2.25	.2	16 springs in 2 groups; travertine deposits.
1.5	1.5	2.25	.2	1 (?) spring discharging 26 lpm and possibly depositing calcite.
1.5	1.5	2.25	.2	
1.5	1.5	2.25	.2	4 springs discharging ∿110 lpm, in extensive travertine deposits.
1.5	1.5	2.25	.2	Spring deposits, not identified; in Mt. Ranier National Park; chemical temperatures not reliable.
1.5	1.5	2.25	.2	Chemical temperatures not reliable.
375	2.5	940	133	Numerous thermal phenomena, largely in Yellowstone caldera; individual areas not itemized; total discharge ∿185,000 lpm; 13 research drill holes with maximum T 237.5°C at 332 m; other geochemical and mixing-model T's indicate 330°C.
∿1414		∿2995	∿371	

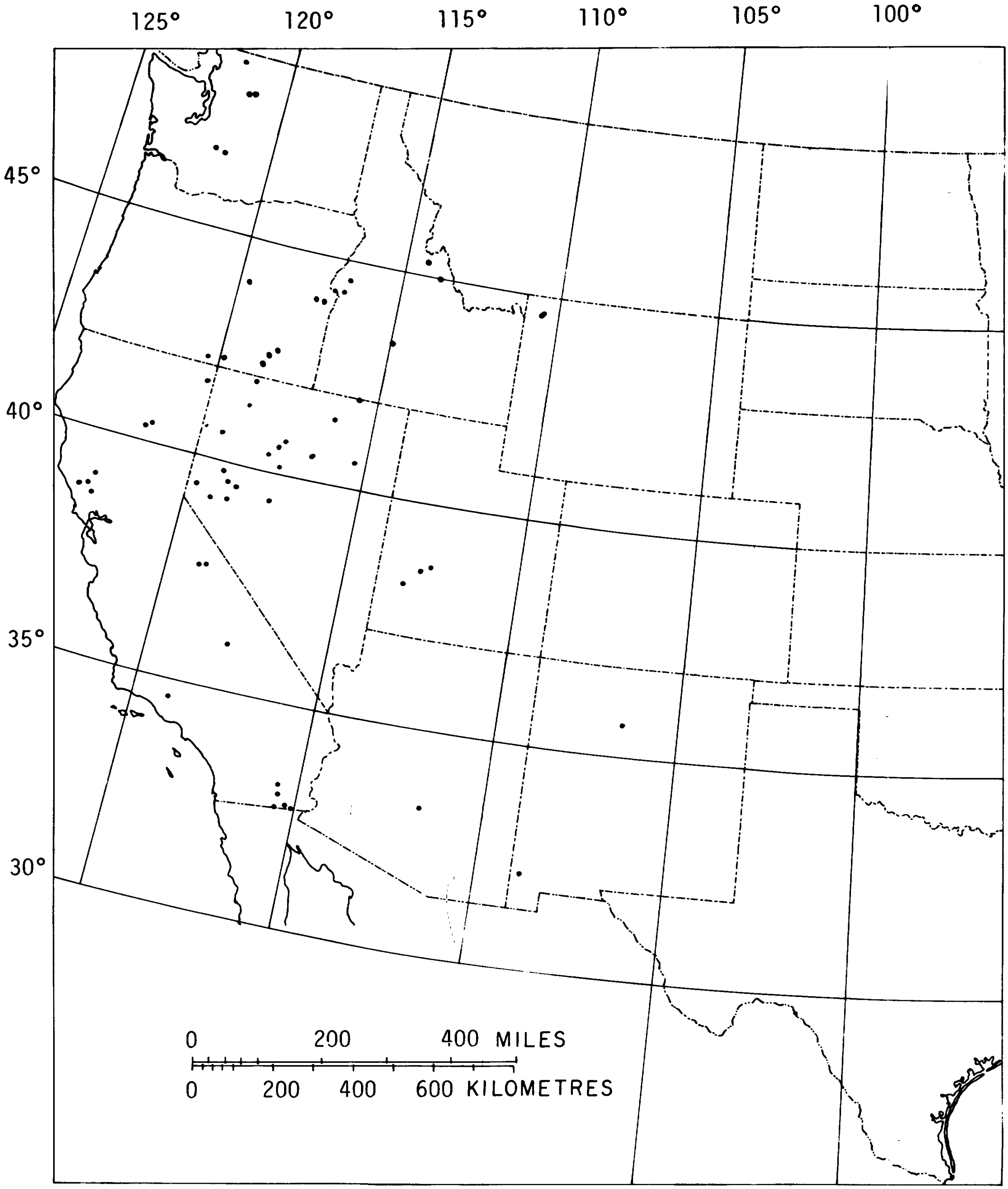

Figure 1.—Location of hydrothermal convection systems in the conterminous United States with indicated subsurface temperatures above 150°C.

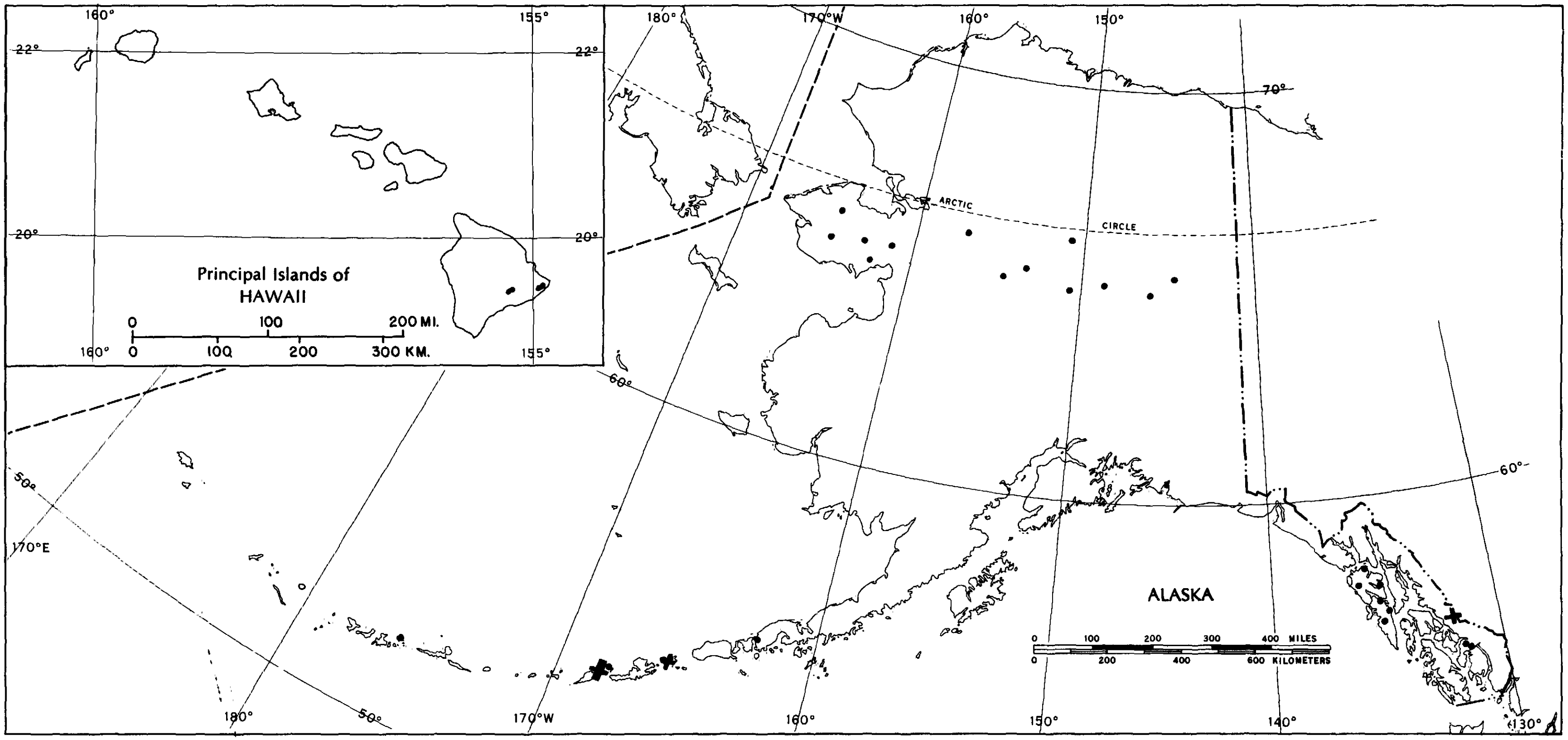

FIGURE 2.—Location of hydrothermal convection systems in Alaska and Hawaii with indicated subsurface temperatures above 150°C (+) and between 90° and 150°C (dots).

Table 5.—Identified hot-water convection systems with

Location			Temperatures °C			
Name	Latitude ° ' N	Longitude ° ' W	Surface 1/	Geochemical SiO₂ 2/	Geochemical Na-K-Ca 2/	Subsurface 3/
ALASKA						
Okmok caldera	53 29	168 06	100	110	75	125
Great Sitkin Is.	52 04	176 05	99			125
Pilgrim H.S.	65 06	164 55	88	137	146	150
Serpentine Sprs.	65 51	164 42	77	132	161	140
Near Lava Creek	65 13	162 54	65	128	91	130
Clear Creek	64 51	162 18	67	119	83	125
Granite Mtn. (Sweepstakes)	65 22	161 15	49	122	75	130
South	66 09	157 07	50	115	72	120
Melozi H.S.	65 08	154 40	55	124		130
Little Melozitna	65 28	153 19	38	126		130
Kanuti	66 20	150 48	66		136	140
Manley (Baker) H.S.	65 00	150 38	59	115	137	140
Tolovana	65 16	148 50	60	122	162	130
Chena	65 03	146 03	57	129	137	140
Circle	65 29	144 39	54	135	143	145
E. Cold Bay	55 13	162 29	54	117	144	145
Near Tenakee Inlet	58 13	135 55	82	147	72	150
Hooniah H.S.	57 48	136 20	44	136		140
Tenakee H.S.	57 47	135 13	43	111	63	115

1/ Maximum surface temperature reported from a spring or fumarole.

2/ Predicted using chemical geothermometers, assuming last equilibration in the reservoir; assumes saturation of SiO_2 with respect to quartz, and no loss of Ca from calcite deposition.

3/ Assumed average reservoir temperature based on data presently available.

4/ From surface manifestations, geophysical data, well records, and geologic inference. Assumes 1.5 km^2 if no data pertinent to size is available.

5/ Top assumed at depth of 1.5 km if no data available. Bottom assumed at 3 km depth for all convection systems.

6/ Calculated from assumed area and thickness.

7/ Calculated as product of assumed volume, volumetric specific heat of 0.6 $cal/cm^3°C$, and temperature in degrees C above 15°C.

indicated subsurface temperatures from 90° to 150°C

Reservoir Assumptions				Comments
Sub-surface area	Thickness	Volume	Heat content 10^{18} cal	
km^2 4/	km 5/	km^3 6/	7/	
3	2	6	.4	About 18 springs near 1945 eruption in Okmok caldera; may be more extensive and higher in temperatures; sinter reported.
1.5	1.5	2.25	.2	12 springs and fumaroles near recent volcanism.
1.5	1.5	2.25	.2	Several hot springs in permanently thawed area of .25 km^2.
1.5	1.5	2.25	.2	2 spring areas 1.3 km apart discharging ∿100 lpm and depositing travertine; Na-K-Ca may be too high.
1.5	1.5	2.25	.2	One main spring.
1.5	1.5	2.25	.2	2 springs discharging ∿1,000 lpm.
1.5	1.5	2.25	.2	Several springs.
1.5	1.5	2.25	.1	Several springs.
1.5	1.5	2.25	.2	One main spring discharging ∿500 lpm; chemical data not reliable.
1.5	1.5	2.25	.2	Hot springs discharging ∿230 lpm.
1.5	1.5	2.25	.2	Several hot springs.
1.5	1.5	2.25	.2	Hot spring discharging ∿560 lpm.
1.5	1.5	2.25	.2	Several hot springs, "small" discharge, possibly depositing travertine.
1.5	1.5	2.25	.2	Hot springs discharging ∿840 lpm, depositing sulfur
1.5	1.5	2.25	.2	11 hot springs discharging ∿500 lpm, depositing travertine.
1.5	1.5	2.25	.2	In recent volcanic rocks.
1.5	1.5	2.25	.2	Discharging ∿40 lpm; chemical data not reliable.
1.5	1.5	2.25	.2	3 hot springs discharging ∿110 lpm; chemical data not reliable.
1.5	1.5	2.25	.1	About 12 hot springs discharging ∿80 lpm.

Table 5.—Identified hot-water convection systems with indicated

Name	Location		Temperatures °C			
	Latitude ° ' N	Longitude ° ' W	Surface 1/	Geochemical SiO₂ 2/	Geochemical Na-K-Ca 2/	Subsurface 3/
ALASKA Con.						
Near Fish Bay	57 22	135 23	47	143		150
Baranof H.S.	57 05	134 50	50	119	68	125
Goddard H.S.	56 50	135 22	67	148	147	150
Bailey H.S.	55 59	131 40	88	158		150
Bell Island H.S.	55 56	131 34	72	140		145
ARIZONA						
Verde H.S.	34 21.5	111 42.5	36	118	146	150
Castle H.S.	33 59.1	112 21.6	50	109	71	110
North of Clifton	33 04.7	109 18.2	59	138	174	140
Clifton H.S.	33 03.2	109 17.8	75	107	161	110
Eagle Creek Spring	33 02.8	109 28.6	36	114	104	115
Gillard H.S.	32 58.5	109 21.0	82	135	138	140
Mt. Graham	32 51.4	109 44.9	42	106	102	110
CALIFORNIA						
Kelley H.S.	41 27.5	120 50	96	144	85	130
Hunt H.S.	41 02.1	122 55.1	58	101	75	105
Big Bend H.S.	41 01.3	122 55.1	82	121	137	140
Salt Springs(1)	40 40.2	122 38.7	20	107	55	110
Wendel-Amedee area	40 18	120 11	95	135	129	140

Reservoir Assumptions				Comments
Sub-surface area km² 4/	Thickness km 5/	Volume km³ 6/	Heat content 10^{18} cal 7/	
1.5	1.5	2.25	.2	Springs discharging ∿95 lpm; chemical data not reliable.
1.5	1.5	2.25	.2	Springs discharging ∿300 lpm.
1.5	1.5	2.25	.2	3 hot springs discharging ∿50 lpm.
1.5	1.5	2.25	.2	9 hot springs discharging ∿315 lpm; chemical data not reliable.
1.5	1.5	2.25	.1	5 hot springs discharging ∿40 lpm; chemical data not reliable.
1.5	1.5	2.25	.2	Several springs; indicated temperatures may be too high.
1.5	1.5	2.25	.1	Two springs.
1.5	1.5	2.25	.2	Two springs; may be depositing calcite
1.5	1.5	2.25	.1	Several springs; may be depositing calcite.
1.5	1.5	2.25	.1	Two springs; indicated geochemical temperature may be too high.
1.5	1.5	2.25	.2	5 springs
1.5	1.5	2.25	.1	1 hot mineral well; geochemical temperatures may be too high.
1.5	2	3	.2	1 spring flowing ∿1,200 lpm; 1,000 m well drilled in 1969, reported 110°C.
1.5	1.5	2.25	.1	2 hot springs flowing 8 lpm
1.5	1.5	2.25	.2	6 hot springs, flowing 38 lpm.
1.5	1.5	2.25	.1	Spring from travertine cone, flowing 20 lpm
7	2	14	1.1	Many flowing 3,500 lpm; 4 wells, deepest 338 m, T=107°C; possibly separate systems at Wendel and Amedee.

Table 5.—Identified hot-water convection systems with indicated

Location			Temperatures °C			
Name	Latitude ° ' N	Longitude ° ' W	Surface 1/	Geochemical SiO_2 2/	Geochemical Na-K-Ca 2/	Subsurface 3/
CALIFORNIA Con.						
Tuscan (Lick) S.	40 14.5	122 08.4	30	137	112	140
Soda Spring	39 24.8	122 58.6	17	148	158	150
Salt Spring(2)	39 25.8	122 32.3	25	157	123	150
Crabtree H.S.	39 17.4	122 49.3	41	163	133	150
Fouts (Redeye) S.	39 21.0	122 40.1	26	150	126	150
Fouts (Champagne) S.	39 20.5	122 39.4	18	117	128	130
Orr's H.S.	39 13.8	123 21.9	40	112	67	115
Vichy Springs	39 09.9	123 09.4	32	132	145	135
Cooks Springs	39 15.2	122 31.4	17	133	187	140
Saratoga Springs	39 10.5	122 58.7	16	137	46	140
Wilbur H.S. area	39 02.2	122 25.2	60	180	240	145
Deadshot Spring	39 05.1	122 27.4	26	135	204	135
Point Arena H.S.	38 52.6	123 30.6	44	105	62	105
Ornbaun Springs	38 54.7	123 18.4	16	126	122	125
Seigler Springs	38 52.5	122 41.3	52	169	188	150
Baker Soda Spring	38 53.6	122 31.9	24	124	202	130
One-Shot Mining Co.	38 50.0	122 21.4	22	135	153	150
Aetna Springs	38 39.5	122 28.7	33	135	94	135
Walter Springs	38 39.2	122 21.4	19	135	82	135
Mark West Springs	38 32.9	122 43.2	31	140	48	140

Reservoir Assumptions				Comments
Sub-surface area km² 4/	Thickness km 5/	Volume km³ 6/	Heat content 10^{18} cal 7/	
1.5	1.5	2.25	.2	20 Springs flowing 190 lpm.
1.5	1.5	2.25	.2	High bicarbonate spring; geothermometry doubtful.
1.5	1.5	2.25	.2	Note: distinct from Salt Springs, above; geothermometry doubtful.
1.5	1.5	2.25	.2	4 springs, flowing 57 lpm; geothermometry doubtful.
1.5	1.5	2.25	.2	4 springs, flow 7.5 lpm; geothermometry doubtful.
1.5	1.5	2.25	.2	4 springs, geothermometry doubtful.
1.5	1.5	2.25	.1	7 springs flowing 95 lpm.
1.5	1.5	2.25	.2	7 springs flowing 113 lpm; Na-K-Ca may be inaccurate due to travertine deposition.
1.5	1.5	2.25	.2	Geothermometry doubtful.
1.5	1.5	2.25	.2	5 springs, flow 9 lpm; geothermometry doubtful.
16	2	32	2.5	12 springs, flow 80 lpm; well drilled to 1,100 m, 141°C; should be in table 4?
1.5	1.5	2.25	.2	4 springs flowing 4 lpm; geothermometry doubtful.
1.5	1.5	2.25	.1	2 springs flowing 19 lpm.
1.5	1.5	2.25	.2	1 spring flowing less than 1 lpm.
2	1.5	3	.2	13 springs flowing 132 lpm; geothermometry doubtful.
1.5	1.5	2.25	.2	Numerous springs; geothermometry doubtful.
1.5	1.5	2.25	.2	Flow 189 lpm; sinter and travertine reported.
1.5	1.5	2.25	.2	6 springs flowing 75 lpm; geothermometry doubtful.
1.5	1.5	2.25	.2	Flow 6 lpm; geothermometry doubtful.
1.5	1.5	2.25	.2	~9 hot springs in a group flowing 113 lpm.

Table 5.—Identified hot-water convection systems with indicated

Location			Temperatures °C			
Name	Latitude ° ' N	Longitude ° ' W	Surface 1/	Geochemical 2/ SiO_2	Geochemical 2/ Na-K-Ca	Subsurface 3/
CALIFORNIA Con.						
Napa Soda S. Rock (Priest)	38 31.1	122 15.6	26	143	81	145
Los Guilicos W.S.	38 23.7	122 33.0	31	129	184	135
(Jackson's) Napa Soda Springs	38 23.4	122 16.7	16	149	60	150
Brockway (Cornelian) H.S.	39 13.5	120 0.4	60	119	94	120
Grovers H.S.	38 41.9	119 51.6	63	135	126	140
Fales H.S.	38 20	119 24	62	147	165	150
Buckeye H.S.	38 14.3	119 19.6	64	122	138	140
Benton H.S.	37 48	118 31.8	57	113	79	115
Travertine H.S.	38 14.8	119 12.1	70	114	172	120
Near Black Pt.	38 2.4	119 5	63	122	124	125
Paoha Island	37 59.8	119 01.2	83	186		125
Mono H.S.	37 19.5	119 01.0	44	110	80	115
Blayney Meadows H.S.	37 14.1	118 53	43	102	57	105
Mercey H.S.	36 42.2	120 51.6	46	122	94	125
Randsburg area	35 23.0	117 32.2	115			125
Arrowhead H.S. area	34 08.6	117.15.2	94	132	147	150
Pilger Estates H.S.	33 26.0	115 41.1	82	125	145	145
Warner H.S.	33 17.0	116 38.4	64	141	100	145
Glamis (E. Brawley)	32 58	115 11				135
Glamis (East)	33 59	115 04				135
Dunes	32 49	115 01				135

Reservoir Assumptions				Comments
Sub-surface area km2 4/	Thick-ness km 5/	Vol-ume km3 6/	Heat con-tent 10^{18} cal 7/	
1.5	1.5	2.25	.2	2 springs flowing 60-85 lpm; geothermometry doubtful.
1.5	1.5	2.25	.2	3 springs flowing 75 lpm; Na-K-Ca may be inaccurate due to travertine deposition.
1.5	1.5	2.25	.2	27 springs; geothermometry doubtful.
1.5	1.5	2.25	.1	6 springs flowing 570 lpm.
1.5	1.5	2.25	.2	12 springs flowing 378 lpm.
1.5	1.5	2.25	.2	20 springs flowing 95 lpm, possibly depositing travertine.
1.5	1.5	2.25	.2	1 spring flowing 75 lpm.
1.5	1.5	2.25	.1	2 springs flowing 1,500 lpm.
1.5	1.5	2.25	.1	3 main springs flowing 38 lpm; extensive travertine.
1.5	1.5	2.25	.1	
1.5	1.5	2.25	.1	Several springs flowing 370 lpm; non-quartz equilibration of SiO_2 likely.
1.5	1.5	2.25	.1	Four springs flowing 95 lpm.
1.5	1.5	2.25	.1	Eight springs flowing 150 lpm.
1.5	1.5	2.25	.2	3 hot springs flowing 23 lpm.
1.5	2.5	3.75	.3	1 well reported 115°C at 235 m.
2	1.5	3	.2	2 groups of hot springs flowing 190 lpm.
1.5	1.5	2.25	.2	Near Salton Sea; possibly more extensive.
1.5	1.5	2.25	.2	6 springs flowing 570 lpm.
2	1.5	3	.2	Estimated using temperature gradient data; a part above 150°C?
4	1.5	6	.4	Temperature gradient data; a part above 150°C?
6	1.5	9	.6	Temperature gradient data: a part above 150°C?

Table 5.—Identified hot-water convection systems with indicated

Name	Location		Temperatures °C			
	Latitude ° ' N	Longitude ° ' W	Surface 1/	Geochemical		Subsurface 3/
				2/ SiO_2	2/ Na-K-Ca	
COLORADO						
Routt H.S.	40 33.6	106 51	64	131	168	135
Steamboat Springs	40 29.1	106 50.3	66	129	195	135
Idaho Springs	39 44.2	105 30.2	50	109	208	115
Glenwood Springs	39 33	107 19.3	66	137	190	140
Avalanche Springs	39 13.9	107 13.5	57	136	125	140
Cottonwood Springs	38 48.7	106 13.5	62	107	83	110
Mt. Princeton S.	38 43.9	106 10.2	66	112	52	115
Poncha H.S.	38 29.9	106 04.5	76	129	143	145
Mineral H.S.	38 10.1	105 55.0	63	103	91	105
Waunita H.S.	38 31.0	106 29.1	71	129	87	130
Cebolla H.S.	38 16.5	107 05.9	46	125	233	130
Orvis H.S.	38 08	107 44	58	109	231	110
Wagon Wheel Gap	37 45	106 49.2	66	129	188	135
Pagosa H.S.	37 15.5	107 00.5	70	165	278	150?

Reservoir Assumptions				Comments
Sub-surface area km^2 4/	Thickness km 5/	Volume km^3 6/	Heat content 10^{18} cal 7/	
1.5	1.5	2.25	.2	Three hot springs; Chemical data not reliable.
1.5	1.5	2.25	.2	Many hot springs; chemical data not reliable; some travertine.
1.5	1.5	2.25	.1	8 springs, total discharge 190 lpm depositing travertine; probably fault-controlled; chemical data not reliable.
1.5	1.5	2.25	.2	11 springs discharging about 11,400 lpm; chemical data not reliable; some travertine.
1.5	1.5	2.25	.2	5 springs discharging ∿54 lpm; chemical data not reliable.
4	1.5	6	.3	5 springs discharging ∿570 lpm; extensive zeolitization.
5	1.5	7.5	.5	4 main springs, 30 others; extensive zeolitization, present deposition of opal, calcite, and phillipsite reported.
1.5	1.5	2.25	.2	3 springs depositing travertine and discharging ∿1,900 lpm; associated with flourite deposits; Na-K-Ca temperature may be too high.
1.5	1.5	2.25	.1	30 springs discharging ∿190 lpm, reported with travertine and sinter (?); wells to 354 m depth and 60°C.
1.5	1.5	2.25	.2	2 groups, more than 100 springs discharging 3,785 lpm.
1.5	1.5	2.25	.2	20 springs discharging ∿380 lpm; travertine reported; chemical data not reliable.
1.5	1.5	2.25	.1	1 spring discharging ∿1,140 lpm; chemical data not reliable.
1.5	1.5	2.25	.2	3 springs depositing travertine and associated with flourite deposits; Na-K-Ca temperature probably too high.
1.5	1.5	2.25	.2	Springs discharging ∿380 lpm and depositing travertine; 1 well for space heating; chemical data not reliable.

Table 5.—Identified hot-water convection systems with indicated

Location			Temperatures °C			
Name	Latitude ° ' N	Longitude ° ' W	Surface 1/	Geochemical SiO$_2$ 2/	Geochemical Na-K-Ca 2/	Subsurface 3/
HAWAII						
Steaming Flats (Sulphur Bank area)	19 26.5	155 16	97	--No Data--		∿150?
Upper Kau area	19 23.7	155 17.3	∿22	--	--	100
1955 eruption area, East Rift	19 26.5	154 57	hot	--No Data--		∿150?
Puulena area, East Rift	19 28.3	154 53	?	--No Data--		∿150?
IDAHO						
Red River H.S.	45 47.3	115 08.8	55	123	80	125
Riggins H.S.	45 24.7	116 28.5	47	120	95	125
Burgdorf H.S.	45 16.7	115 55.2	45	121	57	125
Zim's (Yoghann) H.S.	45 02.6	116 17.0	65	115	85	120
Krigbaum H.S.	44 58.1	116 11.4	43	121	96	125
Starkey H.S	44 51.2	116 25.8	56	108	70	115
White Licks H.S.	44 40.9	116 13.8	65	143	145	150
Near Cove School	44 35.0	116 37.7	70	120	78	125
Near Deer Creek	44 32.4	116 45.0	50	107	63	110
Near Midvale	44 28.3	116 43.9	28	128	243	135
Near Midvale Airprt.	44 28.2	116 45.9	28	121	51	125
Hot Creek Springs	44 38.5	116 02.7	34	111	62	115

Reservoir Assumptions				Comments
Subsurface area km² 4/	Thickness km 5/	Volume km³ 6/	Heat content 10^{18} cal 7/	
1.5	1.5	2.25	.2	Nearly constant fumarolic activity, no water discharge; area may be larger.
5	0.7	3.5	.2	Resistivity anomally drilled by N.S.F. grant to G. V. Keller, 1973; low-temperature convection system identified top at water table, ∿80°C at -490 m; bottom of convection near -1,150 m, ∿100°C, then steep gradient to basaltic magma chamber (?).
2	2	4	.3	Steaming area; three wells drilled 1961, deepest ∿210 m, ∿113°C; NSF grant 1975 to University of Hawaii for deep test.
2	2	4	.3	No surface manifestations; geophysical anomalies identified.
1.5	1.5	2.25	.2	9 springs discharging ∿130 lpm; mixing model T=190°C.
1.5	1.5	2.25	.2	4 springs discharging ∿190 lpm; mixing model T=220°C.
1.5	1.5	2.25	.2	2 springs discharging ∿610 lpm.
1.5	1.5	2.25	.1	Discharging hot well.
1.5	1.5	2.25	.2	2 springs discharging 150 lpm; mixing model T=200°C.
1.5	1.5	2.25	.1	7 hot springs discharging 490 lpm.
1.5	1.5	2.25	.2	Numerous springs discharging 113 lpm; may be part of larger system including hot springs near Cove School; mixing model T=220°C.
1.5	1.5	2.25	.2	Numerous springs discharging 1,630 lpm.
1.5	1.5	2.25	.1	Hot springs discharging 219 lpm.
1.5	1.5	2.25	.2	Flowing well; may be part of single system including Deer Creek and Midvale.
1.5	1.5	2.25	.2	Flowing well; geochemical temperatures unreliable.
1.5	1.5	2.25	.1	Springs discharging ∿3,000 lpm; mixing model suggests 195°C.

Table 5.—Identified hot-water convection systems with indicated

Name	Location		Temperatures °C			
	Latitude ° ' N	Longitude ° ' W	Surface 1/	Geochemical SiO2 2/	Geochemical Na-K-Ca 2/	Subsurface 3/
IDAHO Con.						
Molly's H.S.	44 38.3	115 41.6	59	130	83	135
Vulcan H.S.	44 34.1	115 41.5	87	148	135	150
Cabarton H.S.	44 25	116 01.7	71	124	99	130
Boiling Springs	44 21.9	115 51.4	86	134	89	140
Near Payette River	44 05.1	116 03	80	148	139	150
Near Grimes Pass	44 02.8	115 51.1	55	110	74	115
Kirkham H.S.	44 04.3	115 32.6	65	118	79	120
Bonneville H.S.	44 09.5	115 18.4	85	138	142	145
Stanley H.S.	44 13.5	114 55.6	41	107	47	110
Sunbeam H.S.	44 16.1	114 44.9	76	133	130	140
Slate Creek H.S.	44 10.1	114 37.5	50	129	91	130
Roystone H.S.	43 57.2	116 18	55	148	150	150
N.E. Boise Thermal area	43 36.1	116 09.9	75	124	79	125
Neinmeyer H.S.	43 45.5	115 34.7	76	138	126	140
Dutch Frank Springs	43 47.7	115 25.5	65	120	72	125
Paradise H.S.	43 33.2	115 16.3	56	118	72	120
Worswick H.S.	43 33.5	114 47.2	81	135	93	140

subsurface temperatures from 90° to 150°C—Continued

Reservoir Assumptions				Comments
Sub-surface area km² 4/	Thickness km 5/	Volume km³ 6/	Heat content 10^{18} cal 7/	
1.5	1.5	2.25	.2	7 springs discharging 76 lpm; mixing model suggests 195°C.
1.5	1.5	2.25	.2	13 springs discharging ∿1,900 lpm; sinter reported.
1.5	1.5	2.25	.2	Numerous springs discharing ∿265 lpm; mixing model T = 165°C.
1.5	1.5	2.25	.2	Numerous vents discharging ∿600 lpm and depositing minor zeolites, calcites, and mercury minerals.
1.5	1.5	2.25	.2	One spring discharging ∿75 lpm; mixing model suggests 200°C.
1.5	1.5	2.25	.1	Spring(s?) discharging ∿260 lpm.
1.5	1.5	2.25	.1	Numerous springs discharging ∿950 lpm.
1.5	1.5	2.25	.2	8 springs and seeps discharging ∿1,900 lpm; mixing model suggests 175°C.
4	1.5	6	.3	6 springs discharging ∿420 lpm; south-western of a possible 10-km line extending NE to Sunbeam; mixing T = 180°C.
1.5	1.5	2.25	.2	Numerous vents discharging ∿1,700 lpm.
1.5	1.5	2.25	.2	8 springs and seeps discharging ∿700 lpm; mixing T = 210°C.
2	1.5	3	.2	5 springs discharging ∿75 lpm.
4	2	8	.5	Linear zone of springs and associated thermal wells on the NE edge of Boise; used for space heating.
1.5	1.5	2.25	.2	13 springs discharging ∿1,300 lpm with gas, mixing model suggests 190°C.
1.5	1.5	2.25	.2	Numerous springs, gassy, discharging ∿1,150 lpm.
1.5	1.5	2.25	.1	Several springs.
1.5	1.5	2.25	.2	Numerous springs discharging ∿1,750 lpm.

Table 5.—Identified hot-water convection systems with indicated

Name	Location: Latitude ° ' N	Location: Longitude ° ' W	Temperatures °C: Surface 1/	Temperatures °C: Geochemical SiO_2 2/	Temperatures °C: Geochemical Na-K-Ca 2/	Temperatures °C: Subsurface 3/
IDAHO Con.						
Guyer H.S.	43 40.5	114 24.6	71	129	88	135
Clarendon H.S.	43 33.6	114 24.9	47	125	114	130
Hailey H.S.	43 30.3	114 22.2	63	129	83	135
Near Brockie Airpt	43 32.4	113 30.1	41	107	91	110
Elk Creek H.S.	43 25.4	114 37.6	54	113	80	120
Near Punkin Corner	43 18.1	114 54.4	35	123	71	125
Barron's H.S.	43 18.1	114 54.4	71	124	91	130
Near Magic Reservoir	43 19.7	114 23.2	71	138	163	140
Near Bennett Creek	43 06.9	115 27.9	68	129	71	135
Latty H.S.	43 07.0	115 18.3	55	138	137	140
Near Ryegrass Creek	43 05.8	115 24.6	62	129	81	135
Near Radio Towers	43 02.2	115 27.5	38	129	125	130
White Arrow H.S.	43 02.9	114 57.2	65	136	113	140
Near Chalk Mine	43 02.9	114 55	47	133	98	140
Near Clover Creek	43 01.4	115 00.6	43	113	70	120
Near Gravel Pits	42 54.3	115 29.5	34	109	144	145
Bruneau-Grandview	42 56	115 56	84	138	93	145
Near Banbury	42 41.4	114 50	59	136	108	140

Reservoir Assumptions				Comments
Subsurface area km² 4/	Thickness km 5/	Volume km³ 6/	Heat content 10^{18} cal 7/	
1.5	1.5	2.25	.2	Numerous springs discharging ∿3,800 lpm.
1.5	1.5	2.25	.2	Numerous springs discharging ∿380 lpm; mixing model suggest 215°C.
1.5	1.5	2.25	.2	Numerous springs discharging ∿265 lpm; mixing model suggests 190°C.
1.5	1.5	2.25	.1	1 well flowing ∿45 lpm.
1.5	1.5	2.25	.1	5 springs discharging ∿55 lpm.
1.5	1.5	2.25	.1	Flowing well discharging 15 lpm; may be part of extensive system underlying a large portion of the Cames Prairie, and including Elk Creek, Barrons, and Waldrop.
1.5	1.5	2.25	.2	Numerous springs discharging ∿120 lpm.
1.5	1.5	2.25	.2	One well flowing 51 lpm; mixing models indicate temperatures as high as 275°C.
1.5	1.5	2.25	.2	Flowing well discharging ∿2,600 lpm.
1.5	1.5	2.25	.2	One spring; may be part of extensive system that includes Bennett Creek and Ryegrass Creek; SiO_2 temperature of all may be too high because of equilibration with diatomite.
1.5	1.5	2.25	.2	Flowing well.
1.5	1.5	2.25	.2	1 flowing well discharging 30 lpm.
1.5	1.5	2.25	.2	4 springs discharging ∿3,100 lpm; mixing model indicates 200°C.
1.5	1.5	2.25	.2	1 flowing well.
1.5	1.5	2.25	.1	1 flowing well.
1.5	1.5	2.25	.2	1 flowing well discharging ∿8 lpm. Na-K-Ca temperature may be inaccurate carbonate deposition reported. May be diatomaceous earth at depth.
2250	1.5	3375	263	An extensive area with many warm and hot artesian wells; mixing model temperatures up to 275°C.
8	1.5	12.0	.9	1 flowing well discharging ∿225 lpm; mixing T =215°C; includes Miracle and 1 other spring.

Table 5.—Identified hot-water convection systems with indicated

Location			Temperatures °C			
Name	Latitude ° ' N	Longitude ° ' W	Surface 1/	Geochemical SiO_2 2/	Geochemical Na-K-Ca 2/	Subsurface 3/
IDAHO Con.						
Near Cedar Hill	42 24.9	114 18.1	38	116	65	120
Near Bridger Springs	42 28.7	113 37.5	60	111	89	115
Oakley Warm Springs	42 10.4	113 51.7	47	119	92	120
Raft River thermal area	42 06.1	113 22.8	96	136	139	140
Maple Grove H.S.	42 18.2	111 42.2	76	107	236	110
Near Riverdale	42 09.9	111 50.4	45	126	170	125
Wayland H.S.	42 08.2	111 56.9	77	126	270	130
Near Newdale	43 53.2	111 35.4	36	122	84	125
Ashton Warm Springs	44 05.7	111 27.5	41	143	91	145
MONTANA						
Helena (Broadwater) Hot Spring	46 36.5	112 05	65	136	135	140
White Sulphur Springs	46 32.8	110 54.2	57	103	148	150
Alhambra H.S.	46 27	111 59	59	115	111	120
Boulder H.S.	46 12	112 05.6	76	143	135	145
Gregson (Fairmont) H.S.	46 02.6	112 48.4	74	128	126	130
Pipestone H.S.	45 53.8	112 13.9	61	115	113	120
Barkels (Silver Star) H.S.	45 41.5	112 17.2	72	143	139	145
Norris (Hapgood) H.S.	45 34.6	111 41	52	130	153	150
Jardine (Big Hole or Jackson) H.S.	45 21.8	113 24.7	58	104	148	150

subsurface temperatures from 90° to 150°C—Continued

Reservoir Assumptions				Comments
Sub-surface area km² 4/	Thickness km 5/	Volume km³ 6/	Heat content 10^{18} cal 7/	
6	1.5	9	.6	1 flowing well discharging ~2,050 lpm.
1.5	1.5	2.25	.1	1 flowing well discharging 7,900 lpm; mixing T = 150°C.
1.5	1.5	2.25	.1	1 spring discharging 38 lpm; mixing T = 195°C.
20	1.5	30	2.3	Area of flowing hot wells recently explored by ERDA; 140°C measured at depth of 1,400 m in well flowing ~3,800 lpm.
2	1.5	3	.2	Numerous springs discharging ~1,300 lpm; Na-k-Ca possibly inaccurate due to deposition of carbonate.
1.5	1.5	2.25	.2	1 flowing well; Na-K-Ca possibly inaccurate from deposition of carbonate.
5	1.5	7.5	.5	Numerous springs discharging ~3,400 lpm and depositing travertine; Na-K-Ca thermometry may be inaccurate.
1.5	1.5	2.25	.2	Flowing well.
1.5	1.5	2.25	.2	Springs discharging ~8 lpm from Pleistocene basalt.
1.5	1.5	2.25	.2	2 hot springs discharging 110 lpm.
1.5	1.5	2.25	.2	About 9 springs discharging ~2,000 lpm; mixing model suggests 150°C.
1.5	1.5	2.25	.1	About 22 springs
1.5	1.5	2.25	.2	Many springs in two groups; siliceous sinter; large discharge.
1.5	1.5	2.25	.2	Several springs
1.5	1.5	2.25	.1	Several springs.
1.5	1.5	2.25	.2	4 springs discharging 200 lpm.
1.5	1.5	2.25	.2	5 springs discharging 200 lpm.
1.5	1.5	2.25	.2	About 100 springs ~5,700 lpm; mixing model indicates 150°C.

Table 5.—Identified hot-water convection systems with indicated

Name	Location: Latitude ° ' N	Location: Longitude ° ' W	Temperatures °C: Surface 1/	Geochemical: 2/ SiO_2	Geochemical: 2/ Na-K-Ca	Subsurface 3/
NEVADA						
Bog H.S.	41 55.5	118 48.1	88	108	109	115
Howard H.S.	41 43.3	118 30.3	56	128	81	130
Dyke H.S.	41 34.0	118 33.7	66	129	137	140
Near Soldier Meadow	41 21.5	119 13.2	54	113	65	115
Double H.S.	41 03.0	119 02.8	80	140	127	145
Near Black Rock	40 57	118 58	90	148	116	150
Fly Ranch H.S.	40 52.0	119 20.9	80	127	154	130
Butte Sprs.	40 46	119 07	86	129	120	130
Mineral H.S.	41 47.3	114 43.3	60	127	129	130
Hot Hole (Elko)	40 49.1	115 46.5	89	115	127	115
Near Carlin	40 42.0	116 08.0	79	119	81	120
Hot Sulphur Sprs.	41 9.4	114 59.1	90	128	191	140
Hot Springs Point	40 24.2	116 31.0	54	116	233	125
Walti H.S.	39 54.1	116 35.2	72	117	78	120
Spencer H.S.	39 19	116 51	72	123	210	125
Hot Pot	40 55.3	117 06.5	58	125	195	125

subsurface temperatures from 90° to 150°C—Continued

Reservoir Assumptions				Comments
Sub-surface area km² 4/	Thickness km 5/	Volume km³ 6/	Heat content 10^{18} cal 7/	
2	2	4	.2	2 springs discharging ∿4,000 lpm at 54°C.
1.5	1.5	2.25	.2	Several springs.
1.5	1.5	2.25	.2	1 (?) spring discharging ∿100 lpm.
6	2	12	.7	Several springs in area of ∿6 km² discharging ∿50 lpm.
10	2	20	1.6	Several springs along linear zone 20 km north from Black Rock Point; largest group discharging ∿175 lpm; minor travertine.
1.5	1.5	2.25	.2	
8	2	16	1.1	Area of large spring pools and two abandoned wells discharging ∿500 lpm and depositing travertine, so Na-K-Ca may be too high.
1.5	1.5	2.25	.2	
1.5	1.5	2.25	.2	Several springs and shallow wells.
2	1.5	3	.2	Several springs depositing travertine, so Na-K-Ca temperature may be high.
1.5	1.5	2.25	.1	
1.5	1.5	2.25	.2	3 springs discharging ∿190 lpm; paleozoic limestone at depth; Na-K-Ca geothermometer may be inaccurate; may be part of more extensive area extending 4.8 km along west edge of Snake Mtns.
5	1.5	7.5	.5	Hot springs, discharging ∿125 lpm; depositing travertine; Na-K-Ca may be inaccurate.
2	1.5	3	.2	6 springs discharging 300 lpm and depositing travertine.
1.5	1.5	2.25	.2	Several hot springs discharging 50 lpm and depositing travertine so Na-K-Ca thermometry may be inaccurate.
1.5	1.5	2.25	.2	One spring discharging ∿270 lpm; depositing travertine; Na-K-Ca may be inaccurate.

