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Undiscovered Geothermal Resources of the United States

by

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Resource assessment can be defined as the broad-based estimation of future supplies of minerals and fuels. This assessment includes not only the quantities that could be produced under present economic conditions, but also the quantities not yet discovered or that might be produced with improved technology or under different economic conditions. Resource assessment requires (1) estimation of the amount of a given material in a specified part of the Earth's crust and (2) estimation of the fraction of that material that might be recovered and used under assumed economic, legal, and technological conditions. Geothermal resource assessment is thus primarily the estimation of the thermal energy in the ground, referenced to mean annual temperature, coupled with an estimation of the amount of this energy that might be extracted economically and legally at some reasonable future time.

Estimates of geothermal resources of the United States vary by as much as six orders of magnitude, from a high of 10 billion megawatt-centuries (MWe-c) to several low estimates that cluster around 10,000 MWe-c (Muffler, 1973). Part of the reason for the discrepancy in published geothermal resource estimates lies in a confusion among the terms resource base, reserve, and resource. These terms come from the oil and minerals industries and have been defined as follows:

Resource base: "*** the sum total of a mineral raw material present in the earth's crust within a given geographic area *** whether its existence is known or unknown and regardless of cost considerations and of technological feasibility of extraction." (Netschert, 1958; Schurr and Netschert, 1960, p. 297)

Reserve: "quantities of minerals *** that can be reasonably assumed to exist and which are producible with existing technology and under present economic conditions." (Flawn, 1966, p. 10).

Resource: "*** that part of the resource base (including reserves) which seems likely to become available given certain technologic and economic conditions." (Netschert, 1958; Schurr and Netschert, 1960, p. 297)

For geothermal energy, geothermal resource base was defined by Muffler and Cataldi (1979) as all the heat in the Earth's crust beneath a specific area, measured from local mean annual temperature. Accessible resource base was defined as the thermal energy at depths shallow enough to be tapped by drilling in the foreseeable future. The geothermal resource was defined as that fraction of the accessible resource base that might be extracted economically and legally at some reasonable future time. The geothermal resource can be divided into economic and subeconomic categories depending on present-day economics. This logic can be displayed as the vertical axis of a McKelvey diagram; this diagram defines the terms resource and reserve.

In the United States, geothermal resource assessment is the responsibility of the U.S. Geological Survey. In meeting this responsibility, the USGS has during the past decade produced three geothermal resource assessments (White and Williams, 1975; Muffler, 1979a; Reed, 1983a). Collectively, these national resource assessments evaluated geothermal energy in five categories:

1. Regional conduction-dominated regimes
2. Igneous-related geothermal systems
3. Geopressured-geothermal energy
4. Hydrothermal convection systems at temperatures $\geq 90°C$.
5. Geothermal regimes at tempteratures 90°C.

Geothermal resources were estimated only for categories 3, 4, and 5. Energy calculations made for categories 1 and 2 are not resource estimates, but are useful in giving a conceptual framework for the occurrence of geopressured and hydrothermal convection systems. In particular, for igneous-related geothermal systems, the thermal energy still remaining in silicic intrusions and adjacent country rock was calculated by conductive cooling models using estimates of the size and age of intrusions (Smith and Shaw, 1979). This calculation assumes that cooling of the igneous body by hydrothermal convection was offset by the effects of magmatic preheating and additions of magma after the assumed time of emplacement (Smith and Shaw, 1975).

The figures estimated and calculated in the 1978 USGS geothermal resource assessment of the United States suggest that the energy in igneous-related systems to a depth of 10 km is 100 times greater than the energy in all identified and undiscovered hydrothermal convection systems to a depth of 3 km, and nearly 1000 times greater than the energy in all hydrothermal convection systems identified to date. From this comparison it can be inferred that very large amounts of geothermal energy yet remain to be found. The major geothermal questions facing us for the United States in general and for the Cascade Range in particular are:

-- What is the true value of this igneous-related energy?
-- What is the distribution of this energy among magma, hot dry rock, and hydrothermal convections systems
-- Where is all this energy located?

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PREVIOUS ESTIMATES BY THE U.S. GEOLOGICAL SURVEY OF GEOTHERMAL RESOURCES OF THE CASCADE RANGE

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Introduction

As part of a second assessment by the U.S. Geological Survey of geothermal resources of the United States (Muffler, 1979), estimates were made of thermal energy in igneous-related geothermal systems and in hydrothermal convection systems with reservoir temperatures $\geq 90^{\circ}C$ in the Cascade Range. The assessment is based on data available July 1, 1978, and is a revision of the first national geothermal resource assessment (White and Williams, 1975). This paper summarizes results of the 1978 assessment for the Cascade Range, in order to provide a starting point for a revised regional assessment of the Cascades. Medicine Lake volcano, California, and Newberry volcano, Oregon, are considered as part of the Cascades in this discussion, although in the 1978 assessment they were not included in that province.

Igneous-related systems

Thermal energy stored in magma, solidified pluton, hot country rock and associated hydrothermal convection systems to a depth of 10 km was estimated for young igneous centers in the western U.S. (Smith and Shaw, 1979). This thermal energy was calculated using a conductive cooling model based on estimates of volumes and ages of presumed shallow silicic magma chambers. The primary assumption of the model is that for a given system a single pulse of magma was emplaced instantaneously and cooled conductively from the time represented by the youngest age of the system. Another assumption is that effects of magmatic preheating and resupply to the system are balanced by hydrothermal cooling. As a best guess, it was estimated that about half of igneous-related geothermal energy exists as magma and nearly that much as solidified intrusion and hot surrounding rock, with only a few percent expressed by hydrothermal convection systems (Muffler, 1981).

The thermal energy calculated for 11 igneous-related systems in the Cascades is approximately 3900 x 10^{18} joules (table 1), an amount approximately equal to that calculated for the Clear Lake volcanic center near the The Geysers geothermal field in California. The 11 Cascades systems represent about 8% of the igneous-related energy in the United States, excluding the thermal energy in the very large Yellowstone-Island Park system. Six igneous-related systems in the Cascades were not evaluated for thermal energy -- Mt. Hood, Mt. McLoughlin, Mt. Jefferson, Mt. Baker, Mt. Rainier, and Mt. Adams -- because adequate age or size data were not available or the magma chamber was thought to be too small and/or at a depth greater than 10 km. Of the 11 evaluated systems, eight appear to be at least partially molten and thus may serve as geothermal heat sources. These are Lassen Peak, Medicine Lake, and Shasta in California; Crater Lake, Newberry, and South Sister in Oregon; Glacier Peak and Mt. St. Helens in Washington.

One aspect of a revised assessment of the Cascade geothermal regime might involve a refinement of igneous-related thermal energy estimates. Based on detailed geologic mapping and geochronological and geophysical studies at several volcanic centers in the Cascades, new age and size data may permit revision of existing estimates, as well as calculation of thermal energy at some of the previously unevaluated systems. However, the assumptions of the conductive cooling model may be poorly approximated in the Cascade Range. Although the Cascades has high regional heat flow and extensive Quaternary volcanism, substantial magma chambers in the upper crust have not been detected by various geophysical techniques (Iyer, 1984). What, then, are the geometries and depths of crustal magma reservoirs underlying Cascades volcanic centers? At what rates is magma supplied to these igneous centers of predominantly intermediate composition and how is it stored in the upper crust? How efficient is hydrothermal cooling of Cascades magma bodies? What is the division of thermal energy among magma, solidified intrusion, and hydrothermal systems in the Cascades? What is the nature of magmatic supply and storage to areas between the major stratovolcanoes?

Hydrothermal systems

Fourteen hydrothermal systems in the Cascades (including Newberry volcano) were estimated in 1978 to contain 84 x 10^{18} joules (table 2), about 5% of the hydrothermal energy of all identified systems in the United States. The recoverable thermal energy (resource) in 3 high-temperature systems -- Morgan-Growler H.S. near Lassen Peak, California; Gamma Hot Springs near Glacier Peak, Washington; and Newberry volcano, Oregon -- is equivalent to 883 MW of electricity for 30 years (as a comparison, the current electrical generating capacity at The Geysers is about 1400 MW.) Because the Lassen system and Ohanapecosh Hot Springs are in National Parks and thus not available for exploitation, the resource was not calculated there. Beneficial or usable heat of 0.5x10^{18} joules was estimated for hydrothermal systems with reservoir temperatures between 90^{0} and 150^{0}C.

Determination of the portion of igneous-related energy that is expressed hydrothermally for various Cascades volcanoes is a first step in characterizing the relationship between magmatic heat sources and hydrothermal systems. Unfortunately, assessment data for this purpose are incomplete in the Cascades. Only three systems have both igneous and hydrothermal estimates to permit calculation of the hydrothermal-to-igneous ratio: Lassen, 5%, Glacier Peak, 4%, and

Newberry, 11%. These values are fairly high compared to some other systems such as Long Valley, Valles, and Yellowstone, where the hydrothermal portion is 1-3%. The overall hydrothermal-to-igneous ratio in the Cascade Range is 2%, excluding hydrothermal systems not obviously associated with igneous systems.

Undiscovered thermal energy in hydrothermal systems in the Cascades is estimated to be very large amount, approximately 20 times the amount stored in identified systems. The undiscovered estimate is simply a multiple of the identified component, chosen subjectively on the basis of the favorable geologic setting. The transport of heat into the upper crust by magma movement, as evidenced by the abundance of young volcanic rocks, and the occurrence of hydrothermal systems along the Range, coupled with the possiblilty of groundwater masking, suggest that a large resource may exist. To confirm the large undiscovered estimate would require some combination of the following:

1) Previous thermal energy estimates be revised upward by applying new temperature and volume data. For example, drilling results at Newberry volcano in Oregon (Sammel, 1981) indicate a higher subsurface temperature there than previously thought, 265^{o}C compared to 230^{o}C. Also, the hot-spring systems along the Western/High Cascades boundary may represent larger, hotter systems at depth. Nominal volumes of 3.3 cu km that were assigned to these systems should be re-evaluated in light of new data;

2) Previously unidentified hydrothermal systems be evaluated. For example, no hydrothermal resources have yet been quantified at Medicine Lake or Crater Lake. The discovery of new Lassen-type systems is unlikely, but more "Newberries" might be uncovered beneath the rain curtain at other volcanic edifices. And we need to determine what, if any, concealed hydrothermal systems exist at depth between the major stratovolcanoes.