Table 5.—Identified hot-water convection systems with indicated

Location			Temperatures °C			
Name	Latitude ° ' N	Longitude ° ' W	Surface 1/	Geochemical SiO_2 2/	Geochemical Na-K-Ca 2/	Subsurface 3/
NEVADA Con.						
Buffalo Valley H.S.	40 22.1	117 19.5	79	125	140	130
Hot Springs	41 25.4	117 23.0	58	107	209	110
Golconda H.S.	40 57.7	117 29.6	74	116	201	125
Sou (Gilberts) H.S.	40 05.4	117 43.5	93	115	99	115
Dixie H.S.	39 47.9	118 04.0	72	143	143	150
The Needles	40 08.8	119 40.5	98	137	214	145
Walleys H.S.	38 58.9	119 49.9	71	109	85	110
Nevada H.S.	38 54.0	119 24.7	61	104	86	105
Darrough H.S.	38 49.3	117 10.8	97	136	127	140
Warm Springs	38 11.3	116 22.5	61	111	192	125
Bartholomae H.S.	39 24.3	116 20.8	54	129	72	130
NEW MEXICO						
Jemez (Ojos Calientes) H.S.	35 47	106 41	73	134	197	135
Radium H.S.	32 30	106 55.5	52	124	222	130
Lower Frisco	33 15	108 47	37	128	150	150
Gila H.S.	33 12	108 12	68	121	114	125

Reservoir Assumptions				Comments
Sub-surface area km² 4/	Thickness km 5/	Volume km³ 6/	Heat content 10^{18} cal 7/	
4	2.5	10	.7	More than 200 hot springs with largest discharging 61 lpm; in travertine area so Na-K-Ca thermometry may be inaccurate.
1.5	1.5	2.25	.1	Discharging from travertine so Na-K-Ca thermometry may be inaccurate.
1.5	1.5	2.25	.2	About 12 springs discharging 750 lpm and depositing manganiferous travertine; area may be considerably larger.
1.5	1.5	2.25	.1	Several hot springs depositing travertine.
2	1.5	3	.2	Several hot springs discharging ∿200 lpm.
2	1.5	3	.2	Two lines of springs that have deposited travertine cones in Pyramid Lake; two wells on eastern line, 116°C at 450 and 1,800 m depth; may be considerably larger system.
1.5	1.5	2.25	.1	Many hot springs discharging ∿75 lpm along base of recent faultscarp.
1.5	1.5	2.25	.1	Several springs in travertine area discharging ∿200 lpm.
1.5	1.5	2.25	.2	Several springs and well discharging ∿350 lpm; one well 129°C at 230 m depth discharging ∿4,000 lpm; area may be considerably larger.
1.5	1.5	2.25	.2	2 springs.
1.5	1.5	2.25	.2	Spring discharging ∿400 lpm.
1.5	1.5	2.25	.2	About 10 springs depositing travertine and discharging ∿750 lpm; Na-K-Ca probably not reliable; 9.7 km SSW of Valles Caldera.
1.5	1.5	2.25	.2	
1.5	1.5	2.25	.2	Discharge ∿75 lpm; Na-K-Ca probably not reliable.
1.5	1.5	2.25	.2	Four hot springs discharging ∿3,400 lpm; area may be somewhat larger.

Table 5.—Identified hot-water convection systems with indicated

Location			Temperatures °C			
Name	Latitude ° ' N	Longitude ° ' W	Surface 1/	Geochemical SiO_2 2/	Geochemical Na-K-Ca 2/	Subsurface 3/
OREGON						
Mt. Hood	45 22.5	121 42.5	90	--No Data--		125
Carey (Austin) H.S.	45 01.2	122 00.6	86	126	118	125
Kahneetah H.S.	44 51.9	121 12.9	52	140	103	140
Breitenbush H.S.	44 46.9	121 58.5	92	127	149	150
Belknap H.S.	44 11.6	122 03.2	71	135	114	140
Klamath Falls	42 15	121 45	74	136	130	120
Summer Lake H.S.	42 43.5	120 38.7	43	134	112	140
Radium H.S.	44 55.8	117 56.4	58	124	108	130
Hot Lake (2)	45 14.6	117 57.6	80	100	115	120
Medical H.S.	45 01.1	117 37.5	60	125	125	130
Ritter H.S.	44 53.7	119 08.6	41	119	92	125
Fisher H.S.	42 17.9	119 46.5	68	123	165	130
Blue Mountain H.S.	44 21.3	118 34.4	58	99	126	130
Near Little Valley	43 53.5	117 30.0	70	145	119	150
Beulah H.S.	43 56.7	118 08.2	60	169	86	130
Near Riverside	43 28.0	118 11.3	63	143	138	150
Crane H.S.	43 26.4	118 38.4	78	127	124	130
Near Harney Lake	43 10.9	119 06.2	68	133	130	135
Near Trout Creek	42 11.3	118 09.2	52	140	144	145
Near McDermitt	42 04.1	117 30.0	52	120	100	120

Reservoir Assumptions				Comments
Subsurface area km^2 4/	Thickness km 5/	Volume km^3 6/	Heat content 10^{18} cal 7/	
2	2	4	.3	Many fumaroles but not water discharge; semiactive volcano; temperatures may be higher; area may be larger.
1.5	1.5	2.25	.1	Several hot springs in 0.1 km discharging ∿950 lpm.
1.5	1.5	2.25	.2	Hot spring discharging ∿200 lpm.
1.5	1.5	2.25	.2	40 to 60 springs in 0.1 km area discharging 3,400 lpm.
1.5	1.5	2.25	.2	3 springs discharging ∿300 lpm.
240	2	480	30	Numerous springs and shallow wells discharging from fault zones; largest spring ∿200 lpm; well temperatures 60° to 115°C used for domestic heating; large area indicated.
4	1.5	6.0	.4	3 springs discharging ∿75 lpm.
1.5	1.5	2.25	.2	2 flowing wells discharging ∿1,100 lpm.
1.5	1.5	2.25	.1	1 large spring pool discharging ∿1,500 lpm.
1.5	1.5	2.25	.2	2 springs discharging ∿200 lpm.
1.5	1.5	2.25	.1	1 hot spring discharging ∿130 lpm.
3	1.5	4.5	.3	Hot spring discharging ∿70 lpm; some H_2S.
1.5	1.5	2.25	.2	Several springs discharging ∿250 lpm.
1.5	1.5	2.25	.2	Several springs discharging ∿550 lpm.
1.5	1.5	2.25	.2	1 (?) spring discharging ∿50 lpm from vitric tuff so SiO_2 temperature may not be reliable; sinter and travertine reported.
1.5	1.5	2.25	.2	Several springs discharging ∿200 lpm.
1.5	1.5	2.25	.2	2 springs discharging ∿550 lpm.
3	1.5	4.5	.3	Spring discharging ∿550 lpm.
1.5	1.5	2.25	.2	Several springs discharging ∿200 lpm.
2	1.5	3.0	.2	Hot spring discharging ∿750 lpm.

Table 5.—Identified hot-water convection systems with indicated

Name	Location: Latitude ° ' N	Location: Longitude ° ' W	Temperatures °C: Surface 1/	Temperatures °C: Geochemical SiO_2 2/	Temperatures °C: Geochemical Na-K-Ca 2/	Temperatures °C: Subsurface 3/
UTAH						
Hooper H.S.	41 08	112 11.3	60	101	223	105
Crystal H.S.	40 29	111 54	58	103	135	135
Baker (Abraham, Crater) H.S.	39 36.8	112 43.9	87	118	122	125
Meadow H.S.	38 51.8	112 30	41	100	68	105
Monroe(Cooper) H.S.	38 38.2	112 06.4	76	110	118	120
Joseph H.S.	38 36.7	112 11.2	64	133	141	140
WASHINGTON						
Sol Duc H.S.	47 58.1	123 52.1	56	148	92	150
Olympic H.S.	47 58.9	123 41.2	52	126	87	130
Sulphur Creek H.S.	48 15.3	121 10.8	37	122	113	125
Garland (San Juan)	47 20.5	121 53.4	38	148	185	150
Ohanapecosh H.S.	46 44.2	121 33.6	49	126	164	130
WYOMING						
Huckleberry H.S.	44 07	110 41	71	150	141	150
Auburn H.S.	42 49.5	111 0	62	143	209	150
Totals (224 Systems)						

subsurface temperatures from 90° to 150°C—Continued

Reservoir Assumptions				Comments
Subsurface area	Thickness	Volume	Heat content 10^{18} cal	
km^2 4/	km 5/	km^3 6/	7/	
1.5	1.5	2.25	.1	4 saline hot springs in 2 groups 0.6 km apart; geothermometry may not be reliable.
1.5	1.5	2.25	.2	4 hot springs discharging ∿230 lpm.
1.5	1.5	2.25	.1	4 hot springs depositing travertine and Mn oxides at edge of young basalt flows.
1.5	1.5	2.25	.1	3 springs on 1.6 km trend; includes Hatton Hot Springs (Black Rock or Wiwepa) Hot Springs; analyzed spring discharges 226 lpm
5	1.5	7.5	.5	9 springs in 3 groups on 48 km trend along Sevier fault; includes Red Hill and Johnson Hot Springs; depositing travertine.
1.5	1.5	2.25	.2	Springs depositing travertine and discharging ∿110 lpm.
1.5	1.5	2.25	.2	11 springs discharging ∿500 lpm.
1.5	1.5	2.25	.2	17 springs discharging ∿500 lpm along fault zone.
1.5	1.5	2.25	.1	Springs discharging 15 lpm; minor precipitation (carbonate?).
1.5	1.5	2.25	.2	3 springs discharging ∿95 lpm; extensive travertine; chemical temperatures not reliable.
1.5	1.5	2.25	.2	5 springs discharging ∿225 lpm; extensive precipitation (carbonate?).
1.5	1.5	2.25	.2	2 small groups of hot springs discharging ∿380 lpm.
1.5	1.5	2.25	.2	More than 100 vents; discharging ∿140 lpm and depositing travertine.
∿2938		∿4564	∿345	

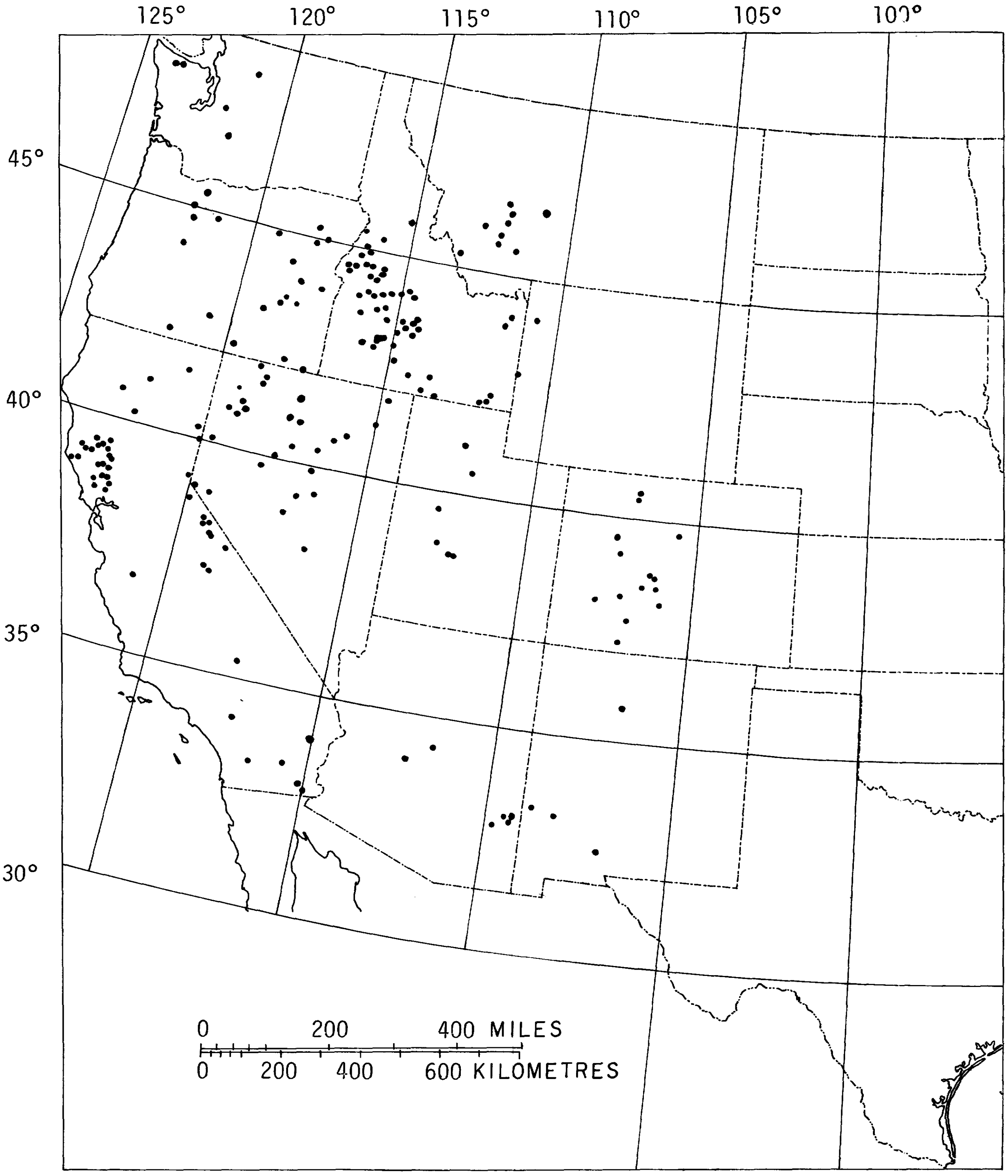

FIGURE 3.—Location of hydrothermal convection systems in the conterminous United States with indicated subsurface temperatures between 90° and 150°C.

dissolved salts (1,000 to 10,000 mg/kg), but a few contain 2 to 3 percent. The Salton Sea geothermal system is especially saline, having about 26 percent dissolved salts at reservoir temperatures exceeding 340°C.

Much attention has been given recently to constituents whose contents are strongly dependent on temperature. A few of these are useful in predicting subsurface temperatures from chemical analyses of water samples from springs or shallow wells. SiO_2 (Fournier and Rowe, 1966) and Na-K-Ca relations (Fournier and Truesdell, 1973) have been especially useful in providing most of the predicted temperatures in this report.

The basic assumptions involved in chemical geothermometers need to be emphasized. The most important (Fournier and others, 1974) are: (1) temperature-dependent reactions exist between constitutents in the water and the rocks of a reservoir; (2) all constituents involved in the reactions are sufficiently abundant so that supply is not a limiting factor; (3) chemical equilibrium is attained at the reservoir temperature; (4) little or no equilibration or change in composition occurs at lower temperatures as the water flows from the reservoir to the surface; and (5) the water from the reservoir does not mix with any other water at intermediate levels. Assumptions 1, 2, and 3 commonly seem to be valid for the SiO_2 and Na-K-Ca geothermometers. Nearly all reservoir rocks contain quartz, and residence times of a few days or weeks are sufficient to saturate the water in SiO_2 with respect to quartz at temperatures much above 150°C. Also, most waters seem to attain equilibrium in Na, K, and Ca with respect to the common clay minerals and feldspars. However, some indicated temperatures of our tabulated data are not reliable, at least in part because waters high in free CO_2 may not have attained equilibrium with the rocks or because they attained equilibrium with mineral assemblages other than those assumed for the geothermometers. In order to gain internal consistency, the SiO_2 temperatures reported in the tables are based on equilibrium with quartz rather than chalcedony or amorphous forms of silica. However, some reported systems, especially those of low temperature, may have equilibrated with one of these more soluble forms of silica. The predicted temperatures of such systems will be too high. Assumption 4, that water flows to the surface without chemical change, is probably never strictly true, but useful minimum temperatures can be predicted. Assumption 5, that no mixing occurs with cool shallow waters, may frequently be invalid. Mixing, formerly considered to be a major obstacle in predicting subsurface temperatures, has recently been utilized to advantage by Fournier and Truesdell (1974). In favorable circumstances, temperatures higher than those indicated by the SiO_2 on Na-K-Ca geothermometers can be predicted at deeper levels in a stacked series of reservoirs (Truesdell and Fournier, 1975). These mixing models are still so new that they have been applied only to a few systems. Other chemical and isotopic methods of temperature prediction are also being developed by Truesdell and others.

Experience has shown that natural geysers and active deposition of siliceous sinter are reliable indicators of subsurface temperatures at least as high as 180°C. On the other hand, travertine deposits ($CaCO_3$) and opaline residues produced by sulfuric acid leaching (from oxidation of H_2S) are commonly identified incorrectly as siliceous sinter but actually have no reliable relation to reservoir temperature.

The origin of the heat has major importance in predicting the geothermal resources of individual convection systems. Two principal origins are considered here: (1) heat directly related to volcanic sources localized as "hotspots" in the shallow crust of the Earth (Smith and Shaw, this circular) and (2) heat related to geothermal gradient, or the general increase in temperature with depth as a consequence of conductive heat flow (Diment and others, this circular). For both types, the ultimate source of most of the heat is from deep within the Earth, probably resulting in large part from natural radioactivity. As indicated by Smith and Shaw, the basalts and andesites that form most volcanoes have probably risen rapidly from the mantle to the surface in volcanic eruption. As a result, their heat is dispersed rather than stored and does not provide useful geothermal concentrations. However, the high-silica varieties of volcanic rocks, perhaps because of their very high viscosities, commonly are associated with magma chambers at shallow levels in the crust (perhaps 2 to 10 km but most commonly about 4 km; Smith and Shaw, this circular) and can sustain high-temperature convection systems for many thousands of years. Many large geothermal systems appear to be associated with young silicic volcanic rocks. Some hot-spring systems that have no direct association with young silicic volcanic systems may derive their heat from older volcanic systems or from very young igneous systems with no surface expression.

Other hot-spring systems are probably not related to silicic volcanic rocks. The heat of their systems is related to the regional geothermal gradient, which is higher in some regions such as the Great Basin than in others (Diment and others, this circular). Many hot springs of the Great Basin emerge from steeply dipping faults that may extend to depths of at least a few kilometres (Hose and Taylor, 1974; Olmsted and others, 1975). The water may be entirely of surface origin, circulating downward, being heated by thermal conduction with consequent decrease in density, and then rising and discharging from surface springs. In such systems, the normal conducted heat is being removed; temperatures immediately adjacent to the deep recharge channels are lower than those at similar depths not affected by convective heat losses. Temperatures should decline with time as rocks adjacent to channels are cooled and as new heat is supplied by conduction through increasing distances from channel walls. In our opinion, the abundant fault-controlled spring systems of low temperature throughout the Great Basin are likely to be of this origin. We suspect, however, that systems such as Beowawe, Leach, and Bradys in Nevada require volcanic heat and are not supplied only by geothermal gradient, even though located within the Battle Mountain high where conductive heat flow is considerably higher than the normal heat flow of the Great Basin (Diment and others, this circular). We, with R. L. Smith (oral commun. 1975), are skeptical that geothermal gradient alone can sustain high temperatures for the long durations of time indicated for these systems.

Identified systems

The accompanying tables are based on the scanty data available to us early in 1975. Sixty-three systems have indicated temperatures above 150°C (table 4 and figs. 1 and 2), and 224 have indicated temperatures between 90°C and 150°C (table 5 and figs. 2 and 3). Numerous hot springs in the range of 50° to 90°C (Waring, 1965) have not been included because geochemical and other evidence is lacking to suggest reservoir temperatures greater than 90°C. As additional data become available,some of these will no doubt qualify for higher temperature categories.

The more prominent systems have well-established names from local usage and literature. In most instances the name appearing on the topographic map of the area or the name given by Waring (1965) is used. If more than one name is available locally or in the literature for a particular spring, the additional names are shown in parentheses in the tables. Other springs or wells without established names are identified by some nearby geographic feature on available maps, which also provide latitude and longitude.

Measured surface temperatures provide minimum reservoir temperatures. Where the chemical temperatures T_{SiO_2} and $T_{Na\text{-}K\text{-}Ca}$ both indicate temperatures above about 125°C, we are confident that most subsurface temperatures will equal or exceed the predicted temperature. The user of these tables, however, should be especially skeptical of temperatures that are below 125°C, as well as temperatures that differ between the two chemical methods by more than about 20°C. Other systems whose predicted temperatures warrant skepticism are those of moderately high discharge (more than about 50 lpm from a single spring or obout 200 lpm from a system) that also have surface temperatures much below boiling (70°C or less). An indicated high subsurface temperature is credible for a cool spring of low discharge where excess heat can be lost by conduction but is much less credible for a system combining a low surface temperature and a high rate of discharge. Geochemical temperatures in most but not all cases provide minimal estimates of subsurface temperatures. Note that we have predicted some reservoir temperatures that are near the average rather than the maximum geochemical temperature. In most cases, our predicted temperature is at least as high as the preferred geochemical temperature (generally T_{SiO_2}); however, in some systems where subsurface temperature projections have been made (most notably by Olmsted and others, 1975), the assumed reservoir volume includes a substantial part that may be less than the indicated geochemical temperature.

The subsurface area assumed to be underlain by a reservoir of the indicated average temperature is derived from all available data. These include, as minimum, the surface area containing springs, spring deposits, and bleaching from attack by sulfuric acid derived from oxidation of H_2S. Geophysical data (Combs and Muffler, 1973), where available, provided the principal means for estimating the area and, in a few cases, the indicated depth of the reservoir, even though sufficient drilling has not yet been done to document carefully the relation between a geophysical anomaly and geothermal potential. Parts or all of some electrical resistivity anomalies may be caused by hydrothermal alteration, rocks rich in clay minerals, or saline ground waters, particularly in many areas of the Basin and Range province. Other types of geophysical surveys may also indicate anomalies that are not closely related to geothermal reservoirs. In most instances where surface expression and geology were used to indicate reservoir dimensions and geophysical data were then examined, the reservoir dimensions either remained the same or, more commonly, were significantly increased.

Although the pattern of industry exploration and drilling activity is viewed as highly significant in indicating the extent of a reservoir in several areas, in general only scanty data are available now from private industry. The lack of reliable data concerning areal extent is a serious constraint in this assessment because many estimates of the subsurface areas shown in tables 3 to 5 differ by more than three orders of magnitude; in contrast, all other parameters vary by less than one order of magnitude. Thus, the areal extent is the most critical single parameter in estimating the heat content of a system. Temperature, however, is of critical importance in determining how a system may be utilized. Systems with minimal surface evidence, such as a single spring, a restricted group of springs, or a single thermal well without other evidence, and systems for which geology or geophysics do not suggest a larger subsurface area are arbitrarily assigned a subsurface area of 1.5 km^2 (assumed to be 1½ km long on the dominant structural trend, even if unknown in direction, and 0.5 km on each side of this trend). Many of the separate systems we have indicated may be interconnected at depths greater than 2 or 3 km.

The heat reservoir of all convection systems is arbitrarily assumed to extend to 3 km in depth, which is the current limit of geothermal drilling. Heat at greater depths in volcanic systems is included in the volcanic system resources (Smith and Shaw, this circular); heat below 3 km in depth in other areas is included in the resource base calculations for conduction-dominated regions (Diment and others, this circular). A convection system in the latter environment has removed heat, relative to surrounding ground, as previously noted.

The top of a convective reservoir is generally not well defined but is generally assumed to have an average depth of 1, 1 ½, or 2 km, depending on assumed shape of the convection system and inferred similarities to drilled areas. Although the differences among our various depth estimates (tables 3 to 5) clearly affect drilling costs, the tables show that assumed thickness introduces much less variation in calculated volumes and heat contents than the assumed areas.

The tabulated volumes are simple multiplications of the assumed areas and thicknesses. Estimated stored heat is then calculated from reservoir temperatures (less 15°C, ambient surface temperature; for simplicity, assumed constant for all of the United States), volume, and volumetric specific heat assumed as 0.6 cal/cm^3°C. Volumetric specific heats are known to differ slightly by rock type, porosity, and water content (Diment and others, this circular), but the assumption of a single volumetric specific heat introduces only slight errors relative to the great uncertainties of other parameters.

Little is known about the specific intermediate-temperature systems of table 5 and figures 2 and 3. Most of these systems are included in this category because of their chemically indicated temperatures but are listed with minimal reservoir areas, volumes, and heat contents. One notable exception is the Bruneau-Grandview area of Idaho, shown on table 5 as having an area of 2,250 km^2 and 263×10^{18} cal of stored heat. This large area in the southwestern part of the Snake River Plain is characterized by hot springs of modest temperature (commonly 35° to 45°C; Waring, 1965) and many shallow thermal wells that discharge at temperatures as high as 84°C. In addition to this broad distribution of thermal springs and wells, the regional heat flow is probably high to very high (Diment and others, this circular), and geophysical surveys show no sharp boundaries for the area known to be anomalous. This geothermal area is likely to be huge, and it may even extend under a large part of the Snake River Plain.

Even less is known about our low-temperature hydrothermal resources (<90°C). Many spring systems tabulated by Waring (1965) are probably in this category, and the warmer ones may be useful in space heating. For example, Iceland and Hungary make extensive use of water at temperatures below 100°C, and 80°C is actually the preferred distribution temperature in Reykjavik, Iceland (Einarsson, 1970).

Pattern of distribution of identified convection systems

Figures 1 and 3 confirm the well-known abundance of thermal systems in the Western United States and their scarcity elsewhere. Most of the high-temperature systems occur in the areas of anomalously high conductive heat flow (Diment and others, this circular, figs. 9 to 11); many of these systems also occur in or near areas of young volcanic rocks (Smith and Shaw, this circular, figs. 5 to 7).

The numerical data of tables 4 and 5 are summarized in table 6, which also divides the systems into two categories, depending on whether the predicted magnitude of their heat reservoirs exceeds the minimum assumed value.

Note that the heat contained in identified hot-water systems is about 30 times that in vapor-dominated systems, and total heat contained in systems with indicated temperatures above 150°C is about the same as that in systems between 90°C and 150°C. Such comparisons of systems of different types must be tempered by the extent of our knowledge of each type; for obvious reasons, much more attention has been given to the more attractive large high-temperature systems. Six of the high-temperature systems (Surprise Valley, Long Valley, Coso Hot Springs, Salton Sea, and Heber, California, and Yellowstone National Park, Wyoming) are each predicted to contain more than 10×10^{18} cal of stored heat; they total about 75 percent of the total estimated heat of all of the identified high-temperature systems. Even more striking is the dominance of a few large systems in the intermediate-temperature range. Only two identified systems are predicted to contain more than 10×10^{18} cal each, and only seven contain more than 1×10^{18} cal. The dominance of the Bruneau-Grandview area of Idaho is especially startling; this may be more a reflection of a lack of adequate data and reliable predictive technique than of fact. However, geothermal convection systems may have the same log-normal relation between grade and frequency that metalliferous deposits and hydrocarbon reservoirs have. If this is so, relatively few systems contain most of the resources.

Undiscovered convective systems

Good reasons exist for optimism that abundant geothermal resources in hot-water convective systems are available for future discovery. Our use of the term "discovery," however, must be defined; a geothermal discovery is considered to result from any of the following:

1. New knowledge of the extent of an already identified system that increases its tabulated volume appreciably; the difference is considered to be the newly discovered part (but this may be offset in part by decreased estimates for individual systems).
2. The temperature of an identified system is found to be higher than first estimated—enough for the system to qualify for a higher temperature category and more valued potential utilization (but increases may also be offset, probably in small part, by decreases).
3. A previously unknown system is discovered, commonly with no obvious surface evidence for its existence.

Most of the tabulated convection systems of this report (tables 4 and 5) should be viewed as targets for future exploration and discovery.

Our reasons for being optimistic that many exploitable hot-water systems exist for future discovery are:

1. Many of the young silicic volcanic systems tabluated by Smith and Shaw (this circular) have no recognized convection systems.
2. Other young silicic systems may still be developing, with no direct evidence for their existence in the shallow crust.
3. With few exceptions, old, deeply eroded volcanic systems are associated with exten-

Table 6.—Summary of identified hydrothermal convection systems

	Number	Subsurface area, km^2	Volume, km^3	Heat Content, 10^{18} cal
Vapor-dominated systems (∿240°C)	3	122	194	26
Hot-water systems, identified				
High-temperature systems (<150°C)				
Systems each with heat content $>0.2 \times 10^{18}$ cal	38	1374	2939	366
Systems each with heat content $\leq 0.2 \times 10^{18}$ cal	25	40	56	5
Total high-temperature systems	63	1414	2995	371
Intermediate-temperature systems (90°-150°C)				
Systems each with heat content $>0.2 \times 10^{18}$ cal	28	2638	4112	311
Systems each with heat content $\leq 0.2 \times 10^{18}$ cal	196	300	452	34
Total intermediate-temperature system	224	2938	4564	345
Total identified hot-water systems	287	4352	7559	714
Total hydrothermal convection systems	290	4474	7753	740

sive hydrothermal alteration. Until recently, such alteration was interpreted as the effect of magmatic fluids, perhaps much different from the large convection systems of Larderello, The Geysers, Wairakei, and the Imperial Valley fields. However, extensive isotope studies of waters and rocks of both the old and the presently active systems have shown that local waters of surface origin are generally the dominant fluid (Taylor, 1974; White, 1974); the active systems are probably the present-day equivalents of old ore-forming systems. The volumes of altered rocks of the ore-forming systems are commonly many tens or hundreds of cubic kilometres. Furthermore, the isotope studies also demonstrate that each volume of altered rock commonly required the flow of 1 to 10 volumes of water through the system. The isotopic and other data also indicate that temperatures of these old systems most frequently ranged from 200° to 400° C at probable depths of 1 to 4 km below the ground surface of the time. If this analogy is correct, many active systems should have similar volumes and temperatures in their deeper parts.

4. Many old volcanic systems probably still sustain moderate- to high-temperature convection systems that may not have surface expression. Most of these volcanic systems are too old or poorly known to be evaluated in detail (Smith and Shaw, this circular).
5. Recent major progress has been made in applying several kinds of chemical, isotopic, and thermodynamic mixing models to convection systems that differ from the simple model (Fournier and Truesdell, 1974; Truesdell and Fournier, 1975). Different levels of mixing with dilute, cool meteoric waters are probably involved. With proper sampling of springs and shallow wells, evidence for high temperatures at deeper levels can be obtained; such evidence is normally lost by re-equilibration in a hot reservoir of a simple system. Reassessment of data from many of the systems of tables 4 and 5 and from other inconspicuous systems of low surface temperature is likely to result in many new discoveries, as we have defined the term.

We are fully aware that some extensively explored areas are better known to some others than to us, especially in light of the recent rapid rate of accumulation of proprietary data by industry. In time, some of these data will become available, and our techniques, estimates, and assumptions will improve enough to justify a new assessment.

We estimate that five times the volume and heat contents of the high- temperature (>150°C) systems of table 4 (excluding Yellowstone Park) are not presently recognized and exist as targets for future discovery. We cannot specifically justify this number other than to emphasize our previously stated reasons for optimism; a factor of 2 is almost certainly too small, and 20 is likely to be too large. We estimate that about three times the volume and heat content of the intermediate-temperature resources of table 5 are unrecognized, but this may be conservative.

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Igneous-Related Geothermal Systems

By R. L. Smith and H. R. Shaw

This preliminary survey of the geothermal resource base associated with igneous-derived thermal anomalies in the upper 10 km of the crust is a tentative first approach to an extremely complex problem. Our results are unavoidably speculative, and they emphasize the paucity of quantitative knowledge in the field of geothermal energy. Many of the data gaps are the same ones that exist in the fields of igneous petrology and volcanology and, to some extent, in the field of igneous-related ore deposits. Relevant research in these fields can go far toward solving many geothermal problems and vice versa.

Our approach to numerical estimates of igneous-related heat contents rests on estimates of the probable volumes of high-level magma chambers [2] and determinations of the radiometric ages of the youngest volcanism from those chambers combined with simple thermal calculations based on these values. In the following discussion these quantities are symbolized V_B (best volume) and Ty (last eruption), respectively, and are listed with the corresponding heat content estimates in table 7. This list contains the most important volcanic systems for which we were able to obtain data as well as many that may be of interest but for which data are lacking.

Mathematically, our thermal calculations contain two major assumptions: (1) heat transfer in rocks surrounding the magma chamber is by solid-state conduction and (2) effects of magmatic preheating and gains of magma after the time Ty are ignored.

The first assumption means that we neglect heat losses caused by mass-transfer mechanisms in surrounding rocks; specifically, we have not attempted to account for heat losses by hydrothermal convection systems. The second assumption means that, within the accuracy of V_B and Ty, our calculations of total magmatic heat yield minimal estimates because both magmatic preheating and gains of magma are additive.

Whereas the present rates of hydrothermal heat transfer are known for a few systems such as those of Long Valley, California, Steamboat Springs, Nevada, and Yellowstone National Park, Wyoming, we defend the first assumption from the standpoint that there are no uniform quantitative criteria for adjustments. The same is true of the second assumption. The net effects of the two assumptions relative to the present igneous-related resource tend to compensate one another, but the proportional amounts are not known with any confidence. Therefore, our calculations can only be used as a reference scale of heat contents on the "dry" basis that must be modified as data accrue on both the hydrothermal and magmatic histories of specific systems.

The calculations presented in this report are based almost entirely on a series of working hypotheses and a general model for volcano evolution developed by us over a period of years, which is still incomplete (Smith and Shaw, 1973, and unpub. data). In its simplest form, the rationale for our model holds that basic rocks (basalts, andesite, and comparable magmas) are formed in the mantle and/or lower crust and rise to the surface through narrow pipes and fissures; the individual magma pulses are volumetrically small, and such systems contribute little stored heat to the upper crust until magma chambers begin to form at high levels. With the exception of the large oceanic volcanoes, basic magmas do not form large high-level storage chambers out of context with derivative [3] silicic magmas (dacites, rhyolites, and comparable derivative magmas). On the other hand, we think that silicic

[2] We use the term "magma chamber" in the most general sense. In our thermal calculations, however, we specifically refer to the region of the crust that, at the time Ty (time-zero of our calculations), is inferred to contain molten or partly molten rock (magma).

[3] The term "derivative" carries no specific connotation as to the ultimate origin of the silicic end of the magma series; that is, crustal melting is not ruled out.

(Text resumes on p. 73)

Table 7.—Magnitudes and heat contents of identified volcanic systems[1]

EXPLANATION OF SYMBOLS

AGE = T

- Ty - Last eruption
- Tys - Youngest silicic eruption
- Tyb - Youngest basic eruption
- Ts - Age (silicic)
- Tc - Age caldera eruption
- Tg - Greatest known age (composition unspecified)
- Tgs - Greatest age (silicic)
- Tgb - Greatest age (basic)
- Tb - Age (basic)

The assumed "time-zero" for calculating the present distribution of heat is underlined in the table; Tys is used wherever data exist.

AREA = A

- Ac - From caldera
- Av - From vent distribution
- As - From shadow
- Af - From fractures
- Au - From uplift
- Ag - From geophysical anomaly (unspecified)
  - Agg - Gravity
  - Agm - Magnetic
  - Ags - Seismic
  - Ago - Other, see remarks
- Ao - Other, see remarks

[1]/Volcanic systems marked by an asterisk in column 1 are known to have some associated hydrothermal activity (see Renner, White, and Williams, this volume). Heat content calculations in columns 11-13 ignore hydrothermal losses. Methods of calculation are outlined below.

VOLUME = V

Vc - From caldera

Vv - From vent distribution

Vs - From shadow

Vf - From fractures

Vu - From uplift

Vg - From geophysical anomaly

Vgg - Gravity

Vgm - Magnetic

Vgs - Seismic

Vgo - Other, see remarks

Vo - Other, see remarks

Vee - From extrapolation of silicic ejecta volume.

Vb - Best estimate

NOTES ON THERMAL CALCULATIONS IN TABLE 7

COLUMN 11, ΔQ, total calories x 10^{18}

Assumptions: Initial temperature = 850°C

Latent heat of crystallization = 65 cal/g

Heat capacity = 0.3 cal/g /°C

Mean density of magma = 2.5 g /cm^3

The above values are approximate averages for the composition and temperature ranges of table 7. From these values the heat liberated between 850°C and 650°C is 125 cal/g. The total heat liberated between 850°C and 300°C is 230 cal/g.

One cubic kilometre of magma represents 2.5 x 10^{15} g. The total heat liberated per cubic kilometre is 0.575 x 10^{18} calories. This number multiplied by the volume V_B in column 9 gives the ΔQ Total of column 11.

Estimates of heat content in the tables are given in units of 10^{18} calories.

COLUMN 12, ΔQ now, calories x 10^{18}

The time required for a change of the original gradient at the Earth's surface to a steady-state gradient between the surface temperature and the magma chamber temperature is given approximately by relations discussed by Jaeger (1964). For the assumed depth of cover of 4 km and a thermal diffusivity of 0.007 cm^2/sec, this time is about 360,000 years. Where Ty is much younger than this time, the total heat remaining in the system now (column 12) is assumed to be about the same as the total value in column 11. Estimates of losses for older systems require detailed calculations of the disturbance of the geothermal gradient.

The value of thermal diffusivity used is an average estimate for crustal rocks. Roof rocks above large caldera systems such as Yellowstone, Idaho-Wyoming (IW-2), Valles, New Mexico (NM-1) and Long Valley, California (C-3) may have smaller values of conductive thermal diffusivity. Hydrothermal convection systems, however, can increase the effective value of thermal diffusivity by a significant amount, depending on average permeabilities of roof rocks.

COLUMN 13, ΔQ out, calories x 10^{18}

The total amount of heat transfer per square centimetre from a magma chamber into roof rocks is given by Carslaw and Jaeger (1959, p.61) and also is discussed by Shaw (1974). Using these relations the total heat transfer (ΔQ_{out}) in column 13 is given by

$$\Delta Q_{out} = 51.6At^{1/2} \text{ (cal)}$$

where A is contact area (from column 7 converted to square centimeters) and t is the time in seconds since Ty. Calculations in column 13 are approximately valid only if the time of solidification is greater than Ty in column 6. The time of solidification is approximated by lines 3 and 4 in figure 4.

If Ty is much greater than 360,000 years and the time for solidification, the calculation of heat content is ambiguous because of the increasing importance of hydrothermal losses. On the basis of conduction models, however, the total time for decay of igneous-related thermal anomalies may be very long. As an example, the time required for the central temperature in a magma chamber of horizontal slab like geometry to decay from the initial magma temperature to nearly ambient temperature is about 2 m.y. for a magma chamber 5 km thick and about 10 m.y. for a 10-km-thick chamber. Even a liberal allowance for hydrothermal losses means that the igneous-related thermal anomalies for the largest systems of table 7 probably are preserved for times of the order 10 m.y. or longer.

Queries in columns 11-13 mean that even though data exist, we are not confident that they pertain even approximately to the assumptions of the calculations. Blank spaces in the table mean that more geological and geochronological study is needed before we are willing to provide estimates.

Table 7.—Magnitudes and heat contents of identified volcanic systems—Continued

ALASKA

1 S-S No. *	2 Name of Area	3 Lat.	4 Long.	5 Composition Last Eruption	6 Age Data	7 Chamber Area Km^2	8 Chamber Volume Range Km^3	9 Chamber Volume V_B Km^3	10 Solidification State °C	11 ΔQ Total Calories $\times 10^{18}$	12 ΔQ Now Calories $\times 10^{18}$	13 ΔQ Out Calories $\times 10^{18}$	14 Remarks
A-1	BULDIR	52°23'N	176°01'E	BASIC	$<2\times10^3$?								>10 KM DEPTH
A-2	KISKA	52°06'N	177°36'E	BASIC	ACTIVE								>10 KM DEPTH
A-3	SEGULA	52°01'N	178°08'E	BASIC SILICIC ?	$<10^4$?								NEED BETTER DATA ON COMPOSITION AND AGE
A-4	DAVIDOF	51°58'N	178°20'E	NO DATA	$<10^4$?	5 A_c	12.5-50 V_c	12.5	>650°	7	7		
A-5	LITTLE SITKIN	51°57'N	178°32'E	BASIC	ACTIVE	17.3 A_c	45-180 V_c	75	>850°	43	43		
A-6	SEMISOPOCHNOI (CERBERUS)	51°56'N	179°35'E	BASIC	ACTIVE	42.4 A_c	106-424 V_c	150	>850°	86	86		
A-7	SUGARLOAF	51°54'N	179°38'E	BASIC	$<10^4$?								>10 KM DEPTH
A-8	GARELOI	51°48'N	178°48'W	BASIC	ACTIVE								>10 KM DEPTH
A-9	TANAGA	51°53'N	178°07'W	BASIC ?	ACTIVE	85.9 A_c	215-860 V_c	400	>850°	230	230		
A-10	TAKAWANGHA	51°52'N	178°00'W	BASIC ?	$<10^4$	8.9 A_c	22.5-90 V_c	>22.5	>650°	13	13		
A-11	BOBROF	51°55'N	177°27'W	BASIC ?	$<10^4$?								>10 KM DEPTH ?
A-12	KANAGA	51°55'N	177°10'W	BASIC ?	ACTIVE	23.0 A_c	57.5-230 V_c	75	>850°	43	43		
A-13	MOFFET	51°56'N	176°45'W	BASIC	$<10^4$?								>10 KM DEPTH
A-14	ADAGDAK	51°59'N	176°36'W	BASIC	$<10^4$?								>10 KM DEPTH
A-15 *	GREAT SITKIN	52°04'N	176°07'W	BASIC	ACTIVE	1.8 A_c	4.5-18 V_c	>5	>850°	3	3		
A-16	KASATOCHI	52°11'N	175°30'W	BASIC ?	ACTIVE ?								>10 KM DEPTH ?
A-1[illegible]	[illegible]ONIUJI	[illegible]	[illegible]	BASI[illegible] ?	AC[illegible] ?								> [illegible] DEPTH ?
A-18	SERGIEF	52°19'N	174°23'W	NO DATA	NO DATA								NEED COMPOSITION AND AGE DATA

Table 7.—Magnitudes and heat contents of identified volcanic systems——Continued

ALASKA

1 S-S No. *	2 Name of Area	3 Lat.	4 Long	5 Composition Last Eruption	6 Age Data	7 Chamber Area Km^2	8 Chamber Volume Range Km^3	9 Chamber Volume V_b Km^3	10 Solidification State °C	11 ΔQ Total Calories $\times 10^{18}$	12 ΔQ Now Calories $\times 10^{18}$	13 ΔQ Out Calories $\times 10^{18}$	14 Remarks
A-19	Korovin	52°23'N	174°10'W	Basic	Active								>10 Km Depth
A-20	Kliuchef	52°19'N	174°09'W	No Data	No Data	28.6 A_c	70-280 V_c	>100		58			Need composition and age data
A-21	Sarichef	52°19'N	174°03'W	Basic ?	Active?								>10 Km Depth
A-22	Seguam	52°19'N	172°23'W	No Data	Active	20(2) A_c	100-400 V_c	200	>850	115	115		Appears to be a double caldera. Not reported. Needs investigation
A-23	Amukta	52°30'N	171°16'W	Basic ?	Active								>10 Km Depth ?
A-24	Chagulak	52°35'N	171°09'W	Basic ?	$<10^4$?								>10 Km Depth ?
A-25	Yunaska	52°39'N	170°39'W	No Data Basic ?	Active	12.1 A_c	30-120 V_c	40	>850	23	23		
A-26	Herbert	52°45'N	170°07'W	Basic?	$<10^4$?								>10 Km Depth ?
A-27	Carlisle	52°54'N	170°04'W	Basic?	Active								>10 Km Depth ?
A-28	Cleveland	52°49'N	169°58'W	Basic ?	Active								>10 Km Depth ?
A-29	Uliaga	53°04'N	169°47'W	Basic ?	$<10^4$?								>10 Km Depth?
A-30	Tana	52°50'N	169°46'W	Basic?	No Data								>10 Km Depth ?
A-31	Kagamil	52°58'N	169°44'W	Basic ?	Active								>10 Km Depth ?
A-32	Vsevidof	53°08'N	168°42'W	Silicic	Active								Need more data
A-33	Recheschnoi	53°09'N	168°33'W	Basic ?	$<10^4$								Need more data
[illegible] *	[illegible]kmok	[illegible]	[illegible]	[illegible]asic	[illegible]ctive 8×10^3 T_c	[illegible] A_c	[illegible]	[illegible]	>[illegible]	[illegible]	[illegible]		
A-35	Tulik	53°23'N	168°03'W	Basic	No Data								>10 Km Depth
A-36	Bogoslof	53°56'N	168°02'W	Basic	Active								>10 Km Depth

Table 7.—Magnitudes and heat contents of identified volcanic systems——Continued

ALASKA

1 S-S No. *	2 Name of Area	3 Lat.	4 Long.	5 Composition Last Eruption	6 Age Data	7 Chamber Area Km^2	8 Chamber Volume Range Km^3	9 Chamber Volume V_b Km^3	10 Solid-ification State °C	11 ΔQ Total Calories $\times 10^{18}$	12 ΔQ Now Calories $\times 10^{18}$	13 ΔQ Out Calories $\times 10^{18}$	14 Remarks
A-37	Makushin	53°52'N	168°56'W	Silicic ?	Active	3.6 A_c	9-36 V_c	>10	>850	6	6		
A-38	Table Top	53°58'N	166°40'W	Basic	No Data								>10 Km Depth
A-39 *	Akutan	54°08'N	166°00'W	Basic ?	Active	3.5 A_c	9-36 V_c	>10	>850	6	6		
A-40	Mt. Gilbert (Akun)	54°16'N	165°39'W	Basic ?	No Data								>10 Km Depth ?
A-41	Pogromni	54°34'N	164°42'W	Basic	Active								>10 Km Depth
A-42	Westdahl	54°31'N	164°39'W	No Data	Active								Need more data
A-43	Fisher	54°38'N	164°25'W	Basic	Active ? <2x10³ T_y	122.6 A_c	300-1200 V_c	600	>850	345	345		
A-44	Shishaldin	54°45'N	163°58'W	Basic	Active								>10 Km Depth
A-45	Isanotski	54°45'N	163°44'W	Basic ?	Active								>10 Km Depth ?
A-46	Roundtop	54°48'N	163°36'W	No Data	Active ? <2x10³ ?								
A-47	Amak	55°25'N	163°09'W	Basic ?	No Data								>10 Km Depth ?
A-48	Frosty	55°04'N	162°51'W	Silicic ? Basic ?	No Data								
A-49	Walrus (Morzhovoi)	55°00'N	162°50'W	Basic	No Data >10⁴ ?								>10 Km Depth
A-50	Dutton	55°11'N	162°16'W	Basic	No Data								>10 Km Depth
A-51	Emmons	55°20'N	162°04'W	Basic	Active	117.3 A_c	300-1200 V_c	600	>650	345	345		
A-52	Hague	55°22'N	161°59'W	Basic	Active								
A-53	Double Crater	55°23'N	161°57'W	Basic	Active ?								
A-54	Pavlof	55°25'N	161°54'W	Basic	Active								>10 Km Depth

Table 7.—Magnitudes and heat contents of identified volcanic systems——Continued

ALASKA

1 S-S No. *	2 NAME OF AREA	3 LAT.	4 LONG.	5 COMPOSITION LAST ERUPTION	6 AGE DATA	7 CHAMBER AREA KM^2	8 CHAMBER VOLUME RANGE KM^3	9 CHAMBER VOLUME V_B KM^3	10 SOLID-IFICATION STATE °C	11 ΔQ TOTAL CALORIES $\times 10^{18}$	12 ΔQ NOW CALORIES $\times 10^{18}$	13 ΔQ OUT CALORIES $\times 10^{18}$	14 REMARKS
A-55	PAVLOF SISTER	55°27'N	161°51'W	BASIC	ACTIVE								>10 KM DEPTH
A-56	DANA	55°37'N	161°12'W	SILICIC	NO DATA $<10^4$?	1.6 A_c	4-16 V_c	>5		>3			NEED MORE DATA
A-57	KUPREANOF	56°01'N	159°48'W	NO DATA BASIC?	ACTIVE								NEED MORE DATA
A-58	VENIAMINOF	56°10'N	159°23'W	BASIC	ACTIVE 3.7×10^3 T_c	50.4 A_c	125-500 V_c	200	>850	115	115		
A-59	BLACK (PURPLE)	56°32'N	158°37'W	SILICIC	$<10^4$?	6.9 A_c	17.5-70 V_c	>20		12	12	6	NEED MORE DATA
A-60	ANIAKCHAK	56°53'N	158°10'W	SILICIC	ACTIVE 3.6×10^3 T_c	55.6 A_c	140-560 V_c	225	>850	129	129		
A-61	CHIGINIGAK	57°08'N	157°00'W	BASIC ? SILICIC ?	ACTIVE								NEED MORE DATA
A-62	KIALAGVIK	57°12'N	156°42'W	NO DATA	NO DATA								NO DATA
A-63	PEULIK	57°45'N	156°21'W	BASIC ? SILICIC	ACTIVE ?	10.5 A_c	25-100 V_c	>30		17	17		SILICIC IN FOCUS OF BOTH CALDERA & MT. PEULIK; PERHAPS YOUNG BASIC ERUPTION ON N. FLANK PEULIK. NEED AGE DATA
A-64	MARTIN	58°09'N	155°24'W	BASIC ?	ACTIVE								>10 KM DEPTH ?
A-65	MAGEIK	58°12'N	155°15'W	?	ACTIVE								SILICIC DOMES ? NEED MORE DATA
A-66	NOVARUPTA	59°17'N	155°10'W	SILICIC	ACTIVE	8.1 A_c	20-80 V_c	50	>850	29	29		
A-67	MT. GRIGGS (KNIFE PEAK)	58°20'N	155°08'W	BASIC ?	ACTIVE								>10 KM DEPTH ?
A-68	TRIDENT	58°14'N	155°07'W	BASIC	ACTIVE								>10 KM DEPTH ?
A-69	KATMAI	58°16'N	154°59'W	BASIC	ACTIVE	8.1 A_c	20-80 V_c	>20	>850	12	12		
A-70	SNOWY	58°20'N	154°41'W	NO DATA	NO DATA								NO DATA
A-71	DENISON	58°25'N	154°27'W	NO DATA	NO DATA								NO DATA
A-72	STELLER	58°26'N	154°24'W	NO DATA	NO DATA								NO DATA

Table 7.—Magnitudes and heat contents of identified volcanic systems——Continued

ALASKA

1 S-S No. *	2 Name of Area	3 Lat.	4 Long.	5 Composition Last Eruption	6 Age Data	7 Chamber Area Km^2	8 Chamber Volume Range Km^3	9 Chamber Volume V_B Km^3	10 Solidification State °C	11 ΔQ Total Calories $\times 10^{18}$	12 ΔQ Now Calories $\times 10^{18}$	13 ΔQ Out Calories $\times 10^{18}$	14 Remarks
A-73	Kukak	58°27'N	154°21'W	No Data	Active?								No data
A-74	Devils Desk	58°29'N	154°17'W	No Data	No Data								No data
A-75	Kaguyak	58°37'N	154°05'W	Silicic	$\le 10^4$?	41 A_c	10-40 V_c	>15		9	9	3	Need age data
A-76	Fourpeaked	58°47'N	153°42'W	No Data	$< 10^4$?								No data
A-77	Douglas	58°52'N	153°33'W	No Data	Active?								No data
A-78	Augustine	59°22'N	153°25'W	Silicic	Active								Need geophysical data?
A-79	Iliamna	60°02'N	153°06'W	Basic	Active								>10 Km depth
A-80	Redoubt	60°28'N	152°45'W	Basic	Active								>10 Km depth
A-81	Double	60°44'N	152°35'W	No Data	No Data								No data
A-82	Black	60°51'N	152°25'W	No Data	No Data								No data
A-83	Spurr	61°18'N	152°15'W	Basic	Active								Caldera? Need more data
A-84	Drum	62°07'N	144°38'W	Silicic	$\le 1.8\times10^5 T_{YS}$ $1.8\times10^5 T_S$ $5\times10^5 T_B$	140 A_V	350-1400 V_V	400		230	~200	>100	
A-85	Sanford	62°13'N	144°07'W	No Data	$<3.5\times10^5 T_B$								Need data on parasitic vents high on flanks of Sanford (inaccessible?)
A-86	Wrangell	62°00'N	144°01'W	Basic?	Active $>1.7\times10^3 T_C$	15 A_c	37.5-150 V_c	50	>850	29	29		
A-87	White River	61°27'N	141°28'W	Silicic	1.5×10^3		80 V_{EE}	80	>850	46	46		
A-88	Edgecumbe	57°01'N	135°46'W	Basic? Silicic	Active? $<9\times10^3$ $9\times10^3 T_S$	74 A_V	185-740 V_V	250		144	144	60	Silicic in focus Basic on flanks

Table 7.—Magnitudes and heat contents of identified volcanic systems[1]—Continued

ARIZONA

1 S-S No. *	2 Name of Area	3 Lat.	4 Long.	5 Composition Last Eruption	6 Age Data	7 Chamber Area Km^2	8 Chamber Volume Range Km^3	9 Chamber Volume V_b Km^3	10 Solidification State °C	11 ΔQ Total Calories $\times 10^{18}$	12 ΔQ Now Calories $\times 10^{18}$	13 ΔQ Out Calories $\times 10^{18}$	14 Remarks
AZ-1	SAN FRANCISCO MOUNTAINS	35°21'N	111°41'W	BASALT RHYOLITE	8×10^2 Tyb 2×10^5 Tys $<3 \times 10^2$ Tyb	250 A_s	625 - 2500 V_s	1250	> 650°	719	719	316	SHADOW AREA AND VOLUMES SO DERIVED HIGHLY SPECULATIVE SEE DISCUSSION IN S-S PAPER MIGRATING SYSTEM
AZ-2	KENDRICK PEAK	35°24'N	111°51'W	SILICIC	1.9×10^6 Ts				< 650°				NEED MORE DATA
AZ-3	SITGREAVES PEAK	35°21'N	112°00'W	RHYOLITE	2.6×10^6 Tys 2.7×10^6 Ts 2.8×10^6 Tes				< 650°				YOUNGEST AGE ON GOVERNMENT MOUNTAIN
AZ-4	BILL WILLIAMS MTN	35°12'N	112°12'W	SILICIC	3.5×10^6 Ts 4.1×10^6 Ts				< 650°				NEED MORE DATA PROBABLY TOO OLD FOR SIZE

CALIFORNIA

1 S-S No. *	2 Name of Area	3 Lat.	4 Long.	5 Composition Last Eruption	6 Age Data	7 Chamber Area Km^2	8 Chamber Volume Range Km^3	9 Chamber Volume V_b Km^3	10 Solidification State °C	11 ΔQ Total Calories $\times 10^{18}$	12 ΔQ Now Calories $\times 10^{18}$	13 ΔQ Out Calories $\times 10^{18}$	14 Remarks
C 1 *	LASSEN PK	40°29'N	121°30'W	RHYODACITE	58 YRS Tys	80 A_v	200 - 800 V_v	400	>650°	230	230	2	
C 2 *	CLEAR LAKE	38°53'N	122°45'W	BASIC?	$<10^4$ Tyb 2×10^4 Tcc 9×10^4 Tys $\sim 10^6$ Tes	256 A_v 560 A_{ss}	640 - 2560 V_v 1390 - 5480 V_{ss}	1500	>650°	863	863	490	NOT CORRECTED FOR PENETRATION OF SHADOW (A_v) A_{ss} (-35 BOUGER CONTOUR)
C 3 *	LONG VALLEY	37°42'N	118°54'W	RHYOLITE	10^5 Tys 7×10^5 Tc 9×10^5 Tes	480 A_c	1200 - 4800 V_c	2400	>650°	1380	1380	430	CONSIDERED HERE AS A SYSTEM INDEPENDENT OF THE MONO-INYO DOMES, EXCEPT FOR SOLIDIFICATION STATE WHICH MAY BE INFLUENCED BY HEAT INPUT FROM INYO FISSURE SYSTEM
C 4 *	SALTON SEA	33°10'N	115°38'W	RHYOLITE	1.6×10^4? $<5.5 \times 10^4$ Tys	50 A_v	125 - 500 V_v	200	>650°	115	115	18	MORE AND BETTER AGE DATA NEEDED
C 5 *	COSO MTS	36°02'N	117°49'W	BASALT	3×10^4 Tyb 4×10^4 Tys 9×10^5 Tes	110 A_v	275 - 1100 V_v	650	>650°	374	374	65	32 RHYOLITE DOMES $< 9 \times 10^5$ yrs
C 6	MONO DOMES	37°53'N	119°00'W	RHYOLITE	$<10^3$ Tys $<7 \times 10^4$ Tes	130 A_{vf}	325 - 1300 V_{vf}	650	>650°	374	374	12	MONO-LONG VALLEY SYSTEMS COMPLICATED BY INYO DOME CHAIN WHICH INTERSECTS BOTH AND IS BEST EVIDENCE FOR VIABLE DEEP HEAT SOURCE
C 7	MEDICINE LAKE	41°35'N	121°37'W	RHYOLITE	$\leq 10^3$ Tys	64 A_c 74 A_v	160 - 640 V_c 185 - 740 V_v	300	>650°	173	173	7	
C 8	SHASTA	41°25'N	122°12'W	DACITE	$(5-9.5) \times 10^3$? Tys	50 A_{oso}	125 - 500 V_{oso}	300	>650°	173	173	10	BY ANALOGY TO CRATER LAKE AND LARGE GRAVITY LOW
C 9	SUTTER BUTTES	39°12'N	121°49'W	Basic?	$1.5 \times 10^?$ Tyb 1.6×10^6 Tys 1.9×10^6 Tes	40 A_v	100 - 400 V_v	100	< 650°	58	< 10	50?	RING OF SILICIC VENTS
C 10 *	MORGAN MTN DOMES	40°25'N	121°30'W	DACITE?	NO DATA PLEIST.?								NEEDS INVESTIGATION
C 11 *	WARNER MTNS (BOYD CREEK-SUGAR HILL DOMES) (SURPRISE VALLEY)	41°43'N	120°17'W	RHYOLITE	7.1×10^6 Tys 7.7×10^6 Ts								NEEDS INVESTIGATION MAY BE PART OF LARGE SUBJACENT PLUTON SEE COUGAR PEAK OREGON (SAME AGE)

Table 7.—Magnitudes and heat contents of identified volcanic systems——Continued

CALIFORNIA-CON.

1 S-S No. *	2 Name of Area	3 Lat.	4 Long.	5 Composition Last Eruption	6 Age Data	7 Chamber Area Km^2	8 Chamber Volume Range Km^3	9 Chamber Volume V_B Km^3	10 Solidification State °C	11 ΔQ Total Calories $\times 10^{18}$	12 ΔQ Now Calories $\times 10^{18}$	13 ΔQ Out Calories $\times 10^{18}$	14 Remarks
C 12 *	Bridgeport-Bodie Volcanic Complex Calif-Nevada	38°15'N ±10'	119°00'W ±10'	Basic	$2.5 \times 10^5 T_{yb}$ $1.1 \times 10^9 T_e$ $2.5 \times 10^6 T_{ys}$								Needs investigation may be large low grade resource
C 13 *	Lava Mountains	35°26'N	117°31'W	Silicic	No data Pleist.? Plio.		120 V_{oc}	>120	<650°	>69			Needs investigation as low grade resource. Volume probably greater by 5-10 times
C 14	Big Pine	37°03'N	118°19'W	Rhyolite	$9.8 \times 10^5 T_{ys}$	42 A_s	105-420 V_s	100	<650°	58	<20	30?	Highly speculative (V_s) elapsed time since last eruption seems too great for this to be a viable system
C 15	Olancha Domes	36°19'N	117°52'W	Rhyolite?	No data								Need age data
C 16	Jackson Buttes	38°20'N	120°43'W	Dacite?	No data Plio?								Need age data
C 17 *	Paoha Island, Mono Lake	38°00'N	119°03'W	Silicic	85 Yrs? Data not definitive								Need more data. Active fumaroles; considered as separate system from Mono Domes

HAWAII

1 S-S No. *	2 Name of Area	3 Lat.	4 Long.	5 Composition Last Eruption	6 Age Data	7 Chamber Area Km^2	8 Chamber Volume Range Km^3	9 Chamber Volume V_B Km^3	10 Solidification State °C	11 ΔQ Total Calories $\times 10^{18}$	12 ΔQ Now Calories $\times 10^{18}$	13 ΔQ Out Calories $\times 10^{18}$	14 Remarks
H-1 *	Kilauea	19°26'N	155°18'W	Basalt	Active T_{yb}	12.5 A_u	37.5-50 V_{ugs}	40	>850	23	23		Chamber probably a plexus of sills & dikes with <5%, of hot rock volume, molten at any one time
H-2	Mauna Loa	19°29'N	155°37'W	Olivine Basalt	25 Yrs Active T_{yb}								>5 Km depth
H-3	Hualalai	19°42'N	155°52'W	Olivine Basalt	1.74×10^2 T_{yb}								Potentially active. >5 Km depth
H-4	Mauna Kea	19°49'N	155°28'W	Hawaiite	Post-glacial T_{yb}								Potentially active? No historic eruptions. >5 Km depth
H-5	Haleakala	20°43'N	156°15'W	Mafic Ol. Basalt	2.25×10^2? T_{yb}								Potentially active. >5 Km depth

Table 7.—Magnitudes and heat contents of identified volcanic systems——Continued

IDAHO - WYOMING

1 S-S No. *	2 Name of Area	3 Lat.	4 Long.	5 Composition Last Eruption	6 Age Data	7 Chamber Area Km^2	8 Chamber Volume Range Km^3	9 Chamber Volume V_B Km^3	10 Solid-ification State °C	11 ΔQ Total Calories $\times 10^{18}$	12 ΔQ Now Calories $\times 10^{18}$	13 ΔQ Out Calories $\times 10^{18}$	14 Remarks
IW1	ISLAND PARK-HUCKLEBERRY RIDGE SYSTEM	44°17'N	111°19'W	BASIC IN WEST	$<10^6$ T_E 1.2×10^6 T_{YS} 2×10^6 T_C	3900 Ac ?	10000 - 40000 V_c	30000	≤ 650°	17250	7456 4026 - WEST PART	12264	COMPOUND SYSTEM WITH YELLOWSTONE CALDERA, TREATED HERE AS SINGLE SYSTEM. WEST PART PENETRATED BY BASALT AT 2×10^5 YRS
IW2 *	YELLOWSTONE CALDERA SYSTEM	44°44'N	110°30'W	RHYOLITE	6.9×10^4 T_{YS} 6×10^5 T_C	2500 Ac	6250 - 25000 V_c	15000	>650°	8625	8625	1900	YELLOWSTONE CALDERA TREATED AS INDEPENDENT FROM OLDER SYSTEM
IW3	BLACKFOOT DOMES	42°48'N	111°36'W	SILICIC?	4×10^4 T_S 8×10^4 T_S	25 A_V	60 - 240 V_V	100	< 650°	58	58	14	NEED GRAVITY SURVEY
IW4	BIG SOUTHERN BUTTE	43°18'N	113°00'W	SILICIC	3×10^5? T_S	NO DATA	100 Vee	100	< 650°	58	<58		NEED GRAVITY SURVEY AND DATA ON AREA

NEVADA

1 S-S No. *	2 Name of Area	3 Lat.	4 Long.	5 Composition Last Eruption	6 Age Data	7 Chamber Area Km^2	8 Chamber Volume Range Km^3	9 Chamber Volume V_B Km^3	10 Solid-ification State	11 ΔQ Total Calories $\times 10^{18}$	12 ΔQ Now Calories $\times 10^{18}$	13 ΔQ Out Calories $\times 10^{18}$	14 Remarks
N1 *	STEAMBOAT SPRINGS	39°23'N	119°45'W	RHYOLITE	1.2×10^6 T_{YS}	18 A_V	45 - 180 V_V	90	<650°	52	?	?	FAULT-CONTROLLED SYSTEM POSSIBLY RELATED TO CHAMBER MUCH LARGER THAN INDICATED
N2	SILVER PEAK	37°45'N	117°51'W	BASIC	4.8×10^6 T_{YB} 6.1×10^6 T_{YS}	40 Ac	100 - 400 V_c	200	<650°	115	?		RECENT BASALT CINDER CONE ON LOWER FLANKS

NEW MEXICO

1 S-S No. *	2 Name of Area	3 Lat.	4 Long.	5 Composition Last Eruption	6 Age Data	7 Chamber Area Km^2	8 Chamber Volume Range Km^3	9 Chamber Volume V_B Km^3	10 Solid-ification State °C	11 ΔQ Total Calories $\times 10^{18}$	12 ΔQ Now Calories $\times 10^{18}$	13 ΔQ Out Calories $\times 10^{18}$	14 Remarks
NM-1 *	VALLES CALDERA	35°52'N	106°34'W	RHYOLITE	$\sim10^5$ T_{YS} 1.1×10^6 T_C 1.4×10^6 T_C	400 Ac	1000 - 4000 V_c	3500	>650°?	2013	2013	~400	> 20 RHYOLITIC ERUPTIONS $<1.1\times10^6$ yrs V_{cc} SUBJECT TO REVISION. PROBABLY CLOSE TO SOLIDUS TEMPERATURE
NM-2	MT TAYLOR	35°14'N	107°36'W	BASALT?	2.74×10^6 T_{YS}								NEED MORE DATA
NM-3	NO AGUA DOMES	36°45'N	105°58'W	RHYOLITE	PLIOCENE?								NEED MORE DATA

Table 7.—Magnitudes and heat contents of identified volcanic systems—Continued

OREGON

1 S-S No. *	2 Name of Area	3 Lat.	4 Long.	5 Composition Last Eruption	6 Age Data	7 Chamber Area Km^2	8 Chamber Volume Range Km^3	9 Chamber Volume V_b Km^3	10 Solid-ification State °C	11 ΔQ Total Calories $\times 10^{18}$	12 ΔQ Now Calories $\times 10^{18}$	13 ΔQ Out Calories $\times 10^{18}$	14 Remarks
O-1	Crater Lake	42°56'N	122°09'W	Silicic ?	$>7\times10^3$? $6.6\times10^3\ T_c$	50 A_c	125-500 V_c	>320	>650	>184	>184	4	
O-2 *	Newberry	43°43'N	121°19'W	Rhyolite	$4\times10^2\ T_{rs}$ $>6.6\times10^3\ T_c$	32 A_c	80-320 V_c	100	>650	58	58	2	
O-3	South Sister	44°06'N	121°46'W	Rhyolite	$<2\times10^3\ T_{rs}$	30 A_v 40 A_s	75-300 V_v	100	>650	58	58	5	
O-4 *	Mt Hood	45°22'N	121°42'W	Andesite	Active								>10 Km Depth ?
O-5	Mt McLoughlin	42°27'N	122°19'W	Andesite	Active?								>10 Km Depth
O-6 *	Mt Jefferson Domes (Breitenbush ?)	44°45'N	121°49'W	Silicic	No Data Pleist.								Need age data
O-7	Melvin-Three Creeks Buttes	44°09'N	121°35'W	Dacite	No Data Pleist.								Need age data
O-8	Cappy-Burn Butte Area	43°19'N	121°59'W	Dacite	No Data Pleist.								Need age data
O-9	Odell Butte	43°28'N	121°52'W	Dacite	No Data Pleist.								Need age data
O-10	Black Butte	44°24'N	121°38'W	Dacite	No Data Pleist.								Need age data
O-11	Rustler Peak	42°37'N	122°21'W	Dacite ?	No Data Pleist.								Need age data
O-12	China Hat and East Butte	43°41'N 43°40'N	121°02'W 120°00'W	Rhyolite	$7.6\times10^5\ T_{rs}$ $8.5\times10^5\ T_{rs}$		85 V_{EE}	85	<650	49			See Quartz Mountain
O-13	Quartz Mountain	43°33'N	120°48'W	Rhyolite	$<1.0\times10^6$ 1.1×10^6		36 V_{EE}	36	<650	21			Need geophysical data. If 12 & 13 are part of same high system the chamber could be $>10^2$ KM^3
O-14	Glass Buttes	43°43'N	120°04'W	Rhyolite	$\leq 4.9\times10^6$		330 V_{EE}	330	<650	190	<10?	180?	Needs investigation ?
O-15	Cougar Mountain	43°25'N	120°53'W	Rhyolite	4.3×10^6		7 V_{EE}	>7	<650	>4	0?	>4?	
O-16	Cougar Peak Area	42°19'N ±30'	120°38'W ±20'	Rhyolite ?	$7.1-7.7\times10^6$ T_s				<650				Needs more detailed investigation including gravity data. May be underlain by pluton. See Calif. C-11, same age.
O-17 *	Harney-Malheur	43°15'N ?	119°08'W ?	Rhyolite ?	$\leq 6.0\times10^6$ 9×10^6		2500 V_{EE}	2500	<650	1438	?		Needs more detailed investigation especially more accurate location of chamber area. Probably is large low grade res.

Table 7.—Magnitudes and heat contents of identified volcanic systems—Continued

UTAH

1 S-S No. *	2 Name of Area	3 Lat.	4 Long.	5 Composition Last Eruption	6 Age Data	7 Chamber Area Km^2	8 Chamber Volume Range Km^3	9 Chamber Volume V_B Km^3	10 Solidification State °C	11 ΔQ Total Calories $\times 10^{18}$	12 ΔQ Now Calories $\times 10^{18}$	13 ΔQ Out Calories $\times 10^{18}$	14 Remarks	
U1 *	MINERAL MTS.	38°26'N	112°48'W	RHYOLITE	PLEIST.?	> 10		> 50		> 29	?		NEED MORE DATA PROBABLY VIABLE GEO-THERMAL SYSTEM	
U2 *	COVE CREEK DOMES	38°45'N	112°44'W	BASALT	PLEIST. T_B PLIO. T_S	94 A_V	235 - 940 V_V	400		230	?		NEED AGE DATA	
U3	WHITE MTN. RHYOLITE	38°55'N	112°30 W	BASALT	$< 10^4 T_{YB}$ $< 10^6 T_{YS}$	NO DATA							NEED AGE DATA	
U4	TUSHAR MOUNTAINS	38°25'N	112°25'W	RHYOLITE ?	$2 \times 10^7 T_S$	NO DATA							NEED MORE DETAILED DATA PROBABLY A LARGE CHAMBER AND COMPLEX SYSTEM	
U5	TOPAZ MOUNTAIN	39°42'N	113°08W	RHYOLITE	$6 \times 10^6 T_{YS}$ $1.6 \times 10^7 T_S$	NO DATA							NEED MORE DATA	
U6	SMELTER KNOLL	39°25'N	112°53'W	RHYOLITE	$3.4 \times 10^6 T_S$	NO DATA							NEED MORE DATA	

WASHINGTON

1 S-S No. *	2 Name of Area	3 Lat.	4 Long.	5 Composition Last Eruption	6 Age Data	7 Chamber Area Km^2	8 Chamber Volume Range Km^3	9 Chamber Volume V_B Km^3	10 Solidification State °C	11 ΔQ Total Calories $\times 10^{18}$	12 ΔQ Now Calories $\times 10^{18}$	13 ΔQ Out Calories $\times 10^{18}$	14 Remarks	
W1 *	MT BAKER	48°47'N	121°49'W	ANDESITE	ACTIVE								> 10 Km DEPTH	
W2 *	GLACIER PEAK	48°07'N	121°07'W	RHYODACITE	$1.2 \times 10^4 T_{YS}$ $< 7 \times 10^5 T_S$	5 ± 5 A_0	12.5 - 50 V_0	12.5 ± 12.5		8?	8?	4?	VOLUME BY ANALOGY TO ST. HELENS AND OTHER CASCADE VOLCANOES	
W3	MT RAINIER	46°51'N	121°45'W	ANDESITE	ACTIVE								> 10 Km DEPTH	
W4	MT ST HELENS	46°12'N	122°11'W	ANDESITE	ACTIVE $1.9 \times 10^2 T_{YB}$ $1.75 \times 10^3 T_{YS}$	5 A_V	12.5 - 50 V_V	> 12.5	> 850	> 8	> 8			
W5	MT ADAMS	46°12'N	121°29'W	ANDESITE?	ACTIVE?								NEED MORE DATA > 10 Km DEPTH ?	

magmas are always erupted from high-level storage chambers, probably in the upper 10 km of the crust.

This reasoning suggests to us that purely basic volcanic systems (most common on a world basis) rarely form thermal anomalies of economic interest, whereas silicic volcanic systems probably always do if they are large enough. In making this statement we are specifically excluding possible economic development of certain oceanic volcanic systems. In some oceanic systems there is the admitted possibility of the existence of high-level basic magma chambers.

The estimates given in table 7 are based on those volcanic systems showing evidence from the presence of silicic eruptives that a high-level magma chamber formed in the recent past or is forming at the present time. In table 7 we have tried to give conservative estimates for those silicic volcanic systems that we think have contributed the greatest original and present heat to the upper 10 km of crust.

The problem of the thermal estimates is many faceted, but our major concern is in estimating the volume of the magma chamber for a time when it is known to have contained magma. The range of volumes of silicic magma chambers of interest spans four orders of magnitude or more, and the volumes of most silicic volcanic systems can be approximated within an order of magnitude by inspection of a geologic map or by analogy with kindred volcanoes that are better known. For well-documented volcanic areas, other constraints allow estimates of chamber volumes that are probably within a factor of 2 or 3 of their true volumes. The age of the last known eruption of magma is taken as evidence for a magma chamber that was then at least partly molten; cooling from the time of that eruption is assumed to take place in a closed system. Because many silicic volcanic systems are not closed but continue to receive subchamber heating, this procedure gives minimum cooling times and heat contents for most systems. Cooling by hydrothermal convection tends to offset continued heating, but in our opinion *the rate of supply of magma from deep crustal or mantle sources is the dominant heat supply for both high-level magmatic and hydrothermal systems.*

For most systems the value of *Ty* used for thermal calculations is the age of the youngest silicic extrusive rocks (*Tys* in table 7). We also give estimates for heat contents of systems lacking known silicic extrusive rocks, but it should be recognized that the controlling concepts and assumptions are less applicable for those systems.

Our scheme for evaluating volcanic areas for their geothermal potential was conceived and developed as a guide for exploration rather than as a rigorous method for quantitative estimation of resources. As more and better data on geothermal areas become available, we hope that the scheme might ultimately evolve into a quantitative method for characterizing both developing and declining thermal anomalies associated with high-level igneous systems. The method as applied in this summary of igneous-related geothermal resources does not permit discussion of the many qualifiers that should accompany any detailed consideration of a given geothermal area. The estimates in table 7 should be viewed as first and incomplete approximations of an igneous-related resource about which little is known with any degree of certainty. If nothing else, however, the table gives a list of areas that may be of geothermal interest and gives some idea of the relative magnitudes of igneous-related heat contents that may exist in these areas. Figure 4 illustrates the criteria that we used to gage the present probable thermal state of a magmatic system.

Some of the fundamental data, from which all estimates are made, are shown in table 7. The basis of specific numbers in columns 6–9 is indicated by symbols as explained in the legend of table 7. Column 7 is based on various surface manifestations of volcanism, volcano-tectonics, geophysics, and silicic ejecta volumes. The volume range (column 8) is calculated by assuming that the thickness of the magma chamber ranges from 2.5 to 10 km. This volume range is reduced in column 9 to a single best estimate to simplify the thermal calculations for columns 11–13. In general we assume that the smaller the area, the smaller the thickness of the chamber. On an order of magnitude basis, this assumption must generally be correct.

Column 10 indicates our best guess on the present thermal state of a magma chamber and, thus, whether or not we think that magma exists in the system now. Many entries are shown as greater

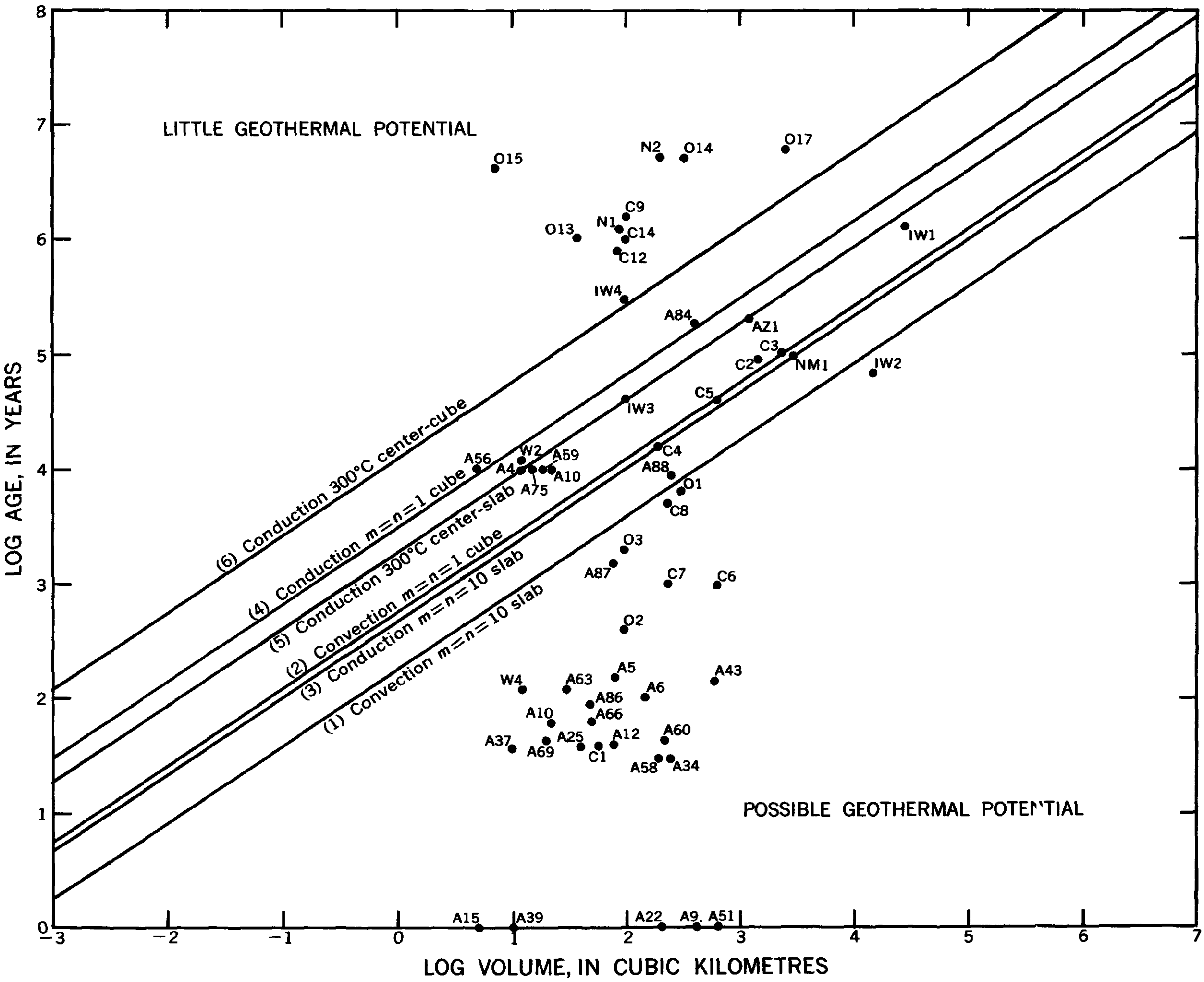

FIGURE 4.—Graph of theoretical cooling time versus volume for magma bodies. Points represent youngest ages (*Ty*) and estimated volumes (V_B) for the best known magmatic systems of table 7. See text for explanation of lines 1–6.

than or less than 650°C, which is the approximate minimum temperature of solidification of granitic melts. The assumed approximate liquidus (all liquid, no crystals) temperature for most silicic magmatic systems is about 850°C. Where the value >850°C is given, as in many of the active Alaskan systems, present ejecta compositions indicate that the chambers contain andesites or dacites of higher liquidus temperatures than rhyolites.

Column 11, designated ΔQ_{total}, is computed directly from column 9 and assumes that the entire heat content of the high-level storage chamber is contained in a fixed volume of magma. This assumption is clearly incorrect if a continuous convective supply of magma from deeper sources has been important in the origin and present state of high-level systems, as we maintain. The assumption leads to a minimum value for the total heat transfer in the upper crust derived from any given volcanic system. However, ΔQ_{total} is the only quantity that can be defended as having existed in the chamber at one time. All of these systems have histories involving preheating, and evaluation of the true total will require much more data on volumes, compositions, depths, shapes, rates of magma convection, differentiation, etc., than are available today; geochronological data are imperative in this context.

The total heat content of column 11 is the "initial" heat content of the magma chamber of volume V_B at the time Ty assumed in the calculations and is based on the total heat that would be liberated if the total chamber volume, assumed to be closed to additional inputs of magma, cooled from an initial average temperature of 850°C to a final average temperature of 300°C. The calculation includes the latent heat of crystallization and the heat capacity integrated over the temperature interval (see table 7, footnote). The lower cutoff is an arbitrary approximation of the average preexisting ambient temperature within the depth zone occupied by magma. In some cases, 850°C is either too low or too high an initial temperature. This uncertainty and the uncertainties of thermal properties mean that some of the listed values may be as much as 50 percent too high or too low. These uncertainties, however, tend to compensate because of the ranges of magma types and also because of the uncertainties of magmatic preheating and continued additions of magma. On this basis we believe that column 11 generally represents minimal estimates of the total igneous-related heat source. Hydrothermal convection does not affect this number because it is determined solely by the data on magma volumes and temperatures.

Column 12 is our estimate of that part of the heat content of column 11 that still remains in the ground, both within and around the original magma chamber. The estimate is based on a calculation of the time that would be required for the geothermal gradient at the Earth's surface to be significantly increased by conductive heat transfer in dry rocks over its original value if a magma chamber was suddenly emplaced with its top at a depth of 4 km. The calculation is sensitive to the assumed depth of the roof and the properties of roof rocks. If the time since the last eruption, Ty, when the total heat content existed mainly in the environs of the magma chamber, is much less than 300,000 years, then little of that initial heat has been lost at the Earth's surface unless there has been active hydrothermal convection in roof rocks for much of that time. In most places there is no way to evaluate this possibility so we have reported heat contents only on the basis of conductive models. Yellowstone is an obvious exception, but the magmatic-hydrothermal heat balances there are not discussed in this brief report.

In detail, the calculation leading to column 12 is very complicated and is strongly subject to all factors influencing heat-transfer mechanisms. More important uncertainties, however, are caused by the longevity of magmatic injection in the same vicinity both before and after the assumed value Ty. Again, we believe this to imply that the estimates of magma-related heat remaining in the ground are underestimates, probably grossly low, for systems with long histories of magmatic injection that have not had comparable hydrothermal losses.

Column 13 (ΔQ_{out}) gives the heat transferred from the magma chamber into the roof rocks between Ty and the present, calculated by the usual methods of conduction theory. This heat flow is a minimal estimate of the total amount of magmatic heat available throughout the life of the igneous event for the support of past and present hydrothermal systems within the assumed 4 km of roof rock. Numerous complications attend this calculation: magmatic heating prior to Ty will greatly increase the estimates, whereas hydrothermal convection losses decrease the estimates. Systems that have been volcanologically active over a significant time span and are still active clearly do not fit the primary assumption of the calculation, namely, that a single pulse of magma was emplaced instantaneously. Accordingly, we are unable to make any estimates of ΔQ_{out} for volcanoes labeled "active" in column 6. Generally speaking, the realistic magnitudes of heat transfer into roof rocks must be far greater than we can calculate from our simple assumptions.

Column 11, ΔQ_{total} is the sum of the heat now in the intrusion, the anomalous heat now in the roof rocks, and the anomalous heat lost to the atmosphere through the life of the igneous anomaly. Column 12 (ΔQ_{now}) is the sum of the heat now in the intrusion and the heat now in the roof rocks, whereas column 13 (ΔQ_{out}) is the sum of the heat in the roof rocks and the heat lost to the atmosphere. For igneous systems younger than about 300,000 years, the heat lost to the atmosphere is assumed to be small from the standpoint of conduction models, and ΔQ_{out} is equal to the anomalous heat now in the roof rocks. For systems older than 300,000 years, there has been significant escape of heat to the atmosphere, so that ΔQ_{now} is less than ΔQ_{total}, and ΔQ_{out} is greater than the heat now stored in the roof rocks. It should be noted that table 7 includes only a few of the many existing volcanic systems old enough for atmospheric losses to be significant. The amount of residual heat in these additional systems and the longevity of igneous intrusion and subchamber heating in all systems could increase our estimates of igneous-related heat contents by at least 2 and possibly up to 10 times. This problem, however, converges with the more general problems of regional heat flow that are not considered here (see Diment, this circular).

The total for the ΔQ_{now} of all magma-related systems of column 12 in the conterminous United States is roughly $23{,}000 \times 10^{18}$ cal or about 30 times the total estimated heat content of all tabulated hydrothermal systems (Renner and others, this volume). Of this total, about half (or $11{,}000 \times 10^{18}$ cal) probably exists as molten or partially molten magma. The total for molten magma in Alaska represents about $2{,}000 \times 10^{18}$ cal, which is nearly equal to its total "present heat content" of column 12. The total magmatic heat released to and presumably still contained by roof rocks in the conterminous United States is more than $6{,}000 \times 10^{18}$ cal (column 13 adjusted for surface losses). In some systems, part of this heat has been dissipated by hydrothermal activity. The real total, because of the residual effects of preheating, probably exceeds the estimates of hydrothermal systems by at least an order of magnitude.

The estimates given in table 7 include nearly all of the high-level systems in the United States known or inferred to contain magma, plus a few postmagmatic systems for which we have enough data to make a tentative estimate. The data on igneous-related heat contents presented in this tentative assessment, however, cannot be considered exhaustive. In table 7, for many of the silicic systems, column 12 indicates that virtually all of the original heat content remains in the ground on the basis of the simplified conduction models for dry rocks. The reason for this is that our reconnaissance was aimed at discovering the youngest silicic systems that may still contain magma, and usually these systems have the best surface exposures and are the most completely documented. Calculations of present heat content (ΔQ_{now}) for post magmatic systems are fewer because of the paucity of data. An exception, for example, is Sutter Buttes, California, for which good volcanological data exist but which is probably too old to represent an important thermal resource, as is indicated by the low present heat content of column 12.

There are undoubtedly systems in the transitional range between those that still retain nearly all of their original heat and those that retain virtually none. These are the systems that also would plot somewhere within the transitional band between lines 1 and 6 in figure 4, between subsolidus systems that are inferred to have pluton temperatures near 300°C and systems that are inferred to have chambers still entirely molten. Some of the transitional systems are represented in table 7 among the entries lacking sufficient data for heat-content calculations, but others may be added as data become available. Inclusion of these systems would probably increase the total heat-content estimate by several times.

Additionally, there are several broad areas in and peripheral to the Basin and Range province where fault-controlled warm and hot springs seem clustered near former silicic volcanic loci. The data necessary to confirm or deny a relation between the thermal water and subjacent igneous bodies are inadequate. We think, however, that estimates of the low-grade geothermal resources found to be related to silicic igneous systems eventually will be revised upward and that old and very large silicic plutons may retain some heat not accounted for by currently proposed thermal models. Deep faulting superimposed on these older plutons allows deep penetration and circulation of ground waters. Three of these areas are listed in table 7 as examples of the kind of systems we have in mind (O17, Harney-Malheur basin, Oregon; C12, Bridgeport-Bodie area, California; O16–C11, Cougar Peak-Surprise Valley area, Oregon-California).

In figure 4 the age-volume (Ty, V_B) data for 54 volcanic systems are plotted to show the approximate present position of each system in relation to its probable solidification state and to the 300°C isotherm. This plot is the essence of our scheme for reconnaissance evaluation of silicic volcanic areas (Smith and Shaw, 1973).