Certainly, the educated guess made in 1978 should be re-evaluated in light of a refined magmatic-hydrothermal model for the Cascade Range. What distinguishes the volcanic centers that have hydrothermal systems from those that do not? To what extent does masking of hydrothermal systems by cool groundwater occur? Where are convenient structures for localization of hydrothermal activity? Can we apply information gained at a few type hydrothermal reservoirs to systems elsewhere in the Cascades? If hydrothermal reservoirs are masked and not expressed at the surface, how do we obtain the necessary data on temperatures and sizes for energy estimates, short of extensive drilling? If the Cascade Range does not have large hydrothermal resources, why not?

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Table 1. Thermal energy calculated for young igneous systems of the Cascade Range. Data are from Smith and others (1978).

		Age, Years	Chamber Area, km^2	Chamber Volume, km^3	Heat Now 10^{18} J
CA	Lassen Peak	61	80	400	960
CA	Medicine Lake	<1000	64-74	300	724
CA	Mt. Shasta	<200; 5-9500 ?	50	300	724
OR	Crater Lake	6600	50	>320	>770
OR	Newberry	1300,>6600	32	100	240
OR	South Sister	<2000	30-40	100	240
OR	Melvin-3 Crks	400,000 ?	10	40	76
OR	Cappy Burn	2,500,000 ?	8	40	26
OR	Bearwallow	<1,000,000 ?	8	30	41
WA	Glacier Peak	12,000;<70,000	5	12.5	35
WA	Mt. St. Helens	Active	5	>12.5	>35
				TOTAL	3900 x 10^{18} J

Table 2. Thermal energy estimates of hydrothermal systems in the Cascade Range. Data are from Brook and others (1979). Values of temperature, volume, and stored energy are mean values; associated standard deviations are omitted here for simplicity of presentation but are given in Brook and others (1979).

		Temp. ^oC	Volume km^3	Stored Energy 10^{18} J	Recoverable Energy 10^{18} J	Electricity or Beneficial Heat
>150°C						
CA	Lassen	237	71	42	Nat. Park	--
CA	Morgan-Growler	217	8.3	4.5	1.1	116
OR	Newberry	230	47	27	6.9	740
WA	Gamma H.S.	165	3.3	1.4	0.3	27
						883 MW-30 years
90°-150°C						
CA	Big Bend H.S.	116	3.3	0.91	0.23	0.055
OR	Mt. Hood area	122	3.3	0.96	0.24	0.058
OR	Carey H.S.	104	3.3	0.80	0.20	0.048
OR	Breitenbush H.S.	125	3.3	0.99	0.25	0.059
OR	Belknap H.S.	113	3.3	0.88	0.22	0.053
OR	Foley H.S.	99	3.3	0.76	0.19	0.046
OR	McCredie H.S.	91	3.3	0.68	0.17	0.041
OR	Umpqua H.S.	112	3.3	0.87	0.22	0.052
WA	Baker H.S.	134	3.3	1.07	0.27	0.064
WA	Ohanapecosh H.S.	127	3.3	1.00	Nat. Park	--
						0.5 x 10^{18} J
			TOTAL	84 x 10^{18} J		

SEGMENTATION OF THE JUAN DE FUCA PLATE: A FIRST-ORDER GEOTHERMAL ASSESSMENT OF THE CASCADE RANGE?

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If the assessment of the geothermal potential of the Cascade Range were a straight-forward application of existing techniques, one could argue rather persuasively that the gross plate tectonic setting, although interesting, was likely of marginal use in the assessment process. That is, the real world of geothermal assessments deals with the nature of magmatic/hydrothermal systems in the shallow crust (0-10 km) in localized areas (horizontal scales of at most a few 10's of km). The location of the seismic Benioff zone, the dip of the Juan de Fuca plate, and the like, although ultimately responsible for the volcanism observed in the Cascade Range, would not be the primary input (and in reality probably not considered at all) to the geothermal assessment.

On the other hand, during the past seven years, it has become increasingly apparent that the Cascade Range does not represent a simple assessment problem that is amenable to application of existing methodology. Part of the problem with attempting to understand the geothermal potential of the range is the tendency to treat the entire range as a single province- in essence we have failed to recognize the significant scale changes that exist throughout the range. As an example, it is widely known that the Cascade Range is the volcanic arc for the subduction zone created by the convergence between the Juan de Fuca and the North American plates. On the other hand, the tectonic setting of the range changes markedly from north to south, and these changes are in large part the result of changes in the geometry of the Juan de Fuca plate and the interaction with the North American plate. In this paper I suggest that the changes in the tectonic framework, when coupled with the distribution of late Cenozoic and Quaternary volcanism in the Cascade Range, provide a simple first order geothermal assessment of the Cascade Range from the American border south to about Crater Lake: 1) north of Mount Rainier- zero potential, 2) Mount Hood to Mount Rainier- slight potential (but probably not much above zero), and 3) Mount Hood to Crater Lake- moderate potential (perhaps locally some areas with high potential). Although I have not yet modeled the southern portion of the range (Crater Lake to Lassen Peak), this entire region probably should have moderate potential (again with some local areas of high potential).

The changes in the tectonic framework that allows this first order geothermal assessment is postulated to be the result of the segmentation of the Juan de Fuca plate as it subducts beneath the North American plate. (In this report, segmentation refers to the change in dip of the Juan de Fuca plate landward of the coastal thrust zone). In subduction zones, the tectonic framework is determined in part by both geometry (including convergence rate, direction, dip angle) and the type of interaction between the plates (aseismic- without large magnitude thrust earthquakes or seismic). The lack of Benioff zone earthquakes along most of the subduction zone has made if difficult to determine those aspects of plate geometry related to the dip of the Juan de Fuca plate, and there remains considerable uncertainty as to the nature of the plate interaction.

Although northwestern Washington is the simplest portion of the Cascade Range from the point of view of geothermal assessments (zero), it does provide the only area in the range where there is a tie between the geometry of the Juan de Fuca plate as deduced from the occurrence of Benioff zone earthquakes, and the distribution of crustal seismicity and volcanism in the overlying North American plate. The lack of Cenozoic volcanism probably is the result of three factors: 1) the shallow dip of the Juan de Fuca plate and the thick crust beneath the Olympic Mountains, 2) the thickness of the continental lithosphere in the North Cascades, and 3) the continuous nature of the slab as it subducts to the northeast. The thick crust beneath the Olympic Mountains allows a significantly increased area for the direct interaction between the Juan de Fuca and the North American plates, and this results in horizontal compression dominating the fore arc. The stresses from the plate interaction produce high rates of seismicity in the Puget Sound basin, and work to minimize volcanism in the arc. The thickness of the lithosphere beneath the North Cascades and the continuous nature of the slab may work together to inhibit volcanism by acting as either a heat sink or preventing hot aesthenosphere from coming in contact with the subducting slab. As a consequence there is little late Cenozoic volcanism within the Cascade Range, and the geothermal potential is zero.

Between Mount Rainier and Mount Hood, the geometry of the Juan de Fuca plate is not as well constrained as to the north, but the few Benioff zone earthquakes suggest that the dip of the subducting plate (landward of the thrust zone) is greater than to the north. The sudden end of the Benioff zone earthquakes has been used to suggest that the Juan de Fuca plate may not be continuous south of Mount Rainier, but has broken off as a result of repeat thrust events on the interface with the North American plate. This breakoff of the Juan de Fuca plate may decrease internal gravitational stresses to the point that seismicity ceases within the slab. However, because the slab (oceanward of the proposed break) is still strongly coupled with the overlying North American plate, horizontal compression again dominates the arc region. Earthquake focal mechanisms are largely strike-slip on vertical fault planes, in good agreement with the horizontal stress direction expected from gross plate convergence.

The poor geothermal prospects in most of this area reflect the lack of volcanism observed at the surface and the compressional nature of the crustal stresses. As an example, Mount St. Helens is located at a left-stepping offset along a regional strike-slip fault zone characterized by right-lateral motion. It is likely that the resulting compression at Mount St. Helens has limited the development of secondary vents away from the immediate vicinity of the cone and is responsible for the very small scale of the upper crustal magmatic system (~10 km^3 in the upper 7-10 km). My designation of a slight potential for this region reflects the existence of strike-slip faulting (which may have a favorable opening geometry in some local areas) and the increase in late Cenozoic volcanism southward toward Mount Hood.

Finally, south of Mount Hood, seismic data offer essentially no constraints on the geothermal potential of the Cascades. As with the Mount Rainier to Mount Hood segment of the Juan de Fuca plate, the lack of Benioff zone earthquakes beneath Oregon may suggest that the plate is broken near the coastal thrust zone. Further, the crust beneath the Oregon Coast Range is significantly thinner than that beneath the Olympic Mountains in Washington, as a consequence the contact between the Juan de Fuca and the North American plates is reduced. Superficially, this geometry might lead to extensional processes dominating the arc region, and the volume of late Cenozoic volcanics found in the Oregon Cascades is consistent with this proposed extension. Simply because the crust is under extension rather than compression, the geothermal potential should be higher than to the north in Washington. In addition, local areas with abnormally high rates of crustal extension may exist, and these areas would be expected to have the highest geothermal potential.

Volcanism in the Cascade Range

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Volcanism in the Cascade Range started about 36 m.y. ago and produced such widespread units as the Ohanapecosh Formation in Washington and the Little Butte Volcanics in Oregon. Since then, volcanic activity has been more or less continuous on a time scale of a few million years, although certain segments of the range have apparently had relatively short periods of increased or decreased activity. For example, in Washington few volcanic rocks were erupted 17-15 m.y. ago at the time of maximum outpouring of the Columbia River Basalt Group from fissure vents farther east. In Oregon, on the other hand, that period produced voluminous flows and tuffs of the Sardine Formation. Despite contrary claims, evidence is weak or lacking for regional volcanic episodicity in the Cascades, a concept whose testing is beset with sampling problems.

There is no compelling evidence that volcanism formed a narrow linear chain before about 3-4 m.y. ago. For example, in central Oregon vents of Oligocene age extend from the western Cascades more than 100 km to central Oregon. Even since 3 m.y. ago, volcanism has not been confined to a narrow belt; Mount Adams is 50 km east of Mount St. Helens, Newberry Volcano 60 km east of the Cascade crest, and Medicine Lake 55 km east of Mount Shasta. Minor vents are even more widely scattered. Nonetheless, most eruptions during the last few million years in Oregon have occurred along a very linear chain forming the High Cascades. This chain is also exemplified by north-trending linear fissures, normal faults, and alignments of cinder cones and widely spaced stratovolcanoes.