Pairs of lines in figure 4 are drawn to represent a spectrum of cooling models that identify igneous systems that are now approaching ambient temperatures (points above lines 5 or 6), systems that may now just be approaching the postmagmatic stage (points between lines 3 and 4), and systems that probably still have magma chambers with a large molten fraction (points below lines 1 or 2). The pairs of lines represent

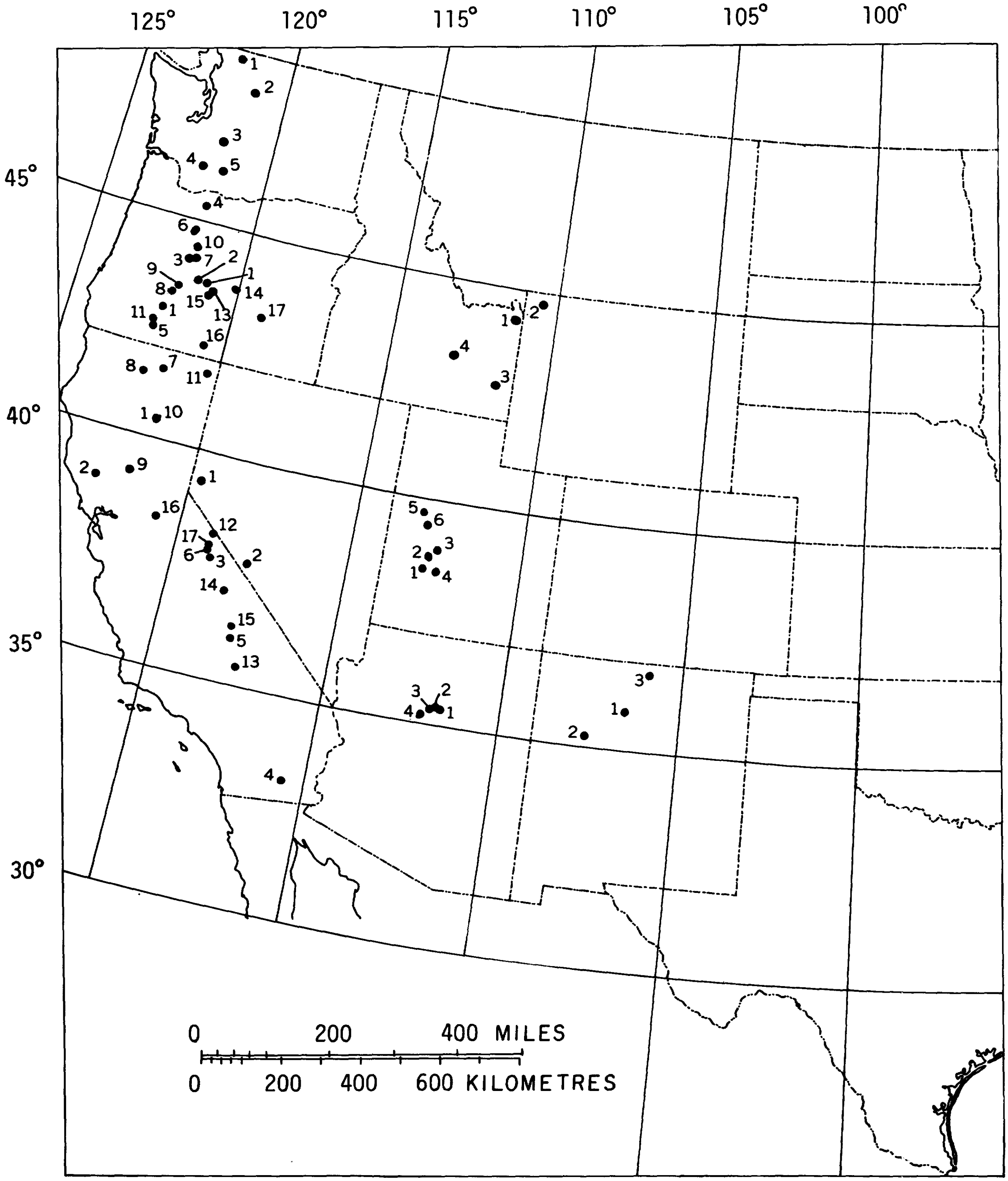

FIGURE 5.—Identified volcanic systems in the conterminous United States. Numbers are the same as in table 7.

the effect of shapes ranging from slablike to equant for each of the different cooling models.

Lines 1 and 2 assume that cooling takes place by internal convection of the magma chamber until solidification is nearly complete. Lines 3 and 4 assume that cooling is entirely by conduction, both inside and outside the magma chamber, until solidification is complete. Even in the presence of convection inside the magma chamber, however, lines 3 and 4 come closest to representing the range of true solidification times for most chambers because magmatic convection tends to stagnate long before solidification is complete (Shaw, 1974). Lines 5 and 6 represent an estimate of the time required before the central temperature of the solidified pluton has fallen to about 300°C. Hydrothermal activity will cause these lines to shift toward lines 3 and 4, respectively, but it is our opinion that in most cases the shift is not large and the positions are fairly realistic.

The points plotted in figure 4 can be compared with the lines only on the assumption that the plotted volume V_B was instantaneously emplaced and cooled from the time represented by the youngest age *Ty*. This assumption is usually questionable, so that the relation of each point to the cooling models must be individually examined in detail. The effects of magmatic preheating and continued supply of magma, of course, tend to shift points lower on the diagram relative to models of instantaneous emplacement of magma chambers that cool as closed systems.

Basic (or basaltic) volcanic systems, except for the special case of Kilauea, Hawaii, are given little emphasis in this resource estimate because they probably do not contain a significant high-level thermal anomaly and because we do not yet have the data necessary to evaluate them even if they did.

Young basic volcanoes, however, are indicators of magma source regions in the mantle and under some conditions are potential indicators of buried high-level silicic bodies with no obvious surface manifestations. Future investigations may reveal these hidden silicic bodies by geophysical studies, studies of xenoliths in the basic rocks, or by other means. The common association of silicic domes and lavas with basaltic lava fields shows that basaltic systems should not be automatically rejected for geothermal exploration. However, completely hidden silicic magma bodies associated with basaltic lava fields at the present time probably are few and small. Exceptions to this generalization may exist, such as the "Lassen-Shasta rectangle" of table 8, but in those the possibility is also indicated by abundant volcanological evidence. Figures 5, 6, and 7 illustrate the geographic distributions of the silicic and basaltic systems of tables 7 and 8 for the United States.

Table 8 contains a partial list of known basic lava fields and cinder cones known or inferred to be less than about 10,000 years old. Whereas table 7 identifies systems by principal eruptive centers, table 8 identifies broader regions of distributed basaltic volcanism. Ultimately these distributed basaltic systems, and perhaps older ones not listed in table 8, may form some small part of the resource base, but they cannot be evaluated in the present context.

A similar line of reasoning applies to a large number of simple andesitic stratovolcanoes of Alaska and the Cascade Mountains that is, the ones for which no volume estimates are given in table 7. These are listed because they are active or potentially active volcanoes and clearly reflect viable thermal sources. In our opinion, however, the significant thermal anomaly for most of these volcanoes lies below 10 km and is so indicated in the table. Future studies may show that some of these volcanoes have evolved or are evolving higher level storage chambers, but we think that few of them will contribute greatly to the total resource base above the 10-km level, though they may be important locally.

We wish to thank R. G. Luedke for help in the final drafting of table 7 and figure 4 and Manuel Nathenson for preparations of figures 5, 6, and 7. We also thank T. P. Miller for reviewing table 7 for errors of fact for the Alaskan volcanic systems. Time and space do not permit listing source references for tables 7 and 8.

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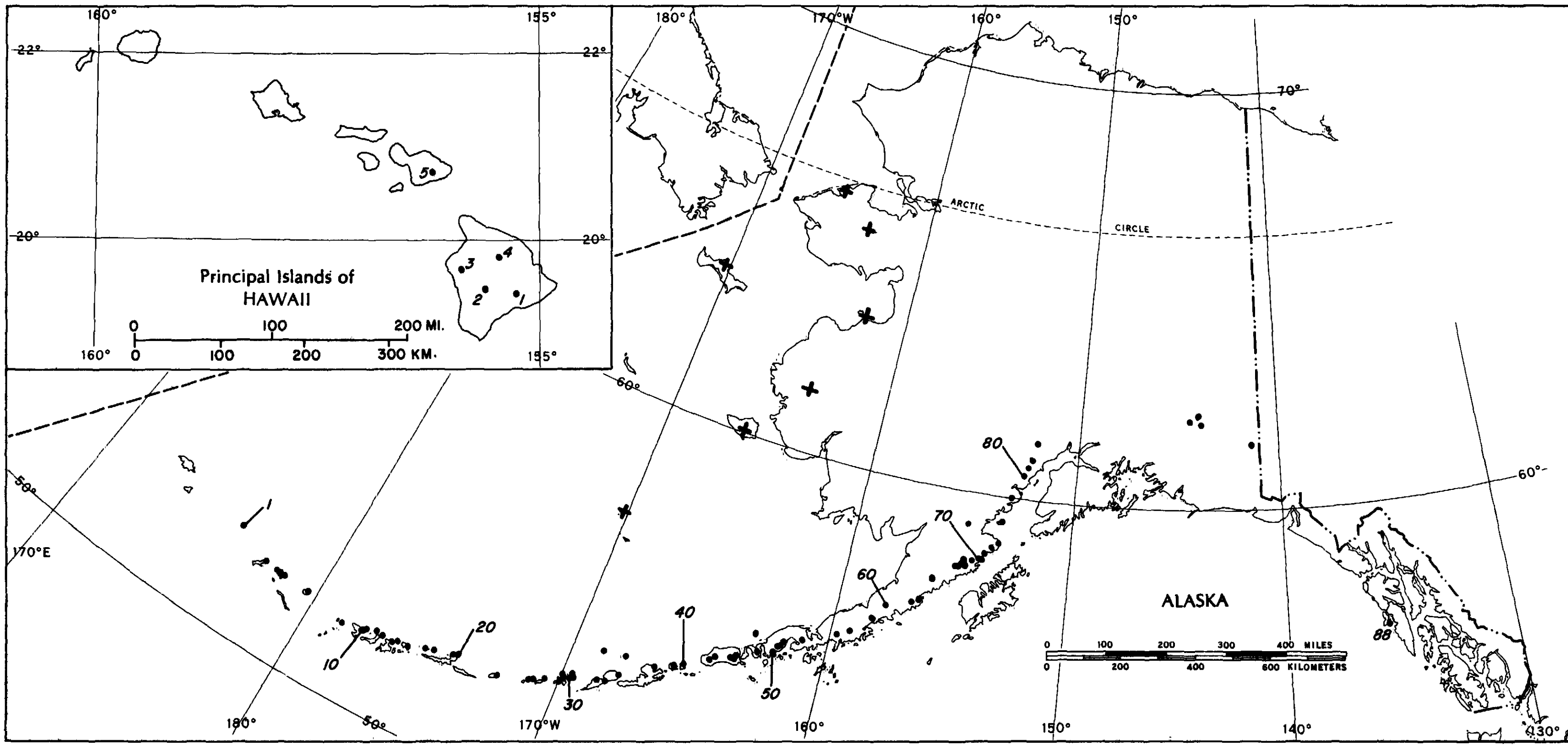

Figure 6.—Identified volcanic systems (small dots) and basaltic lava fields (+) known to be less than 10,000 years old in Alaska and Hawaii.

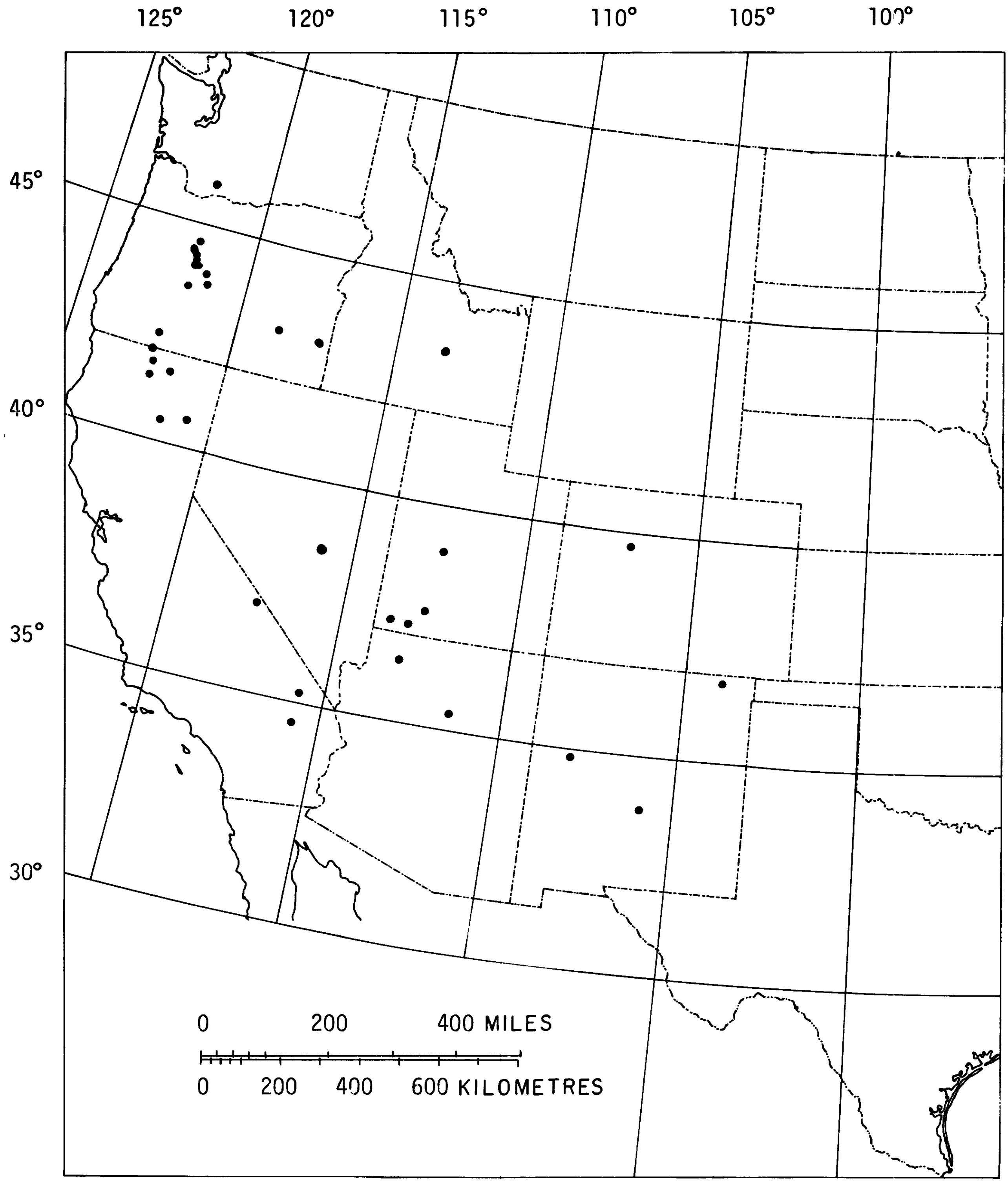

FIGURE 7.—Basaltic lava fields known to be less than 10,000 years old in the conterminous United States.

Table 8.—Basic volcanic fields probably less than 10,000 years old

	Lat N	Long W
ALASKA		
Devil Mountain field	66°18'	165°31'
Imuruk Lake field	65°32'	163°30'
Kookooligit Mountains (St. Lawrence I)	63°36'	170°24'
St. Michaels "	63°28'	162°10'
Ingakslugwat Hills "	61°23'	164°00'
Nunivak Island "	60°05'	166°30'
St. Paul Island "	57°11'	170°18'
WASHINGTON		
Red Mountain-Big Lava Bed area	45°56'	121°49'
OREGON		
North Cinder Peak	44°37'	121°48'
Nash Crater	44°25'	121°57'
Sand Crater	44°23'	121°56
Belknap Craters	44°17'	121°50'
North Sisters area	44°11'	121°47'
Leconte Crater	44°03'	121°48'
Cayuse Cone	44°04'	121°40'
Bachelor Butte	43°59'	121°41'
Lava Butte	43°55'	121°21'
Newberry Shield (lower flank)	43°43'	121°14'
Davis Lake to Black Rock Butte	43°33'	121°49'
Brown Mountain	42°22'	122°16'
Diamond Craters	43°06'	118°42'
Jordan Craters	43°02'	117°25'

Table 8.—Basic volcanic fields probably less than 10,000 years old—Continued

	Lat N	Long W
IDAHO		
Craters of the Moon	43°24'	113°30'
NEVADA		
Lunar Crater field (?)	38°29'	115°58'
ARIZONA		
Unikaret	36°23'	113°08'
Sunset	35°22'	111°30'
UTAH		
Ice Springs field	38°58'	112°30'
Crescent, Miter, Terrace		
Santa Clara	37°15'	113°38'
Crater Hill	37°13'	113°06'
Markagunt field	37°34'	112°42'
NEW MEXICO		
Carrizozo	33°47'	105°56'
McCartys	34°48'	108°00'
Capulin	36°47'	103°58'
COLORADO		
Dotsero	39°40'	107°02'
CALIFORNIA		
Copco Lake area	41°59'	122°20'
Goosenest	41°43'	122°13'
Lassen-Shasta rectangle		

An area approximately cornered by Lassen-Shasta-Medicine Lake-Eagle Lake is about 40 x 80 miles (3200 sq. mi.). This area contains many Holocene cones and basic lava flows and must represent a profound thermal anomaly in the mantle. Future volcanic activity may be expected in any part of this area at any time. The possibility of hidden shallow silicic reservoirs should be considered, and the entire area should be investigated in detail. Cinder Cone (1851), Hat Creek Lava flow, Burnt Lava flow and many other recent events are located in this area.

Ubehebe Craters	37°01'	117°27'
Cima field	35°15'	115°45'
Amboy-Pisgah field	34°33'	115°47'

Temperatures and Heat Contents Based on Conductive Transpo*t of Heat

By W. H. Diment, T. C. Urban, J. H. Sass, B. V. Marshall, R. J. Munroe, and A. H. Lachenbruch

Our objectives here are (1) to describe the heat-flow provinces of the United States as they are presently known, (2) to present estimates of temperatures to a depth of 10 km in these provinces and of the range of temperature that might be expected within each province, (3) to summarize the physical data and assumptions that constrain these estimates, and (4) to integrate the temperature distributions with area and depth so as to give an estimate of the sensible heat stored in the Earth to a depth of 10 km. This quantity represents an upper (and quite unapproachable) limit to the thermal energy that can be extracted from the top 10 km of the solid earth.

Various facets of these objectives have been addressed before, often with a rigor beyond the scope of this summary. However, a new summary seems warranted because (1) additional data are available and notions regarding their interpretation are evolving, (2) existing data occasionally have been distorted to achieve an unrealistically favorable view of the potential of geothermal energy, and (3) the data base upon which all estimates are based remains poor, and a summary of this sort is an opportunity to point out how it could be improved.

In the petroleum-producing provinces, temperatures are reasonably well known from measurements of temperature along, or at the bottom of, holes drilled for petroleum. In some provinces, such as the gulf coast, the information extends to depths of 7 km or more. A principal source for this information is the "Geothermal Survey of North America" (for example, Kehle and others, 1970), a project recently completed under the aegis of the American Association of Petroleum Geologists. The products of this survey include: (1) A data bank (tape or cards) containing bottom-hole temperatures for more than 25,000 wells along with appropriate mean annual surface temperatures; (2) two maps of North America at a scale of 1:5,000,000; the first is the Geothermal Gradient Map of North America (Am. Assoc. Petroleum Geologists-U.S. Geol. Survey, 1975); the second shows the depths to the 70°, 100°, and 150°C isotherms in those areas where sufficient information is available; and (3) a series of computer-drawn temperature gradient maps at a scale of 1:1,000,000. Reports of compilation procedures and interpretation are available now only in abstract form (Gould, 1974; Kehle and Schoepel, 1974; Shelton and others, 1974).

Deep holes are scarce outside of the petroleum-producing provinces. However, temperatures at depth may be estimated from a knowledge of the near-surface heat flow and the thermal properties of rocks.

Birch, Roy, and Decker (1968) measured heat flow q in holes drilled into plutonic rocks at various sites in the northeastern United States. They also measured the radioactive heat generation A of the rocks at each site and found a linear relation between q and A:

$$q=q^*+DA.$$

In a general way the term DA is the component of heat flow due to the radioactive heat production of the upper crust, and the q^* or "reduced heat flow" is the component that originates from the lower crust and mantle. Subsequent measurements in plutonic rocks of the Basin and Range (Roy and others, 1968b) and the Sierra Nevada (Lachenbruch, 1968; Roy and others, 1968b) also showed a linear relation. The parameters for the three provinces, in which extensive q–A studies have been made, are shown in the following table:

Region	q^* (HFU)	D (km)	DA_{max} (HFU)
Eastern United States ----	0.8	7.5	~1.5
Basin and Range province	1.4	10	~1
Sierra Nevada section ----	.4	10	~1

1 heat-flow unit (HFU) = 1×10^{6} cal/cm² s.

The slope (D) varies little among the regions, but differences among the intercept values (q^*) are large.

These observations lead to the notion that continents can be divided into heat-flow provinces, each typified mainly by its own q^*, and that the variation of heat flow within a province is a consequence of the variation of the heat generation of upper crustal rocks. Indeed, q–A points from Australia (Jaeger, 1970) and the Precambrian shield areas of the world (see summary by Rao and Jessup, 1975) seem to fall close to either the Northeastern United States or the Basin and Range lines. Moreover, the regions of high q^* are those of recent tectonism, and those of low q^* are those of either ancient tectonism (Northeastern United States and the Precambrian shields) or regions where a cold slab subducted into the mantle appears to give rise to anomalously low q^*, as in the Sierra Nevada (Lachenbruch, 1968; Roy and others, 1968b, 1972; Blackwell, 1971).

These observations have also been taken to mean (Roy and others, 1968b, 1972; Blackwell, 1971) that: (1) q^* for the Eastern United States may be typical of all stable continental regions older than a few hundred million years, and (2) q^* for the Basin and Range province is close to the maximum that might be found over broad regions of the continental crust, the rationale being that the base of the crust is near melting and further increase of temperature would lead to melting that would buffer additional heat input into the crust. Both of these generalizations are attractive. They form a framework from which to view the many anomalies that are now becoming apparent.

It is also evident from the preceding table that the second term DA in the equation can be a large fraction of the observed heat flow. Consequently, a detailed map of heat flow (were the data available) within a heat-flow province would be both highly variable and highly intricate because of rapid lateral variations in heat production. This is best illustrated by the studies in New England (Birch and others, 1968; Roy and others, 1968b; Roy and others, 1972), where a comparatively large number of heat-flow measurements have been made in a variety of basement rock types and a large amount of heat production information is available both from measurements on cores and from surveys of gamma activity. It is also evident from the relatively detailed investigations in the Sierra Nevada (Lachenbruch, 1968).

TEMPERATURE CALCULATIONS

Types of models

The q–A relation provides a basis for calculation of temperature T at depth z, assuming that the variation with depth of heat production A and thermal conductivity K are reasonably well known. The two most commonly used models for the decrease in A with depth are: (1) A is constant to the depth H, and (2) A decreases exponentially with depth $A = A_0 \exp -(z/D)$ where D is a constant and A_0 is the radioactive heat generation at the surface.

Assuming the conductivity is constant, the temperatures at depth for the two models are (for example, Jaeger, 1965; Lachenbruch, 1968):

$$T = T_o + \frac{AH^2}{2K}\left[\frac{2z}{H} - \left(\frac{z^2}{H}\right)\right] + \frac{q^* z}{K},$$

$$T = T_o + \frac{A_0 D^2}{K}\left(1 - e^{\frac{-z}{D}}\right) + \frac{q^* z}{K},$$

where T_o is the mean surface temperature and hereafter is taken as zero because we are interested primarily in temperature above mean surface temperature.

If the two models are to yield equal heat flow at the surface, D must equal H. If D is about 10 km, as suggested by the q–A relation, A will decrease to A_o/e or $0.37A_o$ at 10 km and to $0.05A_o$ at 30 km (roughly the thickness of the crust). In a general way, then, the first model (constant A) represents the case where crustal radioactivity is concentrated near the surface, and the second model (exponential A) represents the case where it decreases with depth throughout the crust. Consequently, q^* can be thought of as the heat coming from the lower crust and mantle in the first case and as the heat coming from the mantle in the second.

For the same q the temperatures at depth for the exponential model are somewhat higher than for the constant A model. However, at 3 km the difference is about 1°, and at 10 km it does not exceed a few tens of degrees even for the highest values of heat generation. For purposes of the present discussion, the difference between the models is unimportant.

Thermal conductivity

A constant conductivity of 6 mcal/cm s°C was chosen for the models because it is appropriate for igneous rock of felsic to intermediate composition, gneisses, and schists, which typify the major part of the metamorphosed and plutonized basement. Because the thermal conductivity of typical basement rocks decreases with temperature, a value of 6.5 to 7.5 mcal/cm s°C might be more appropriate for the lower temperature region and 5 to 6 for the higher temperature regions.

The conductivities of most igneous and metamorphic rocks are reasonably well known (Birch and Clark, 1940; Clark, 1966), as are the conductivities of their constituent minerals (see also Horai and Simmons, 1969).

The dependence of conductivity upon temperature is also generally known for igneous and metamorphic rocks and some of the rock-forming minerals (Birch and Clark, 1940). It is important to recognize that, although the conductivities of most rock-forming minerals decrease with temperature, the conductivity of the feldspars (the principal constituents of rocks within the crust) increases slightly with temperature. Consequently, it is difficult to envision an average conductivity of a large volume of rock much less than 5 (see, for example, Birch and Clark, 1940, figs. 4 and 5) even at the higher temperature levels.

Conductivities of individual rock types may differ widely from the average value assumed. Some of these types may occur in sufficient thickness that they result in temperature distributions significantly different than the ones given. On the high-conductivity side we have dolomite (~12), quartzose sandstone (10–16, depending on porosity), salt (12), and dunite-peridotite-pyroxenite-eclogite (10–14), and, on the low side, anorthosite (4.5), gabbro-basalt-diabase (~5), serpentinite (~5), and shale (2.5–4, depending on water content).

Water (K~1.5 at 30°C) is an important constituent of many sediments. Thus, recently deposited muds may have conductivities less than 2 that increase on compaction to 3–4 as the mud is transformed into a shale. Some volcanic rocks, such as ash-flow tuffs, may have porosities approaching 50 percent and conductivities less than 3. Bituminous materials such as coal, oil, and oil shale (Clark, 1966) have very low conductivities but rarely are thick enough to cause large distortions in the thermal regime.

Clearly, if a region is blanketed by low-conductivity sediments, the temperatures at depth will be higher than those in our models, which assume a uniform conductivity of 6. Let us assume a conductivity of 3 for the "thermal blanket," which is about the minimum permissible conductivity for a substantial thickness of sediment. The difference in temperature ΔT below the blanket as compared with the $K=6$ model will be:

$$\Delta T=q[(6-3)/(6\times3)]z$$

where the numerical values are the conductivities and z is thickness of the low-conductivity blanket. The excess temperatures are given in table 9.

The effect of the "blanket" is evidently small for low heat flow and thin blankets. It is substantial, however, for a thick blanket and a high heat flow. The upper limit ($z=2$ km, $q=3.0$, $\Delta T=100$) is difficult to envision but might be possible in some localities.

If the conductivities are significantly higher than the 6 assumed in the models, the gradients will be lower and the temperatures below the higher conductivity zone will be lower. Following the table above, and assuming a conductivity of 9, we have:

$$\Delta T=q[(6-9)/(6\times9)]z$$

and the effect of the high-conductivity layer is given in table 10.

Evidently, the assumption of a high-conductivity layer, such as might be represented by a thick section of dolomite and quartzose sandstone along with accompanying interbeds of shale, produces a small distortion of the tempera-

Table 9.—Excess temperature below a low-conductivity layer

Thickness (z) of Low conductivity Layer (km)	ΔT(°C) relative to K=6 models					
	q=0.5	q=1.0	q=1.5	q=2.0	q=2.5	q=3.0
0.5	4	8	13	17	21	25
1.0	8	16	25	33	42	50
2.0	16	33	50	67	84	100

ture distribution as represented by $K=6$, except for the highest q's and the greatest thicknesses.

Radioactive heat generation

The mean value of radioactive heat generation due to radioactive decay of uranium, thorium, and potassium for upper crustal rocks is probably close to 5.0 HGU (heat-generation units; see for an example Birch, 1954; Clark, 1966; Roy and others, 1968b; Blackwell, 1971). The range for rocks exposed at the surface is considerable: much less than 1 HGU for most ultramafic rocks and anorthosites, about 1 HGU for mafic rocks, and from 2 to 20 HGU for felsic and intermediate plutonic rocks, gneisses, and schists. The last group comprises the bulk of the basement rock of the upper crust and is the one that is of primary interest here.

In figure 8 all A values obtained in connection with q–A studies in the United States are plotted as a histogram. The range of A is large, and it is real. The lowest values represent such rocks as the Precambrian anorthosites of the Adirondack Mountains of New York, and the highest represent the rocks of the Mesozoic White Mountain Plutonic Series of New Hampshire. The large number of relatively low values for the Basin and Range, northern Rocky Mountain, and San Andreas provinces represent Mesozoic plutonic rocks of intermediate composition such as granodiorite, quartz diorite, and diorite.

This histogram reveals little about the average distribution of heat generation of near-surface crustal rocks; metasedimentary rocks, and gneisses are poorly known and are not represented. However, it indicates the extremes to be expected, and it suggests that large volumes of rock of $A>10$ are probably rare, at least in some geologic provinces.

Average temperature distributions and heat contents by province

In order to estimate the heat stored in the crust, we resort to some rather sweeping assumptions:

1. Three provinces are selected as "typical" of all provinces in the United States. Each physiographic province is assigned a parameter set (q^*, A, D, and K) of one of the three "typical" provinces. For some provinces there is sufficient information for such assignment; for others it is based upon somewhat arbitrary judgements. Moreover, we do not know whether provinces having characteristics intermediate between the "typical" provinces are present, or whether extremes are represented by the "typical" provinces.

Table 10.—Deficit of temperature below a high-conductivity layer

Thickness of high-conductivity layer (km)	ΔT(°C) relative to K=6 models					
	q=0.5	q=1.0	q=1.5	q=2.0	q=2.5	q=3.0
0.5	-1	-3	-4	-6	-7	-8
1.0	-3	-6	-8	-11	-14	-17
2.0	-6	-11	-17	-22	-28	-33
3.0	-9	-18	-24	-33	-28	-54
4.0	-12	-22	-34	-44	-56	-66
5.0	-15	-30	-40	-55	-70	-85

2. An average A of 5 HGU has been assumed for all provinces, although regions of large lateral extent are known for which significantly higher or lower values would be appropriate. The significance of this assumption can be judged from tables 11 and 12 and figure 8.
3. An exponential model for decreases of heat generation has been adopted. This model gives somewhat higher temperatures than the constant A model at depth (see tables 11–14).

The heat contents per unit area ($\bar{Q}$) of the two depth intervals (0–3 and 3–10 km) are calculated for the typical provinces (table 12). The total heat content of the intervals in each province (table 13) is then obtained by multiplying by the appropriate area (S). The sum (table 13) for the two intervals (over the whole of the conterminous United States) is designated the basic calculation. In table 14 we show the effect of varying some of the parameters used to construct the models from the basic calculations.

One of these (increase all q^*'s by 0.2 HFU) requires an explanatory note. Birch (1948) showed that a gradient correction for Pleistocene climatic effects of about 3°C/km might be appropriate for holes of shallow depths (~300 m). Diment, Urban, and Revetta (1972) suggested that application of such a correction would help reduce differences of heat-flow measurements made in high- and low-conductivity rocks in the Eastern United States and that q^* in the Eastern United States might be raised by about 0.2 HFU. This view is not universally held (for example, Sass, Lachenbruch, and Jessop, 1971; Slack, 1974), particularly with respect to its applicability to all regions. This detail illustrates the type of uncertainty that may arise in the measurement of heat flow in shallow holes.

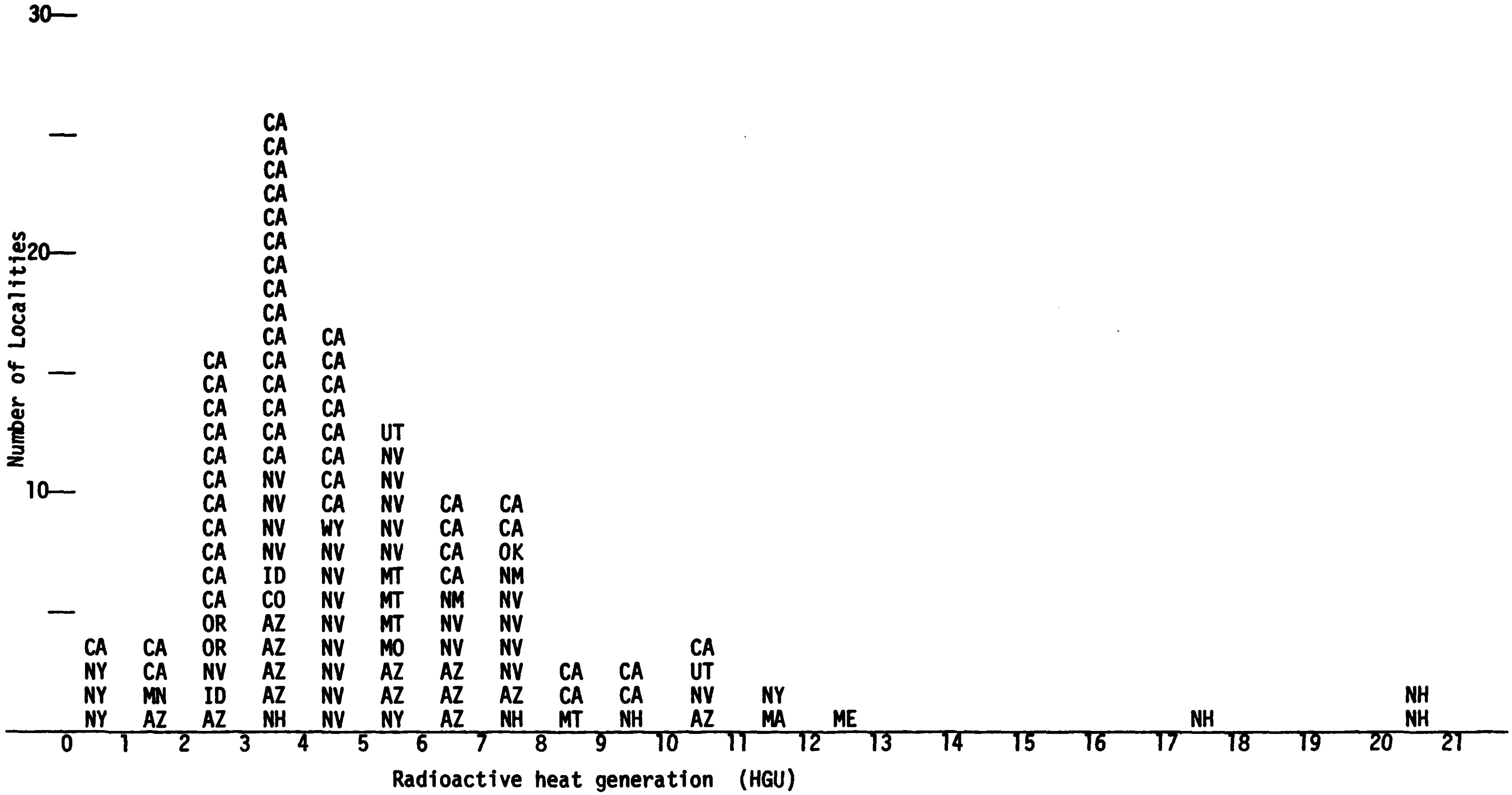

FIGURE 8.—Histogram of heat-generation values obtained in connection with *q–A* investigations. Data are given by State. From Henyey (1968), Lachenbruch (1968), Lachenbruch and Bunker (1971), Roy and others (1968b, 1972), Blackwell (1969, 1971, 1974b), and U.S. Geological Survey unpublished data.

Table 11.—Basic calculations: temperatures Tz (above mean annual surface temperature) at depths z of 3, 10, and 30 km for various assumed values of q* and A

	Exponential A					Constant A				
	A=0	A=5	A=10	A=15	A=20	A=0	A=5	A=10	A=15	A=20
q*=0.4 HFU										
D=10 km										
T_3 =	20	42	63	85	106	20	41	63	84	105
T_{10}=	67	119	172	225	277	67	108	150	192	233
T_{30}=	200	279	358	438	517	200	241	283	325	366
q*=0.8 HFU										
D=7.5 km										
T_3 =	40	55	71	86	102	40	55	70	85	100
T_{10}=	133	168	202	237	271	133	154	175	196	217
T_{30}=	400	446	492	538	584	400	421	442	463	484
q*=1.4 HFU										
D=10 km										
T_3 =	70	92	113	135	156	70	91	112	134	155
T_{10}=	233	286	339	391	444	233	275	317	358	400
T_{30}=	700	779	858	937	1017	700	742	784	825	867

Table 12.-Heat stored[1] ($\bar{Q}$ in megacal/cm^2) in the depth intervals 0-3 km and 3-10 km for various assumed values of q* and A

	Exponential A					Constant A				
	A=0	A=5	A=10	A=15	A=20	A=0	A=5	A=10	A=15	A=20
q*=0.4 HFU										
D=10 km										
$\bar{Q}_{0-3}$	1.8	3.8	5.9	7.9	10.0	1.8	3.8	5.9	7.9	9.9
$\bar{Q}_{3-10}$	18.2	34.6	50.9	67.3	83.6	18.2	32.8	47.5	62.1	76.8
q*=0.8 HFU										
D=7.5 km										
$\bar{Q}_{0-3}$	3.6	5.1	6.6	8.1	9.5	3.6	5.1	6.5	8.0	9.5
$\bar{Q}_{3-10}$	36.4	47.5	58.6	69.7	80.8	36.4	45.4	54.3	63.3	72.2
q*=1.4 HFU										
D=10 km										
$\bar{Q}_{0-3}$	6.3	8.3	10.4	12.4	14.5	6.3	8.3	10.4	12.4	14.4
$\bar{Q}_{3-10}$	63.7	80.5	96.4	112.8	129.1	63.7	78.3	93.0	107.6	122.3

1/ $\bar{Q}_{0-3}=c\int_0^3 T\,dz$, $\bar{Q}_{3-10}=c\int_3^{10} T\,dz$, where the volumetric specific heat (c) is assumed constant and equal to 0.6 cal/cm^3 °C, and z is in units of km, and T is temperature above 0°C.

Table 13.—Heat stored beneath provinces in depth intervals 0-3 km and 3-10 km

Model	Province	Area (S) $km^2 \times 10^6$	S(%)	$S\bar{Q}_{0-3}$ cal $\times 10^{22}$	$S\bar{Q}_{3-10}$ cal $\times 10^{22}$	$S\bar{Q}_{0-10}$ cal $\times 10^{22}$
q*=0.4 HFU D =10 km A =5 HGU	Sierra Nevada	0.081	0.87	0.31	2.80	3.11
q*=0.8 HFU D =7.5 km A =5 HGU	Eastern	3.375	36.08	17.2	160.4	177.6
	Coastal Plain	1.313	14.03	6.70	62.4	69.1
	Great Plains	1.633	17.45	8.30	77.6	85.9
	Wyoming Basin	0.123	1.31	0.625	5.84	6.47
	Peninsular Ranges	0.009	0.10	0.046	0.428	0.474
	Pacific Northwest	0.092	0.98	0.468	4.37	4.84
	Klamath Mountains	0.050	0.53	0.254	2.38	2.63
	Great Valley	0.058	0.62	0.295	2.76	3.06
	Colorado Plateau	0.392	4.19	1.99	18.6	20.6
	Subtotal	7.046	75.29	35.9	334.8	370.7
q*=1.4 HFU D =10 km A =5 HGU	Basin and Range	1.045	11.16	8.72	83.7	92.4
	Northern Rocky Mountains	0.298	3.18	2.49	23.9	26.4
	Central Rocky Mountains	0.151	1.61	1.26	12.1	13.4
	Southern Rocky Mountains	0.136	1.45	1.13	10.9	12.0
	Columbia Plateaus	0.324	3.46	2.70	25.9	28.6
	Cascade Range	0.128	1.37	1.07	10.3	11.4
	San Andreas fault zone	0.151	1.61	1.26	12.1	13.4
	Subtotal	2.233	23.84	18.6	178.9	197.5
	TOTAL (Conterminous United States)	9.360	100.00	54.8	516.9	571.3
	Alaska	1.519	--	10.2	97.2	107.4
	Hawaii	0.017	--	0.141	1.37	1.5
	TOTAL (United States)	10.896		65.141	615.47	680.2

Table 14.—Heat content above mean annual surface temperature of the continental crust of the conterminous United States in the depth intervals 0-3 km and 3-10 km

Model	Heat content (10^{24} calories)		
	$S\bar{Q}_{0-3}$	$S\bar{Q}_{3-10}$	$S\bar{Q}_{0-10}$
1. Basic calculation (assume sets of q*, D, K=6, A=5 as in table 13, fig. 10)	0.55	5.17	5.72
2. Basic calculation (except K=5)	0.66	6.20	6.86
3. Basic calculation (except K=7)	0.47	4.43	4.90
4. Basic calculation (except all q*'s increased by 0.2 HFU)	0.63	6.01	6.64
5. Basic calculation (except A=2.5 HGU)	0.48	4.60	5.08
6. Basic calculation (except A=7.5 HGU)	0.63	5.74	6.37
Best estimate (conterminous United States)	0.7 ± 0.1	6 ± 1	7 ± 1
Best estimate (United States)	0.8 ± 0.1	7 ± 1	8 ± 1

The heat content of the crust as represented by the basic calculation (table 14) is probably too low for several reasons:

1. The thermal blanket of low-conductivity sediments of the coastal plain and other areas of recent sedimentation have not been taken into account. However, depression of the thermal gradient by subsidence would tend to offset the effect of the low-conductivity sediments (Grossling, 1956; Jaeger, 1965).
2. The geopressured region of the gulf coast (Papadopulos and others, this circular) exhibits anomalously high temperatures.
3. Several provinces have rather arbitrarily been assigned to the normal or eastern category. The Great Plains, Colorado Plateau, and Cascades (Smith and Shaw, this circular) may well exhibit temperatures intermediate between the eastern and Basin and Range types.
4. Much of the observed high heat flow in the California Coast Ranges may be due to mechanical generation of heat in the upper 10 to 15 km of the crust (Lachenbruch and Sass, 1973) also resulting in temperatures intermediate between the eastern and Basin and Range types.
5. The hot spots discussed in other reports in this circular have not been added. However, their volumes, as presently estimated, are so small that their contribution to the value represented by the basic calculation is not significant.
6. The possible existence of regions of anomalously high heat flow (with respect to the Basin and Range models) has not been taken into account (see subsequent discussion).

Hawaii and Alaska have been listed separately in table 13 because so little is known of their thermal regimes. We have rather arbitrarily assigned Hawaii and half of Alaska to the Basin and Range (hot) type and the other half of Alaska to the eastern (normal) type.

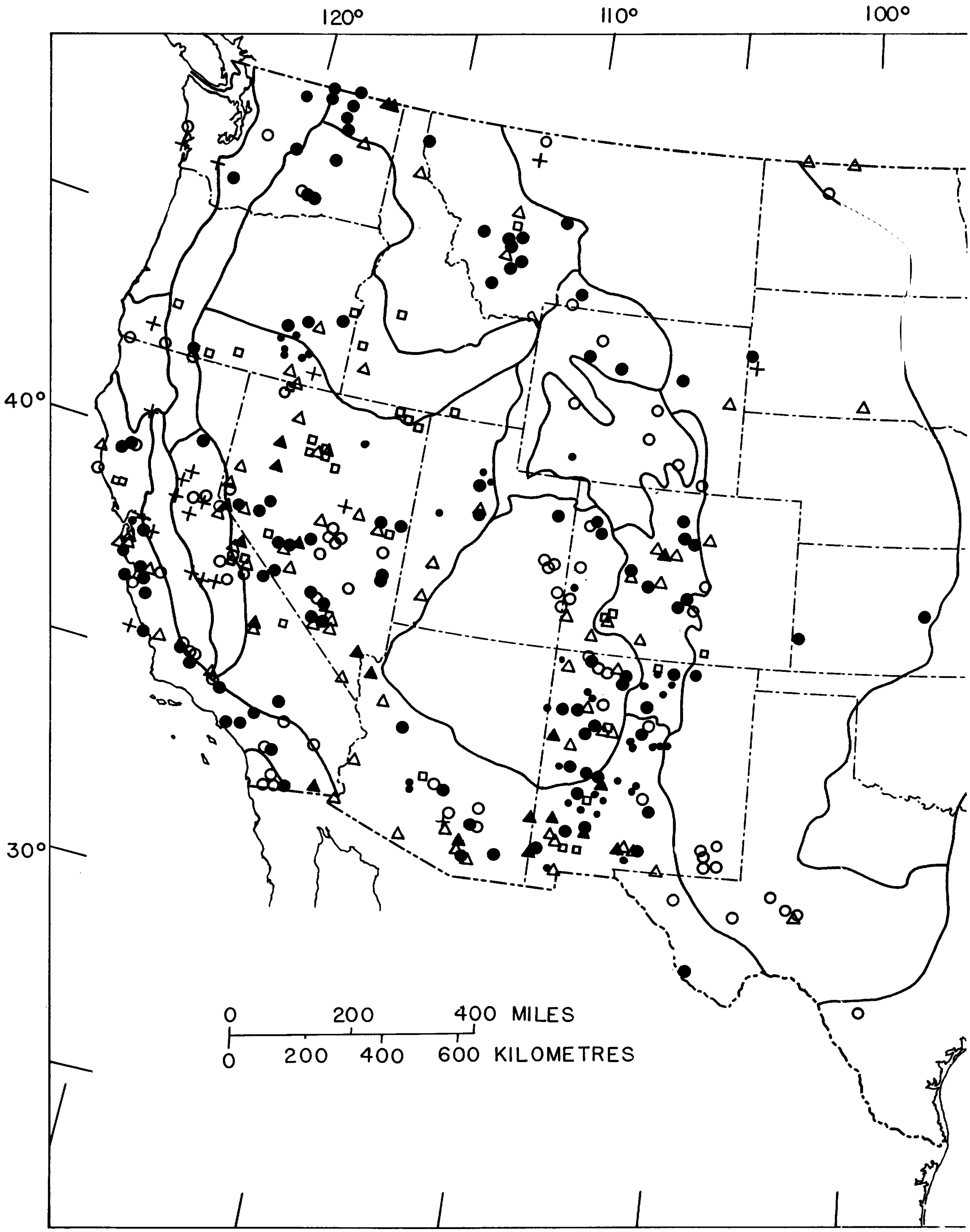

FIGURE 9.—Observed heat-flow (q) measurements in the United States. Measurements obtained in holes less than 150 m deep (indicated by dots) are not considered reliable as a group and therefore are not coded as to value. References to earlier data may be found in a compilation by Lee and Uyeda (1965). References for later data are the same as those in figure 8 plus those cited in references and U.S. Geological Survey unpublished data.

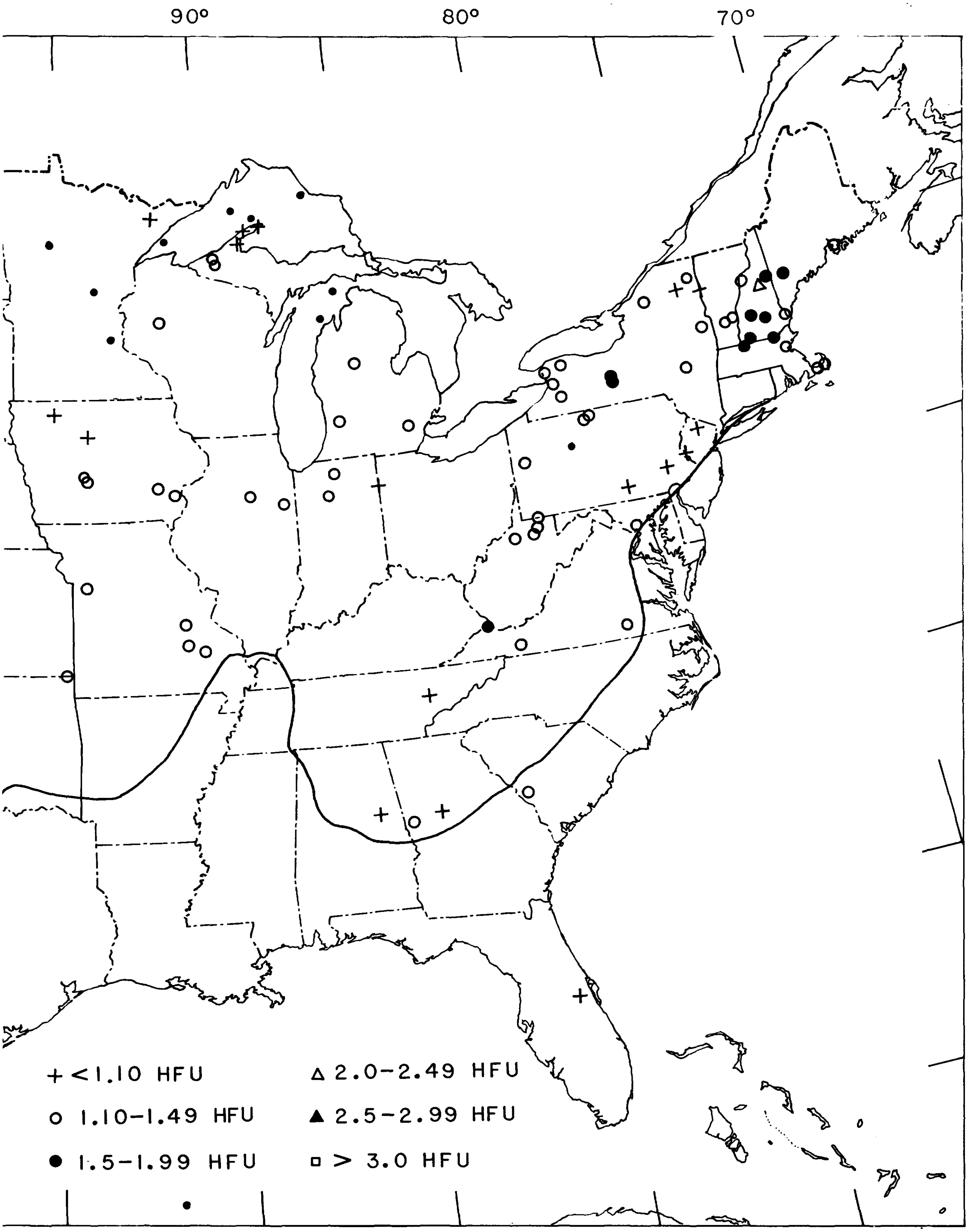

FIGURE 9.—Continued.

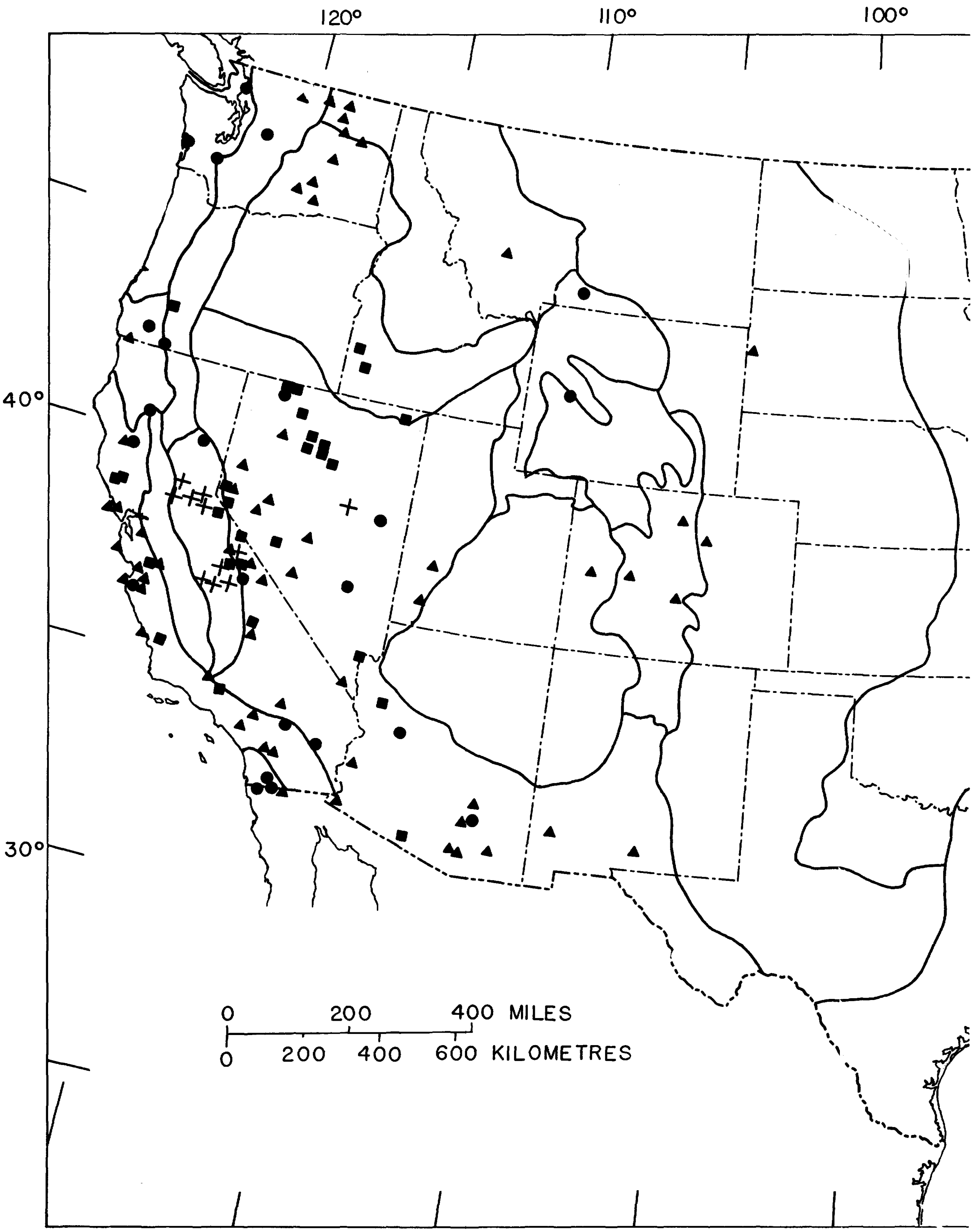

FIGURE 10.—Reduced heat-flow (q^*) measurements in the United States. References are the same as those in figure 8.

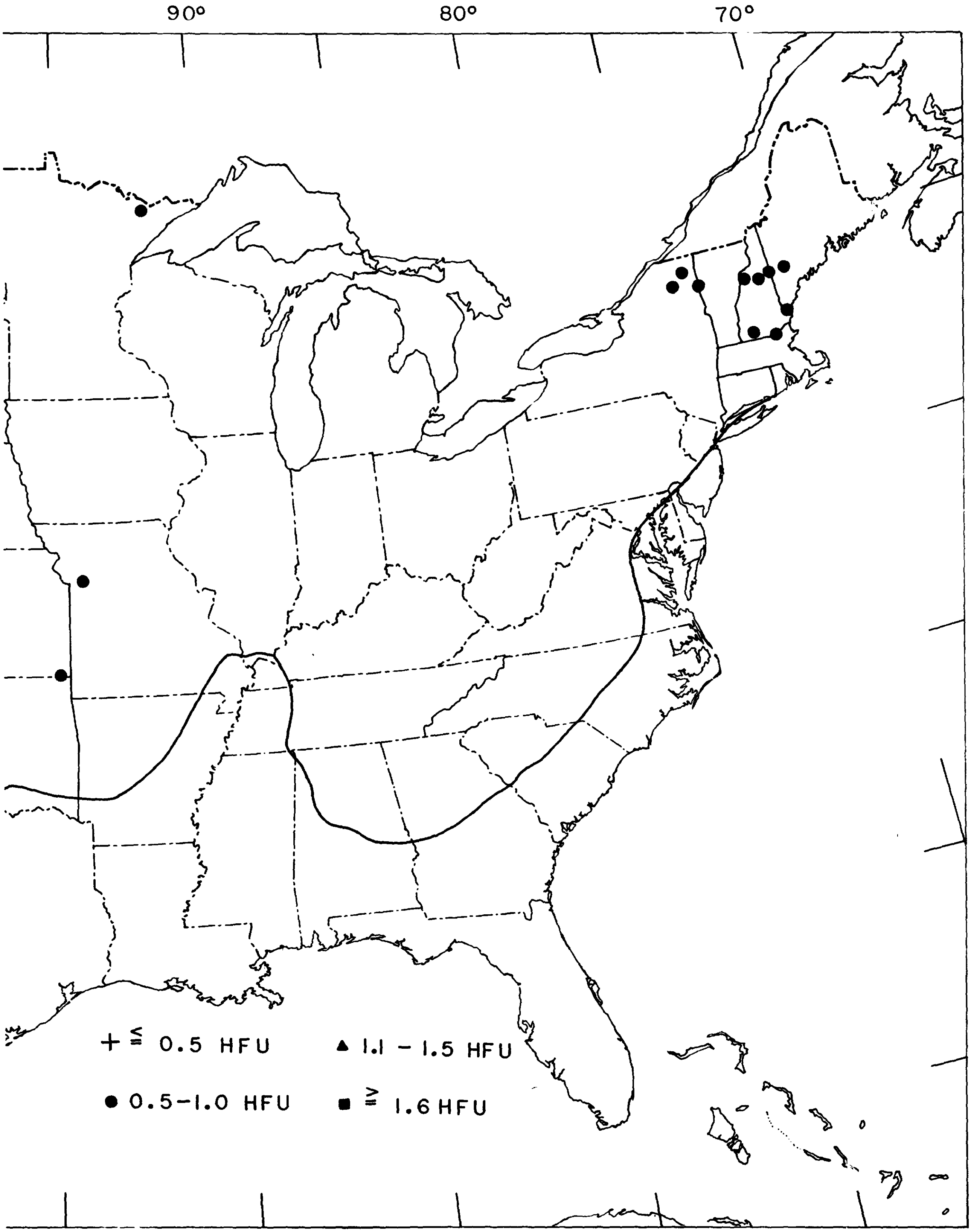

FIGURE 10.—Continued

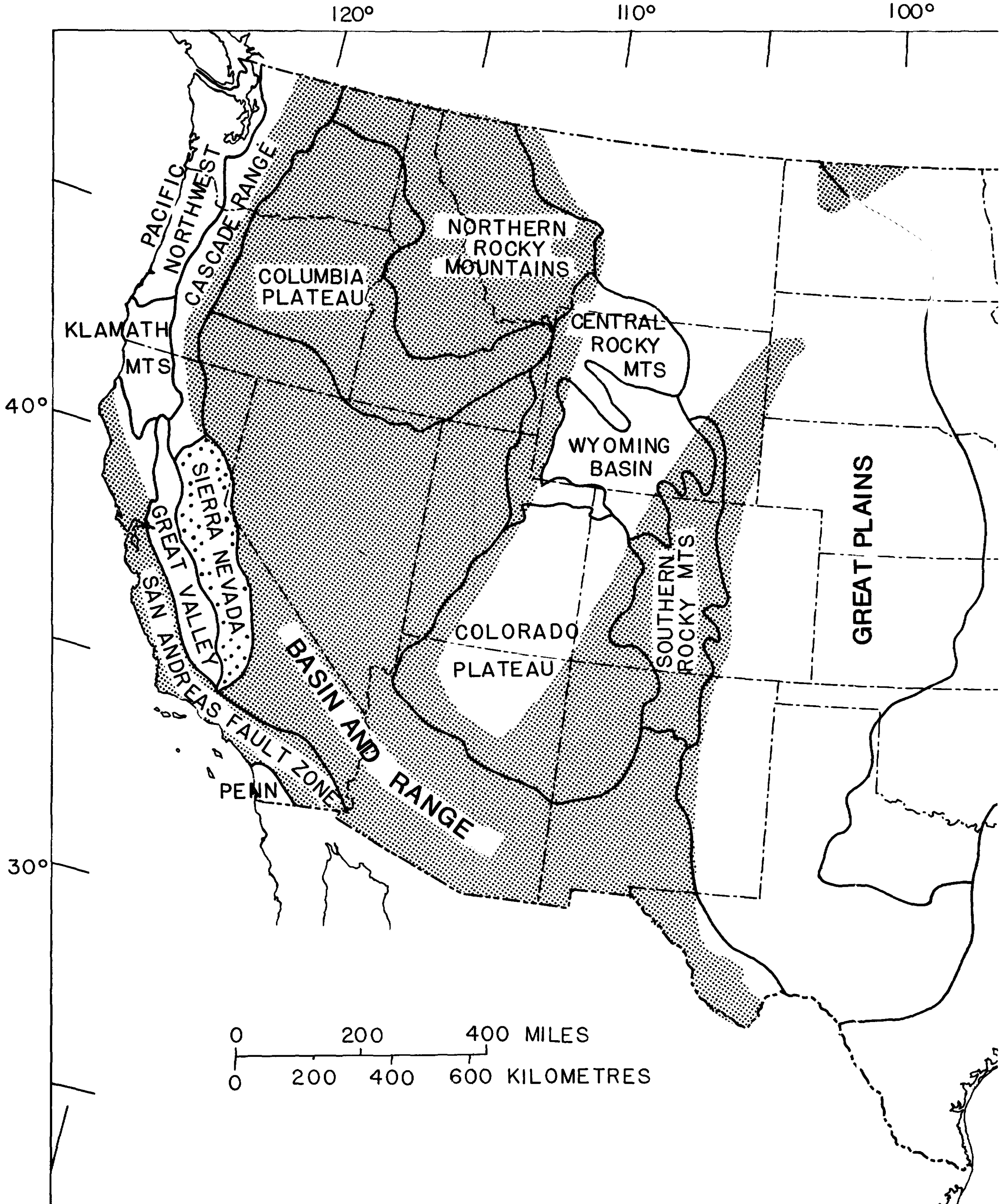

FIGURE 11.—Map showing probable extent of hot (stippled), normal (white), and cold (dotted) crustal regions of the United States based on data in figures 9 and 10. Physiographic provinces (largely generalized from Fenneman (1946)) do not necessarily represent heat-flow provinces.

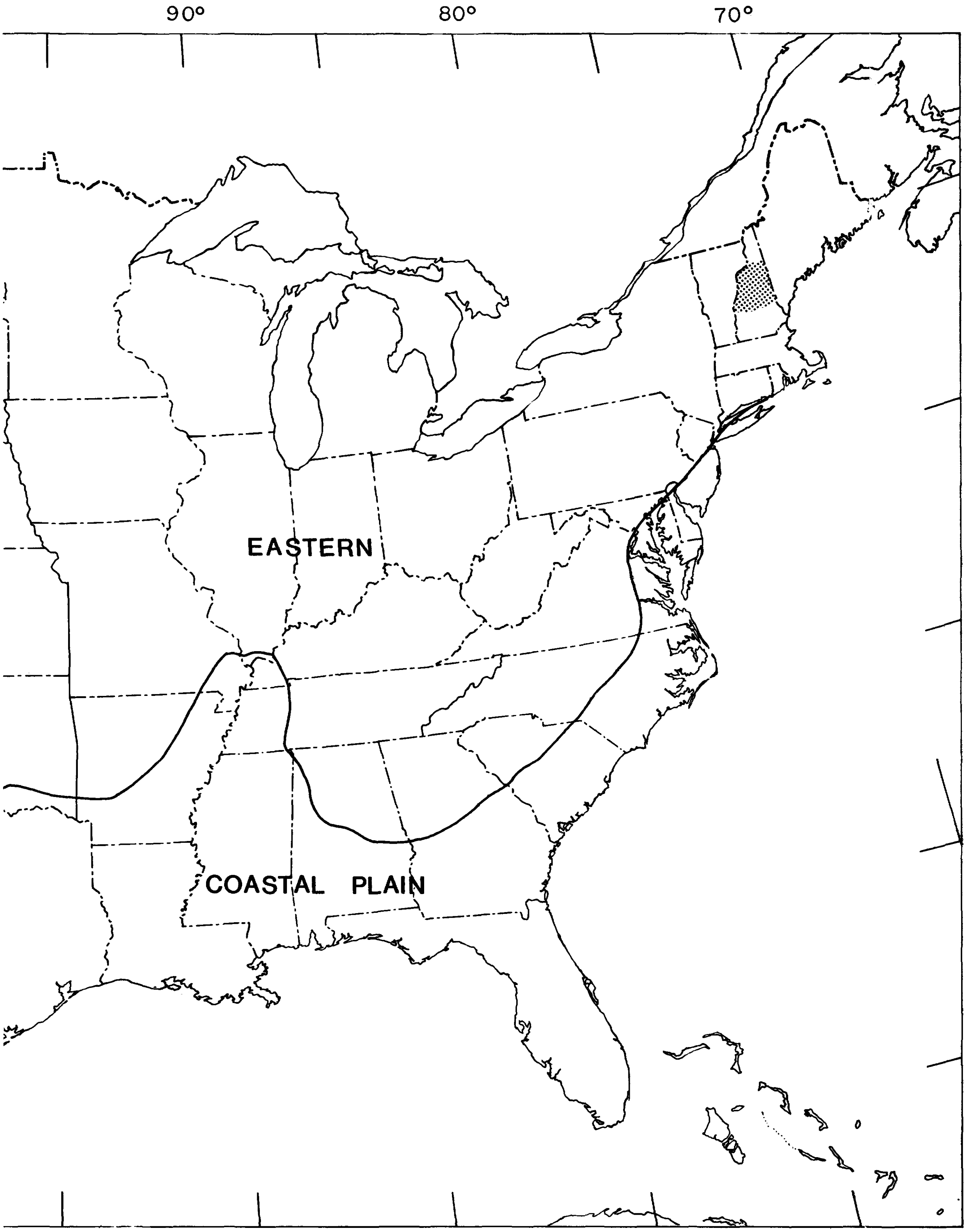

FIGURE 11.—Continued

DISCUSSION

The maps of observed heat flow (fig. 9) and reduced heat flow (fig. 10) are the basis for construction of the map (fig. 11) showing the location of the hot, normal, and cold crustal regions of the United States. Clearly, the data are not sufficiently numerous to define accurately the limits of the regions. Moreover, many of the observations themselves are weak in that they are based on assumed conductivities or on temperatures obtained under less than ideal conditions. Undue reliance should not be placed on individual values without reference to the original sources.

Temperature profiles (fig. 12) based on the exponential model with $K=6$ illustrate the range of temperatures that might be expected within each type of province. If one wishes to adjust the temperatures for slightly different parameters, it is useful to recall that: (1) Temperatures are inversely proportional to conductivity ($K=6$) for a given heat flow and that a temperature corresponding to a different conductivity K can be obtained by multiplying by $6/K$, and (2) an increase of 0.1 HFU in q^* results in an increase in temperature of 1.67°C per km of depth.

The temperature profiles (fig. 12*A* and B) for the Sierra Nevada type (cold) and the eastern type (normal) indicate that temperatures of economic interest are deep even for regions of high radioactive heat generation ($A=20$), which, as previously indicated, are limited in occurrence and in areal extent. To be sure, temperatures under a thick blanket of low-conductivity sediments might be 50°C or so higher, but the geographical extent of such regions is small. With current drilling technology, provinces of the cold or normal type hold little promise for geothermal exploitation except conceivably in very local areas.

The temperatures at depths less than 5 km in the Basin and Range (hot) type are not particularly attractive either, at least as expressed in figure 12*C* and 12*D*, especially in view of the fact that the values of $A>10$ are practically unknown in this region. However, the temperatures and heat flows are sufficiently high that the existence of a thick low-conductivity blanket in a given locality could raise the temperatures to levels of economic interest.

The temperatures in figure 12*C* and 12*D* ($A=5$) may not represent the maxima to be expected on a regional scale in the high-heat-flow provinces. Indeed, heat flows much higher ($q^*>1.8$, $q>2.5$) than those used to construct the temperature profiles for $A=5$ have been observed (Blackwell, 1971; Sass and others, 1971; Roy and others, 1968a, 1972; Decker and Smithson, 1975; Reiter and others, 1975; Urban and Diment, 1975; USGS, unpub. data). If the thermal regime is entirely conductive, such values require partial melting near the base of the crust (~30 km) or within the lower crust, or intrusion of magma into the crust. On the other hand, the anomalously high values could result from shallow hydrologic disturbance (not evident in the interval of measurement ~300 m) or from deep hydrothermal convection.

Recent and more detailed investigations in northern Nevada and southern Idaho strengthen the notion that the Battle Mountain heat-flow high, as originally proposed by Sass and others (1971), is indeed a region of anomalously high heat flow and not merely a consequence of aberrations involving local hydrothermal activity. Perhaps some anomalously high heat flows observed in other regions are also indicative of high temperatures (with respect to fig. 12*C* and 12*D* on a regional scale. The data are not adequate to prove or disprove such speculation.

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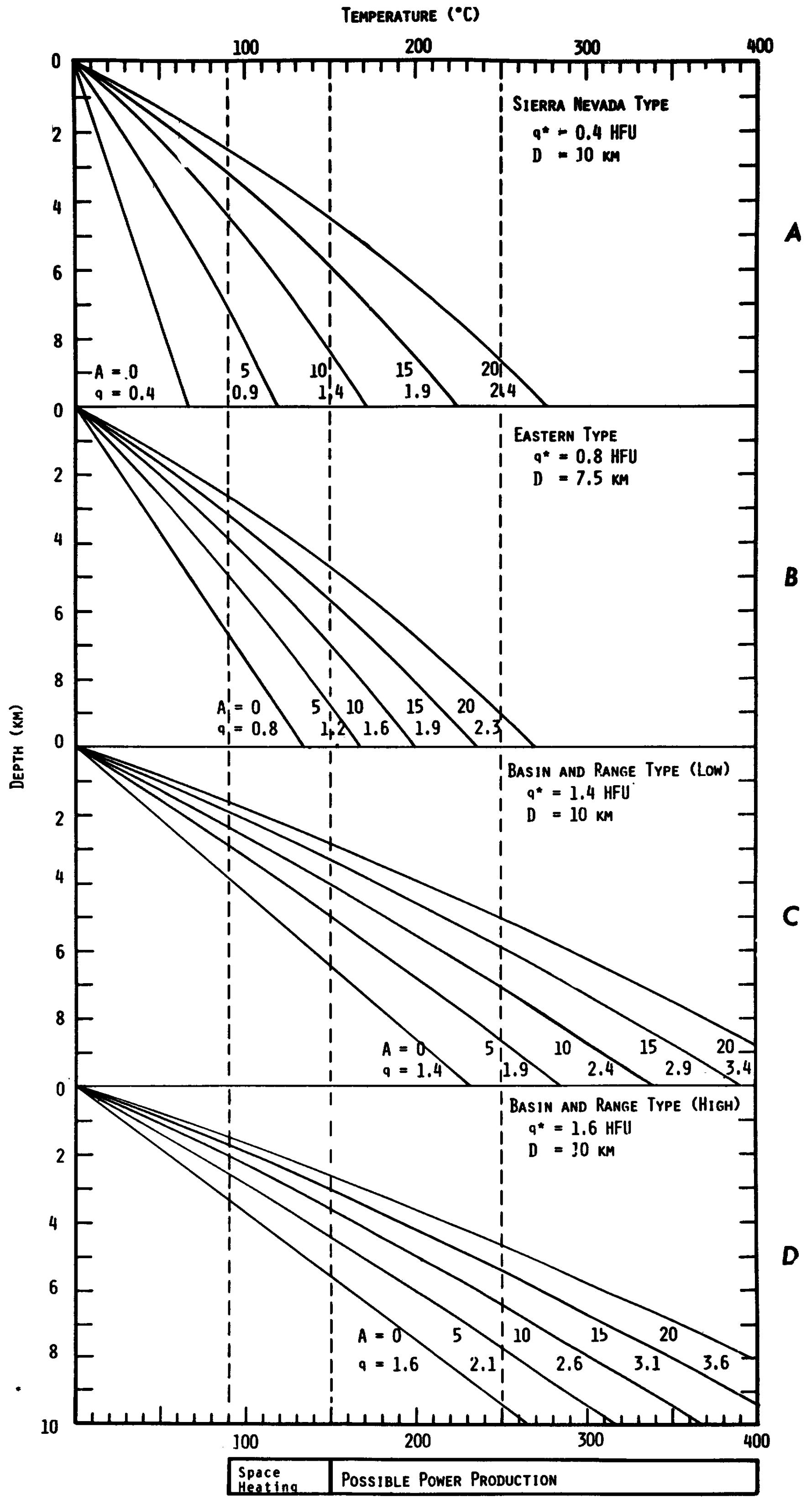

FIGURE 12.—Temperature profiles for various types of provinces in the United States based on an exponential decrease of radioactive heat generation (A) with depth and uniform conductivity (K=6 mcal/cm s°C). Units of q and q^* are HFU (1 HFU=1×10^{-6} cal/cm^2 s), and units of A are HGU (1 HGU=1×10^{-13} cal/cm^3 s).

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Geothermal Resources in Hydrothermal Convection Systems and Conduction-Dominated Areas

By Manuel Nathenson and L. J. P. Muffler

Reported here are estimates of the parts of the resource base of heat in hydrothermal convection systems and conduction-dominated areas that can be considered as resources or reserves. These estimates involve analysis of physical recoverability, conversion or utilization efficiency, and economics.

The definition of terms used in this report and elsewhere in this circular deserves emphasis and reiteration here. The *geothermal resource base* is all of the stored heat above 15°C to a 10-km depth (Muffler, 1973). *Geothermal resources* are defined as stored heat, both identified and undiscovered, that is recoverable by using current or near-current technology, regardless of cost. Geothermal resources are further divided into three categories based on cost of recovery:

1. *Submarginal geothermal resources*, recoverable only at a cost that is more than two times the current price of competitive energy systems,
2. *Paramarginal geothermal resources*, recoverable at a cost between one and two times the current price of competitive energy, and
3. *Geothermal reserves*, those identified resources recoverable at a cost that is competitive now with other commercial energy resources.

Undiscovered resources that are economically recoverable are not differentiated in this report but would be the economic equivalent of reserves.

This report considers only hydrothermal convection systems and conduction-dominated areas. Recoverability of geopressured resources is considered by Papadopulos and others (this circular), and recoverability of magma resources is treated by Peck (this circular).

HIGH-TEMPERATURE HYDROTHERMAL CONVECTION SYSTEMS

Renner, White, and Williams (this circular) have calculated that the heat stored in identified high-temperature (>150°C) hot-water convection systems outside of national parks is 238×10^{18} cal and that perhaps five times as much heat occurs in undiscovered hot-water convection systems, again excluding national parks. The sum of these two numbers, $1{,}400\times10^{18}$ cal, is the geothermal energy from hot-water convection systems potentially available for utilization. In the following discussion, we apply generalized recoverability factors and conversion efficiencies presented by Nathenson (1975b) to estimate the potential for electrical generation from various types and grades of hydrothermal convection systems. An economic analysis defines the variables that most strongly affect power production and emphasizes the sensitivity of economics to flow rates from individual wells.

Recoverable electrical energy

The calculation of recoverable electrical energy from a high-temperature hydrothermal convection system involves three major steps:

1. Estimation of what part of the hydrothermal convection system is porous and permeable rock,
2. Estimation of the fraction of stored heat in the porous and permeable volume that can be recovered at the surface,

3. Calculation of the efficiency with which thermal energy at the wellhead can be converted to electrical energy in the power plant.

The recoverability factors used for hydrothermal convection systems are based on techniques for extracting energy from porous, permeable rock. The volumes tabulated by Renner, White, and Williams (this circular), however, are the volumes of the heat reservoirs. The porous, permeable parts of the heat reservoir can range from only a small fraction to nearly all of the heat reservoir. In the resource calculations presented here, we assume that on the average only 50 percent of the heat reservoir is porous and permeable.

Our calculation of hydrothermal convection resources is based entirely on extracting the stored heat from a volume of porous and permeable rock, neglecting recharge of heat by either conduction or movement of water. The potential for heat recharge by conduction is neglected because it is very small compared to expected rates of production from any volume of rock greater than a few cubic kilometres. Likewise, for most of the hot-water systems of the United States, the natural discharge of thermal waters is low compared to reasonable production rates, and, accordingly, the potential for heat recharge by upflow of hot water to most reservoirs is probably low and can be neglected. The validity of this assumption can be assessed only after extensive production histories have been obtained for a reservoir. In those systems in which heat recharge by upflow of hot water is shown to be important, our resource estimate will have to be raised accordingly. Although the recharge potential of heat is neglected in our resource calculations, the potential and, in fact, the need for cold water recharge are not.

We have analyzed two possible methods for extracting energy from a liquid-fllled volume of porous and permeable rocks. The first method assumes that the porous, permeable volume is virtually closed to inflow of water and is produced by boiling to steam by using the energy in the rock. The second method assumes that natural and artifical recharge of cold water is used to recover much of the heat from the reservoir by means of a sweep process.

The fraction of stored energy recovered in the process of boiling the water in a porous volume of rock depends on the amount and pressure of the produced steam, which in turn are determined by the porosity and the initial temperature of the system. The pressure of the produced steam must be high enough to drive the steam through the porous medium and up the well at a significant rate; a reasonable assumption is that the pressure of the steam must be at least 8 bars (Nathenson, 1975b). At a given reservoir temperature, this restriction constrains the range of porosity for which boiling is a viable recovery scheme. At 200°C, the upper limit for the porosity is about 0.05, and the fraction of stored energy obtained is about 0.2. At 250°C, the upper limit for the porosity is about 0.12, and the fraction of stored energy obtained is about 0.4. At porosities below this limit the fraction of stored energy obtained decreases with decreasing porosity in a nearly linear fashion. This production scheme is severely limited if there is significant recharge of water to the reservoir; recovery by boiling is then possible only if steam in the dried zone and water in the recharge zone are produced simultaneously in order to keep the zone of boiling moving into new regions of the reservoir. In summary, the restricted range of porosity, temperature, and the recharge over which the boiling method will work limits its application to rather special circumstances, in particular to vapor-dominated systems (see below).

The second production scheme involves the use of natural and/or artificial recharge of cold water to drive hot water in a reservoir to the producing wells. As the water sweeps through the hot rock, its temperature is raised by removing energy from the rock. The influence of heat conduction on this process takes place on two length scales. On the microscale of pores filled with water in a rock matrix, conduction makes the temperature of the rock and the pores come to equilibrium in a matter of a few minutes. On the scale of a volume of rock several hundred metres on a side having one zone of cold water and rock and a second zone of hot water and rock, conduction with no fluid movement spreads out an initially sharp change in temperature to a smooth transition of only 60 m thickness in a period of a decade (Nathenson, 1975b). As cold water sweeps into a hot reservoir, conduction may be analyzed to a first approximation by superposition onto the movement of the temperature front, resulting in the premature breakthrough of cooler water into the hot zone. Another factor in the sweep process is the rotation of an initially vertical interface between cold water and hot water in a porous medium, owing to the difference in hydrostatic pressure on the two sides of the interface. Although this rotation is retarded by the energy stored in the rock, it also tends to cause premature breakthrough of cold water into the hot zone. These processes cannot be evaluated rigorously but can be combined qualitatively to yield an estimate that one-half of the energy stored in a reservoir of porous, permeable rock in a hot-water system can be recovered. Taking into account that only one-half of the heat reservoir is likely to be porous and permeable, the fraction of the resource base estimates of Renner, White, and Williams (this circular) that can be recovered as thermal energy at the surface is 0.25.

Vapor-dominated reservoirs are assumed to contain steam as the pressure-controlling phase, with liquid water immobilized in the pores by surface forces (Truesdell and White, 1973). Production results primarily from the boiling of this pore water to steam, although in later stages there may be some boiling from an inferred deep water table. Because the liquid fraction in a vapor-dominated reservoir is small, the pressure and temperature of steam produced in the boiling process are generally close enough to the initial values for the system that ample pressure remains to drive the steam to and up the well. The fraction of stored energy that may be recovered, calculated by considering an energy balance for the boiling process, is critically dependent on the average liquid saturation, as indicated by the values in table 15.

Efficiencies for converting thermal energies to electrical energies for a number of different surface technologies have been calculated by Nathenson (1975b). The choice of the optimum cycle is dependent on resource temperature, efficiency of conversion, cost, and special reservoir problems such as high dissolved solids, high gas contents, deposition of scale, and environmental factors. Current technology for hot-water systems consists of a single-stage flash to a steam-water mixture, separation of the steam and water, and use of the steam in a condensing turbine. A two-stage flash plant is being installed at the Hatchobaru system in Japan (Aikawa and Soda, 1975), and other technologies are in development. The choice and availability of a particular technology may determine whether a given system is economically viable or not, but this choice requires detailed knowledge of the reservoir characteristics of the system and of the economics of alternative conversion technologies. Since detailed reservoir knowledge is generally unavailable and the development of conversion technologies is in a rapid state of flux, no attempt is made here to analyze individual systems. Instead, for hot-water systems we use the the representative conversion efficiencies given in table 15 for the fraction of heat above 15°C converted to electricity (Nathenson, 1975b). Vapor-dominated systems tend to produce steam in the range of 180° to 240°C temperature; the conversion efficiency corresponding to this temperature range is approximately 0.2 (Nathenson, 1975b).

Defining Q as the heat stored above 15°C and e_r as the recovery factor, the electrical energy produced is

$$E = Qe_r \qquad (1)$$

(for Q in multiples of 10^{18} cal, E in megawatts electrical times centuries (MWe·cent) is 1,327 Qe_r). The recovery factors given in table 15 are used to convert stored heat into potential electrical energy for all identified convective hydrothermal systems with temperatures above 150°C outside of the national parks (table 16). Only those with sufficient data to give an estimated volume greater than the minimum volume used by Renner and others (this circular) appear by name in table 16. This minimum volume corresponds to a stored heat of about 0.2×10^{18} cal. The total electrical energy from the identified convective systems in table 16 is 8,000 MWe·cent, or 27,000 MWe for 30 years.* The locations of the systems identified by name in table 16 are plotted on maps of the United States (fig. 13) as large dots.

The Geysers is the only vapor-dominated system in table 16, and it also contains the only measured geothermal reserves for electrical energy. Since the recoverable energy of a vapor-dominated system depends so critically on the average water content (which can only be roughly estimated from the limited public data), a reserve estimate cannot be determined reliably. Average

* The conversion to a 30 year period assumes that each MW·cent of electricity can be produced at a rate of 3.33 MW for 30 years. This assumption applies to all similar conversions throughout this circular.

Table 15.—Recovery factors (e_r) for electric power generation

Vapor-dominated systems (above 200°C)				
Water content as fraction of total volume	0.01	0.02	0.03	0.05
Fraction of stored energy recovered as heat*	0.020	0.039	0.059	0.097
e_r = recovery factor = fraction of stored energy recoverable as electrical energy at a conversion efficiency of 0.2*	0.004	0.008	0.012	0.019

*Volume of porous, permeable part assumed to be one-half of heat reservoir.

Hot-water systems			
Temperature range °C	150-200	200-250	250-300
Conversion efficiency	0.08	0.10	0.12
e_r = recovery factor = fraction of stored energy recoverable as electrical energy**	0.02	0.025	0.03

**Volume of porous, permeable part assumed to be one-half of heat reservoir. One-half of the thermal energy in the porous, permeable part assumed recoverable for a net recovery of 25% of the thermal energy.

Table 16.—Estimated potential electric energy from identified high-temperature hydrothermal convection system, each with estimated stored heat equal to or greater than 0.3 x 10^{18} cal (from Renner and others, this circular)

	Subsurface Temperature °C	Volume km^3	Stored Heat 10^{18} cal	Recovery Factor e_r	Electrical Potential MW e ·cent
Alaska					
Geyser Bight	210	8	0.9	0.025	30
Hot Springs Cove	155	4	0.3	0.02	8
California					
The Geysers	240	140	18.9	0.019	477
Surprise Valley	175	250	24	0.02	637
Morgan Springs	210	10	1.2	0.025	40
Sulphur Bank mine	185	3.75	0.4	0.02	11
Calistoga	160	9	0.8	0.02	21
Skagg's H.S.	155	3	0.3	0.02	8
Long Valley	220	450	55	0.025	1825
Coso H.S.	220	336	41	0.025	1360
Salton Sea	340	108	21	0.03	836
Brawley	200	27	3	0.025	100
Heber	190	100	11	0.02	292
East Mesa	180	56	5.5	0.02	146
Idaho					
Big Creek H.S.	175	3	0.3	0.02	8
Sharkey H.S.	175	3	0.3	0.02	8
Weiser area	160	70	6.1	0.02	162
Crane Creek	180	60	5.9	0.02	157

Table 16.—Estimated potential electric energy from identified high-temperature hydrothermal convection systems, each with estimated stored heat equal to or greater than 0.3 x 10^{18} cal (from Renner and others, this circular) —Continued

	Subsurface Temperature °C	Volume km^3	Stored Heat 10^{18} cal	Recovery Factor e_r	Electrical Potential MW e ·cent
Nevada					
Baltazor H.S.	170	3	0.3	0.02	8
Pinto H.S.	165	7.5	0.7	0.02	19
Great Boiling (Gerlach) S.	170	25	2.3	0.02	61
Sulphur H.S.	190	10	1.1	0.02	29
Beowawe H.S.	240	42	5.7	0.025	189
Leach H.S.	170	10	0.9	0.02	24
Stillwater area	160	25	2.2	0.02	58
Soda Lake	165	12.5	1.1	0.02	29
Brady H.S.	214	30	3.6	0.025	119
Steamboat Springs	210	16	1.9	0.025	63
New Mexico					
Valles caldera	240	130	18	0.025	597
Oregon					
Mickey H.S.	210	12	1.4	0.025	46
Alvord H.S.	200	4.5	0.5	0.025	17
Hot Lake	180	12	1.2	0.02	32
Vale H.S.	160	100	8.7	0.02	231
Neal H.S.	180	4	0.4	0.02	11
Lakeview	160	16	1.4	0.02	37
Crumps Spring	180	8	0.8	0.02	21

Table 16.—Estimated potential electric energy from identified high-temperature hydrothermal convection systems, each with estimated stored heat equal to or greater than 0.3 x 10^{18} cal (from Renner and others, this circular) —Continued

	Subsurface Temperature °C	Volume km^3	Stored Heat 10^{18} cal	Recovery Factor e_r	Electrical Potential MW e ·cent
Utah					
Roosevelt (McKean) H.S.	230	8	1.0	0.025	33
Cove Fort-Sulphurdale	200	22.5	2.5	0.025	83
Total:			252 x 10^{18} cal		7833 MW·cent
Other systems estimated to have minimum volume			5 x 10^{18} cal		133 MW·cent
Identified Resources:			257 x 10^{18} cal		7966 MW·cent

water content at Larderello, Italy, appears to be about 1 percent of total volume for a reservoir 2 km thick (Nathenson, 1975a). Assuming this value for The Geysers, equation 1 with the value for 1 percent (table 15) yields a reserve estimate of only 100 MWe·cent or 333 MWe for 30 years. On thermodynamic grounds, Truesdell and White (1973) maintain that the average water content at The Geysers is considerably greater than that at Larderello and may be as much as 9 percent. More complete calculations for the conductive heat losses from the steam in the well to the surrounding rocks using the method of Nathenson (1975a) show that the water content estimated from the theory of Truesdell and White (1973) is closer to 5 percent than 9 percent. Using the value of 5 percent, the potential of The Geysers is 477 MWe·cent or 1,590 MWe for 30 years, and this value appears in table 16.