Most volcanism in the Cascades during the last few million years has been south of Mount Rainier. Only a few volcanoes occur farther north, and all of these fall along a northwest-trending linear belt between Glacier Peak, Washington, and Meager Mountain, British Columbia, that is strongly discordant to the chain that trends south from Mount Rainier. The Glacier Peak-Meager Mountain belt projects southeastward through several known Tertiary structures to the western Snake River Plain and thus may be partly controlled by an old crustal weakness. The belt is nearly perpendicular to the direction of plate convergence, and moderately high seismicity outboard of the belt has been interpreted to define a Benioff zone with low dip. In contrast, the area west of the Oregon and northern California Cascades has low seismicity except along spreading ridges and related transform faults, and no Benioff zone. In that area the Cascade crest is oblique to the convergence direction, normal faults and fissure vents indicate east-west extension, and volcanism is much more voluminous than farther north.

More than 2,000 volcanoes of late Pliocene or Quaternary age have been identified in the Cascade Range; many others lie buried by younger flows. About 50 percent of the volcanoes are in Oregon, 35 percent in California, and 15 percent in Washington and British Columbia. The vast majority of these volcanoes are monogenetic cinder cones or fissure vents that erupted basalt or basaltic andesite flows. In some parts of the range the cinder cones and fissure vents are randomly dispersed, but in other parts they occur in distinct clusters. One such cluster of late Holocene basalt and basaltic-andesite vents occurs between the Three Sisters and Three Fingered Jack in Oregon, and another somewhat older cluster occurs in the Indian Heaven area between Mount St. Helens and Mount Adams, Washington.

Small stratovolcanoes and relatively steep-sided lava shields form perhaps 5 to 10 percent of the total vents. Many appear to be of basaltic-andesite composition, with lesser basalt and andesite, and they rise as much as 500 m above the surrounding lava fields. Most of these volcanoes are probably short-lived, perhaps being formed in a few years to as much as a few thousand years.

Large stratovolcanoes form only about 1 percent of the total volcano population, yet their imposing size leads to popular identification of the range with stratovolcanoes. The largest volcanoes, Mount Rainier in Washington and Mount Shasta in California, lie near the ends of the chain of greatest volcanism in the Cascades. The deeply eroded remains of several large Pliocene and Pleistocene stratovolcanoes have been identified along the chain; few stratovolcanoes persist long as imposing edifices because of erosion, especially by glaciers.

Clusters of stratovolcanoes occur at Three Sisters, Crater Lake, and near Lassen Peak. These clusters represent long-lived volcanic systems that have produced relatively large volumes of intermediate and silicic magma; the cluster at Crater Lake was largely destroyed by the cataclysmic eruption that formed Crater Lake caldera.

Large shield-shaped volcanoes or volcanic fields lie in a north-trending belt along the east side of the Cascade Range. These include Newberry volcano in Oregon and Medicine Lake volcano in northern California; Lassen Peak may lie along this trend. Similar volcanoes or volcanic fields several million years older occur along this same alignment at Yamsay Mountain and Bald Mountain, Oregon, and in the Simcoe, Washington, area. All of these eastern vent areas have erupted silicic as well as mafic lava, and several have summit calderas.

The petrochemistry of many of the stratovolcanoes has been studied, but most of the lava fields between stratovolcanoes have received no such attention and few generalizations can be made about them. Basaltic andesite and basalt appear to dominate in areas between stratovolcanoes, and even some of the stratovolcanoes consist mainly of basaltic andesite rather than andesite; calc-alkaline rocks dominate, but tholeiitic types also occur. Andesite is locally abundant, especially on some of the stratovolcanoes, and in a few areas dacite, rhyodacite, and rhyolite are common. With the possible exception of youthful Mount St. Helens, most areas in which silicic rocks are abundant appear to be products of long-lived magmatic systems from which eruptions have occurred over periods of hundreds of thousands of years. These include Lassen Peak, Shasta, and Medicine Lake in California, Crater Lake, Newberry Volcano, and the Three Sisters-Broken Top highland in Oregon, possibly Glacier Peak in Washington, and Mount Garibaldi and Meager Mountain in British Columbia. Local centers of silicic volcanism that have generated voluminous pyroclastic flows occur at Crater Lake, Medicine Lake, and Newberry Volcano, all of which have calderas; a caldera may lie buried in the Three Sisters-Broken Top highland, judging by the distribution of ash-flow tuffs around that area. This paucity of Quaternary calderas suggests that large high-level silicic magma bodies are not common.

Geothermal resources have been identified at Meager Mountain, B. C., high temperatures have been encountered in shallow holes on Newberry volcano, and several of the volcanoes have fumaroles. Geothermal resources are presumed to be most likely to occur at long-lived volcanic systems that have produced silicic rocks. Some andesite and basaltic-andesite stratovolcanoes may have geothermal resources, although drilling at Mount Hood was not encouraging. A background of relatively high heat flow in the Cascades as a consequence of long-lived volcanism throughout most of the chain may indicate an extensive deep resource.

Quaternary Volcanism in the Southernmost Cascade Range, California

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Volcanism in the southernmost Cascade Range can be characterized on two scales. On a regional scale volcanism is basaltic to andesitic. Individual volcanoes have small volumes and relatively short lifetimes. Many eruptive centers coalesce to form the crest of the Cascade Range. Individual volcanoes range from monogenetic basalt and basaltic andesite cinder cones to larger lava cone and shield volcanoes of basaltic to andesitic composition. Rocks with greater than approximately 60 per cent SiO_2 are sparse. Superimposed on the regional mafic volcanism are long-lived, larger volume volcanic centers that have produced eruption products spanning a wide range of composition from mafic andesite to rhyodacite or rhyolite.

Each volcanic center consists of a large andesitic composite cone and flanking silicic domes and flows. Three of these volcanic centers younger than about 3 Ma have been recognized in the Lassen area. Each has had a similar history, consisting of three stages: (1) an initial cone-building period of mafic andesite and andesite lava flows and pyroclastics; (2) a later cone-building period characterized by thick andesite and silicic andesite lava flows, and (3) silicic volcanism flanking the main cone. The silicic magma chamber of the third stage provides a heat source for development of a hydrothermal system within the core of the main cone. Alteration of permeable rocks of the cone facilitates increased glacial and fluvial erosion of the central area of the volcano. The result is preservation of a resistant rim of the thick later cone-building lava flows and flanking silicic rocks surrounding a depression in the altered and eroded core of the composite cone.

The two older volcanic centers have reached this stage, and their hydrothermal systems are extinct. The younger Lassen volcanic center (LVC) however, hosts active silicic volcanism and a well developed hydrothermal system, including the thermal features in Lassen Volcanic National Park (LVNP) and Mill Canyon.

The LVC developed in three stages. Stages I and II produced the Brokeoff Volcano, an 80 km^3 andesitic stratocone. Stage I deposits consist of olivine-augite and olivine-hypersthene-augite andesite lava flows and pyroclastic rocks erupted from a central vent during the period from about 0.7 to 0.55 Ma. Stage I culminated in eruption of a small volume of hornblende-pyroxene dacite. An interval of erosion lasting 0.1 Ma followed stage I. During stage II, which lasted 0.1 Ma, thick flows of porphyritic augite-hypersthene silicic andesite, generally lacking interbedded pyroclastic material, were erupted. Volcanism then shifted to the north and east and became more silicic. Eruption of a minimum of 50 km^3 of rhyolitic magma at 0.35 Ma as air-fall tephra and ash flows initiated stage III. This eruption probably produced a collapse caldera now filled by a dacite domefield. The dacite domefield consists of two groups of rocks totaling on the order of 30-50 km^3. During the period from 0.25 to 0.2 Ma pyroxene-hornblende dacite magma produced a group of six domes and flows. Hornblende-biotite rhyodacite has been erupted as domes, lava flows and pyroclastic flows in at least ten episodes over the past 0.1 Ma. At least six times during the past 0.1 m.y. hybrid andesite totaling 10 km^3 and consisting of thoroughly mixed mafic and silicic magma has been erupted from the margins of the dacite domefield.

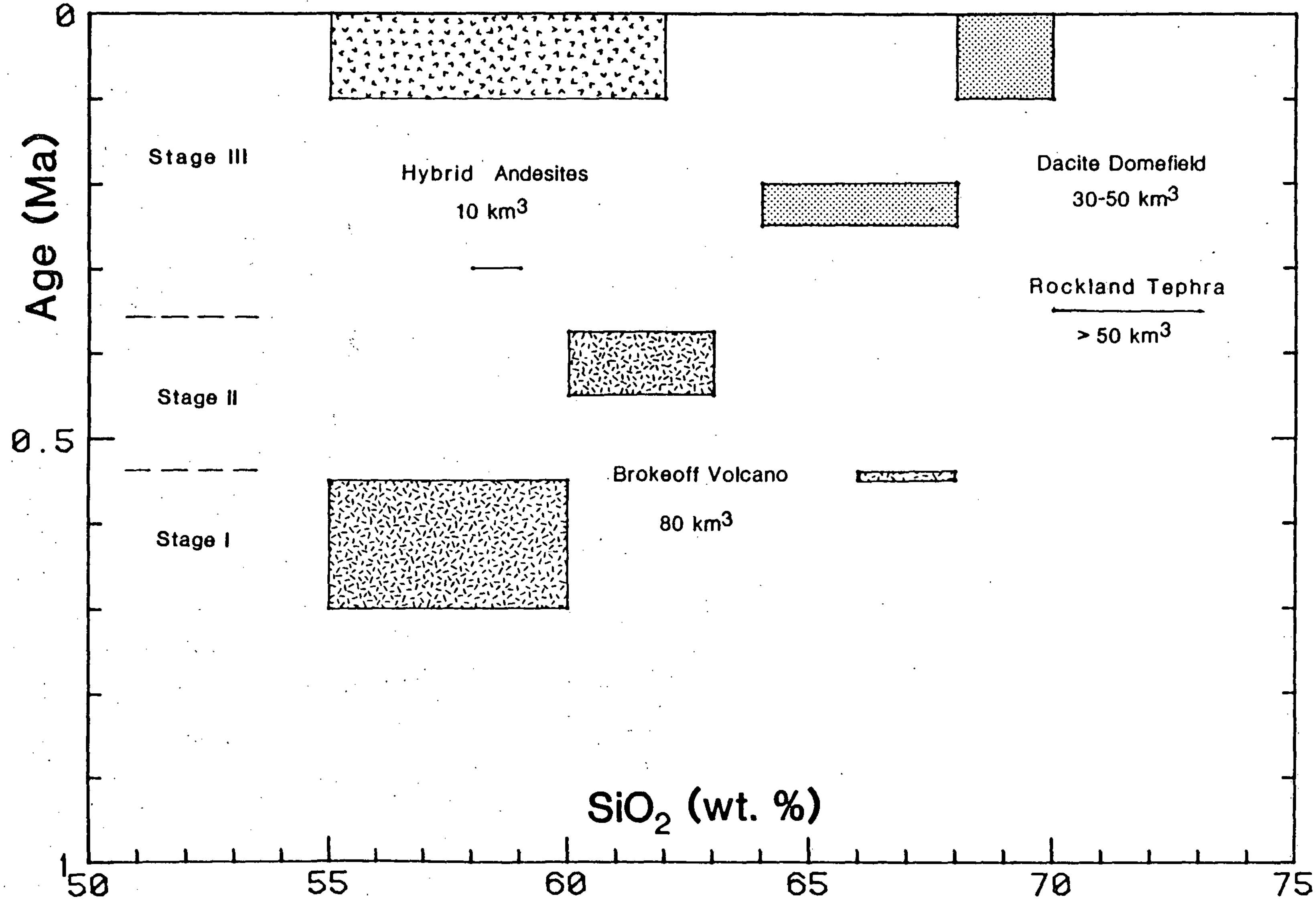
Porphyritic andesite and dacite with high Al_2O_3, low TiO_2, and medium K_2O contents and FeO/MgO ratios of 1.5-2.0 are the most abundant rock types in the LVC. Early mafic andesite, late rhyodacite, and hybrid andesite are subordinate in abundance. Rocks of the LVC resemble other calc-alkaline volcanic rocks emplaced on a continental margin overlying sialic crust. Harker variation diagrams of the major element chemistry of rocks from the LVC show smooth trends from 55-73 per cent SiO_2. The overall evolution is from mafic to silicic. However the evolution is not strictly sequential (see fig). Sparse isotope data suggest that crustal interaction is not significant in the development of the silicic rocks. However, partial melting of young mafic crust may play a role in the origin of the silicic magmas at LVC. Smooth geochemical trends, increasing silica content with time, and isotope data all suggest the origin of LVC magmas by crystal fractionation of a mafic parent. The long time span of activity at LVC and the presence of a hydrothermal system imply the presence of an evolving magma chamber. However, the current conceptual model of the magmatic system of the LVC as outlined below is poorly constrained and somewhat speculative.