Using an alternate method, wells at The Geysers initially produce an average of about 150,000 lb/hr (19 kg/s); 2×10^6 lb/hr (252 kg/s) are needed to generate 110 MWe (Finney, 1973). Initial well spacings are on the order of 40 acres (0.16 km^2) with infill wells drilled as needed to maintain declining flows. Renner and others (this circular) estimate the reservoir area to be 70 km^2. Assuming that a 40-acre spacing for new zones can be maintained, the potential of The Geysers is 3,570 MWe for a 30-year life or 1,070 MWe·cent, which is more than double the figure if a 5-percent water content is assumed. The assumptions in this second calculation are such that 1,070 MWe·cent seems likely to be in error on the high side. Accordingly, we have chosen to use the results of the first calculation (477 MWe·cent) in table 16. The resolution of this discrepancy may lie in such factors as a reservoir extending below the arbitrary bottom at a 3-km depth (Renner and others, this circular), a water content higher than 5 percent, a ratio of volume of porous and permeable parts to volume of heat reservoir that is higher than 0.5, well spacing that should be larger than the assumed 40 acres, or steam production that results partly from boiling from a deep brine zone.

The high-temperature hydrothermal convection systems are technologically exploitable, and therefore their recoverable heat contents are a resource. If the heat of a given system cannot be recovered because there is not a porous, permeable reservoir, then that system is considered to be part of the resource base rather than a resource until a recovery technology is available.

Economic constraints

The cost of geothermal power can be broken down into various components: exploration costs, land-acquisition costs, costs of production wells, cost of fluid-transmission facilities, plant costs, fluid-disposal costs, and cost of money. Although the cost of production wells is a small fraction of the total cost, it is the cost factor that has the

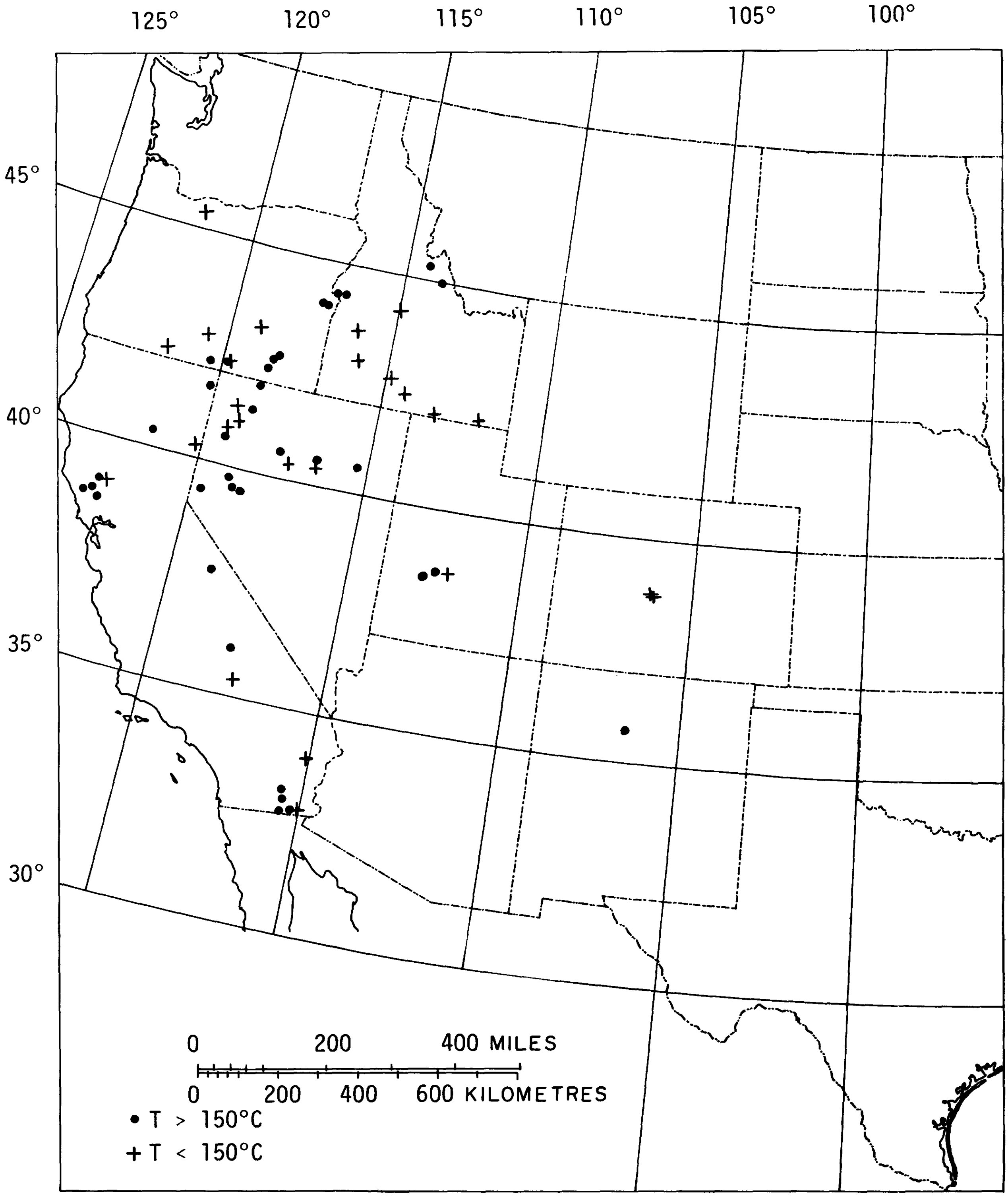

Figure 134.—Locations of hydrothermal convection systems with temperatures above 90°C and stored heat greater than or equal to 0.3×10^{18} cal.

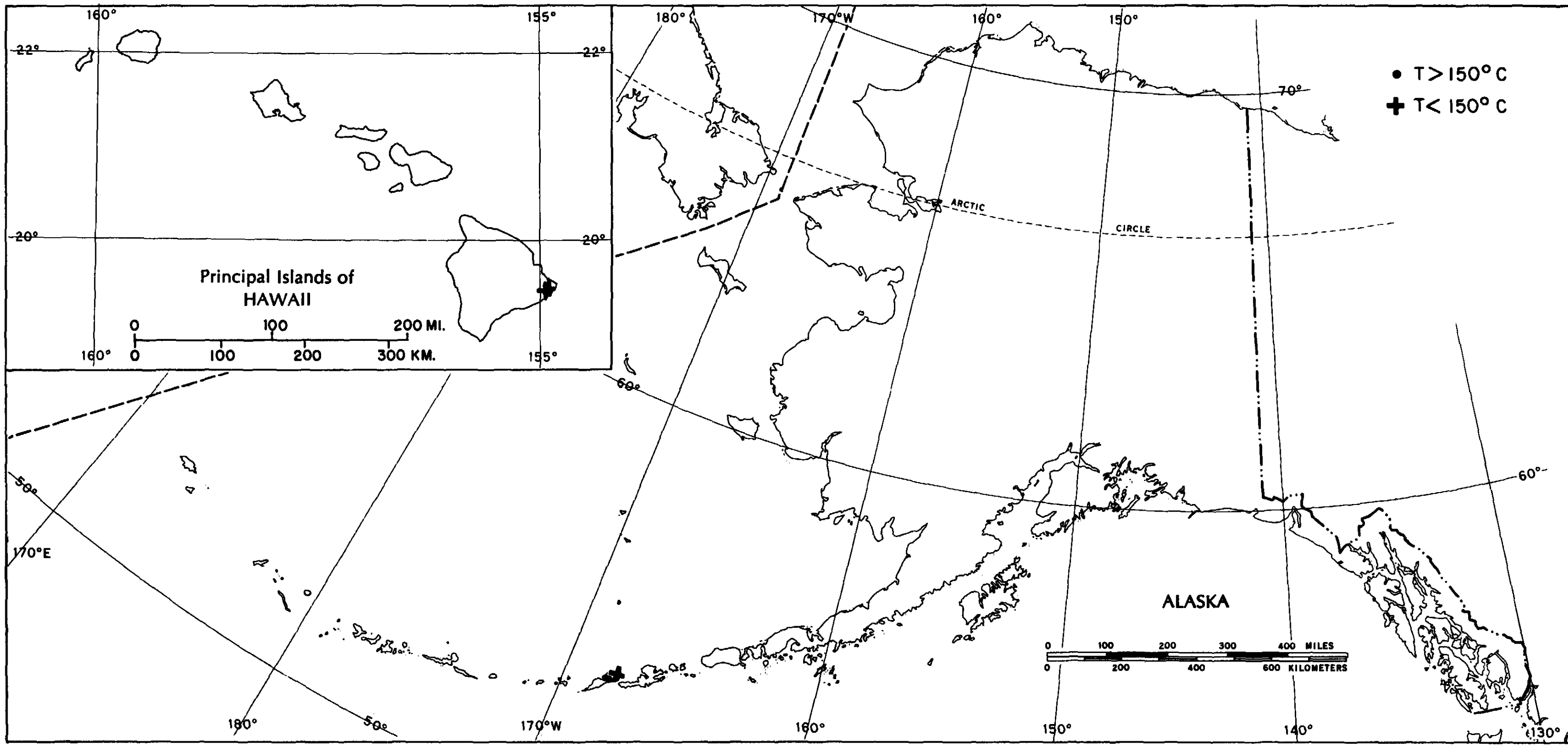

FIGURE 13B.—Locations of hydrothermal convection systems with temperatures above 90°C and stored heat greater than or equal to 0.3×10^{18} cal.

greatest range and the most uncertainty and that is therefore critical in determining whether development of a given field is likely to be economically feasible. Although each production well can be assigned an average cost, the total cost of production wells is a direct function of the number of wells that need to be drilled to supply sufficient steam for the generating plant under consideration. This number in turn depends on the reservoir temperature and the average well productivity. Although the minimum reservoir temperature commonly can be estimated prior to drilling by using chemical geothermometers, average well productivity can be determined only after extensive investment in drilling and testing. Accordingly, the uncertainty in average well productivity is a major financial uncertainty in any geothermal development. Inasmuch as the division of geothermal resources into reserves, paramarginal resources, and submarginal resources is based on economics, these uncertainties impact any attempt to evaluate geothermal resources.

The following analysis indicates the degree to which the cost attributable to production wells is sensitive to well productivity. Consider a well of average mass flow M that is productive for a time T and costs C dollars to drill and case. This well can produce a certain amount of electricity per kilogram of produced fluids [w_{act}]. Representative values for w_{act} for vapor-dominated and hot-water systems are given in table 17. The rate of energy flow (power) produced by the well is

$$\dot{E}=Mw_{\mathrm{act}}. \quad (2)$$

The part of the power cost in mils per kilowatt-hour (mils/kWh) that pays for the well is called c. Assuming that the well flow is constant and produces revenue evenly over the period T, then the original cost of the well must be escalated by a factor $1+\varepsilon$ to allow for the cost of money. The value of the produced energy needed to pay for the well is then $\dot{E}Tc$, and the escalated cost of the well is $C(1+\epsilon)$. Equating value and cost and rearranging, we obtain

$$c=\frac{C(1+\epsilon)}{\dot{E}T}. \quad (3)$$

Some estimates of drilling cost C as a function of depth are given in table 18 where the values for 2 and 3 km are based on recent geothermal experience (C. H. Bloomster, oral commun., 1975), and the other values are based on estimates made for crystalline rocks (Shoemaker, 1975). Taking the well life T as 20 years and the interest rate as 10 percent, the cost-of-money factor $1+\epsilon$ is 20/8.514=2.349. The cost-of-money factor together with the data in tables 17 and 18 has been used to prepare figure 14.

Figure 14A shows the electric power production per well as a function of mass flow for several reservoir temperatures and for vapor-dominated and hot-water systems, using the data of table 17 and equation 2 . Points are shown for Wairakei from the values of Axtmann (1975) and for units 5 and 6 of The Geysers from the data of Finney (1973). The 250°C line for hot-water systems was calculated by assuming a two-stage flash system. The Wairakei point plots below this line because its efficiency is low owing to its single-flash system. Figure 14B shows the electric power per well as a function of the value of the energy needed to pay for the well for several well costs from the data in table 4 and equation 3. The electric power per well for The Geysers and Wairakei are the same as in figure 14A. The value needed to pay for the well at Wairakei was caluculated on the basis of 20-year life and $150,000 drilling cost for depths of about 1 km. The value needed to pay for a well at The Geysers was calculated on the basis of a 15-year life and $500,000 drilling cost for wells drilled to near 3 km depth. This shorter lifetime was assumed because of the decline in rate of flow at The Geysers.

Two points should be made about the calculations for The Geysers and Wairakei. First, although they are very different systems, the return needed to pay for a well works out to be close to the same (0.7 and 0.85 mils/kWh, respectively). Second, the total cost of power involves many additional costs above the value needed to pay for a well. Calculations by Clarence Bloomster of Battelle Pacific Northwest Laboratories, discussed in Lucking (1975), show total power cost for a system like The Geysers to be 10 mils/kWh 4.7 being the cost of producing steam from the reservoir, of which only 0.7 is needed to pay for an individual well by equation 3. Calculations for an analog of Wairakei single flash but with reinjection) yield a total power cost of 20 mils/kWh, with 13 due to costs of producing steam from the reservoir, of which only 0.85 is needed to pay for an individual well by equation 3.

Table 17.—Typical values for electric energy produced per kilogram of reservoir fluid (w_{act}) as a function of source temperatures

Hot water				
Temperature °C	150	200	250	300
w_{act} in $\frac{kW\cdot s}{kg}$	33	70	120	177

Vapor-dominated, saturated steam			
Temperature °C	150	200	250
w_{act} in $\frac{kW\cdot s}{kg}$	430	550	620

Although the calculated values in figure 14 are only a part of the total, an important trend is shown by the diagrams. Since geothermal power must compete with many other power sources, the total value of the power is fixed by market forces at a particular time and place. If we assume that only a certain number of mils per kilowatt-hour is available to pay for the well, then we can use figure 14 to estimate the minimum flow needed to make a particular resource competitive. Taking the value needed to pay for a well of 1.5 mils/kWh in figure 14*B* and projecting it vertically to a $300,000 well cost, the power level needed from each well is about 2.7 MWe as determined by horizontal projection. The same horizontal projection may then be extended into figure 14*A* to show how the required mass flow from each well varies with temperature and resource type. Required flows in the vapor-dominated systems are relatively low. Because of this, more expensive wells can be drilled, but still at a profit. For the hot-water systems, the required minimum flows become very large at the lower temperatures. As the value needed to pay for a well and the cost of the well vary with time and specific circumstances, figure 14 may be used to estimate required well flows. For high well costs such as $1 to $5 million, the value needed to pay for a well must be increased if the minimum well flow for economic operation is not to be unreasonably large. Note that the use of figure 14 is only semiquantitative. Specific systems and technology should be evaluated by making a detailed cost study such as the program of Bloomster and others (1975).

Table 18.—Drilling cost model

Depth km	1	2	3	5	10
Average cost per metre	$150	$150	$167	$200	$500
Drilling time days	15-25	30-50	75-150	150-250	750-1100
Cost (10^3)	$150	$300	$500	$1000	$5000

The interpretation of figure 14*A* requires consideration of the factors influencing the flow of a well produced by flashing discharge (Nathenson, 1974). For flash eruptions, a limit is imposed by the frictional pressure drop in the well at very high rates of flow. At 200° to 250°C in a 25-cm-diameter well, flowing friction limits the rate to about 200 to 250 kg/s. From figure 14, a constant value of 1.5 mils/kWh needed to pay for a well at these assumed rates of flow yields possible total well costs of $1.5 to $2.5 million. In many cases, however, flows are limited not by flowing friction in the well but by the available permeability and permissible pressure drop in the reservoir, resulting in a much lower value of well flow for a field. For example, before it was perforated at a higher level, Mesa 6–1 well had a flow of approximately 14 kg/s with a flashing level at about 1,300 m below ground level. Thus, a pressure drop of 110 bars could yield only 14 kg/s from the formation because of restricted permeability. On the other end of the spectrum, because of very high permeabilities, flows in some wells at Wairakei are 130 kg/s in only a 20-cm casing. At temperatures close to 150°C, and in some high-temperature fields, downhole pumps are likely to be used to provide additional drive and to maintain pressures that prevent the boiling of water and deposition of scale. The pumps are likely to provide on the order of 10 bars increased pressure (Mathews and McBee, 1974), which could be of great assistance in low-temperature systems where it may be desirable to keep the fluid all liquid. However, this would not greatly improve a high-temperature flashing well in rocks of low permeability where stimulation may be more advantageous.

Reserves and resources

To determine where each of the systems listed in table 16 fits into reserves, paramarginal resources, and submarginal resources, the next step is to apply figure 14 to each system. Although the temperature is known for each of these systems at least by geochemical methods, the average well flows are known for only a few of these systems. In lieu of an objective analysis, subjective decisions were made as to the most likely divisions between the various categories. The most important single factor is temperature; reservoirs above 200°C are most likely to contain reserves. Other data that are utilized to yield the division shown in table 19 of the resources of hydrothermal convection systems for the generation of electricity are significant size and lack of severe problems such as high salinity or fluid supply suspected to be inadequate. The reserves (known and inferred) are about equal to the paramarginal resources. The submarginal resources are smaller than either of the higher categories because the 150°C lower limit assumed by Renner, White, and Williams (this circular) removes many systems that would be submarginal if included. Also, the exploration of the systems at temperatures above but near to 150°C has not been as systematic as at the higher temperatures.

The undiscovered resources given in table 19 have been estimated on the basis that one times the identified resources of vapor-dominated systems will be found in the United States outside of the national parks and that five times the identified resources of hot-water systems will be found (Renner and others, this circular).

INTERMEDIATE-TEMPERATURE HYDROTHERMAL CONVECTION SYSTEMS

The identified hydrothermal convection systems with predicted temperatures from 90° to 150°C have been discussed and their estimated heat contents tabulated by Renner, White, and Williams (this circular). The factors affecting physical recoverability are much the same as those for the high-temperature systems except that boiling in the reservoir is unlikely to be important. Accordingly, the previous estimate of one-fourth of the stored energy recoverable as thermal energy at the surface from high-temperature systems should also apply to the systems of intermediate temperature. Table 20 shows the identified hydrothermal convection systems of intermediate temperature, those above the minimum volume of Renner, White, and Williams (this circular) being listed by name. The data on size of most of these systems are very limited, and many may be significantly larger than the minimum estimates. The sum of the heat stored in the identified systems in table 20 is 345×10^{18} cal, of which 263×10^{18} is predicted for Bruneau-Grandview, Idaho, and 30×10^{18} cal for Klamath Falls, Oregon. If the estimates of volume and temperature by Renner, White, and Williams

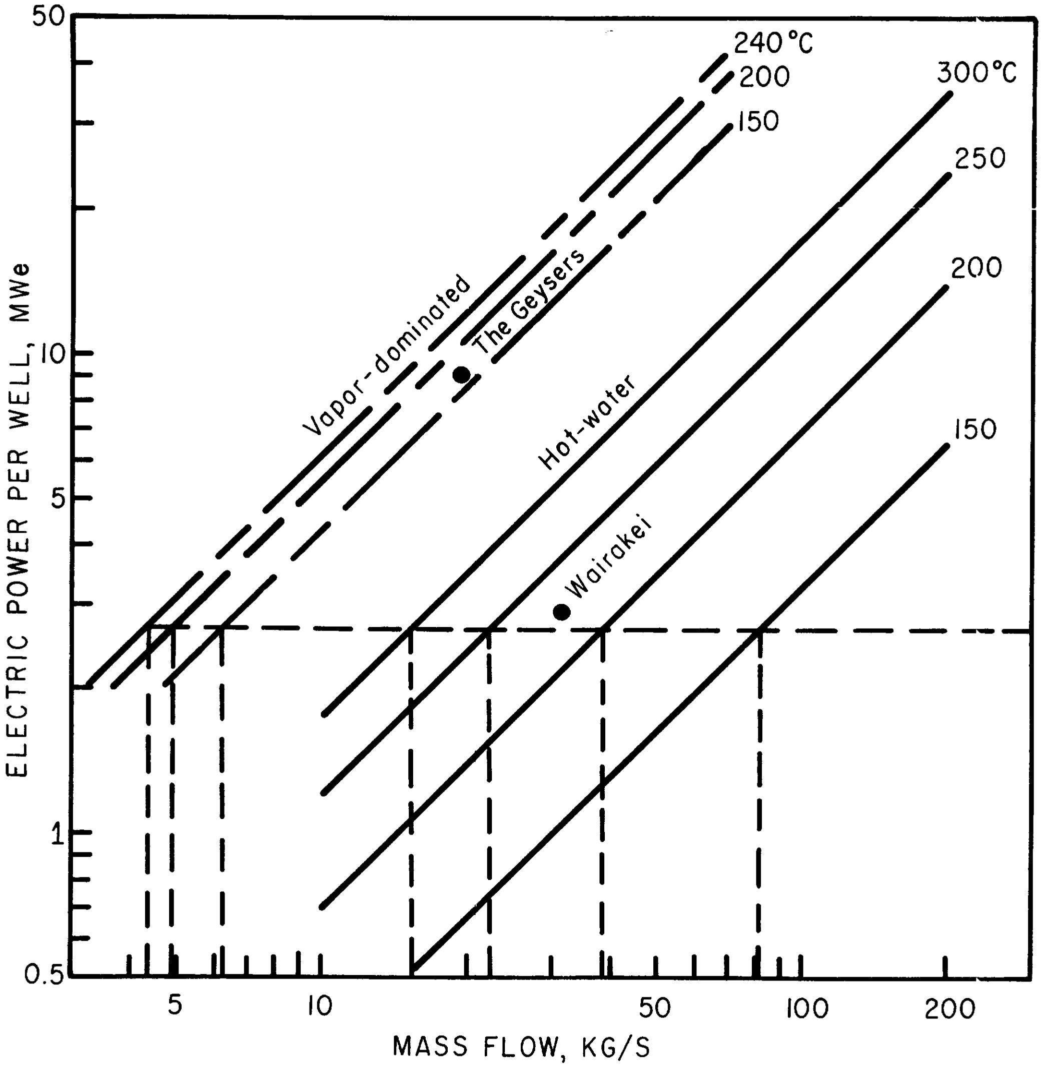

FIGURE 14A.—Electric power per well as a function of mass flow for various temperatures of hot-water and vapor-dominated systems.

(this circular) for Bruneau-Grandview are confirmed by drilling, this system is indeed immense. Until a specific use is chosen for such systems, the fraction of the energy that can be applied to beneficial use cannot be specified. As discussed in Nathenson(1975b), energy equivalent to a temperature drop of 10° to 20°C can readily be extracted for direct use; this energy is more than can be obtained from 150°C water in converting thermal energy into electricity.

In order to obtain some perspective on the predicted 345×10^{18} cal of stored heat in the identified systems in table 20, the amount of beneficial heat may be calculated. The term beneficial heat as used in this report is thermal energy that can be applied directly to its intended nonelectrical use; beneficial heat plus waste heat equals initial available heat. Assuming that one-fourth of the stored energy is recoverable at the surface and that the efficiency of utilization is 0.24 (20°C temperature drop at 100°C, 32°C at 150°C), the recovery factor is $(0.24) \times (0.25) = 0.06$, and the beneficial heat is $(0.06)(345 \times 10^{18})$ cal $= 20.7 \times 10^{18}$ cal. If this usable heat were to be supplied by electrical energy, it would require (20.7) (1327) = 27,000 MW·cent. Thus, if the institutional problems can be resolved, this resource has significant potential.

The economics of process heating present several complications. The first of these is that the value per calorie depends on the temperature of the resource. As the resource temperature becomes

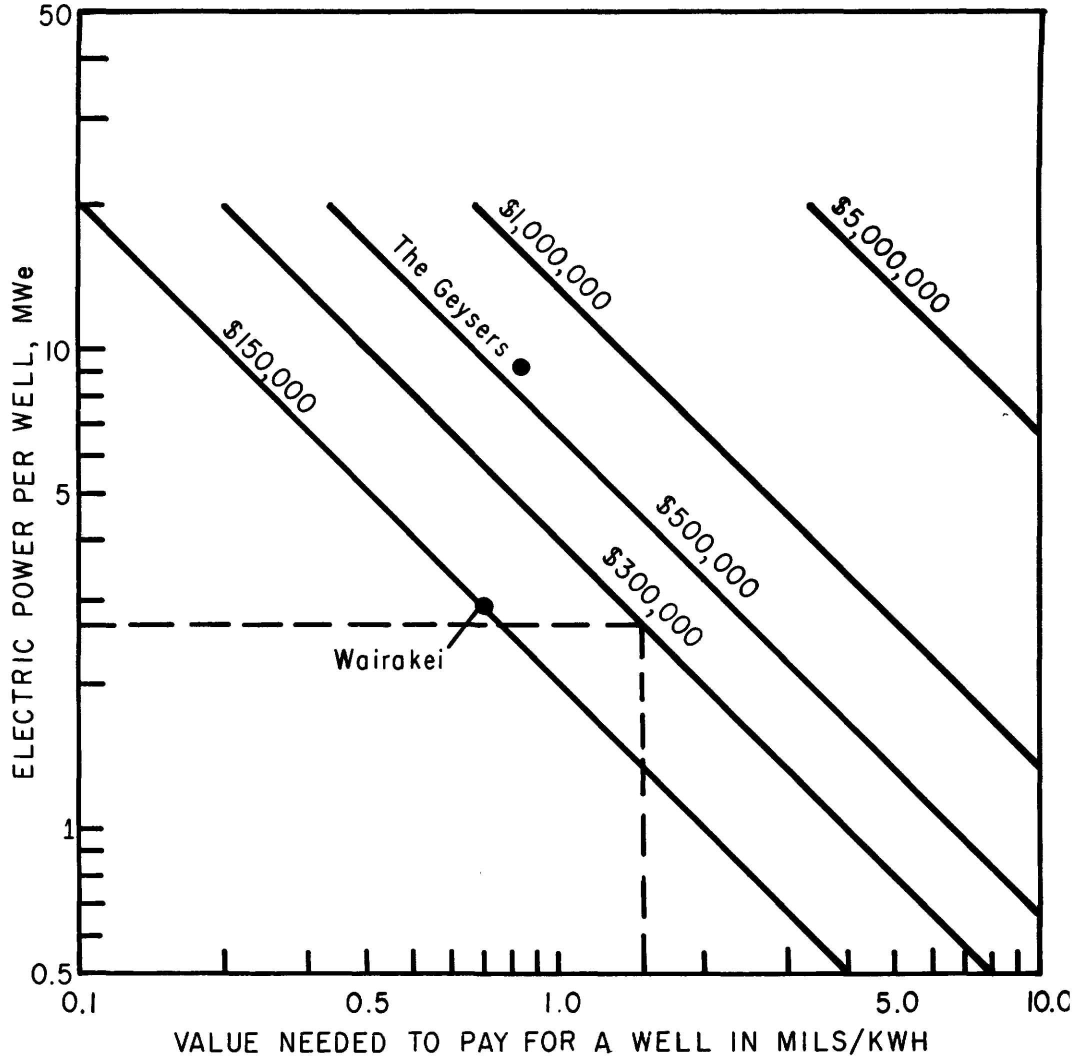

FIGURE 14B.—Electric power per well as a function of value needed to pay for the well for various well costs.

lower, various end uses are no longer possible because some minimum temperature is needed, and the market is thus diminished. Hot water can only be transmitted a few tens of kilometres, so a user must either be available at the site or be induced to move to the site. If no potential user is already in the area, one must be located and persuaded to come to the area (in electricity production, the local power company is always the potential customer). The exploitation of geothermal resources for direct heating is similar to electricity production in that exploration and development of the field can only be done by an organization designed for high-risk operations, such as a resource exploration company or an oil company.

Assuming that these organizational problems are solved as potential direct uses become attractive, the chances for improved economics are high. The costs of this kind of exploitation involve exploration, field development, and construction of transmission system. If the transmission costs can be kept low by near-site use, the lack of an expensive power plant permits a relatively low investment per energy unit.

Because of the limited knowledge in this country of the economics of using geothermal energy for direct heating, the division of the resource into reserves and paramarginal and submarginal resources is not feasible. The sum of the identified systems in table 20 is reported as a total identified resource of 20.7×10^{18} cal. The undiscovered re-

Table 19.—Resources of hydrothermal convection systems for the generation of electrical power in megawatt centuries (not including national parks)

Identified		Undiscovered	
Reserves (Measured and inferred)	3500 MW·cent		
Paramarginal Resources	3500 MW·cent	Resources	38000 MW·cent
Submarginal Resources	>1000* MW·cent		

*The small amount of submarginal as compared to paramarginal resources results from the 150°C lower temperature limit assumed by Renner and others (this volume); this lower limit removes many systems which would be submarginal if included. Also, exploration of systems at temperature above but near to 150°C has not been as systematic as of systems at the higher temperatures.

Table 20.—Intermediate-temperature hydrothermal convection systems with stored heat greater than or equal to 0.3×10^{18} cal (from Renner and others, this circular)

	Subsurface temperature °C	Volume km^3	Stored heat 10^{18} cal
ALASKA			
Okmok caldera	125	6	0.4
CALIFORNIA			
Wendel-Amedee area	140	14	1.1
Wilbur H.S. area	145	32	2.5
Randsburg Steam Well	125	3.75	0.3
Glamis (East)	135	6	0.4
Dunes	135	9	0.6
COLORADO			
Cottonwood Springs	110	6	0.3
Mt. Princeton Springs	115	7.5	0.5
HAWAII			
1955 Eruption area, East Rift	150?	4	0.3
Puulena area, East Rift	150?	4	0.3
IDAHO			
Stanley Hot Spring	110	6	0.3
N.E. Boise Thermal Area	125	8	0.5
Bruneau-Grandview	145	3375	263
Near Banbury	140	12	0.9
Near Cedar Hill	120	9	0.6
Raft River Thermal Area	140	30	2.3
Wayland Hot Spring	130	7.5	0.5
NEVADA			
Near Soldier Meadow	115	12	0.7
Double Hot Spring	145	20	1.6
Fly Ranch Hot Spring	130	16	1.1
Hot Springs Point	125	7.5	0.5
Buffalo Valley Hot Spring	130	10	0.7
OREGON			
Mt. Hood	125	4	0.3
Klamath Falls	120	480	30
Summer Lake Hot Spring	140	6	0.4
Fisher Hot Spring	130	4.5	0.3
Near Harney Lake	135	4.5	0.3
UTAH			
Monroe (Cooper) Hot Spring	120	7.5	0.5
TOTAL:			311×10^{18} cal
Other systems estimated to have minimum volume			34×10^{18} cal
Total, all systems:			345×10^{18} cal

sources are estimated to be three times the known resources (Renner and others, this circular) or 62.1×10^{18} cal. If these predicted quantities are confirmed in future investigations, they are likely to become increasingly attractive for a variety of uses as alternative sources of energy become more expensive and unreliable.

SOLIDIFIED IGNEOUS SYSTEMS

The resource base in igneous systems has been discussed and their estimated heat contents and solidification states tabulated by Smith and Shaw (this circular). The problems and possibilities of obtaining energy from the molten parts of igneous systems are discussed by Peck (this circular). With current technology, the heat stored in the molten parts is not recoverable, and this large quanity of heat represents an attractive target for research. Much heat is also stored on the margins of the molten systems and in the now-solidified systems. Some of these systems without associated hydrothermal convection systems may still have temperatures significantly above the temperatures indicated by regional heat flow. These areas constitute the most favorable targets for recovery using hydraulic fractures to produce circulation loops in a body of low-permeability rock (the "hot dry rock" of Smith and others (1973). Inasmuch as many problems need to be solved before this is a proven concept, the stored heat cannot be considered a resource until the technology has been proven.

The basic scheme for recovering energy from hot dry rock involves drilling a hole several kilometres into hot rock, creating a large hydraulic fracture about 1 km in diameter, and casing the hole nearly to its bottom. A second hole is then drilled to intersect the top of the fracture. Cold fluid is injected into the deep hole as hot fluid is produced from the shallow hole. Assuming that such a circulation loop can be created, the power output is strongly dependent on the behavior of the circulation loop. Harlow and Pracht (1972) suggest that permeability and porosity will grow from cracking due to stress caused by thermal contraction. If this happens, the volume of rock through which the water is flowing will grow with time, and nearly constant power levels can be maintained. The fraction of stored energy so obtained at the surface is likely to be similar to that for hydrothermal convection system. Well flows in this case are determined by crack thickness and length and available pressure drive due to buoyancy and any pumping. It is also possible that porosity will not be created. In this case, heat transfer from the rock to the water flowing in the crack is by conduction. As shown by Nathenson (1975b), the fraction of stored energy obtained at the surface is then likely to be one-half of the value for hydrothermal convection systems. The temperature of produced fluids declines with time, and most utilization schemes do not tolerate much variation in temperature. The well flows in this case are determined by crack size and the chosen values for abandonment time and temperature. Answers to the various questions about this technique should be provided by research programs currently underway.

CONDUCTION-DOMINATED AREAS

The conductive thermal regime of the upper 10 km of the Earth has been discussed and heat contents tabulated by physiographic province (Diment and others, this circular). Because of the limited knowledge of radioactive heat generation and thermal conductivities, temperatures at depth cannot be specified completely, but broad limits can be put on them. The only parts of the resource base that are currently recoverable in addition to geopressured areas are the sedimentary basins of adequate permeability. Low-temperature waters for direct heating can be recovered by the sweep process discussed by Nathenson (1975b). Possible areas of higher-than-normal heat flow include parts of the Salton Trough not associated with hydrothermal convection systems, individual basins of the Basin and Range province, and the Rio Grande rift. Other basins in the normal heat-flow provinces are also possible, but temperatures do not increase rapidly unless conductivities are low, thus providing a "thermal blanket" (Diment and others, this circular). The recovery of such energy is likely to be generally submarginal, but more data are required.

Most of the heat stored in the conductive areas of the United States is in rocks too low in permeability to be considered recoverable. Except in unusually favorable combinations of high reduced heat flow, high radioactive heat production, and/or a low-conductivity blanket, temperatures are not high (Diment and others, this circular). If a hydrofracturing scheme is proven in the especially favorable areas associated with igneous systems, the same scheme may be applicable to other less favorable areas of the United States.

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Recoverability of Geothermal Energy Directly from Molten Igneous Systems

By D. L. Peck

A very large quantity of heat in molten igneous magma systems at depths less than 10 km has been identified by Smith and Shaw (this circular). On the basis of existing geological and geophysical data, they have listed 17 inferred molten bodies of silicic and intermediate composition in the conterminous United States, 24 bodies of mainly intermediate composition in Alaska, and 1 basaltic body in Hawaii. The total estimated heat energy in these systems is at least $25{,}000 \times 10^{18}$ cal, 30 or more times the estimated heat content of all hydrothermal systems in the United States at depths less than 3 km. Only a fraction of the estimated energy is in the molten bodies themselves—much of it is in the solidified margins of the bodies and the adjacent country rocks. The heat content of the molten or partly molten bodies is estimated to be $13{,}000 \times 10^{18}$ cal, contained in magma at temperatures between 650° and 1,200°C. The large inferred volumes and cross-sectional areas of a number of these bodies make them suitable targets for geophysical exploration.

Unfortunately, this large quantity of heat is not presently recoverable and may never be so. The tops of the inferred magma bodies lie at estimated depths that are all greater than 3 km and mostly are still deeper. Formidable technological problems will have to be solved before the resource can be tapped. The technology for drilling at these temperatures and pressures, for example, is not available at present and may not be developed for many years. The feasibility of utilizing the energy contained in magma is presently being evaluated by Sandia Laboratories under a project that was started in 1974 (Colp, 1974) with the support of the Energy Research and Development Administration. This report is based on the results of that project to date, particularly a workshop on magma (March 1975) sponsored by Sandia Laboratories and the U.S. Geological Survey.

The critical problems involved in recovering energy directly from molten igneous systems can be analyzed in terms of the three major stages in developing the energy of such a body: (1) Finding a magma body and determining its size, depth, composition, heat content, etc., by remote techniques; (2) drilling into the body and emplacing a heat-transfer device sufficiently strong and corrosion resistant to last an appreciable period; and (3) extracting energy from the body at a sufficient rate to amortize development and production costs. The critical technical problems involved in each stage are discussed in the following sections. Because the solution of several of these problems, particularly the critical problem of drilling, lies some years in the future, no attempt is made to analyze the economics of direct utilization of magma energy.

EXPLORATION

To extract energy directly from magma, we must first find magma. Because of the very large costs of drilling such a body, the location, depth, and shape of the body and the heat content and heat-transfer properties of the magma need to be determined by remote techniques. Geological and geophysical techniques provide powerful tools for these tasks and have led to the listing of the 41 inferred bodies in table 8 (Smith and Shaw, this circular). Although these methods are not unequivocal, magma bodies have been identified with considerable probability beneath Yellowstone National Park, Long Valley in California, and Kilauea Volcano in Hawaii.

A large body of data on the outer margins and internal structures of shallow intrusive bodies has been gathered in geologic study of older exhumed plutons, such as those associated with mining districts throughout the Western United States and in many other areas of the world. The depths and temperatures of intrusion can be estimated by geologic analysis, comparison of rock compositions with experimentally studied systems, and mineral geothermometers and geobarometers. The geology at the surface above a specific body, such as the structure and the composition, distribution, and age of young volcanic rocks, provides clues to the shape, depth, composition, and stage of evolution of the intrusion. Geophysical techniques—particularly seismic, electrical, gravimetric, and magnetic methods—can be used to identify the body if its characteristic dimensions are greater than half its depth.

Improvement of several of the more promising techniques would lead to a greater ability to determine the critical physical and chemical parameters of a magma body by remote methods. Exhumed magma bodies need to be reexamined with a view toward reconstructing their structure and dynamics during emplacement and early cooling. Various geophysical techniques need to be tested on known magma bodies, such as the Kilauean lava lakes and the inferred magma below Yellowstone Park. The physical properties of magma, such as compressional and shear wave velocities, electrical conductivities, and density, need to be determined as a function of temperature (650° to 1,200°C), pressure (1 to 2,000 bars), oxygen fugacity, and water content.

DEVELOPMENT

Development of the energy of a molten igneous system entails drilling into the magma, probably at depths of 3 to 6 km (pressures of 1 to 2 kb) and temperatures of 650° to 1,200°C, and extracting heat. During the drilling and throughout the exploitation stage, the borehole must remain undeformed, and the materials of the heat-extraction system must resist corrosion.

Drilling technology appears at the present time to be the most critical limiting factor in the technological feasibility of direct development of magma energy. Magma, at least degassed molten basaltic lava, has been penetrated many times by drilling through the shallow crust of stagnant Hawaiian lava lakes. However, present drilling technology at the pressures of 1 to 2 kb relevant to this discussion is limited to temperatures below 250°C. Major innovations in technology will be needed to reach the magma.

A related problem is the development of equipment for in-hole sampling and measurement of in situ physical parameters at these pressures and temperatures. The eventual development of a magma body will probably require a series of successively deeper drill holes, with detailed sampling at each stage of the pore fluids for chemical analysis (particularly corrosive components) and in situ measurement of such parameters as temperature, pressure, stress, and bulk properties including permeability cation diffusivity, and density. Such measurements are essential to sharpen geophysical interpretations, to determine the strength and ductility of the rocks, and to develop drilling techniques and strategy.

The stability of the borehole during drilling, emplacement of heat-extraction devices, and energy production may be serious problems, depending on the strength and the stress field of the country rocks and solidified margin of the magma body in this hot, high-pressure environment. An experimental rock-deformation program is in progress under the Sandia project to determine the answers. The nature of the local stress field around molten bodies can be studied in general by petrofabric studies of exposed wall-rock of ancient plutons, but the stress field around a specific body under development can be determined only by in situ measurements.

The performance of the heat-recovery system needs to be determined by in-hole testing of equipment before exploitation to insure that the inserted materials retain their mechanical and chemical integrity. Current knowledge about the behavior of alloys and ceramics and the chemical composition of magmas and hydrothermal fluids is sufficient for a general approach to the problem. For a specific magma body, however, a number of physical and chemical parameters of the magma and any related hydrothermal activity above the magma must be determined in order to select the proper materials and predict their reliable performance.

HEAT EXTRACTION

Development of an efficient and durable heat-extraction system is, of course, fundamental for the successful development of magma energy. Essential to the economic exploitation of a specific body of magma is a favorable heat-transfer coefficient for the magma over a 10- to 30-year plant life.

Sandia Laboratories is currently investigating several heat-extraction systems. The initial concept is to use a closed system with a long, heat-exchanger tube in the magma, water for steam generation as a working fluid, and a conventional turbine generator. Other techniques under consideration include the use of gas as a working fluid and solid-electrolyte fuel cells to increase the efficiency of energy extraction. Plans are to extend laboratory tests of a single heat-exchanger tube and boiler to field tests in a Hawaiian lava lake or other suitable locality.

The heat-transfer coefficient of magmas of different compositions is a significant factor in evaluating the feasibility of magma-energy utilization. Will magma in a body under development convect at a sufficient rate near the heat exchanger to increase the rate of extraction of heat significantly over the conductive rate? If not, could this increased rate be induced by the injection of aqueous solutions? If the natural or induced convection is not appreciable, then extraction of heat directly from a molten body probably would not be economically feasible. Preliminary calculations by Hardee (1974) of long-term heat-extraction rates from basaltic magma using a long tube heat exchanger with H_2O as the working fluid yielded rates of only 1 kW/m^2 for nonconvecting magma but rates several orders of magnitude higher for vigorously convecting magma. Most magma bodies are very probably below liquidus temperatures and consist of mixtures of melt and crystals with the possible addition of a gas phase. Relevant critical properties that need to be determined are (1) the distribution of liquid, crystals, gas, and temperature in typical bodies; (2) the viscosity and density of crystal-liquid-gas mixtures; and (3) the heat-transfer coefficients over a range of temperatures for magmas of different compositions.

CONCLUSIONS

Molten igneous systems represent a large part of our geothermal resource base, but that energy cannot be recovered directly at the present time or in the foreseeable future. Whether direct utilization is feasible is not certain. The most critical problems appear to be as follows:

1. Improvement of techniques and development of technology for locating, evaluating, drilling, sampling, in situ measuring of physical parameters, and extracting heat at the temperatures (650° to 1,200°C) and pressures (1 to 2 kb) in the hostile chemical environment anticipated in the vicinity of magma bodies;
2. Keeping drill holes at these depths and temperatures open and stable during drilling and long periods of heat extraction;
3. Will natural convection in the magma appreciably increase heat extraction rates over thermal conduction? If not, can convection be induced?
4. Can these problems be solved with sufficient assurance to justify the expenses of drilling and exploiting an inferred body?

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Assessment of Onshore Geopressured-Geothermal Resources in the Northern Gulf of Mexico Basin

By S. S. Papadopulos, R. H. Wallace, Jr., J. B. Wesselman, and R. E. Taylor

Geopressured zones in the northern Gulf of Mexico basin are known to occur in Tertiary sediments beneath an area of over 278,500 km^2, extending from the Rio Grande in Texas northeastward to the vicinity of the mouth of the Pearl River in Louisiana and from the landward boundary of Eocene growth faulting southeastward to the edge of the Continental Shelf (see fig. 15). During the course of this study, it was found that geopressured zones occur also in unmapped Cretaceous sediments underlying the Tertiary sediments and extending further inland under an additional area of at least 52,000 km^2.

This study assesses in a general way the resource potential of geopressured-geothermal reservoirs within the onshore part of Tertiary sediments under an area of more than 145,000 km^2 along the Texas and Louisiana gulf coast. This assessment covers only the pore fluids of sediments that lie in the interval between the top of the geopressured zones and the maximum depth of well control, that is, a depth of 6 km in Texas and 7 km in Louisiana.

The resource potential of geopressured reservoirs within (1) onshore Tertiary sediments in the interval between the depth of maximum well control and 10 km, (2) offshore Tertiary sediments, and (3) Cretaceous sediments were not included in this assessment. It is estimated, however, that the potential of these additional geopressured reservoirs is about 1 ½ to 2 ½ times that of those assessed in this study.

The resource potential of waters in geopressured-geothermal reservoirs was first brought to public attention by Hottman (1966, 1967). Since that time, Jones (1969, 1970) and Wallace (1970) have discussed and attempted to explain subsurface physical conditions that combine to produce geopressured-geothermal reservoirs. Several analyses of the potential use of geopressured-geothermal energy for electrical power generation and estimates of the magnitude of the resource have been presented (Parmigiano, 1973; Herrin, 1973; Wilson and others, 1974; Durham, 1974; Myers and others, 1974; Dorfman and Kehle, 1974; House and others, 1975).

Unlike other geothermal areas that are being considered for the development of energy, the energy potential of the waters in the geopressured-geothermal areas of the northern Gulf of Mexico is not limited to thermal energy. The abnormally high fluid pressures that have resulted from the compartmentalization of the sand and shale beds that contain these hot waters are a potential source for the development of mechanical (hydraulic) energy. In addition, dissolved natural gas, primarily methane, contributes significantly to the energy potential of these waters.

In contrast to geothermal areas of the Western United States, available subsurface information is abundant for the geopressured-geothermal area of the northern Gulf of Mexico basin. The area has been actively explored for oil and gas since the 1920's, and probably more than 300,000 wells have been drilled in search of petroleum deposits in the Texas and Louisiana gulf coast. A large amount of the information obtained from these wells has been made available to the U.S. Geological Survey for use in ongoing projects

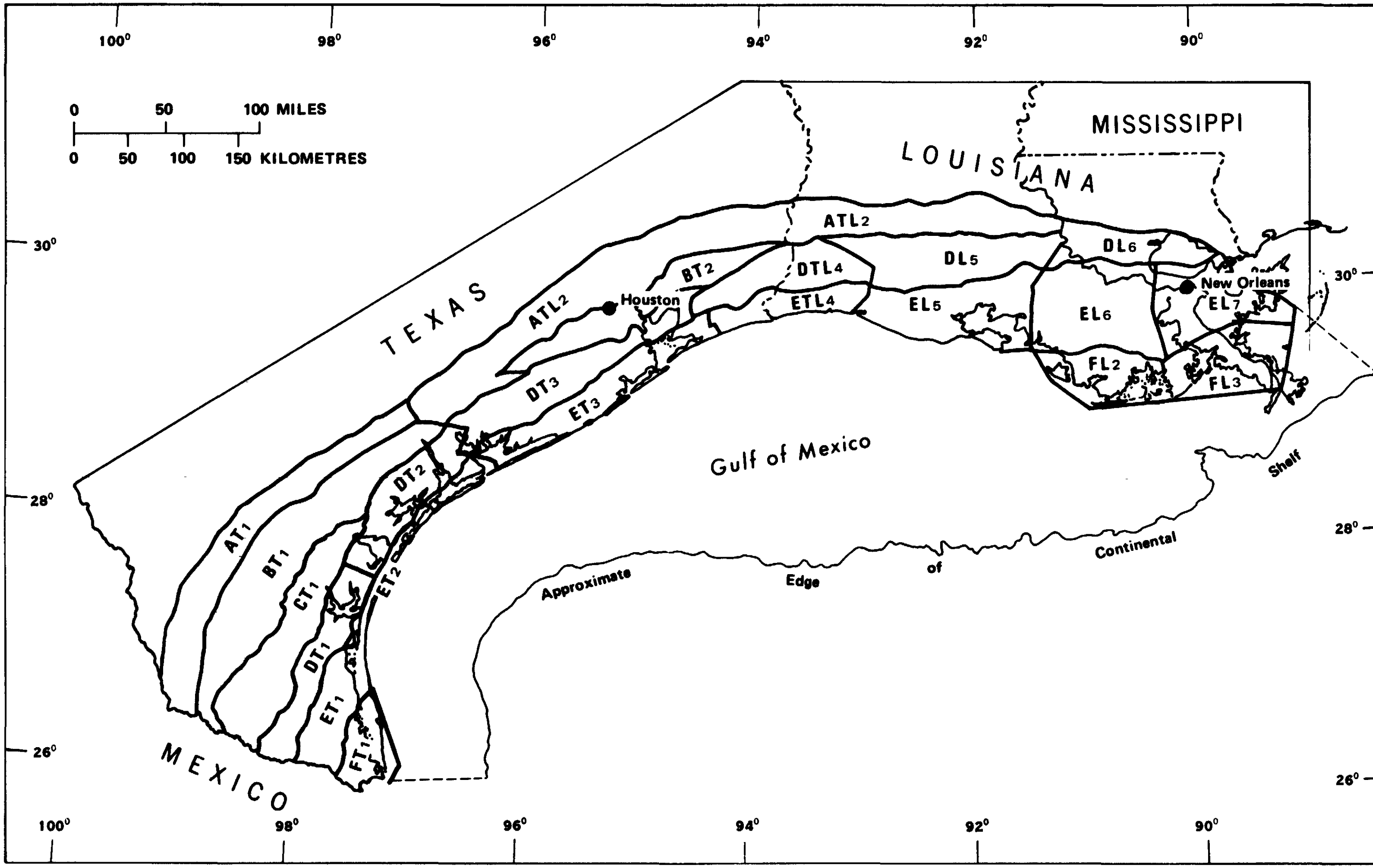


Figure 15.—Location map showing the extent of the assessed geopressured zones and their division into subareas (AT_1, BT_1, and so on).

examining the hydrogeology of the northern Gulf of Mexico basin. Of the vast quantities of data on subsurface conditions that are in the files of these projects, only a small fraction was used in this study. As noted, many promising areas were not included. The data presented here represent general conditions in the various regions outlined. Information on geologic structure, sand thickness, temperature, and pressure were adequate for the purpose of this study. On the other hand, sufficient data on porosity, permeability, and salinity were not available. Note that conditions presented here are idealized and represent best estimates based on the rather small sample of the total available data that were examined within the limited period of time available for this preliminary assessment. The basis on which various data presented in this chapter were determined, calculated, or assumed is discussed in the "Appendix" to this report.

DIVISION OF THE STUDY AREA INTO SUBAREAS

For this investigation, the study area, which encompasses 89,038 km² in Texas and 56,227 km² in Louisiana, was divided into 21 subareas (fig. 15). The boundaries of these subareas were selected on the basis of the geologic history and structure of the study area.

Since Mesozoic time, great river systems have been dumping large volumes of sand and clay along the northern Gulf of Mexico shoreline and shifting its locus gulfward. The process is continuing today. Interconnected fault trends or flexure faults developed along the gulfward margin of each sedimentary sequence as it was overridden by its successor. Compartmented sand and shale beds between these fault trends lie roughly parallel to the shoreline of the Gulf of Mexico. Age of the deposits, as well as fault trends, decreases coastward. The "A" trend (fig. 15) generally coincides with the subsurface distribution of Eocene Wilcox sediments; the "B" trend with Jackson-Yegua beds; the "C" trend with Frio-Vicksburg; the "D" trend with Frio; the "E" trend with Frio-Anahuac; and the "F" trend with Miocene and younger deposits.

These features provided the basic concept for subdivision of the study area. Structural geologic maps of excellent quality prepared for the U.S. Geological Survey by Peppard, Souders, and Associates were used to delineate faults that bound these trends and form the boundaries of the subareas along the trends. Massive shale sequences, although expected to contribute significant amounts of water to adjacent sand beds, are assumed to form the subarea boundaries perpendicular to the various trends.

The geopressured zone to be assessed within each subarea was assumed to extend from the average depth at which pressure exceeds hydrostatic down to the maximum depth of well control. On the basis of an examination of 5 to about 95 wells in each subarea, depending on its size, the depths of 6 km in Texas and 7 km in Louisiana were chosen as reasonable lower limits to which available well control could be extrapolated.

The areal extent of each subarea, as determined by planimetric methods, the average depth to the top of geopressure, and the average pressure, temperature, and salinity at or near the midpoint between the assumed upper and lower limits of the geopressured zone in each subarea are presented in table 21. These average values, which were determined as explained in the "Appendix," are assumed to be representative of conditions in each subarea.

IDEALIZED "CONCEPTUAL RESERVOIRS"

The sediments within the geopressured zones of the study area form a very complex hydrogeologic system. They consist of interbedded sand and shale layers that range in thickness from less than 1 to more than 1,000 m. A thorough analysis of such a system requires data and an interpretation of a complexity that is beyond the scope of this study. To simplify the analysis, the geopressured zone within each subarea was idealized as a "conceptual reservoir." The effect of using these idealized conceptual reservoirs on the energy assessments that are presented in this report is discussed after each assessment.

The conceptual reservoirs were assumed to have a total thickness equal to the assumed thickness of the geopressured zone in each subarea and to consist of a single sand aquifer underlain and overlain by two single confining shale beds. Both the sand aquifer and the confining shale beds were assumed to be continuous and to exist throughout each subarea. The assumed thicknesses and properties of the sand aquifers and of the

Table 21.—Areal extent and average pressure, temperature, and salinity conditions in each subarea

Subarea	Areal Extent km^2	Average depth to top of geopressure km	Average depth to midpoint km	Average pressure MN/m^2	Average temperature °C	Average salinity No. of samples	Average salinity TDS g/l
AT_1	8,948	2.36	4.18	82.4	186	-	16*
ATL_2	20,965	2.47	4.23	78.5	156	-	20*
BT_1	13,588	1.82	3.91	74.6	170	362	30
BT_2	5,595	2.32	4.16	81.4	150	50	38
CT_1	8,230	2.47	4.23	81.4	172	20	24†
DT_1	4,861	2.92	4.46	88.3	172	95	23
DT_2	5,155	2.68	4.34	86.3	169	2	33
DT_3	7,425	2.41	4.20	80.4	153	67	28
DTL_4	5,102	2.62	4.31	83.4	141	18	46
DL_5	7,015	3.01	5.01	102.0	164	222	65**
DL_6	3,729	3.05	5.02	104.0	160	14	55††
ET_1	5,400	2.96	4.48	83.4	168	34	34†
ET_2	1,938	2.63	4.32	83.4	166	24	17
ET_3	7,496	2.37	4.19	80.4	146	65	27
ETL_4	3,461	2.63	4.31	87.3	140	14	52
EL_5	8,144	3.66	5.33	106.9	165	39	90**
EL_6	8,849	3.32	5.16	105.0	159	87	83
EL_7	6,249	2.99	5.00	100.1	146	9	26
FT_1	2,269	3.11	4.55	86.3	171	5	15†
FL_2	4,707	3.76	5.38	105.0	148	46	84††
FL_3	6,139	3.88	5.44	110.8	151	11	45††
Total	145,265						

* Estimated; samples not available.
† Salinity as NaCl calculated from spontaneous potential of well logs; number of samples refers to number of well logs.
** Only few samples from deep zones are included.
†† No samples from deep zones are included.

Table 22.—Assumed thickness and properties of sand and shale beds in idealized "conceptual reservoirs"

Reservoir	Total reservoir thickness km	Sand aquifer				Upper shale bed		Lower shale bed	
		No. of wells	Extrapolated thickness km	Average permeability md	Average porosity %	Assumed thickness km	Assumed porosity %	Assumed thickness km	Assumed porosity %
AT_1	3.64	13	0.63	20	18	1.80	16	1.21	12
ATL_1	3.53	33	0.91	35	21	1.40	17	1.22	11
BT_1	4.18	9	0.56	15	18	1.89	18	1.73	12
BT_2	3.68	6	0.55	20	20	1.82	17	1.31	11
CT_1	3.53	5	1.14	20	19	1.37	17	1.01	11
DT_1	3.08	9	1.22	25	19	1.19	16	0.67	11
DT_2	3.32	4	0.54	20	20	1.74	16	1.04	11
DT_3	3.59	7	0.55	25	20	1.71	16	1.33	12
DTL_4	3.38	8	0.41	30	21	1.75	16	1.22	11
DL_5	3.99	15	0.56	25	20	2.22	14	1.21	10
DL_6	3.95	4	0.76	25	20	2.14	14	1.05	10
ET_1	3.04	4	1.50	20	19	0.88	16	0.66	11
ET_2	3.37	4	0.61	30	20	1.63	16	1.13	11
ET_3	3.63	12	0.74	35	21	1.63	17	1.26	11
ETL_4	3.37	9	0.42	30	21	1.90	16	1.05	11
EL_5	3.34	13	0.67	40	21	1.78	13	0.89	10
EL_6	3.68	12	0.75	40	21	1.95	14	0.98	10
EL_7	4.01	7	1.02	40	21	1.87	15	1.12	9
FT_1	2.89	4	0.93	25	20	1.18	15	0.78	11
FL_2	3.24	6	0.90	50	22	1.49	13	0.85	10
FL_3	3.12	9	0.66	40	22	1.73	13	0.73	9

confining shale beds for each of the idealized conceptual reservoirs are presented in table 22. Also shown on this table are the number of wells used in each subarea to extrapolate the cumulative thickness of sand beds observed in wells to the assumed 6- or 7-km bottom of the geopressured zones. The permeability of both the upper and lower shale beds in all the conceptual reservoirs was assumed to have a uniform value of 0.0001 millidarcy (md) and therefore is not given in this table. This permeability for the confining shale beds is low enough to provide an effective barrier for the maintenance of high pressures in the sand aquifers but also high enough to account for the significant amounts of water that these undercompacted shale beds are expected to contribute from storage in response to pressure declines in the sand aquifers. Other calculated aquifer parameters and water properties used in the evaluation of these idealized conceptual reservoirs are given in table 23. The storage coefficients presented in this table are based on an assumed uniform specific storage of 3.3×10^{-6} per metre for the sand aquifers. Similarly, a uniform "short-term" specific storage of 3.3×10^{-4} per metre was assumed for the confining shale beds. For further details on the assumptions and the methods used in calculating the parameters in table 23, and on other parameters used in the analysis, the reader is referred to the "Appendix."

ASSESSMENT OF THE "FLUID RESOURCE BASE"

The term "fluid resource base" as used in this report refers to the energy contained in the waters stored in the sand and shale beds of geopressured reservoirs. These waters are hot, confined under high pressures, and assumed to contain dissolved methane. Thus, the fluid resource base consists of the thermal energy, the mechanical energy, and the methane contained in these waters. The fluid resource base as defined above was assessed for each of the 21 conceptual reservoirs in the study area. The results of this assessment are presented in table 24.

The volume of water stored in the sand and shale beds of each reservoir was calculated by using the assumed sand and shale bed thicknesses and porosities shown in table 22. The thermal energy in this volume of stored water was taken as the heat content above 15°C. "In situ" densities (see table 23)—that is, densities corrected to the average pressure, temperature, and salinity conditions given in table 21—and an average specific heat of 4,100 joules per kilogram per degree Celsius (J/kg/°C), corresponding to the average pressures and temperatures, were used to assess the thermal energy of each reservoir. The volume of dissolved methane was obtained by using the calculated methane content (see table 23), converted to thermal energy by assuming a heat equivalent of 3.77×10^{7} J/standard m^3 of methane.

The mechanical energy in each reservoir was calculated as the energy that could be produced by the volume of water that would be released from storage in the sand and shale beds if the hydraulic head throughout the conceptual reservoir declines to the land surface. No time limitations were placed on the decline, but it was assumed that the decline is linear with time and that, therefore, the average operating head would be equal to one-half the initial head.

To provide a basis for comparison with the thermal energy in other geothermal systems assessed in the various reports of this circular, the totals of the estimates of table 24 are summarized here in terms of their equivalent:

	Calories
Thermal energy	$10{,}920\times10^{18}$
Methane:	
Volume: 6.7×10^{14} standard m^3	
Thermal equivalent	$6{,}030\times10^{18}$
Mechanical energy:	
Thermal equivalent	50×10^{18}
Total thermal equivalent	$17{,}000\times10^{18}$

This preliminary assessment of the fluid resource base of geopressured-geothermal reservoirs within the onshore Tertiary sediments of the northern Gulf of Mexico basin indicates that the energy stored in the waters of these reservoirs is very large. However, this assessment is based on idealized conceptual reservoirs and on several assumptions about the physical properties of the reservoirs and of the waters stored in them. The possible effects of this idealization of the reservoirs and of the assumptions made in the assessment are discussed briefly below.

Discussion of the reliability of fluid resource base assessment

The mechanical energy component of the fluid resource base is so small compared to the other two components that any discussion of possible errors in its assessment is not warranted. There-

Table 23.—Calculated aquifer parameters and water properties used in this study

Reservoir	Aquifer Parameters			Water Properties		
	Hydraulic head* (LSD)	Transmissivity	Storage coefficient	Average Density: In situ†	Average Density: At surface**	Methane content††
	m	10^{-4} m^2/s	10^{-3}	kg/m^3	kg/m^3	std. m^3/m^3
AT_1	4,750	6.1	2.1	941	893	11.0
ATL_2	4,060	15.0	3.0	966	925	8.4
BT_1	4,010	4.1	1.8	961	920	9.5
BT_2	4,310	5.3	1.8	980	944	7.5
CT_1	4,450	11.0	3.8	957	913	8.9
DT_1	4,920	15.0	4.0	960	913	9.1
DT_2	4,740	5.2	1.8	969	923	9.2
DT_3	4,210	6.7	1.8	975	934	8.0
DTL_4	4,220	6.0	1.4	997	957	7.3
DL_5	5,360	6.8	1.8	1,003	949	8.0
DL_6	5,590	9.2	2.5	1,000	946	8.3
ET_1	4,300	15.0	5.0	969	925	8.8
ET_2	4,560	8.9	2.0	958	914	9.4
ET_3	4,170	13.0	2.4	981	940	7.8
ETL_4	4,550	6.1	1.4	1,005	962	6.7
EL_5	5,340	13.0	2.2	1,022	964	8.0
EL_6	5,340	15.0	2.5	1,020	965	7.7
EL_7	5,330	20.0	3.4	988	939	8.5
FT_1	4,590	11.0	3.1	963	908	10.2
FL_2	5,040	22.0	3.0	1,028	975	7.3
FL_3	5,830	13.0	2.2	1,003	948	8.2

* Based on average pressure and in situ density.
† At average pressure, temperature, and salinity.
** At saturation pressure and average temperature and salinity.
†† Based on solubility of methane at average pressure, temperature, and salinity.

Table 24.—Assessment of the "fluid resource base"

Reservoir	Volume of water in storage	Thermal energy	Methane Energy		Mechanical energy
			Volume	Thermal equivalent	
	10^{12} m^3	10^{20} J	10^{12}std. m^3	10^{20} J	10^{20} J
AT_1	4.91	32.4	54.0	20.4	0.1
ATL_2	11.80	66.1	99.4	37.5	0.2
BT_1	8.81	53.8	83.7	31.6	0.2
BT_2	3.14	17.0	23.6	8.9	0.1
CT_1	4.62	28.5	41.1	15.5	0.1
DT_1	2.39	14.8	21.7	8.2	0.1
DT_2	2.56	15.6	23.5	8.9	0.1
DT_3	4.04	22.3	32.3	12.2	0.1
DTL_4	2.55	13.1	18.6	7.0	0.1
DL_5	3.82	23.4	30.6	11.5	0.2
DL_6	2.06	12.2	17.1	6.5	0.1
ET_1	2.69	16.3	23.7	8.9	0.1
ET_2	0.99	5.9	9.3	3.5	0.0
ET_3	4.28	22.6	33.4	12.6	0.1
ETL_4	1.73	8.9	11.6	4.4	0.1
EL_5	3.75	23.6	30.0	11.3	0.1
EL_6	4.61	27.7	35.5	13.4	0.2
EL_7	3.70	19.6	31.5	11.9	0.1
FT_1	1.02	6.3	10.4	3.9	0.0
FL_2	2.25	12.6	16.4	6.2	0.1
FL_3	2.67	14.9	21.9	8.3	0.1
Totals:	78.39	457.6	669.3	252.6	2.3

fore, the discussion is limited to the thermal and methane energy components. The variables directly involved in the assessment of the thermal and methane energy components of the fluid resource base are the volume of water stored in the geopressured zone and its density, specific heat, temperature, and methane content.

The volume of the stored water depends on the thickness and porosity of the sand and shale beds. The estimates of cumulative sand and shale bed thicknesses are reliable (see the "Appendix"). The fact that the sand and shale consist of several interbedded units rather than the massive single beds that are assumed for the conceptual reservoirs does not affect the volume of water stored in them. The porosities of the sand beds were estimated from data above or just below the top of the geopressured zone, and that of the shale beds from porosity/depth-of-burial relations for the gulf coast shale beds (Dickinson, 1953). The porosities given in table 22 could be in error by two or three percentage points. This would result in an error of about 20 percent in the estimated volume of stored water. The estimates of average temperature and the calculations of the in situ density and specific heat are fairly reliable, and their combined error could not be more than 5 percent. In contrast, the estimates of methane content are not very reliable. No actual measurements of the methane content of waters from the geopressured zone were available. The assumed methane content is based on the solubility of methane at the average temperature, pressure, and salinity of the reservoir waters, as deduced from the knowledge that waters in nongeopressured zones of the gulf coast contain dissolved gas, primarily methane, at or near saturation (see the "Appendix").

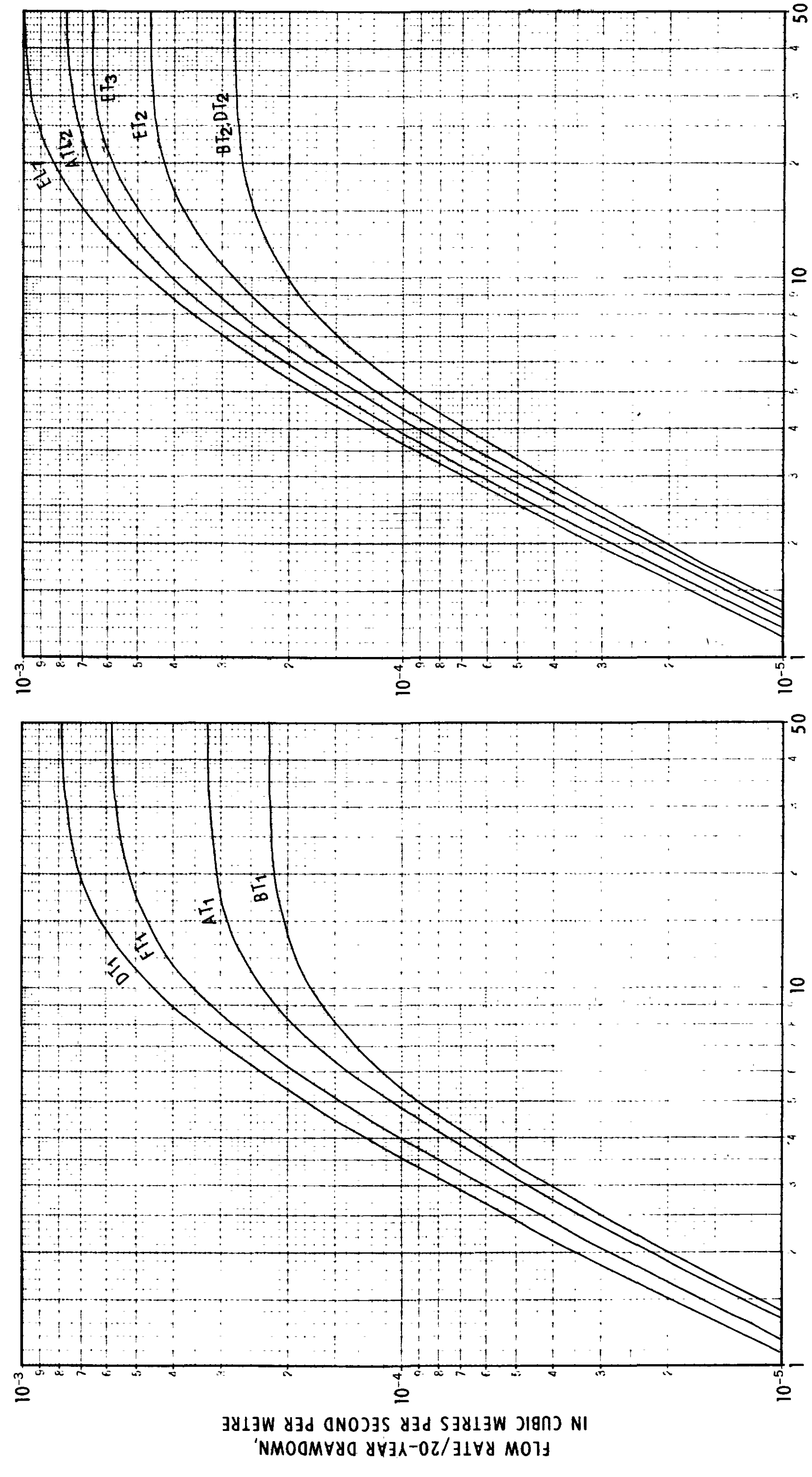
A lower bound for the fluid resource base assessment can be obtained by completely neglecting the mechanical energy component, conservatively assuming that the methane content is one-fourth of that used in the assessment and assuming that the possible errors discussed above are all in the direction of reducing rather than increasing the energy. This lower bound is 398.3×10^{20} J ($9{,}500 \times 10^{18}$ cal), or about 56 percent of the estimate otherwise obtained—still a very large amount.

ASSESSMENT OF RECOVERABLE ENERGY

The recoverability of energy from geopressured reservoirs ultimately depends on the amount of water than can be produced by wells tapping these reservoirs. Therefore, to assess the recoverable energy, a development plan that specifies the production period, the desired flow rate of wells, and the allowable drawdown (or wellhead pressure) needs to be selected. The number of wells that can be placed under this plan and the total production of water are then determined.

The flow rate and the number of wells are interdependent (fig. 16). Note that for a given set of reservoir and well conditions and for a specified drawdown and production period, the flow rate varies with the well spacing and consequently with the number of wells that are placed in a reservoir. As the well spacing increases, the flow rate per well also increases until a maximum flow rate is reached. This maximum flow rate corresponds to a spacing at which interference between wells becomes negligible, and it is controlled mainly by the transmissivity of the reservoir (Papadopulos, 1975). At a smaller well spacing, well interference causes the flow rate per well to decrease. However, as the well spacing decreases, the number of wells that can be placed in a reservoir increases faster than the rate at which the flow rate per well decreases. Therefore, the total production from the reservoir, that is, the product of the number of wells and their flow rate, also increases with decreasing well spacing.

A development plan for recovering energy from geopressured reservoirs cannot be selected only on the basis of hydrogeologic factors. Economic and environmental factors also have to be considered and will probably govern the selection of the development plan. An optimum plan would be one that maximizes economic benefits and minimizes the environmental effects of the development. Distinction should be made here between a plan that is economically feasible and an optimum plan. Any plan that has a benefit-cost ratio larger than one (including tangible or intangible environmental costs) is economically feasible. Among several economically feasible plans, the one providing for the maximum net



FLOW RATE/20-YEAR DRAWDOWN,
IN CUBIC METRES PER SECOND PER METRE
WELL SPACING, IN KILOMETRES
DT1
FT1
AT1
BT1
WELL SPACING, IN KILOMETRES
ET1
AT2
FT3
ET2
BT2;DT2

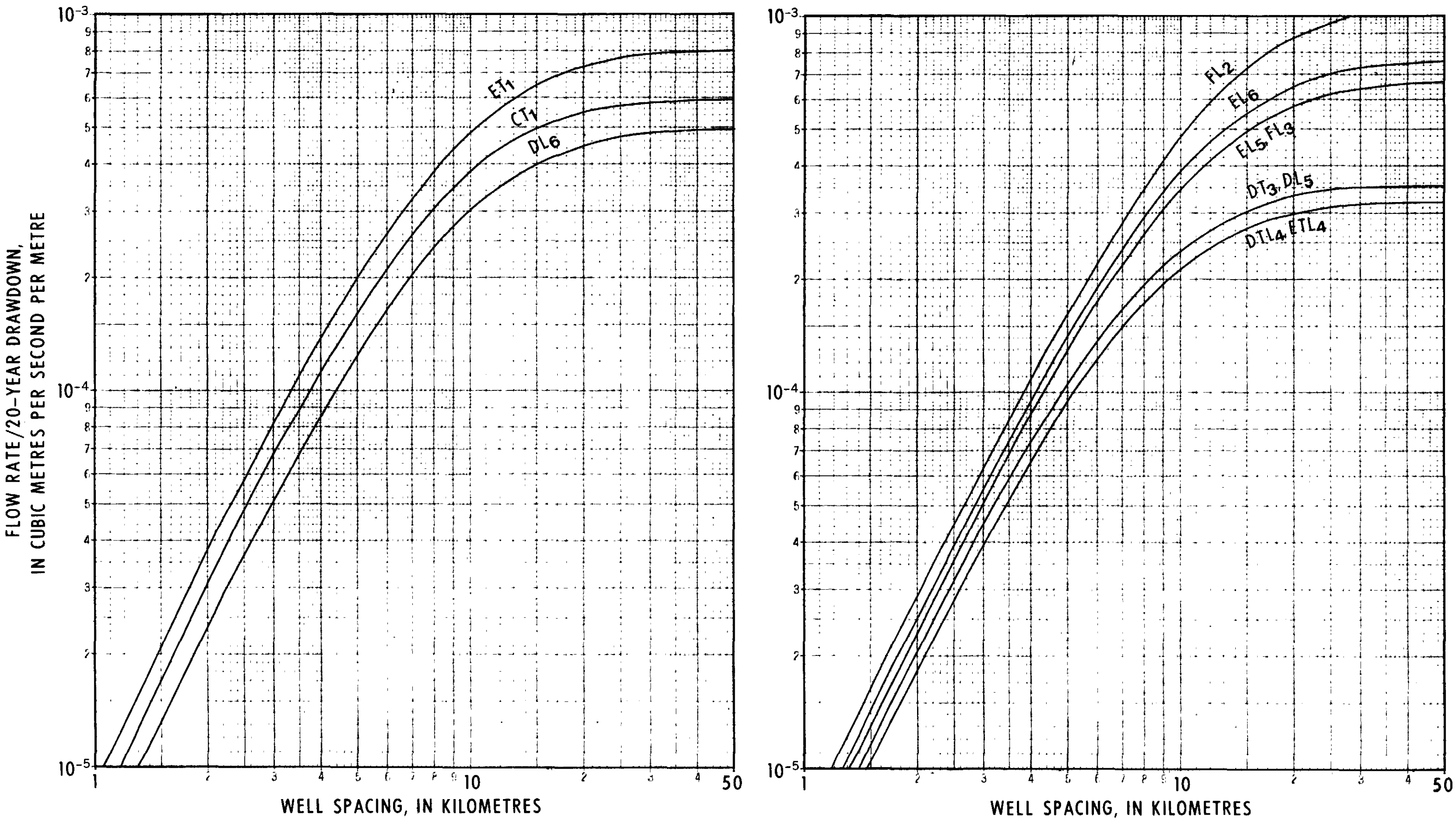


Figure 16.—Variation of well spacing with the ratio of flow rate to the 20-year formation drawdown for each of the conceptual reservoirs.

benefits is the optimum plan. Intuitively, because of the high cost of drilling deep wells in geopressured reservoirs, a plan that develops the reservoirs with wells having the maximum possible flow rate may appear to be the most desirable. However, such a plan is the least likely to be the optimum for the development of the reservoir. For example, in the conceptual reservoir AT_1, wells placed at a spacing of about 40 km would be producing at the maximum possible flow rate for wells in this reservoir, which is 3.3×10^{-4} m^3/s per metre of drawdown (see fig. 16). One such well can be placed in every 1,600 km^2 of area. On the other hand, if wells are placed at one-half this spacing, that is, 20 km apart, the flow rate decreases only slightly to 3.1×10^{-4} m^3/s per metre of drawdown. At this spacing, four wells can be placed in the same 1,600-km^2 area. Assuming that both of these development plans are economically feasible and that costs per well are nearly the same, the one-well plan would have a slightly higher benefit-cost ratio than the four-well plan. However, the net benefits from the four-well plan would be almost four times those of the one-well plan. Thus, if only these two plans were to be considered, the four-well plan would be the optimum.

Selection of an optimum plan of development obviously requires complex studies of economic and environmental factors. In the absence of such economic and environmental studies, which are beyond the scope of this report, three assumed development plans were selected to provide "order-of-magnitude"-type assessments of the energy recoverable from the conceptual reservoirs. The assumed development plans consist of a basic plan, to be referred to as Plan 1, and of two modifications of this basic plan, referred to as Plan 2 and Plan 3.

Plan 1 is the following: The conceptual reservoirs are to be developed with 0.23-m-diameter wells, each capable of sustaining a flow rate of 0.15 m^3/s for a period of 20 years at a wellhead pressure that gradually declines to a minimum value of 14 meganewtons per square metre (MN/m^2), or 2,000 lb/in^2, at the end of the 20-year production period. This basic plan has been used in previous studies (Parmigiano, 1973; Papadopulos, 1975). However, the selection of this plan, and of its modifications discussed below, has been completely arbitrary. The data provided in this report and figure 16 can be used to evaluate possible development plans other than those considered here. Examples illustrating the use of figure 16 for evaluating various plans are given in the "Appendix."

The results of the assessment of recoverable energy from each of the conceptual reservoirs, under the assumed Plan 1, are presented in table 25. These results are summarized here in terms of their equivalent:

	Calories
Thermal energy	229.4×10^{18}
Methane:	
Volume: 13.8×10^{12} standard m^3	
Thermal equivalent	124.2×10^{18}
Mechanical energy:	
Thermal equivalent	9.4×10^{18}
Total thermal equivalent	363.0×10^{18}

Recovery of this energy, which is about 2.1 percent of the total fluid resource base, requires 17,160 wells at a spacing ranging from 2.3 km in subarea EL_7 to 3.9 km in subarea BT_1.

The mechanical energy component of this assessment is only 2.6 percent of the total recoverable energy. To recover this small percentage of energy, the wellhead pressure is restricted to a minimum of 14 MN/m^2. The head equivalent of this pressure is about 1,500 m, or about one-third of the initial head in most of the conceptual reservoirs.

In Plan 2, recovery of the mechanical energy is neglected by removing this restriction on wellhead pressure and allowing the wells to use almost the entire initial head over the 20-year production period. The energy recoverable under Plan 2 consists of only thermal and methane energy. The results of this assessment are:

	Calories
Thermal energy	358.8×10^{18}
Methane:	
Volume: 21.6×10^{12} standard m^3	
Thermal equivalent	193.6×10^{18}
Total thermal equivalent	552.4×10^{18}

The number of wells required to recover this energy, which is about 3.3 percent of the fluid resource base, is 25,840 with well spacings that range from 1.9 to 2.9 km.

Plans 1 and 2 do not consider any restrictions that might be imposed by environmental factors. Subsidence, for example, could pose major re-

Table 25.—Assessment of "recoverable energy" under the assumed basic development plan, Plan 1

Reservoir	Available formation drawdown	Flow rate-drawdown ratio	Well spacing	Number of wells	Volume of water produced	Thermal energy	Methane Energy		Mechanical energy
							Volume	Thermal equivalent	
	m	10^{-5} m^2/s	km		10^{10} m^3	10^{18} J	10^{10}std. m^3	10^{18} J	10^{18} J
AT_1	3,060	4.9	3.1	930	8.80	58.5	96.7	36.5	2.1
ATL_2	2,410	6.2	3.1	2,180	20.64	117.1	173.3	65.4	4.5
BT_1	2,370	6.3	3.9	890	8.43	52.2	80.1	30.2	1.8
BT_2	2,690	5.6	3.5	450	4.20	23.6	32.0	12.1	1.0
CT_1	2,780	5.4	2.6	1,210	11.46	71.5	102.0	38.5	2.6
DT_1	3,250	4.6	2.4	840	7.95	49.6	72.4	27.3	1.9
DT_2	3,090	4.9	3.2	500	4.73	29.3	43.6	16.5	1.1
DT_3	2,580	5.8	3.5	600	5.68	31.9	45.5	17.2	1.3
DTL_4	2,620	5.7	3.7	370	3.50	18.4	25.6	9.7	0.8
DL_5	3,730	4.0	2.9	830	7.86	48.4	62.9	23.7	2.1
DL_6	3,950	3.8	2.5	590	5.59	33.3	46.4	17.5	1.5
ET_1	2,640	5.7	2.4	930	8.80	54.2	77.5	29.2	2.0
ET_2	2,900	5.2	3.2	190	1.80	10.8	16.9	6.4	0.4
ET_3	2,550	5.9	3.2	730	6.91	37.0	53.9	20.4	1.5
ETL_4	2.950	5.1	3.5	280	2.65	13.9	17.8	6.7	0.6
EL_5	3,730	4.0	2.7	1,110	10.51	66.2	84.1	31.7	2.8
EL_6	3,730	4.0	2.6	1,310	12.40	75.1	95.5	36.0	3.3
EL_7	3.680	4.1	2.3	1,180	11.17	59.8	95.0	35.8	2.9
FT_1	2,920	5.1	2.7	310	2.93	18.1	29.9	11.3	0.7
FL_2	3,430	4.4	2.5	750	7.10	40.1	51.8	19.6	1.8
FL_3	4,180	3.6	2.5	980	9.28	52.0	76.1	28.7	2.6
Totals				17,160	162.33	961.0	1,379.2	520.4	39.3

strictions to the maximum drawdowns that can be used in developing geopressured reservoirs, especially near coastal areas of low elevation. Crude estimates, obtained by calculating the volume of water contributed from storage in the shale beds of the conceptual reservoirs (see the "Appendix") and assuming that this will result in equal compaction of the shale beds and that this compaction, in turn, will be transmitted to the land surface without reduction, indicate that average subsidence under Plan 1 could range from about 5 to over 7 m.

In Plan 3, the maximum drawdown is restricted, or the wellhead pressure increased, such that the average subsidence, calculated as described above, does not exceed 1 m. The energy recoverable under Plan 3 is:

	Calories
Thermal energy	53.1×10^{18}
Methane:	
Volume: 3.2×10^{12} standard m^3	
Thermal equivalent	28.8×10^{18}
Mechanical energy, thermal equivalent	3.3×10^{18}
Total thermal equivalent	85.2×10^{18}

Recovery of this energy, which is about 0.5 percent of the total fluid resource base, requires about 3,970 wells placed at spacings ranging from 4.8 to 7.2 km.

These assessments consisted mainly of determining the well spacing and consequently the number of wells that can be developed in each conceptual reservoir under the assumed plans. Techniques developed in the evaluation of the energy potential of a "typical" gulf coast geopressured reservoir (Papadopulos, 1975) were used to prepare for each conceptual reservoir a plot of the variation of the flow rate per unit 20-year formation drawdown, that is, the drawdown in the aquifer just outside the well, which is equal to the drawdown in the well less the pipe friction and other losses, against well spacing for wells placed on a square grid. These plots are shown in figure 16. To prepare these plots, the transmissivity and storage coefficients for each sand aquifer (see table 23) and the assumed hydraulic properties of the confining shale beds (see the "Appendix") were used in well-flow equations that take into account the effects of leakage from storage in the shale beds (Hantush, 1960). For reservoirs where the hydraulic properties of the sand aquifers were the same or very similar, as for example BT_2 and DT_2, only one plot was prepared.

The 20-year available formation drawdown for each reservoir was calculated by subtracting from the initial hydraulic head of the reservoir the assumed maximum drawdown, or the head corresponding to the assumed minimum wellhead pressure, and the pipe friction losses (see the "Appendix") for the assumed well diameter and flow rate. The ratio of the flow rate to the available drawdown was then used to determine well spacing from figure 16. The area of influence of each well was assumed to be equal to the square of the well spacing. The areal extent of each reservoir (see table 21) was then divided by this area of influence to determine the number of wells that can be placed in a conceptual reservoir under the assumed development plan.

The thermal and methane energy in the volume of water produced over the 20-year production period were calculated in a manner similar to that used in the assessment of the fluid resource base, except that density at saturation pressure (see table 23) and an average saturation pressure specific heat of 4,350 J/kg/°C (see the "Appendix") were used in the thermal energy calculations.

On the basis of previous studies of the time-drawdown behavior of wells in idealized geopressured reservoirs having similar properties (Papadopulos, 1975), the "average operating head" for the production of mechanical energy over the 20-year period was assumed to be the initial head less the pipe friction losses and less two-thirds of the 20-year formation drawdown.

The recoverable energy assessments presented above are only energy at the wellhead. Although the methane is a resource that can be directly marketed, thermal and mechanical energy probably will have to be converted to a more readily usable form of energy such as electrical energy. Additional large energy losses will result from this conversion. The efficiency of converting thermal energy at the wellhead to electrical energy has been estimated to be 8 percent by Nathenson and Muffler (this circular). The efficiency of converting mechanical energy to electrical energy is assumed to be 80 percent, although efficiencies as high as 90 percent have been used by House and others (1975). Applying these conversion efficiencies to the thermal and mechanical energy components of the recoverable energy assessments results in the following estimates of electrical energy in megawatt·centuries and of electrical power in megawatts for the 20-year period:

Development plan and energy resource	Electrical energy (MW·cent)	Electrical power (MW (20 yr))
Plan 1:		
Thermal energy	24,380	121,900
Mechanical energy	9,970	49,850
Total	34,350	171,750
Plan 2:		
Thermal energy	38,140	190,700
Plan 3:		
Thermal energy	5,690	28,450
Mechanical energy	3,560	17,800
Total	9,250	46,250

Note that, because of the one-order-of-magnitude difference in the conversion efficiencies for the two kinds of energy resource, the mechanical energy, which constituted only a very small percentage of recoverable energy at the wellhead, under both Plan 1 and Plan 3, provides 29 percent and 38.5 percent, respectively, of the recoverable electrical energy.

The variation in the recoverable energy with the conditions imposed in different development plans further indicates the importance of detailed economic and environmental studies in selecting an optimum development plan. It should be emphasized again that the assessments presented here (1) are not based on such economic and environmental studies, (2) are of the order-of-magnitude type, and (3) could be completely different than assessments based on an optimum plan.