The Lassen volcanic center differs from other Cascade volcanoes by having a larger volume of silicic rocks. The stress regime in the southern Cascades is moderate east-west extension, which favors retention of mafic magma in the crust and development of zoned systems with large volumes of derivative silicic magma. Systematic geochemical variation of the rocks of the Lassen volcanic center and the mineralogic and geochemical homogeneity of the youngest rocks suggest derivation from a single source. The magma system of the Lassen volcanic center can be envisaged as an evolving pluton-sized body of magma in the middle crust (10-20 km depth). The pattern of young vents suggests that an 5-8 km diameter magma body underlies the northwestern corner of LVNP. The young volcanic vents and hydrothermal system may indicate the position of a cupola on the magma chamber, which may be as shallow as 6 km beneath Lassen Peak. The hybrid andesites are produced by mixing of mafic magma intruded into silicic magma and may indicate a sill-like projection of the magma chamber beneath the central plateau of LVNP. Alternatively the hybrid andesites may be erupted by tapping the main chamber at depth.

The upper portion of the chamber contains crystal-rich rhyodacite, and has varied little in composition over at least the last 50,000 years. The system is probably zoned to more mafic compositions at depth. Regional basalt provides heat and material input to maintain the system in the partially molten state. There is abundant evidence for the interaction of mafic and silicic magma. A small volatile-rich cap has periodically developed at the top of the magma chamber but is removed by venting before increasing to substantial volume. Assuming that the vertical extent of the magma chamber is 10 km results in volume estimates of 250-500 km^3.

Despite the volcanologic and petrologic evidence of a magma system beneath the LVC, geophysical evidence for the presence of a magma chamber is lacking. A 25 km oval, -50 mgal gravity anomaly is centered on the dacite domefield and central plateau (hybrid andesites) of LVC. The gravity low is probably an expression of light volcanic rocks in the near surface and the presence of Quaternary plutonic rocks below the volcanic field.

Teleseismic and seismic refraction studies have failed to show the presence of a magma chamber beneath LVC. However the resolution of these studies is such that magma chambers smaller than 5 km in diameter could not be seen. A compilation of regional seismicity reveals an area lacking crustal earthquakes that corresponds to the gravity low. Future geophysical studies will have to be of higher resolution to detect the magmatic system of LVC.



Evolution of the Lassen Volcanic Center

The Mount Shasta Magmatic system

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The still-active compound stratocone of Mount Shasta, 14,162 feet high and more than half a million years old, has grown mainly during several distinct episodes of cone building from single central vents, each of which was active only briefly--possibly only a few hundred years (Christiansen and Miller, 1976; Christiansen and others, 1977; Christiansen, 1982). Four such cone-building episodes younger than about 250,000 years have been recognized, two of them Holocene; major cone building was separated by longer, predominantly erosional periods during which smaller volumes of lava were added to the cones. In addition, an older but similar edifice at the site of Mount Shasta was largely destroyed by a huge volcanic sector avalanche about 300,000 years ago (Crandell and others, 1984). The oldest exposed rocks of Mount Shasta, on its west flank and having a K/Ar age of about 590,000 years (G. B. Dalrymple, written commun. 1979), probably are a remnant of that earlier edifice.

The bulk of the stratocone consists of silicic andesites to mafic dacites, probably averaging just over 60% SiO_2. More or less uniform silicic 2-pyroxene andesites constitute most of each cone of the compound volcano, probably aggregating about 80-85% of the whole. More silicic rocks, including hornblende-bearing andesites, dacites (mainly 63-66% SiO_2), and rare rhyolites, occur principally as domes or relatively small lava flows, erupted both within the earlier central craters and on the flanks of the edifice. Significant volumes of dacitic pumice occur largely as fallout to the northeast. Numerous pyroclastic flows were emplaced during early Holocene time, especially on the west flank of the volcano (Miller, 1978); older pyroclastic flows occur on all sectors. In the past few thousand years the system has erupted every few hundred years (Miller, 1980).

Mount Shasta also has erupted relatively small volumes of more femic magma, generally as calc-alkalic basalts to mafic andesites, from cinder-cone or small-shield vents on its flanks. The highest such vent was noted by Anderson (1974) on the south side at about 9,000 feet elevation. No magmas similar to regional high-alumina olivine tholeiites (Hart and others, 1984) are known to have vented from Mount Shasta. Preliminary petrologic studies suggest that much of the range of compositions constituting the bulk of the volcano may have evolved by fractional crystallization and mixing of variously evolved magmas from a calc-alkalic basaltic parent like some of those erupted from Mount Shasta's flank vents (Baker and others, 1984).

Both the andesite-dacite stratocone of Mount Shasta and the more or less bimodal shield volcano of Medicine Lake, probably at least a million years old (Donnelley-Nolan, this volume), are voluminous long-lived systems near the margins of a large negative gravity anomaly. This gravity low extends across the High Cascade axis, Mount Shasta lying distinctly on the west and Medicine Lake on the east side of the nearly north-south axis at the ends of a linear system of young vents that trends east-northeast, suggesting a locally anomalous stress-field control. Between the Klamath and McCloud Rivers, the Pliocene to Quaternary volcanoes of the main Cascade axis are predominantly rather simple mafic-andesite shields. Within this sector of the chain, substantial volumes of silicic andesites to dacites have been produced only near the intersection of this main axis with the Shasta-Medicine Lake trend.

Taken together, the foregoing relationships suggest a tentative view of the Mount Shasta magmatic system. The areally extensive gravity anomaly, local stress-field control of adjacent volcanic systems, and the systematic relation of two major long-lived and highly evolved systems to the main Cascade axis suggest that the Shasta-Medicine Lake zone represents an anomaly in this sector of the Cascade subduction zone, perhaps related to excess production of femic magma in the mantle. Growth of the Mount Shasta system in this excess-production region during a few rapid increments may represent distinct upper-crustal emplacements of magma that had evolved at deeper levels to silicic pyroxene andesite from parental compositions like the common mafic calc-alkalic lavas of the region. During the intervening much longer times between these major emplacement events continued mantle magmatism probably drives relatively slow evolution of the Shasta magmas in upper-crustal chambers by open-system differentiation processes, including fractional crystallization accompanied by wallrock assimilation and mixing of variously evolved magmatic batches.

It is during these extended periods of upper-crustal magmatic evolution, during which dacitic and even rhyolitic magmas are produced, that any extensive hydrothermal convection might develop in the vicinity of the magmatic system. Based on both the slow eruptive rate and the lack of major cone building during the past several thousand years and on the abundance of dacites among its Holocene volcanic products, Mount Shasta may now be in such a magmatic state. No evidence is known for the existence of a cryptic hydrothermal system beneath Mount Shasta, but the high downward flux of cold water through the volcanic edifice from its heavy winter snowpack and extensive summer melting could mask hydrothermal convection at shallow crustal depths.

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GEOTHERMAL POTENTIAL OF MEDICINE LAKE VOLCANO

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Medicine Lake Volcano (MLV) lies east of the main Cascades trend about 50 km ENE of Mt. Shasta. The volume of MLV is estimated to be 600 km^3, larger even than Mt. Shasta which is the largest of the Cascade stratovolcanoes. Several lines of reasoning suggest that MLV has geothermal potential, including the suggestion by Eichelberger (1981) that a large silicic magma chamber underlies the volcano. Several late Holocene silicic units are present across the top of the volcano. Two of these, the rhyolite of Glass Mountain on the east side of the central caldera and the rhyolite of Little Glass Mountain and Crater Glass flows 16 km to the west, have nearly identical major- and trace-element chemistries. The close similarity of composition suggests that a single large silicic magma body is the source for these two units. Heiken (1978) explained the Glass Mountain-Little Glass Mountain chemical similarity by proposing tapping of a central small magma body by ring dikes. Other silicic lavas have erupted throughout the history of the volcano although the total volume of such eruptions appears to be relatively small. The earliest such lava is a rhyodacite ash flow about 1.25 million years old. Growth of the volcano apparently followed eruption of large volumes of high-alumina basalt that form the principal rock type of the Modoc Plateau. No obvious pattern can be discerned in the chemical evolution of the volcano except that higher silica eruptions occur on the higher parts of the volcano while primitive high-alumina basalts have continued to erupt around the lower flanks. MLV consists of these parental tholeiitic basalts plus calc-alkaline derivatives. The average composition of MLV is probably andesitic, and andesitic lavas are abundant especially on the upper slopes of the volcano. Anderson (1941) suggested that the caldera probably formed by eruption of voluminous andesite flows. Andesitic lavas at MLV are typically crystal-poor and fountain-fed, both characteristics suggesting that they were very hot when they erupted. Most of the volcano's lavas are crystal-poor suggesting relatively high temperatures or short residence times before eruption. Rather than a single large magma body, this may suggest small, short-lived bodies of magma that erupt relatively soon after they form.

Geophysical evidence provides no support for a large silicic magma body although a large mafic intrusion can be inferred from seismic refraction and gravity studies. Electrical and teleseismic studies are not definitive about the presence of magma at depth; small bodies less than 2 km across could be present. Intrusion of primitive basalt appears to be intruded into shallow crustal levels at the volcano, followed by fractional crystallization and assimilation, resulting in small, relatively silicic magma bodies. Small bodies of rhyolitic magma were probably produced by fractional crystallization of andesitic magma (Grove and Donnelly-Nolan, 1983). Given similar bodies of andesitic magma, derivative rhyolites of nearly identical composition might be produced by the same set of processes. Thus, nearly identical rhyolitic products such as those seen at Glass Mountain and Little Glass Mountain might be produced. Both lavas contain a variety of inclusions ranging from granitic (probable country rock) to andesitic (liquid when incorporated) to cumulate hornblende gabbros. The variety and abundance of these inclusions argues against the presence of a single large silicic magma chamber. The granites would have been quickly digested and the cumulate hornblende gabbros should represent material off the bottom of the chamber. Also, none of these young silicic lavas are truly high-silica rhyolites. None contains more than 74% SiO_2, compared to the older rhyolites at MLV that contain 75-77% SiO_2 and are inclusion-free. The younger rhyolites may have erupted after disruption by intruded mafic magma before the silicic magma body had completed its evolution.

Structurally, MLV lies in an E-W extensional environment. It is located at the intersection of a major lineament trending ENE across the Cascades and the westernmost N-S trending normal faults of basin and range type. One of these faults forms Gillem Bluff at the northern edge of the volcano. Farther south, this fault is buried by younger lavas of MLV, but alinements of vents as well as the relatively abrupt eastern topographic margin of MLV strongly suggest that the fault bends SSE and continues S at least 35 km through the vents for the Glass Mountain flow and domes. A strong NE structural trend is manifested high on the NW side of the volcano by open ground cracks, the Little Glass Mountain-Crater Glass flows and domes, and late Holocene basaltic eruptions. A similar direction is evident at a set of Holocene andesitic pit craters on the SE side of the caldera. This evidence together with the E-W extension and dominant N-S alinement of vents over 50 km on MLV, the ring faults of the volcano's caldera, and other alinements of vents and open ground cracks, provide evidence of fracture permeability at depth.

Is there enough fluid for power production? Although springs are rare on the volcano, very large volumes of cold water emerge at the southern edge of MLV lavas to form the Fall River, suggesting that adequate sources of water are available. Whether water is stored and heated directly under MLV is unknown. Hot springs are absent, but this does not preclude the existence of hot water at depth.

Whether or not a large silicic magma body exists under MLV, the volcano still has the potential to be a major geothermal resource. Sufficient heat could be supplied by continued intrusion of mafic magma into shallow crustal levels. Numerous Holocene vents and intermittent rhyolitic volcanism for at least 1.25 m.y. suggest that MLV is a prime target for geothermal energy exploration. Several companies are actively engaged in exploration. One deep hole was drilled during the summer of 1984 and more are planned. Results of the drilling by private companies are proprietary but the amount of interest shown suggests that the companies believe there is geothermal potential at MLV.

Outflow of hot water from MLV might also be possible, particularly in the dominant N-S structural direction. Only 30-50 km N of MLV, geothermal fluids in and near Klamath Falls, Oregon, are used for low-temperature applications such as space heating. Klamath Falls lies equidistant between MLV and Crater Lake Volcano, at the lowest point in the Klamath graben that lies between these two major young volcanoes. The Klamath Falls geothermal system, unrelated to any recent volcanism at the surface, may be at least partly supplied by hot water from MLV.

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Mount Mazama and Crater Lake Caldera

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Geologic mapping and K-Ar dating (Bacon and Lanphere, 1983) provide the data to define the eruptive history of Mount Mazama (Bacon, 1983). Events that led up to the catastrophic, caldera-forming eruption have been reconstructed through detailed field studies of late products (Bacon, 1983, 1985), including paleomagnetic determinations (Champion, 1983). Chemical analyses of rocks from all periods in the growth of Mount Mazama and microprobe analyses of phenocrysts in late products (C. R. Bacon and T. H. Druitt, unpub. data, 1985) help to suggest models for the magmatic system beneath the volcano.

Mount Mazama was composed of overlapping short-lived shields and stratocones and thick lava flows, constructed west and north of a voluminous field of rhyodacite domes and flows. The rhyodacite evidently was extruded in a few pulses, probably of short duration, at various times between ~750 and ~500 ka (thousand years B.P.). It seems likely that basalt lies beneath the rhyodacite, the thickness of which is unknown.

The oldest exposed stratovolcanoes of the Mount Mazama cluster are Phantom Cone and Mount Scott, ~420-395 ka. These are overlapped by andesitic and dacitic lavas erupted ~355-230 ka. Mafic andesitic shields were built near Cloudcap ~290-220 ka and near Llao Rock ~185-110 ka. Andesitic to dacitic lavas were erupted near Llao Rock ~130-110 ka, followed by rapid growth of a complex andesitic stratocone near Hillman Peak ~70 ka. Shortly after the Hillman Peak cone was constructed, dacite pyroclastic flows descended the southwest flank of Mount Mazama and dacite airfall pumice was erupted from vents near Llao Rock and Cloudcap. Provisional Fe-Ti oxide equilibration temperatures for the three sets of dacitic pyroclastic rocks are ~930, 990, and 960°C, respectively. Dacite lava flows, one of which is dated at ~70 ka, were associated with the pyroclasitic activity. Although there is no evidence preserved for caldera formation, several km^3 of dacite magma must have been erupted. Following these eruptions, the existing evidence suggests that Mount Mazama reverted to producing only andesite for ~20 ka.

About 50 ka the dacitic Watchman flow was erupted from a vent now exposed in the west caldera wall, adjacent to the south side of Hillman Peak. Like many earlier dacite and silicic andesite flows of Mount Mazama, the Watchman flow contains inclusions of andesite that were largely molten when incorporated into the host lava. Their presence indicates that the underlying magmatic system was compositionally zoned, containing magma at least as mafic as andesite that was incompletely mixed into the more silicic magma before eruption. Similar inclusions are present in dacitic blocks found in monolithologic avalanche breccias or lithic-pyroclastic-flow deposits on the south and southwest flanks of Mount Mazama. These deposits were formed by collapse of lava domes near the summit of the volcano ~40 ka. Stratigraphic evidence and K-Ar dating indicate that andesite lava flows on the north and west caldera rims were erupted between ~50 and 40 ka. Known younger precaldera lavas were almost exclusively silicic in composition.

A small hornblende rhyodacite dome on the north caldera rim was emplaced ~30 ka. Similar lava domes of the Sharp Peak group northeast of the caldera may be of similar age. Williams Crater, formerly "Forgotten Crater", west of Hillman Peak may be also approximately the same age. Williams Crater and a vent about 1 km to the west erupted basalt and caused the emplacement of an extensive lava flow and several domes of commingled andesite and dacite. Thus, Williams Crater probably marks the west margin of the silicic magma reservoir at this time. The Redcloud Cliff flow, a thick hornblende rhyodacite flow on the east caldera wall, evidently is latest Pleistocene in age, and may be as old as the ~30 ka dome on the north rim. There is no evidence preserved for any volcanic activity after eruption of the above units and before Holocene volcanism began.

About 7,000 ka the shallow chamber began to leak significant volumes of rhyodacitic magma in the form of three lava flows: Llao Rock, Grouse Hill, and Cleetwood. Other rhyodacitic flows may have been present within the area that collapsed to form the caldera. Paleomagnetic evidence suggests that the Llao Rock and Grouse Hill flows were erupted in short succession. About 140 yr. later, the Cleetwood flow was erupted. The bulk composition (on a volatile-free basis) and Fe-Ti oxide phenocryst equilibration temperature (~940°C) of this lava are virtually identical with those of the pumice of the climactic eruption, except that the lava had a much lower water content upon eruption. Surprisingly, the climactic eruption began before the Cleetwood flow had cooled. Although most, if not all, of the precursory lava eruptions began with an explosive phase, flow of magma to the surface apparently was sufficiently sluggish to allow the magma to de-gas during ascent so that conduits eventually were plugged with viscous rhyodacite. Only the Llao flow was preceded by a voluminous pyroclastic phase, but this spread tephra over parts of several western states. Loss of several km^3 of magma from the top of the chamber over a comparatively short period to form the lava flows and related pyroclastic rocks may have depressurized the system, without accompanying loss of volatile pressure, so that vapor saturation occurred at greater depth than before and the climactic eruption could be sustained.

The climactic eruption took place in two steps: (1) a single-vent phase and (2) a ring-vent phase. The first phase began with eruption of the widespread airfall deposit from a high Plinian column, which later collapsed to produce pyroclastic flows as the vent became enlarged. The pyroclastic flows deposited the Wineglass Welded Tuff in valleys on the north and east of the caldera, its distribution being limited by the summit of Mount Mazama which lay to the south of the vent. Phase (1) ended when the caldera began to collapse, initiating phase (2) which produced much more mobile pyroclastic flows from higher eruption columns. These flows deposited nonwelded to partly welded tuff in the valleys all around Mount Mazama and left a lag deposit of lithic breccia near the rim and on high ground, including even the summit of Mount Scott. These deposits show spectacular compositional zonation of homogeneous rhyodacite (70-71% SiO_2) below andesite that decreases fairly regularly in silica content with stratigraphic height (~62-51% SiO_2). Equilibration temperatures for Fe-Ti oxide phenocrysts in the silicic part of the eruption group tightly near ~950°C and, like other "climactic" rocks, are more oxidized than the ~70 ka dacites. Over 50 km^3 of magma were erupted in the climactic eruption, most being rhyodacite. Although the exact depth is unknown, the existence of the caldera suggests that the top of the magma chamber must have been only a few km below the surface.