In addition to subsidence, briefly discussed above, studies are also needed to determine the feasibility of disposing of the waste water by injecting it into shallower saline aquifers, as suggested by Wilson, Shepherd, and Kaufman (1974) and House, Johnson, and Towse (1975), or by discharging it into the Gulf of Mexico. The geochemistry of the waters and the shale beds is another factor that needs to be studied. The membrane properties of the shale beds could play an important role in determining the amounts and quality of the waters that are contributed from the shale beds. The chemistry of the water could also create corrosion problems in the wells that would affect their flow rates. Another factor that could influence the economics of development and that needs to be investigated is the feasibility of converting unsuccessful or depleted oil and gas wells into production wells from the geopressured reservoirs. This list is by no means complete, but it points out the complexity of arriving at an optimum development plan for recovering energy from the geopressured reservoirs.

Discussion of the reliability of the recoverable energy assessments

The most important factor in the recoverable energy assessments is the total volume of water produced by wells. The effect of selecting different development plans on the volume of produced water has already been emphasized. Therefore the discussion here is limited to the reliability of the presented assessments. Under the selected development plans that specified the flow rate, the reliability of the produced volume estimates depends on the effects of the idealization of the reservoirs and on the reliability of the well-spacing determinations.

The conceptual reservoirs were assumed to consist of a single sand bed underlain and overlain by two single shale beds that are continuous and exist throughout each subarea. The assumption of two single confining shale beds would normally result in a smaller contribution of water from storage than that expected in a system of interbedded sand and shale. However, this effect has been accounted for by assuming a larger than usual permeability for the shale beds (see the "Appendix"). Theoretically, assuming a single thick sand bed instead of multiple sand beds having the same total thickness would not have any effect since the transmissivity (hydraulic conductivity times thickness) is the same for both cases. However, underlying this theory is the assumption that in both cases the wells are open throughout the sand beds. Multiple completion of the wells in every sand bed would not be practically possible. Therefore, in the actual multiple-sand-bed systems, wells would probably be completed only in the thick sand beds. Assuming that thick sand beds constitute at least 50 percent of the cumulative sand thickness, the transmissivities of the conceptual reservoirs could be in error by about 50 percent.

Existence of the reservoirs throughout each subarea should not be confused with continuity. Since the assumed development plans require a large number of wells in each subarea, artificial boundaries would be created between wells. Thus, the effect of having compartmentalized reservoirs instead of one continuous reservoir would not be so critical, provided these compartments exist throughout the area and are larger than the area of influence of each well, that is, the square of the well spacing. The wells from the logs of which sand-bed thicknesses were determined were rather uniformly distributed over the study area. These logs as well as others examined during the course of this study and of previous studies of the hydrogeology of the gulf coast indicate that sand-bed sequences exist throughout most of the study area, except in the areas where massive shale sequences were chosen as the boundaries of the subareas. Therefore the assumption that the reservoirs exist throughout the study area could result, at most, in an error of about 20 percent in the number of wells.

The well spacing at a specified flow rate depends on the available drawdown and on the hydraulic parameters of the reservoir. The available drawdowns that were used in the assessments were calculated from the initial hydraulic head and the imposed wellhead conditions. The initial hydraulic heads were based on the average reservoir pressures, which are fairly reliable (see the "Appendix"). Therefore large errors in the assumed drawdowns are not likely. Furthermore, the available drawdowns are so large that, even if they are in error by 50 or 100 m, the flow rate/drawdown ratio does not change by much. Consequently, the well spacing that is determined from figure 16 by using this ratio will also not change appreciably.

Permeabilities, which were used to calculate transmissivities, are believed to be the least reliable of the reservoir data used in this study. If permeabilities are assumed to be overestimated by 50 percent, and this error is combined with the 50-percent error assumed to be due to the idealization of the reservoirs, the transmissivities will be reduced to one-fourth of those assumed in table 23. Therefore the effect of having one-fourth of the assumed transmissivities was examined. Under Plans 1 and 2, the drawdowns are large, ranging from 3,000 to 5,000 m, and cause the flow rate/drawdown ratios to be small, from about 3×10^{-5} to 6×10^{-5} m^3/s/m. At this range of flow rate/drawdown ratios, the curves in figure 16, each of which is for a different set of transmissivities and storage coefficients, are close to each other; therefore the well spacing does not vary much with the hydraulic parameters of the reservoirs. On the average, a 75-percent reduction in transmissivities was found to increase well spacing under these two plans by about 0.5 km and to cause a reduction of about 25 percent in the number of wells. Under Plan 3, the drawdowns are small, ranging from 750 to 1,200 m, and result in flow rate/drawdown ratios from 1.25×10^{-4} to 2.0×10^{-4} m^3/s/m. Within this range, well spacing varies considerably with the hydraulic parameters of the reservoir. Under this plan, a reduction of 75 percent in transmissivities reduces the number of wells by about 60 percent.

Lower bounds for the recoverable energy assessments are obtained by combining the effects of the two factors discussed above on the volume of produced water, by assuming that methane content may be only one-fourth of saturation, and by assuming a combined error of about 5 percent in temperature, density, and specific heat. These lower bounds are:

	Lower bounds (10^{18} cal)	*Percent of original assessments*
Plan 1	155.0	43
Plan 2	233.5	43
Plan 3	19.5	23

The corresponding lower bounds for conversion of thermal and mechanical energy to electrical energy and to power are:

	Lower bounds		
	Electrical energy (MW·cent)	*Electrical power* (MW (20 yr))	*Percent of original assessments*
Plan 1	19,850	99,250	58
Plan 2	21,730	108,650	58
Plan 3	2,880	14,400	31

Note that, because the greatest reduction is in the estimates of methane, the reduction in the recoverable electrical energy is not as large as the reduction in energy at the wellhead. Also, if all the possible errors assumed in obtaining these lower bounds are valid, the resulting reservoir conditions may require completely different development plans for assessing the recoverable energy.

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APPENDIX

Discussion of Data Used in Assessment of Onshore Geopressured-Geothermal Resources in the Northern Gulf of Mexico Basin

This appendix discusses the basis on which data presented in the various tables of this report were determined, calculated, or assumed. Other untabulated data and equations used in the assessments presented in this report are also discussed. Examples illustrate the use of the data for evaluating energy recovery plans other than those presented in the main text of the report.

DATA ON AVERAGE CONDITIONS IN SUBAREAS

Average depth to top of geopressure

Geopressure, as used in this report, refers to a fluid pressure that is higher than hydrostatic. For the average salinities of 60 to 80 g/l that are found in the hydrostatic, or normally pressured, zones of the gulf coast, this corresponds to a vertical pressure gradient greater than 10.5 kilonewtons per square metre per metre ($kN/m^2/m$), or 0.465 $lb/in^2/ft$. In the southern half of the Texas coastal plains, a "top geopressure" map furnished by a major oil company was used to determine the average depth to the top of geopressure within each subarea. Pressure measurements available within the mapped area numbered from 22 to 432 determinations per subarea. Outside the mapped area, in the northeastern Texas coastal plain and in coastal Louisiana, average depth to the top of geopressure was determined by using computer-derived shale density plots, well records, and geophysical log examination. Less control, therefore, was used in these areas because of the time required to process information.

Average pressures

Pressures at the midpoint depth of each subarea were obtained by averaging pressure data on shale density determinations from geophysical logs and on mud weights recorded within various depth intervals. Selected wells that approached or penetrated the midpoint depth were used. The resulting pressure determinations were compared with bottom-hole shut-in pressure measurements made during wireline or drill-stem testing and initial reservoir pressure measurements recorded with the Texas Railroad Commission or the U.S. Bureau of Natural Gas. Agreement between the data sets was very good.

Average temperatures

Temperature data were derived from measurements made during electrical well logging operations. These measurements were adjusted to equilibrium conditions using a correction equation developed by the American Association of Petroleum Geologists Geothermal Survey of North America. Mean depths of occurrence for six isogeothermal surfaces (70°C, 100°C, 120°C, 150°C, 180°C, and 250°C) and average midpoint temperatures were calculated for each county (or parish) in the study area. County determinations were averaged to obtain midpoint temperatures within each subarea. Temperature control was most abundant within 20 southern Texas counties (Rio Grande embayment area). In this area, approximately 1,150 wells from a project data file of 11,000 reached or exceeded midpoint depths. Control was limited to about five wells per county in the northeastern half of the Texas coastal plain and in southern Louisiana.

Salinity

Salinity estimates were based upon an evaluation of all chemical analyses from the geopressure interval available to the U.S. Geological Survey. In some instances, data from outside the interval may have been included. In the southern Texas coastal plain (areas CT_1, DT_1, and ET_1), salinities computed from electric logs (total dissolved solids and NaCl) were also available. Computed salinities in table 21 represent an average of the salinity of each significant sand bed occurring in the interval from the top of the geopressure zone to the total depth of the well. In areas DT_1 and ET_1, calculated values were used as best available data. In area CT_1, sufficient data were available for use of both sampled and calculated salinities. Results were very similar (23,000 mg/l versus 22,000 mg/l total dissolved solids). Most sampled data and many computed values used in these estimates were from the upper part of geopressured reservoirs. Within the geopressured zone, an irregular decrease in salinity with depth is usually observed. Thus, average salinity values shown on table 21 are probably high for the average midpoint depth.

DATA ON PROPERTIES OF A CONCEPTUAL RESERVOIR

Thickness

Sand aquifer.—The thickness of the sand aquifer was determined by averaging cumlative sand-bed thicknesses and percentage of sand for a number of wells within each subarea, as shown in table 22, in the interval between the top of the geopressure zone at the well location and the total depth of the well. This percentage was then applied to the interval between the average top of geopressure and the 6- or 7-km depth (that is, the total thickness of the conceptual reservoir) to obtain an extrapolated thickness for the idealized aquifer.

Confining shale beds.—The total thickness of the shale beds was calculated as the difference between the total thickness of the conceptual reservoir and the extrapolated sand-bed thickness. This total shale-bed thickness was distributed between an upper shale bed and a lower shale bed by assuming the following: (1) the vertical flow in an undeveloped reservoir is steady; therefore, if the permeability of both the underlying and overlying shale beds is assumed to be the same, the hydraulic gradient across them should be also the same, and (2) the fluid pressure at the bottom of the lower shale bed should not exceed 95 percent of the lithostatic pressure.

Porosity and permeability

Sand aquifer.—Porosity measurements from sidewall and conventional cores, representing thousands of wells drilled in the gulf coast, show a general decreasing trend with depth. Permeability values also tend to decrease with depth, the lowest values being measured in the vicinity of the top of the geopressure zone. A highly variable but general increase in permeability occurs at intermediate depths within the zone of geopressure. Data available for the sand aquifer porosity-permeability estimates in table 22 were heavily weighed by samples from the uppermost sand beds of the geopressured zone and lowermost sand beds of the hydrostatic zone (that is, the "tightest" interval). Data from project files, published reports, regulatory agencies, and petroleum industry and industry service companies we re organized and analyzed on an area-by-area basis. Observed permeabilities ranged from 0.1 to 1,000 md and porosities from 12 to 30 percent. Real data from horizons selected for the idealized aquifers were scarce or absent from available records. The averages presented are generally not representative of the idealized aquifer zone. From the range of values observed and from conversations with petroleum industry scientists, it is probable that permeabilities of 100 to 400 md are found in the more massive sand beds at depths greater than 4 km. The porosity of these sand beds could also be larger by as much as 5 percent.

Confining shale beds.—As in the case of the sand beds, the porosity and permeability of shale beds tend to decrease with depth, but a general increase of these parameters occur within the geopressured zone. Schmidt (1973) reports that, in a well at the Manchester Field in Louisiana, shale-bed porosity decreased to 12 percent at a depth of 3.0 km, just above the top of geopressure, and then suddenly increased to 17 percent at a depth of 3.4 km, within the geopressured zone, gradually decreasing again to 12 percent at a depth of 4.5 km. In the absence of sufficient specific data on shale-bed porosities within the study area, porosities for the shale beds were estimated by using (1) a plot of burial depth against porosity of the gulf coast shale beds, presented by Dickinson (1953), and (2) the assumed midpoint depths for the upper and lower shale beds. As stated in the report, the permeability for all shale beds of the conceptual reservoirs was assumed to be 0.0001 md. This value of permeability is about two to three orders of magnitude higher than the permeability of normally compacted shale beds. Shale beds within the geopressured zone are undercompacted however, resulting in permeabilities larger than those of compacted shale beds. Furthermore, under actual reservoir conditions of interbedded shale and sand beds, several shale beds would be draining from both sides to adjacent sand beds that are developed, contributing a much larger amount of water to the sand beds than the shale beds of the idealized conceptual reservoir, which consist of only two beds, each draining on only one side. Thus, it is believed that for the analyses of this study, which are based on the conceptual reservoirs, the assumed shale-bed permeability is reasonable.

Hydraulic head

The hydraulic head above the midpoint of each conceptual reservoir was calculated by dividing the average pressure by the in situ density and the gravitational acceleration. This head was converted to head above land surface by subtracting the average depth to the midpoint. Within the sand aquifer, this hydraulic head was assumed to be uniform. Within the upper confining shale beds, the head was assumed to vary uniformly from zero at the top, that is, hydrostatic, to the head of the sand aquifer at the bottom. The same hydraulic gradient was assumed to occur in the lower shale bed. As explained earlier, this assumption was also used in determining the shale-bed thicknesses.

Hydraulic conductivity

The estimated or assumed sand- and shale-bed permeabilities in millidarcys were converted to permeabilities in square metres. These permeabilities were then multiplied by the gravitational acceleration and divided by the average viscosity of the reservoir waters to obtain the hydraulic conductivities of the sand aquifer and of the confining shale beds.

Specific storage

The specific storage—that is, the volume of water that a unit volume of the reservoir would release from storage under a unit decline of head—had to be assumed for both the sand aquifer and the confining shale beds. A specific storage coefficient of 3.3×10^{-6} per metre was assumed for the sand beds in all of the conceptual reservoirs.

The confining shale beds, which are undercompacted and expected to have a high compressibility, were assumed to have a specific storage of 3.3×10^{-4} per metre. This high specific storage, however, was used only for the "short-term" effects of the recoverability calculations. In determining the total volume of storage that could be released from storage, needed for the assessment of the mechanical energy component of the resource base, the average decline of head in the shale beds was first converted to an equivalent increase in burial depth by assuming a lithostatic pressure gradient of 22.6 $kN/m^2/m$ (1.0 $lb/in^2/ft$). The porosity corresponding to this equivalent burial depth was determined from the previously mentioned plot of burial depth and porosity (Dickinson, 1953). The difference between the original porosity and the new porosity was then used to calculate the volume released from storage.

Transmissivity and storage coefficients

The transmissivity and storage coefficients for the sand aquifers were obtained by multiplying the calculated hydraulic conductivities and the assumed specific storage for the sand aquifers by their extrapolated thickness.

DATA ON WATER PROPERTIES

Density

The "at surface" density of the geothermal waters—that is, the density at saturation pressure and at the average temperature and salinity—was calculated from data given by Haas (1970). The "in situ" density—that is, the density at the average pressure, temperature, and salinity—was calculated as follows. Densities for freshwater at the average temperature were calculated both at saturation pressure and at the average pressure using the 1967 ASME Steam Tables (Meyer and others, 1968). The percent change of the freshwater density from saturation to average pressure was then applied to the at-surface density to calculate the in situ density.

Methane content

Gases dissolved in normally pressured gulf coast formation waters consist primarily of methane. Minor amounts of other hydrocarbons plus carbon dioxide and nitrogen are also present. Methane usually represents more than 95 percent of the total volume. Assuming all of the dissolved gas to be methane and correcting for salinity, data from dissolved gas analyses of 134 water samples taken from Tertiary sands in 79 Texas wells were found to approximate saturated conditions as described by Culberson and McKetta (1951, fig. 5, p. 226). These samples were from normally pressured horizons. The highest pressure recorded was about 22 MN/m^2 (3,200 lb/in^2), the highest temperature was about 102°C, and the highest dissolved gas content was 3.064 standard m^3/m^3 recorded from a water sample with 35,000 mg/l total dissolved solids.

Although the dissolved gas content of waters from geopressured zones has not been measured, on the basis of the above observations these waters were assumed to be methane saturated. The methane content was determined by using curves of methane solubility in freshwater presented by Culberson and McKetta (1951, fig. 5, p. 226), which were extended and modified to correct for salinity by using an extension of a solubility table from O'Sullivan and Smith (1970, p. 1461). These extensions and modifications were in accord with trends shown and are nonlinear.

Kinematic viscosity

For the range of the average pressures and temperatures of the reservoirs, the kinematic viscosity of freshwater varies from 1.75×10^{-7} to 2.25×10^{-7} m^2/s (1967 ASME Steam Tables, Meyer and others, 1968). On this basis, an average kinematic viscosity of 2×10^{-7} m^2/s was assumed for the calculations of transmissivities and hydraulic conductivities, and of the Reynold's number needed in the determination of pipe friction losses.

Specific heat

The specific heat of freshwater at the average temperatures found in the geopressured reservoirs ranges from 4,060 to 4,160 J/kg/°C at the average pressures and from 4,290 to 4,430 J/kg/°C at saturation pressure (1967 ASME Steam Tables, Meyer and others, 1968). In this study, average values of 4,100 and 4,350 J/kg/°C were used for the specific heat at the average reservoir pressures and at saturation pressures, respectively.

PIPE FRICTION AND WELL-FLOW EQUATIONS

The head loss, h_l, due to pipe friction in wells was calculated by using the Darcy-Weisbach equation (Streeter, 1962), which, in terms of flow rate Q, well radius r_w, and well depth L, is expressed as

$$h_l = f\frac{Q^2 L}{4\pi^2 r_w^5 g}$$

where f is a friction factor and g is a gravitational acceleration. The Reynold's number for the assumed flow rate, well radius, and kinematic viscosity was calculated to be 4.2×10^6, a roughness coefficient of 1.65×10^{-5} per metre was assumed for the well casing, and the friction factor corresponding to this Reynold's number and roughness coefficient was found to be 0.0118 by interpolating in a Moody diagram (Streeter, 1962).

The well-flow equation used in the preparation of figure 16 is the "modified leaky aquifer" equation presented by Hantush (1960). The equation, which describes the drawdown distribution s around a well producing at a constant rate Q from an infinite confined aquifer, allows for the effects of leakage from storage in both the upper and lower confining beds. The equation has the form

$$s = \frac{Q}{4\pi T}H(u, \beta)$$

where

$$u = r^2 S/4Tt,$$
$$\beta = r\lambda/4,$$

and

$$\lambda = (K'S'_s/TS)^{1/2} + (K''S''_s/TS)^{1/2}.$$

and in which $H(u, \beta)$ is a tabulated function, r is distance from the well, T and S are the transmissivity and storage coefficients of the aquifer, respectively, t is time since the production started, and K', S'_s and K'', S''_s are the hydraulic conductivity and specific storage of the upper and lower confining beds, respectively. The equation is valid for "relatively small times" of t less than both $(b')^2 S'_s/10K'$ and $(b'')^2 S''_s/10K''$. For the smallest shale-bed thickness used in this study this relatively small time covers a period up to about 100,000(!) years.

In the assumed recovery plans in which wells are placed on a square grid, each well can be assumed to be located at the center of a square reservoir having the dimensions of the well spacing. To apply the above equation, which is for an infinite aquifer, to a bounded square aquifer, the method of images was used. The formation drawdown at the well face was calculated for a unit flow rate and different well spacings, and the results were expressed in terms of the flow rate/drawdown ratio as shown in figure 16.

The volume of leakage—that is, the volume of water contributed from storage in the confining beds—which was used to obtain crude estimates of possible subsidence, as stated in the main text of this section, was calculated from the following equation (Hantush, 1960):

$$V_L = V\left\{1 - \frac{2}{\sqrt{\eta\pi t}} + \frac{1}{\eta t}[1 - e^{\eta t} erfc(\sqrt{\eta t})]\right\}$$

where $\eta = T\lambda^2/S,$

$V = Qt =$ volume of produced water,

and in which e^x is the exponential function and $erfc\ (x)$, the complementary error function, which are both tabulated functions. Other symbols are as previously defined. Although Hantush (1960) presents this equation for aquifers of infinite extent, it can be shown that the equation is equally applicable to bounded aquifers without modification.

ILLUSTRATIVE EXAMPLES

The following three examples illustrate the use of figure 16 for evaluating development plans other than those considered in this report. It is assumed that all alternate plans will have a 20-year production period. Well diameters are also assumed to be of 0.23-m diameter. However, the curves in figure 16 are very insensitive to well diameter, and as such they can be used for wells of different diameter. Pipe friction losses, on the other hand, would increase very rapidly with decreasing well diameter.

In the examples, the following symbols are used:

Q = Flow rate of wells,
S_T = Drawdown in wells,
r_w = Well radius,
N_w = Number of wells,
D_w = Depth of wells,
T_s = Sand aquifer thickness,
T_{shu} = Upper shale-bed thickness,
D_{gp} = Depth to top of geopressure,
h_L = Pipe friction losses,
S_f = 20-year formation drawdown,
l = Well spacing,
A_w = Area of influence of wells,
A_T = Areal extent of reservoirs,
$\overline{C}_{20}$ = Average subsidence at the end of the 20-year production period, and
V_L = Volume of leakage from storage in the confining shale beds.

Example I

Given: $Q=0.20$ m³/s and
$S_T=2{,}000$ m,

find: N_w in reservoir AT_1 for the specified Q and S_T.

Solution: $D_w=D_{gp}+T_s+T_{shu}$
From tables 21 and 22:
$D_W=2{,}360+630+1{,}800=4{,}790$ m.
From pipe friction loss equation given in this "Appendix":
$h_L=300$ m,
$S_f=S_T-h_L$
$=2{,}000-300=1{,}700$ m, and
$Q/S_f=1.18\times10^{-4}$ m²/s.
From figure 16:
$l=5.3$ km, and
$A_w=l^2=28.1$ km².
From table 21:
$A_T=8{,}948$ km², and
$N_w=A_T/A_w=318$.

Example II

Given: $l=10$ km, and
$S_T=2{,}000$ m,

find: Q in reservoir AT_1 for the specified l and S_T.

Solution: From figure 16:
$Q/S_f=2.35\times10^{-4}$ m²/s, and
$S_f=Q/2.35\times10^{-4}$
$=4{,}250\ Q$.
From the pipe friction loss equation with $f=0.0118$, $D_w=4{,}790$ m, and $r_w=0.115$ m.
$h_L=7{,}260\ Q^2$
$S_T=S_f+h_L=2{,}000$ m,
or
$7{,}260\ Q^2+4{,}250\ Q-2{,}000=0.$
Solving this quadratic equation and taking the positive root:
$Q=0.31$ m³/s

Example III

Given: $Q=0.20$ m³/s, and
$\overline{C}_{20}=1.5$ m

find: N_w, l, and S_T in reservoir AT_1 for the specified Q and $\overline{C}_{20}$.

Solution: From leakage equation given in this "Appendix":
$V_L=5.52\times10^7$ m³,
$A_w=V_L/\overline{C}_{20}$,
$=36.8$ km²,
$N_w=A_T/A_w$,
$=243$, and
$l=\sqrt{A_w}$,
$=6.06$ km.
From figure 16:
$Q/S_f=1.42\times10^{-4}$ m³/s/m, and
$S_f=0.20/1.42\times10^{-4}$
$=1{,}410$ m
From pipe friction equation given in this "Appendix":
$h_L=300$ m, and
$S_T=S_f+h_L$,
$=1{,}410+300$,
$=1{,}710$ m.

Once the numbers of wells or the flow rate has been determined, the volume of water that can be produced from the reservoir, and hence the energy recoverable under the development plans of the above examples, can be calculated.

Summary and Conclusions

By D. E. White and D. L. Williams

The appraisal of the geothermal resources of the United States presented here is as factual as we can provide from available data. Much effort has been made in each individual chapter to specify the uncertainties and assumptions involved in each estimate; we urge that these uncertainties be kept in mind. The estimates should be regarded as first attempts that will need to be updated as new information becomes available.

This assessment consists of two major parts: (1) estimates of total heat in the ground to a depth of 10 km and (2) estimates of the part of this total heat that is recoverable with present technology, regardless of price. No attempt has been made to consider most aspects of the legal, environmental, and institutional limitations in exploiting these resources.

DEFINITIONS

Resource-related terms used in this circular are defined as follows: *Geothermal resource base* includes all of the stored heat above 15°C to 10 km depth (under all 50 States). *Geothermal resources* are defined as the stored heat, both identified and undiscovered, that is recoverable using current or near-current technology, regardless of cost. Geothermal resources are further divided into three categories based on cost of recovery: (1) *submarginal geothermal resources*, recoverable only at a cost that is more than two times the current price of competitive energy systems; (2) *paramarginal geothermal resources*, recoverable at a cost between 1 and 2 times the current price of competitive energy; and (3) *geothermal reserves* are those identified resources recoverable at a cost that is competitive now with other energy resources. The distinction between resource base and resources is technologic, in contrast to the distinctions between submarginal resources, paramarginal resources, and reserves, which are economic.

RESOURCE BASE

The three major categories of the resource base are shown in table 26.

The hydrothermal convection systems of category 1 (Renner and others, this circular) occur where circulating water and steam are transferring heat from depth to the near surface; they tend to occur in areas of unusually great heat supply and favorable hydrology. These systems are relatively favorable for geothermal development because high temperatures occur near the ground surface and drilling costs are low. We have a sound basis for optimism that many concealed hydrothermal systems exist, and that they can be discovered (see Renner and others, this circular).

The hot, young igneous (volcanic) systems of category 2 (Smith and Shaw, this circular) occur in regions where molten magma has been generated deep in the Earth's crust or mantle and has intruded upward into the shallow crust. Silicic magma (equivalent to granite where crystallized) commonly comes to rest as large masses at depths of a few kilometres, thus conserving its heat; basaltic magma, being much more fluid, is commonly erupted at the surface, where its heat is rapidly dispersed. Many young igneous systems and a few older still-hot systems

Table 26.—Estimated heat content of geothermal resource base of the United States (heat in the ground, without regard to recoverability)

	Identified systems		Identified + estimate for undiscovered
	Number	Heat Content, 10^{18} cal[1/]	Heat Content 10^{18} cal[1/]
1. Hydrothermal convection systems (to 3 km depth, ∿10,000 ft, near the maximum depth drilled in geothermal areas).			
Vapor-dominated (steam) systems	3	26	∿50
High-temperature hot-water systems (over 150°C)	63	370	∿1,600
Intermediate-temperature hot-water systems (90° to 150°C)	224	345	∿1,400
Total	290	∿741	∿3,050
2. Hot igneous systems (0 to 10 km)			
Molten parts of 48 best known, including Alaska and Hawaii		∿13,000	
Crystallized parts and hot margins of same 48		∿12,000	
Total		∿25,000	∿100,000
3. Regional conductive environments (0 to 10 km; all 50 states subdivided into 19 heat-flow provinces of 3 basic types, Eastern, Basin-and-Range, and Sierra Nevada).			
Total, all states		∿8,000,000	∿8,000,000
Overall total (as reported, without regard to significant figures and uncertainties)		8,025,741	8,103,050

1/ 10^{18} calories equivalent to heat of combustion of ∿690 million barrels of petroleum or ∿154 million short tons of coal.

are identified but are not yet evaluated in detail. Estimates of the heat content of hot igneous systems, both evaluated and unevaluated, are included in table 26. All these are favorable target areas in exploring for concealed hydrothermal convection systems.

The stored heat of the conduction-dominated environments, category 3, is huge in quantity, even though temperatures are low, because so much area and volume are involved (Diment and others, this circular). Most of the heat is transferred from the deep, hot interior by thermal conduction through solid rocks, but some is generated by normal radioactivity of rocks, mainly in the upper crust. The entire United States is subdivided into 19 heat-flow provinces that, with present limited data, are classified into three basic types, each with characteristic trends in temperature with depth. The Basin and Range type has the highest temperature gradients; the eastern and Sierra Nevada types have much lower gradients except in special areas, such as the gulf coast, which constitutes a special part of the resource base. Three kinds of potential energy are available from the geopressured pore fluids, including geothermal energy, mechanical energy from the overpressured fluids, and methane dissolved in the pore waters. Heat flows of the gulf coast are presumed to be similar to the eastern type, but adequate data are lacking. Temperature gradients however, are higher than in most of the eastern region because the high-porosity sediments of the gulf coast have low thermal conductivities.

In general, the average heat content of rocks is considerably higher in the Western United States than in the East. This also helps to explain why the most favorable hydrothermal convection systems and the hot young igneous systems also occur in the West.

The anamolous heat of the hydrothermal convection and the hot igneous systems can be considered as "hotspots" superimposed on regional conduction-dominated environments. About 0.01 percent of the total heat stored beneath the United States to a depth of 10 km is in identified hydrothermal convection systems, and about 0.3 percent is in the best known of the hot igneous systems. If our estimates of the undiscovered and unevaluated "hotspots" are valid (table 26), the corresponding percentages are 0.04 and 1.2.

RECOVERABILITY

The useful heat recoverable from identified systems with present or near-current technology and prices (=reserves) and at as much as double present prices (=paramarginal resources) exists almost entirely in the hydrothermal convection systems of the Western States (table 27) and the geopressured sedimentary environment of the gulf coast (table 28).

Resources of the most attractive identified convection systems (excluding national parks) with predicted reservoir temperatures above 150°C (~300°F) have an estimated electrical production potential of about 8,000 MW·cent, or about 26,000 MW for 30 years (Nathenson and Muffler, this circular)[4] Assumptions in this conversion are: (1) one-half of the volume of the heat reservoirs is porous and permeable, (2) one-half of the heat of the porous, permeable parts is recoverable in fluids at the wellheads, and (3) the conversion efficiency of heat in wellhead fluids to electricity ranges from about 8 to 20 percent, depending on temperature and kind of fluid (hot water or steam). The estimated overall efficiency of conversion of heat in the ground to electrical energy generally ranges from less than 2 to 5 percent, depending on type of system and reservoir temperature.

In order to divide the resources of the high-temperature convection systems into reserves and paramarginal and submarginal resources, each system should have been analyzed individually for economic and physical recoverability. In general, the necessary physical data are not available; few systems have been drilled or tested extensively, and the necessary economic data are not well known. No hot-water system in the United States has yet been produced extensively. Thus, in lieu of an objective analysis, subjective evaluations were made for the three resource categories. The most important single factor is temperature; reservoirs above 200°C are most likely to contain reserves. Other utilized data include indicated magnitude of the reservoir and indicated lack of severe problems, such as high salinity and inadequate fluid supply.

[4] A megawatt-century of electricity is a unit of energy equivalent to 1 MW (1,000 kW) of power being produced for 100 years (or 3.33 MW for 30 years). Approximately 1,000 MW (the capacity of many modern nuclear power plants) is required to satisfy the electrical needs of an average city of 1 million people.

Table 27.—Geothermal resources of hydrothermal convection systems assumed recoverable with present and near-current technology and without regard to cost (Nathenson and Muffler, this circular)

	Heat in ground 10^{18}cal 1/	Heat at well-head 10^{18}cal 2/	Conversion efficiency	Beneficial heat 3/ 10^{18}cal	Electrical energy MW·cent 4/	MW for 30 years 5/
High-temperature systems (>150°C; for generation of electricity)						
Identified resources	257	64	0.08 to 0.2			
Reserves					3,500	11,700
Paramarginal resources					3,500	11,700
Submarginal resources					>1,000 6/	> 3,300 6/
Undiscovered resources	1,200	300	0.08 to 0.2		38,000 7/	126,700 7/
Intermediate-temperature systems (90° to 150°C; mainly non-electrical uses)						
Identified resources	345	86	0.24	20.7		
Undiscovered resources	1,035	260	0.24	62.1		
TOTAL	2,837	710		82.8	46,000	153,400

1/ 10^{18}cal (a billion-billion calories) is equivalent to heat of combustion of 690 million barrels of oil or 154 million short tons of coal ; these estimates exclude the national parks.

2/ Assumed recovery factor 0.25 for all convective resources.

3/ Thermal energy applied directly to its intended thermal (non-electrical) use; 10^{18}cal of beneficial heat, if supplied by electrical energy, would require at least 1,330 MW·cent (or 4,400 MW for 30 years); however, a user of this geotherma energy must be located or must relocate close to the potential supply; insufficient data available to predict demand or to subdivide into reserves, paramarginal, and submarginal resources.

4/ Unit of electrical energy; 1 MW·cent is equivalent to 1000 KW produced continuously for 100 years.

5/ Assumes that each MW·cent of electricity can be produced at rate of 3.33 MW for 30 years.

6/ Small because of exclusion of systems with temperatures below 150°C.

7/ Perhaps as much as 60 percent will be reserves and paramarginal resources; costs of discovery and development are more speculative than for identified resources.

Table 28.—Geothermal resources of geopressured sedimentary environments assumed recoverable with present and near-current technology and without regard to cost (Papadopulos and others, this circular)

	Heat in pore fluids, 10^{18}cal 1/	Percent recovery (heat only)	Heat equivalent at well-head 2/ 10^{18}cal	Conversion efficiency	Electrical energy, MW·cent 3/	MW for 30 years 4/
Gulf Coast geopressured fluids in sediments of Tertiary age; assessed on-shore parts only, to depth ranging up to 7 km.	10,920					
Plan 1, maximizes total recovery over 20-year period; no pressure decline below 2,000 psi; 17,160 wells; subsidence estimated 5 to ∿7 m.						
Thermal energy		0.021	229.4	0.08	24,380	81,260
Methane (thermal equivalent)			124.2		5/	5/
Mechanical energy (thermal equivalent)			9.4	0.80	9,970	33,230
TOTAL			363.0		34,350 6/	114,490 6/

1/ Thermal energy only; 10^{18}cal is equivalent to heat of combustion of 690 million barrels of oil.
2/ All plans assume 0.15 m^3/sec flow rate per well and saturation of water with methane, but reliable data lacking.
3/ Unit of electrical energy; 1 MW·cent is equivalent to 1000 kw produced continuously for 100 yrs.
4/ Estimates made for 20 yr. production period; converted to 30 yrs. to be consistent with other estimates of this circular.
5/ Methane assumed recovered but not used locally for electricity.
6/ Perhaps in part reserves but mostly paramarginal, depending on environmental and other costs.
7/ Thermal equivalent of methane included in heat at well-head but excluded from electrical energy; recoverable part highly speculative because of unknown porosities and permeabilities, but probably largely submarginal; note that the recoverable fraction for these environments is assumed to be lower than that for the assessed resources.
8/ No detailed assessment but considered likely to exist in California and other states.

Table 28.—Geothermal resources of geopressured sedimentary environments assumed recoverable with present and near-current technology and without regard to cost (Papadopulos and others, this circular)—Continued

	Heat in pore fluids, 10^{18}cal 1/	Percent recovery (heat only)	Heat equivalent at wellhead 2/ 10^{18}cal	Conversion efficiency	Electrical energy, MW-cent 3/	MW for 30 years 4/
Plan 2, assumes drawdown to land surface, unrestricted subsidence, and mechanical energy not utilized; 25,850 wells; estimated average subsidence >7 m.						
Thermal energy		0.033	358.8	0.08	38,140	127,100
Methane (thermal equivalent)			193.6		5/	5/
Mechanical energy (thermal equivalent)			---		---	---
TOTAL			552.4		38,140 6/	127,100 6/
Plan 3, limits estimated average subsidence to ~1 m; 4,000 wells.						
Thermal energy		0.005	53.1	0.08	5,690	19,000
Methane (thermal equivalent)			28.8		5/	5/
Mechanical energy (thermal equivalent)			3.3	0.80	3,560	11,900
TOTAL			85.2		9,250 6/	30,900 6/
Other unassessed parts of Gulf Coast geopressured environment, on-shore and off-shore to 10 km. 7/	22,000		>500		>50,000	>166,700
Other geopressured environments to 10 km. 7,8/	11,000		>250		>25,000	> 83,300

Nearly one-half of the production potential from the identified systems (3,500 MW·cent or nearly 12,000 MW for 30 years) is considered to be reserves, recoverable with present prices and technology. Paramarginal resources recoverable at as much as twice present prices and with existting and near-current technology are also estimated to be 3,500 MW·cent or about 12,000 MW for 30 years. In addition, high-temperature resources in undiscovered convection systems, using the estimates of Renner, White, and Williams (this circular) and the conversion efficiencies expected of these systems (Nathenson and Muffler, this circular), are estimated to be 38,000 MW·cent or about 5 times that of the identified hot-water systems, excluding the national parks. Of the undiscovered resources, a considerable fraction is likely to be recoverable at present prices and technology, but a larger part will probably be paramarginal.

All of the intermediate-temperature convective resources (90° to 150°C) are submarginal for the generation of electricity, but, under favorable conditions, some are utilizable now for space heating and industrial uses. The potential for nonelectrical uses may attract new industry in many places because the supply is relatively dependable and because the overall efficiency of the direct use of the geothermal energy for heating is greater than for generating electricity for the same purposes (Nathenson and Muffler, this circular). The beneficial heat that can be recovered in favorable circumstances, assuming that a need occurs near the same locality as the potential supply, totals 20.7×10^{18}cal in identified systems (table 27); this is equivalent to about 14.3 billion barrels of oil.

The heat content of pore fluids of the assessed onshore geopressured parts of the gulf coast to depths up to 7 km (Papadopulos and others, this circular) is shown in table 28. This heat component excludes all heat contained in rocks and minerals and also excludes the potential energy of dissolved methane and the mechanical energy from excess pressure. The recoverable part of the total fluid resource base depends critically on the specific plan (or plans) selected for reservoir development. Factors that can be emphasized include: (1) maximizing total recovery from the reservoirs, (2) maximizing production from individual wells, (3) establishing some minimum pressure decline that limits subsidence of the land surface resulting from production, and (4) varying the utilization of methane and mechanical energy relative to thermal energy. A major uncertainty concerns the actual content of methane in geopressured fluids. Saturation of methane at reservoir temperatures, pressures, and salinities is assumed, but reliable data are not yet available. Three different production plans (of the many possible plans) are summarized in table 28.

Assuming that mechanical energy is convertible into electricity with 80 percent efficiency and thermal energy at reservoir temperatures is convertible into electricity with 8 percent efficiency (Papadopulos and others, this circular), Plan 1 can recover 34,400 MW·cent of electricity; Plan 2, 38,100 MW·cent; and Plan 3, 9,250 MW·cent. Each projection is only for the assessed onshore part of the gulf coast. The energy available from methane, which is not limited to utilization at the wellhead and is best considered in this geothermal assessment as a valuable byproduct, is not included. The potential value of the methane, if present in the assumed contents, is somewhat greater than that of the geothermal and mechanical energies. For any one kind of energy alone, the geopressured fluids are probably paramarginal or submarginal resources; when all kinds of potential energy are recovered, a small but significant part of the total resource may be considered as reserves, with much of the remainder as paramarginal resources.

This assessment of recoverability of geopressured-geothermal resources of the gulf coast is necessarily focused on physical recoverability regardless of cost. Economic assessments are greatly complicated by the three kinds of potentially available energy, as well as by possible environmental problems of subsidence and wastewater disposal. Populated areas near the shoreline are relatively sensitive to subsidence, so their potential geothermal resources may not be utilized, at least not until the subsidence effects can be quantified in other less sensitive areas. Although three alternative development plans were used in evaluating all of the 21 subareas of Texas and Louisiana (Papadopulous and others, this circular), different production plans can be chosen for each subarea, tailored to its tolerance of subsidence and other factors.

The very large heat contents of magma, hot dry rocks near volcanic centers, and most conduction-dominated parts of the resource base are not now geothermal resources because the necessary recovery technology has not yet been developed. Recovery technology for the hot dry rocks is now receiving much attention and may be developed in the future.

CONCLUSIONS

1. The geopressured fluids of the gulf coast have a huge geothermal potential. The energy deliverable at the wellhead in the assessed onshore part of the gulf coast varies according to production plan but is likely to range from 9,000 to 35,000 MW·cent (31,000 to 115,000 MW for 30 years). This range excludes the energy equivalent of recoverable methane, which is probably at least equal in value, but reliable data are lacking. Other geopressured parts of the gulf coast and the United States probably have at least 3 times more potential energy in pore fluids than the evaluated part, but the recoverable fraction may be considerably less because of lower average porosity and permeability to be expected in older and more deeply buried sedimentary rocks. Cost analyses of these huge geopressured resources are not attempted here. Such analyses must consider alternate plans for reservoir development but require better data than are now available on reservoir permeabilities, the quantity and value of recoverable methane, the environmental costs related to compaction of reservoirs and disposal of effluent, the value of any incidental petroleum recovered from geothermal production, and possible utilization of wells already drilled for oil and gas. Much of this resource is probably paramarginal under present conditions, but some part is probably recoverable now from areas where the environmental impact is low.
2. The high-temperature convection systems (>150°C) of interest for the generation of electricity are dominantly in the western conterminous United States, but some are in Alaska and Hawaii. The identified systems are estimated to have reserves of 3,500 MW·cent of electricity, producible at present prices and technology, and about equal potential in paramarginal resources. Undiscovered resources of hot-water systems are predicted to be about 5 times greater than identified resources (excluding Yellowstone Park), and undiscovered vapor-dominated systems may be as much as the identified resources. Sixty-three specific localities are identified that probably have reservoir temperatures above 150°C; these and other favorable volcanic areas are available for detailed exploration and assessment. The total energy recoverable with present technology from undiscovered resources is estimated to be about 38,000 MW·cent, with perhaps 60 percent recoverable at prices as much as double present prices.
3. The intermediate-temperature convection systems (90° to 150°C) have much potential for supplying thermal energy for home and industrial heating, thereby releasing oil and gas for more critical uses. If 25 percent of the stored energy is recoverable at the surface and the efficiency of utilization of wellhead energy is 0.24 (20°C temperature drop at 100°C, and 32°C at 150°C), the overall recovery factor is 0.06, and the potential beneficial heat is 20.7×10^{18} cal (table 27). If this heat were to be supplied by electrical energy, the equivalent of 27,500 MW·cent would be required (efficiency of electrical to thermal assumed 100 percent). This resource clearly has significant potential wherever the demand for thermal energy can be located or relocated close to the potential supply. Two hundred and twenty-four identified systems are tentatively included in this category, but most of the predicted resources are in only a few large systems.
4. The hot igneous magma systems and some areas of high regional temperature gradients provide relatively favorable areas for utilizing the heat of hot dry rock, provided that satisfactory methods

of recovery can be developed. The potential resources are huge; temperatures near or above 315°C at 10 km and 200°C at 6 km are likely to characterize parts of the favorable conduction-dominated environments such as much of the Basin and Range province; most of the hot igneous systems are likely to have even higher temperatures.

5. Disregarding cost, the total magnitude of geothermal resources that can be recovered by present and near-current technology is very large. The identified high-temperature convection systems and the evaluated onshore parts of the gulf coast geopressured system are estimated to have a production potential of about 42,000 MW·cent (using Plan 1; 140,000 MW for 30 years) of recoverable electricity, equivalent to 140 Hoover Dams or 140 average modern nuclear power plants. The undiscovered convection systems and the unassessed geopressured parts of the gulf coast and other sedimentary basins have a production potential that is not yet known but may be at least 100,000 MW·cent (or 330,000 MW for 30 years). Perhaps half of this total can be recovered with existing technology at prices that range up to double present costs.
6. These assessments represent our best estimates of the Nation's geothermal resources as of June 1975. More precise estimates will require detailed investigations of the areas tabulated in this report as well as other areas that will be discovered. These investigations must include extensive drilling, reservoir evaluations, and research to attain a better understanding of the characteristics of these systems.

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