A provisional model of the magmatic system beneath early Holocene Mount Mazama would have relatively differentiated (up to ~72% SiO_2) and cooler ($\geq$870°C) porphyritic magma overlying well-mixed, convecting rhyodacite magma, which was in turn underlain by zoned, crystal-rich andesite. Presumably, basalt underlay the entire reservoir, because basaltic eruptions from monogenetic cones and small shields took place beyond and on the lower flanks of Mount Mazama thoughout its history. Such a long-lived crustal magmatic system must have had hydrothermal phenomena associated with it. The uppermost deposits on the caldera rim were formed by phreatic explosions that took place when this hydorthermal system was disrupted by caldera collapse. Postcaldera volcanism, consisting of andesitic activity outboard of rhyodacitic eruptions, suggests that the remains of the old magmatic system--or a new one--are still viable. High heat flow in the floor of Crater Lake, elevated chloride concentration in the lake water, and the well mixed nature of the lake suggest that hydrothermal convection continues beneath the lake floor (Williams and Von Herzen, 1983). Conceivably, lateral flow of thermal waters might produce geothermal reservoirs beyond the boundaries of Crater Lake National Park where they might be utilized for extraction of geothermal energy.

Beyond the potential for geothermal resources, Crater Lake caldera provides a natural laboratory in which to study not only the evolution of a crustal magma body, but also hydrothermal alteration within an andesitic stratovolcano. Alteration of the older rocks of the caldera walls is ubiquitous, and the lithic breccias of the climactic eruption contain partly-fused plutonic rocks and highly altered fragments from deep within Mount Mazama. These may offer clues to the precaldera hydrothermal history of this long-lived volcanic system.

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The Magmatic System of Newberry Volcano, Oregon

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Newberry Volcano, 60 km east of the crest of the Cascade Range in central Oregon, has been active for more than 500,000 yr and erupted most recently about 1350 yr ago. The surfaces of its flanks are formed mostly of basaltic-andesite and basalt flows, which issued from hundreds of cinder cones and fissure vents. Widespread ash-flow tuffs exposed on some flanks probably underlie the basaltic flows; dacite to rhyolite domes and flows are common on the flanks. The summit caldera is filled by rhyolitic ash-flow and air-fall tuff, by rhyodacite and rhyolite domes and flows, and by a few mafic flows and vent deposits. Drill cores show that fragmental debris fills the caldera to about 500 m below the surface; it overlies rhyodacite to basalt flows that extend to at least 932 m. The caldera resulted from several periods of collapse during voluminous eruptions that formed the ash-flow tuffs on the flanks.

The most recent eruption cycle was 10,000 to 1350 ^{14}C yr ago. Widespread basaltic-andesite flows on the flanks, estimated to be about 10,000 to 8000 yr old, are approximately coeval with several rhyolitic flows and domes in the caldera and on the upper southeast flank. Rhyolitic eruptions in the caldera produced domes, flows, and tephra deposits between 6845 ^{14}C yr (approximate age of Mazama ash, a regional stratigraphic marker) and 6100 ^{14}C yr ago. Numerous basaltic-andesite flows issued from vents on the northwest and southeast flank about 6100 ^{14}C yr ago; eruptions since then have been rhyolitic. Eruptions 4000-3000 yr ago produced obsidian flows and pumice deposits, and 1350 yr ago air-fall tuff, ash-flow tuff, and an obsidian flow were erupted. Although 10,000-8000 yr and 6100 yr basaltic flows are common on the flanks, they are nearly absent in the caldera; those that are present were erupted from fissures on the caldera walls. Mixed basalt-rhyodacite lavas that erupted from near the top of the east wall suggest that basaltic magma may have intercepted silicic magma beneath the caldera.

Analysed rhyolitic rocks younger than about 10,000 yr are chemically similar. They are readily distinguished from abundant older silicic rocks at Newberry by their relatively evolved major- and trace-element composition, especially their higher Rb/Sr ratio (2, compared to 1 in older rhyolitic rocks). Also, even though the range in composition is quite small, successively younger rocks show a progressively more evolved composition. The young rhyolites are all very low in phenocryst content and successively younger rocks have fewer phenocrysts; the youngest obsidian flow contains less than 0.01 percent phenocrysts. This suggests that the rhyolites were erupted from a magma chamber whose upper part was near the liquidus and that the tapped part of the magma did not crystallize during a much longer period than that since the most recent eruption. We attribute the decrease in phenocryst content to increasing water content in the magma or to influxes of heat owing to underplating of a silicic magma chamber by hot basaltic magma 10,000-8000 and 6100 yr ago. The basaltic flows erupted during the last 10,000 yr are presumed to be from a source much deeper than the inferred silicic magma body.

The silicic body beneath Newberry (whether still magma or a solidified pluton) appears to be a single body, because the chemical compositions and petrography of the young rhyolites are so similar. The restriction of the vents for the young rhyolites to the eastern part of the caldera and the absence of mafic vents there (a "shadow effect") suggest that the magma body was centered under the eastern half of the caldera. A 932-m-deep hole drilled by the USGS in the caldera encountered temperatures as high as 265°C and high heat flow, such as would exist above a magma body or a hot but solidified intrusion. Hydrothermal convection beneath the drill hole, however, could extend an unknown distance downward to a deep hot body. If magma is still present, the body apparently is small, deep, or complex in shape inasmuch as it has not been definitely detected by geophysical methods.

THREE SISTERS TO MOUNT JEFFERSON: MAGMATIC SYSTEMS AS HEAT SOURCES

U.S.G.S. Workshop on Geothermal Resources of the Cascade Range

E.M. Taylor, Oregon State University, May 22, 1985

A summary of central High Cascade volcanism, magmatism, and tectonism between Three Sisters and Mount Jefferson, as related to sources of geothermal heat, should include their pre-Pleistocene antecedents because modern heat flow is probably influenced by structural features produced prior to development of the modern High Cascades and by the long-continued evolution of a few isolated silicic magma systems in a mafic volcanic province. The most complete records of early High Cascade volcanism are found in the Deschutes Formation on the east flank of the range and in correlative deposits exposed in the Western Cascades. An episode of crustal tension and production of associated mafic lavas was initiated over a broad part of the eastern Cascades, 8-10 m.a. A combined tholeiitic and calc-alkaline volcanic arc was constructed in the position of the modern High Cascade axis. Shields and composite cones of basalt and basaltic andesite predominated. However, many of the volcanoes produced silicic lavas; in the distal deposits there is a record of hundreds of discrete eruptions of air-fall pumice and pyroclastic flows ranging in composition from mafic andesites to rhyolite. Tensional stresses in the crust apparently increased to a maximal condition by 4-5 m.y., at which time the early High Cascade volcanic arc subsided along discontinuous north-south normal faults, leading by intermittent stages to a broad, graben-like depression of the volcanic axis. Explosive volcanism was enhanced at this time and was 4-5 m.y. characterized by the appearance of andesitic ash-flow tuffs and abundant, inhomogeneous, mixed magmas of diverse compositions. Much of the volcaniclastic output was probably trapped within segments of the subsiding graben.

After 3 m.a. the intensity of volcanism decreased and the magmas brought to the surface were almost entirely basalt and basaltic andesite. Consequently, the modern central High Cascade Range is a broad platform of mafic volcanoes ranging from large composite cones and shields such as North Sister, Mount Washington, and Three Fingered Jack, to small cones and cone groups with associated lava flows such as Belknap Crater, Sand Mountain, and South Cinder Peak. Late High Cascade silicic volcanism has been restricted to the vicinity of Three Sisters and Mount Jefferson. The Three Sisters volcanic group has been intermittently active for several 10^5 years and has produced basalt, basaltic andesite, andesite, dacite, and rhyodacite in the form of lavas, domes, and volcaniclastic deposits representing three clearly distinct magmatic series. All of these compositional types have been erupted during Holocene time; the most recent example is the 1900-year-old rhyodacite of Rock Mesa at the southwest base of South Sister. The development of Mount Jefferson is less well known; it is a composite cone of basaltic andesite surmounted by younger andesite lavas and dacite domes. The most recent eruptions produced dacitic air-fall pumice and pyroclastic-flow deposits that appear to have issued from vents on the northeast flank, while late Pleistocene glaciers were retreating.

The central High Cascade province has been a locus of intrusion by mafic dikes, sills, and plugs during the last 9 million years; this has probably contributed to a regional background of geothermal heat. Perturbations in this background are to be anticipated where hydrothermal circulation is influenced by buried graben faults and thick accumulations of volcaniclastic material. Of particular significance in this context is the prolonged localization of silicic volcanism in the vicinity of Three Sisters and Broken Top. Ash-flow tuffs were produced from source volcanoes near the sites of modern Three Sisters 4-5 m.a. Later, an extensive highland of silicic domes was constructed and several late High Cascade ash-flow sheets spread as far east as Bend. The modern Three Sisters group is the latest manifestation of a long-lived source that has been contributing silicic magmas to the shallow crust. It would be surprising indeed if young, hot, plutonic bodies were not present beneath this area.

MOUNT ADAMS: Eruptive history of an andesite-dacite stratovolcano at the focus of a fundamentally basaltic volcanic field.

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Mount Adams (3742 m) is a late Quaternary stratovolcano 50 km due E of Mount St. Helens, 75 km SSE of Mount Rainier, and 90 km NNE of Mount Hood. At 350 km^3, its volume is exceeded in the High Cascades only by Mount Shasta, but the Simcoe, Newberry, and Medicine Lake back-arc systems are all larger. Mount Adams stands upon the Cascade crest and is drained radially by many streams fed by glaciers radiating from its summit icecap. Only 2.5% of its 650-km^2 area is glacier-mantled today, but as much as 90% was ice-covered in latest Pleistocene time.

There have been no recorded eruptions of Mount Adams and probably just one in the past 3500 years, but at least 7 Holocene eruptions and the persistence of sulfurous summit fumaroles indicate that the volcano remains potentially active. The 7 Holocene eruptions on the stratocone took place at flank vents 2000-2500 m in elevation and produced a wide range of compositions, 49-61% SiO_2.

Nearly all of the high cone above 2300 m in elevation was constructed during latest Pleistocene time, probably between 20 and 10 Ka, explaining the abundance of late-glacial till and the scarcity of older till. Products of this eruptive episode range from 54 to 62% SiO_2 on the main cone. Contemporaneous and younger <u>peripheral</u> vents yielded lavas and scoriae in the range 48-57% SiO_2, and, along with Mount Adams, they define a recently active N-S eruptive alignment 40 km long and only 5 km wide. Basalts within this zone are extremely varied compositionally, containing, e.g., 0.16 to 1.6%

K_2O at 48-48.5% SiO_2.

Beneath the young edifice, deeply eroded elements of a long-lived compound stratovolcano are as old as 500 Ka, representing several eruptive episodes and an unbroken spectrum of rock compositions from 52 to 69% SiO_2. Peripheral to the stratovolcano complex, coalescing shields of basalt (49-53% SiO_2) underlie and interleave with lavas from the andesite-dacite focus of the overall magmatic system.

Persistent eruption of peripheral mafic shields and cinder cones throughout the lifetime of the stratovolcano suggests their fundamental role in transport of heat and magma into the crust. The term "parasitic" should be abandoned for such vents, as it promotes the view that mafic flank eruptions are leaks from a central chamber and implies dependence on the main cone. For large andesite-dacite systems, nearly the opposite is true: The stratocones are the more derivative features, built up of magmas that evolve by concurrent fractionation, crustal melting, assimilation, and recurrent internal mixing within the foci of the larger domains injected by the various mantle-derived basalts that also yield variable arrays of peripheral eruptions.

Dacitic lavas and a few block-and-ash flows (63-69% SiO_2) erupted several times from the focal area, but probably none are younger than 100 Ka. A single rhyolite (72% SiO_2), as yet undated, is also old. The antiquity of all known silicic units, in conjunction with the andesitic compositions of the late Pleistocene summit cone and the Holocene lavas that erupted high around that cone, make it appear unlikely that Mount Adams is now underlain by a magma reservoir large enough to support either a major hydrothermal convection system or fractionation of voluminous silicic differentiates.

The high precipitation on the stratocone, estimated to be 3.5m ~3500 mm/a, makes Mount Adams an important site of ground water recharge. The high permeability of its rubbly lava flows suggests that much of this ground water moves rapidly downward and outward from the stratocone, failing to remain in the warmer central region of the volcano long enough to develop a hydrothermal convection pattern but, instead, dispersing and dissipating whatever heat is supplied from depth to the fumarolically altered core. The weak and diffuse fumarolic emissions on the summit are the only manifestations of possible hydrothermal activity anywhere on the stratocone. The warmest spring that we have found on the (Mt Adams) stratovolcano measured 3°C, on a summer day. Nonetheless, in surrounding areas, springs as warm as 24°C and gradients of 50°C/km in heat-flow holes offer some hope that ground water sufficiently warm for local space heating or agricultural uses might be found by exploring the lowland periphery of the volcano.

Cascades Workshop, Menlo Park, May 22-23, 1985

Characteristics of Cascades magmatic systems determined from seismic studies

H. M. Iyer

Introduction

Seismic techniques (refraction, reflection, seismicity mapping, and velocity and attenuation measurements using seismic waves from local earthquakes, regional earthquakes, and teleseisms, have been used to delineate magma chambers in several silicic volcanic centers in western U.S.A. (Iyer, 1984). However so far, such techniques have only been marginally successful in detecting magma chambers in the Cascade volcanoes. Seismic refraction and seismicity data have provided valuable information on the volcano-tectonic setting of the Cascades (California: Berge and Stauber, 1985; Zucca et al., 1985; Oregon: Leaver et al., 1982; Washington: Rohay, 1979; Taber and Smith, 1985). The most exciting result from teleseismic residuals is the detection of the subducting Juan de Fuca plate beneath Washington and Oregon (Michaelson and Weaver, 1985; Michaelson, personal communication). Teleseismic-residual data at the andesitic stratovolcanoes of the Cascades do not reveal detectable magma chambers. Examples are Mt. Hood, Mt. Shasta, and Mt. Lassen. On the other hand the Newberry and Medicine Lake volcanoes show evidence for massive high-velocity intrusions into the crust.

In this talk, I discuss P-wave residual modelling of the subducting Juan de Fuca plate beneath the Cascades, and results from teleseismic-residual experiments in Cascade stratovolcanoes and shield volcanoes. I also present preliminary results from a high-resolution experiment at the Newberry volcano to detect small (diameter 1 km or more) magmatic intrusions.

Subduction Zone

Unlike in other major subduction zones of the world, due to the lack of a Benioff zone of deep earthquakes, it has been difficult to delineate the subducting lithosphere beneath the continental margin in western north America. At least in Washington and northern California, intermediate depth earthquakes seem to define a zone that looks like the shallow segment of the subducting plate. In Oregon, on the other hand, subduction seems to be taking place completely aseismically.

Michaelson and Weaver (1985) used teleseismic P-wave arrivals from earthquakes recorded by a dense network of seismic stations in Washington and northern Oregon, and using inverse and forward modelling techniques, delineated an eastward dipping high-velocity anomaly in the upper mantle, inferred to be the subducting plate. It is about 50 km thick, has P-velocity higher than the surrounding region by 3-7%, and extends from 45 to 300 km depth. Its dip angle is of the order of 60-70 Degrees. Michaelson and Weaver also found evidence for segmentation of the plate into three different configurations between northern Washington and northern Oregon.

Evidence for the slab beneath Oregon comes from the inversion of P-residual data from a USGS seismic network operated for two years in the Oregon Cascades and from a linear, northwesterly, profile of stations in southern part of the state. Both data sets reveal the presence of an east-dipping, tabular, high-velocity anomaly in the upper mantle. The inversion and forward modelling of residual data from the profile (Caryl Michaelson, personal communication), show an eastward dipping 50-km thick slab in the depth range of 40-200 km. The dip angle is about 50 degrees and the velocity contrast in the slab from the surrounding rock is about 10% at the top and about 4% at depth.

Stratovolcano

We have looked at the three-dimensional crustal structure of three stratovolcanoes in the Cascades, Mt. Hood in Oregon (Weaver et al., 1982), and Mt. Shasta and Mt. Lassen in California, using teleseismic residuals collected by dense seismic networks. In all these cases the results show only regional heterogeneities in crustal structure with no clear evidence for the presence of magma chambers of horizontal or vertical dimensions greater than about 5 km, the resolution limit of the teleseismic technique.

Shield Volcano

Seismic and other geophysical data show the presence of high-velocity intrusions in the crust in the Newberry volcano, Oregon, and in the Medicine Lake volcano, California. Three-dimensional modeling of teleseismic P-wave residuals collected using a dense seismic array over Newberry volcano delineate a column of high-velocity material extending from within 10 km of the surface below the volcano summit to mid-crustal depths near 25 km (Stauber et al., 1985). The P-wave velocity in the column is about 6-14% higher than in the surrounding crustal rocks. Stauber et al., (1985) interpret this intrusive body to be an expression of a swarm of predominantly sub-solidus gabbroic dikes and sills which were intruded as the volcano was built. A similar high-velocity body has been tentatively identified beneath the Medicine Lake volcano by Evans (1982).

High-resolution experiment

Neither the teleseismic results nor results from other geophysical experiments, however, can explain the high temperatures encountered in two shallow drill-holes drilled in the Newberry caldera. To investigate if magma intrusions undetectable due to the limited resolution of the teleseimic method are responsible for these high temperatures, we carried out a high-resolution version of the teleseismic-residual experiment by recording 8 large shots using a dense array of 120 seismographs deployed over a 12 km diameter area centered on the caldera. Preliminary interpretation of the traveltime residuals reveals a ring of high velocity material coinciding with the inner ring fault system of the caldera in the upper 2 km and a zone of lower velocity extending deeper than 2 km in the center of the caldera (Stauber et al., 1985). IS THE LOW-VELOCITY ZONE THE MAGMA CHAMBER WE HAVE BEEN SEARCHING FOR IN THE CASCADES?

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GRAVITY AND MAGNETIC STUDIES IN THE CASCADE RANGE

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ABSTRACT

A compatible set of gravity data for the entire Cascade Range has been compiled. From this data set I prepared a series of interpretive color gravity maps including a free air anomaly map, Bouguer anomaly map at a principle (2.67 g/cm^3) and an alternate (2.43 g/cm^3) reduction density, and filtered and derivative versions of the Bouguer anomaly map. The set is accompanied by a color terrain map at the same scale.

The regional anomaly pattern and gradients outline the various geological provinces adjacent to the Cascade Range and deliniate major structural elements in the range. The more local anomalies and gradients may delineate low density basin and caldera fill, faults, and shallow plutons.

Introduction

The U.S. Geological Survey (USGS) has supported gravity studies as part of its Cascades Geothermal Program. Gravity data are particularly useful in areas where there are large lateral density contrasts such as in volcanic regions. These density contrasts cause variations in the earth's gravity field which can often be related to subsurface geologic features such as faults, intrusions, ore bodies, etc.

The purpose of this paper is to provide preliminary interpretations of some of the local gravity and magnetic features in the Cascades.

The Pacific Northwest, the location of the Cascade Range, has a complex tectonic history involving changes in plate tectonic setting from a passive Atlantic-type margin to an active subduction zone (Dickinson, 1976; Hamilton, 1978; and Hammond, 1979), changes in the location of the subduction zone and its asscociated magmatic arc, accretion of various terranes and micro-plate rotations. The Cascade Range is a volcanic arc extending from Lassen Peak, California to Mt. Garibaldi, British Columbia and is related to subduction of the Juan de Fuca and Gorda plates. The Cascades consist of a wide range of rock types and ages ranging in composition from basalt through rhyolite and in age from Miocene to recent (they do, however, overlie Mesozoic and Paleozoic rocks in Washington and Canada). The diversity of rock types results in numerous density contrasts that produce a wide variety of gravity anomaly patterns.

Local Gravity Anomalies

Local gravity anomalies in the Cascades are generally caused by sediment filled basins (gravity lows), an incorrect Bouguer reduction density (can cause either gravity highs or lows, but lows are more common when a Bouguer reduction density of 2.67 g/cm^3 is used), intrusions (gravity highs) and caldera fill (gravity lows). Since the Cascades south of Mt. Rainier are mantled by a layer of low density volcanics, gravity anomalies in this region usually deliniate rocks which differ from the density of these volcanic rocks. Although the proper Bouguer reduction density for much of the High Cascades is between 2.67 and 2.43 g/cm^3, most of the young, recently active volcanoes in the range have a bulk density of 2.2 g/cm^3. When this density is used to reduce the gravity data for most individual volcanoes, gravity highs, not lows, are commonly observed (Finn and Williams, 1982; Williams and Finn, in press a). These positive anomalies are due to shallow, dense intrusions that probably range in composition from intermediate to mafic. Many other local gravity highs have been identified with old, eroded volcanic centers. Examples are the Goat Rock volcano in the Goat Rocks Wilderness, Washington and the Still Creek and Laurel Hill plutons just southwest of Mt. Hood (Williams and Finn, 1983).

Aeromagnetic Data

All of the stratocone volcanoes in the Cascades have had aeromagnetic surveys flown over them. Studies of the data from Mt. Shasta (Blakely and Christiansen, 1978), Mt. Hood (Flanigan and Williams, 1982), Medicine Lake (Finn, 1982), Newberry (Griscom and Roberts, 1984) Goat Rocks Volcano (Williams and Finn, in press b), and Mt. St. Helens have been made. In most cases, i.e. at Medicine Lake, Newberry, and Mt. Shasta, the magnetic anomaly has been interpreted to be soley due to the terrain. At Mt. Hood, an old reversely-magnetized volcano upon which the present mountain was built was delineated with the magnetic data. Preliminary results from a combined gravity and magnetic study of Mt. St. Helens indicate that a magnetic and gravity high, unrelated to the terrain is associated with Mt. St. Helens. They may be related to a pluton under the volcano. More interesting results, in terms of delineation of geothermal resources, come from a study of gravity and magnetic data from Goat Rocks Volcano (Williams and Finn, in press (b)). Two and one-half-dimensional modeling of the positive gravity and magnetic anomalies delineated a shallow pluton. In most places the gravity and magnetic anomalies defined the pluton. However, some differences exist.

There are places where a magnetic low occurs where there is the gravity high associated with the pluton. These lows correspond to mapped areas of alteration, indicating that the zones of alteration is not just surficial, but must be meters thick. Other areas of mapped alteration do not have magnetic lows associated them, showing that the alteration is surficial. I hope to apply a similar approach to other volcanoes.

Geothermal Resources

The delineation of structural features like faults and local features such as intrusions is useful for geothermal exploration. Faults can be zones of increased permeability. Intrusions under active volcanoes are often asymmetric to the volcanic cone and could be reached at a shallower depth if drill holes are sited with the benefit of the gravity data.

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GEOCHEMICAL FEATURES OF CASCADES HYDROTHERMAL SYSTEMS

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Thermal waters of the Cascade Range include at least 4 distinct types: Na-Cl, Na-SO_4, CO_2-charged, and dilute Na-HCO_3. The most interesting, from a geothermal point of view, are the Na-Cl waters which occur in the Oregon Cascades and at Lassen Volcanic National Park. The large chloride content of these waters, up to 3,350 mg/L, must be derived from rock deposited in a marine environment, and in most of the Cascades this requires circulation to considerable depth. Most thermal springs in the central Oregon Cascades discharge Na-Cl waters which have a Ca-Cl component. In "Ca-Cl component" waters, such as Belknap Springs, some dissolved calcium must be used to electrically balance the chloride in the chemical analysis because insufficient sodium is present.

Belknap Springs

	t°C	pH	SiO_2	Ca	Mg	Na	K	HCO_3	SO_4	Cl
mg/L	86	7.64	91	210	.34	660	15	20	150	1,200
meq/L				10.5		28.7				33.8

Although Ca-Cl thermal waters do not occur in the Cascades, several Ca-Cl mineral springs occur in western Oregon and Washington. Pigeon Springs, a cold mineral spring in western Washington, is an example of a Ca-Cl water:

Pigeon Springs

	t°C	pH	SiO_2	Ca	Mg	Na	K	HCO_3	SO_4	Cl
mg/L	11	8.23	9.2	7,450	5.4	6,400	7.2	250	250	22,500
meq/L				372		278				635

Ca-Cl waters develop from Na-Cl waters by extensive albitization of plagioclase (Hardie, 1983). In the albitization of plagioclase, sodium and silica are consumed, and calcium is released:

$$\underset{\text{anorthite component}}{CaAl_2Si_2O_8} + 2\ Na^+ + 4\ H_4SiO_4 = 2\ \underset{\text{albite}}{NaAlSi_3O_8} + Ca^{++} + 8\ H_2O.$$

A secondary reaction, brought about by the increase in calcium concentration is precipitation of anhydrite:

$$Ca^{++} + SO_4^{=} = \underset{\text{anhydrite}}{CaSO_4}.$$

Provided that little or no water-rock reaction has taken place after the thermal water left the aquifer at depth, and that mixing of thermal and cold groundwater are not important, we can estimate the aquifer-temperature at depth by increasing the water temperature until the activity product constant for the reaction of interest equals the equilibrium constant. At Belknap Springs, equilibrium with respect to the albite-anorthite pair would be attained at 152°C. A similar calculation with respect to anhydrite, also produces an equilibrium temperature of 152°C.

Testing the "Ca-Cl" component waters of the Oregon Cascades for equilibrium with the mineral pair albite-anorthite and anhydrite, produces temperatures similar to those estimated from the SO_4-H_2O isotopic geothermometer (table 1). Quartz geothermometry does not work in this environment, probably because quartz is not present in the deep aquifer. Dissolved silica concentrations are limited by the albitization reaction not the dissolution of quartz. With the exception of Breitenbush, $t_{Na\text{-}K\text{-}Ca}$ indicates temperatures within 25°C of the measured spring temperatures.

Albitization of plagioclase should be a common process in volcanic terranes such as the Cascades. However, before trying these equations in other areas, the effects of dilution with a low TDS water should be discussed. As an example, Belknap Hot Spring water was "diluted" with 10%, 25% and 50% distilled water. The apparent temperatues of equilibrium of these waters with the mineral pair albite-anorthite and anhydrite were calculated.

Percent dilution of Belknap water:	0%	10%	25%	50%
$t_{anhydrite}$	152	158	166	185
$t_{albite-anorthite}$	152	148	140	125

Dilution increases the apparent anhydrite saturation temperature but decreases the apparent albite-anorthite equilibrium temperature.

At Lassen, Growler and Morgan hot springs have anhydrite saturation and albite-anorthite equilibrium temperatures which range from 197 to 226°C. SO_4-H_2O isotopic equilibrium temperatures indicate 225 or 210°C, depending on whether conductive or adiabatic cooling is assumed. Gas geothermometer calculations indicate possible equilibrium temperatures of 242°C (Muffler and others, 1982).

Na-SO_4 waters of the Modoc Plateau and Klamath Falls are also low in magnesium. In the Modoc Plateau, t_{Anhy} and t_{Ab-An} give temperatures of 150 to 200°C, generally slightly cooler than the 190 to 205°C indicated by $t_{SO_4-H_2O}$. At Klamath Falls, t_{Anhy} and t_{Ab-An} indicate 165 to 175°C, while $t_{SO_4-H_2O}$ indicates 185 to 200°C. This may indicate that chemical equilibrium last occured in an aquifer at about 170°C but that SO_4-H_2O isotopic compositions preserve evidence of a hotter aquifer at about 190°C.

$Na-SO_4$ waters (up to 88°C) occur at Mt. St. Helens (M. J. Thompson, unpub. data) and have t_{quartz}, $t_{Na-K-Ca}$, and t_{Ab-An} values which range from 183_{Qtz} to 169_{Ab-An}. The anhydrite saturation temperatues are lower, about 130°C probably due to dissolution of a sulfate mineral after the water left the thermal aquifer but before it discharged at the spring site. Magnesium concentrations are about 15 mg/L indicating that low temperature water-rock reaction has taken place. $Na-SO_4$ waters were also encountered in the Pucci well on Mount Hood (Robison, unpub. data), where t_{Anhy} and t_{Ab-An} indicate 140 to 150°C.

The CO_2-charged waters of the Cascades are generally of moderate to low discharge temperature. Most carry several hundred mg/L chloride and up to a hundred mg/L magnesium. Except for springs in the Meager Mountain area, none have hight SO_4-H_2O isotopic equilibrium temperatures. The springs at Meager Mountain ($t_{SO_4-H_2O}$ of 220 to 290°C) are also higher in dissolved silica than the other CO_2-charged thermal waters of the Cascades. Virtually all CO_2-charged waters of the Cascades have t_{Anhy} and t_{Ab-An} values in the 170 to 275°C range. However, the SO_4-H_2O isotopic equilibrium temperature range is much wider, 10-290°C. It appears that the albitization reaction and anhydrite saturation temperatures are meaningless in CO_2-charged water. The cation geothermometer (Na-K-Ca ± magnesium correction) gives temperatures near the spring temperatures for most CO_2-charged waters. Although high temperatures are required for CO_2 generation, there is no proof that it is being generated in the thermal aquifer. It could just as well be generated at greater depth and dissolved in the water as the gas rises towards the land surface. The depth at which the CO_2 is being generated could be considerably beyond economic drilling depth.

Discharge rates for individual spring systems in the Oregon Cascades were determined from the chloride content and discharge rates of the streams at U.S.G.S. gaging sites. Total discharges (L/S) for individual spring groups, in the Oregon Cascades, correlate with the maximum spring temperature at the respective sites (figure 1). This correlation indicates that conductive cooling is the primary factor which determines spring temperature. A major entry of chloride water not associated with a hot spring occurs at Oakridge on the North Fork of the Middle Fork of the Willamette River. Both a cool saline well (22°C) and a saline spring (18°C) occur in Oakridge (Brown and others, 1980).

Chloride data for streams on the east side of the Cascades in central Oregon indicate that two large discharge cold springs contribute most of the chloride. Discharge rate measurements for the day and site we sampled are not yet available, but using discharge measurements for previous years, the headwater springs of the Metolius River seem to carry about 25 L/S of 1,200 mg/L chloride equivalent waters (Breitenbush equivalent), while the headwater springs of the Spring River carry about 10 L/S of 1,200 mg/L chloride equivalent water.

Deuterium compositions of thermal waters of the Oregon Cascades match the composition of cold springs along the crest of the Cascade Range. If the Cascade crest is the source of recharge of the thermal water, this requires that the circulation time is very short or that, fortuitously, the isotopic distribution of precipitation is the same now as when recharge occurred. Development of "Ca-Cl component" waters is thought to be a very slow process, so circulation times of these "Ca-Cl component" thermal waters could be longer than for most thermal waters. Dissolved helium concentrations provide evidence for circulation times of at least 25,000 years, and perhaps much longer.

Gases discharged from the "Ca-Cl component" waters are principally nitrogen (>92%). They are peculiar, for geothermal water at least, in that their N_2/Ar ratios correlate with estimated aquifer-temperatures ($t_{SO_4-H_2O}$). He versus N_2 plots show two distinct trend lines, one including Belknap, Bigelson, Breitenbush, Foley, and perhaps Austin hot springs, while the other includes McCredie and Wall Creek hot springs.

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Table 1.--Estimated Aquifer Temperatures (°C) - Central Oregon Cascades

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Name	t°C	$t_{SO_4-H_2O}$	t_{Anhy}	t_{Ab-An}	t_{Quartz}	$t_{Na-K-Ca}$
Austin Hot Springs	86	181	186	155	126	84
Breitenbush Hot Springs	84	176	174	190	166	148
Bigelow Hot Springs	59	157	155	147	120	83
Belknap Hot Springs	86	148	152	155	132	82
Foley Hot Springs	79	--	100	130	113	45
Rider Creek Hot Springs	46	136	135	123	99	49
Wall Creek Hot Springs	40	--	160	137	115	69
McCredie Hot Springs	74	118	130	146	120	78
Kitson Hot Springs	44	--	134	133	96	82

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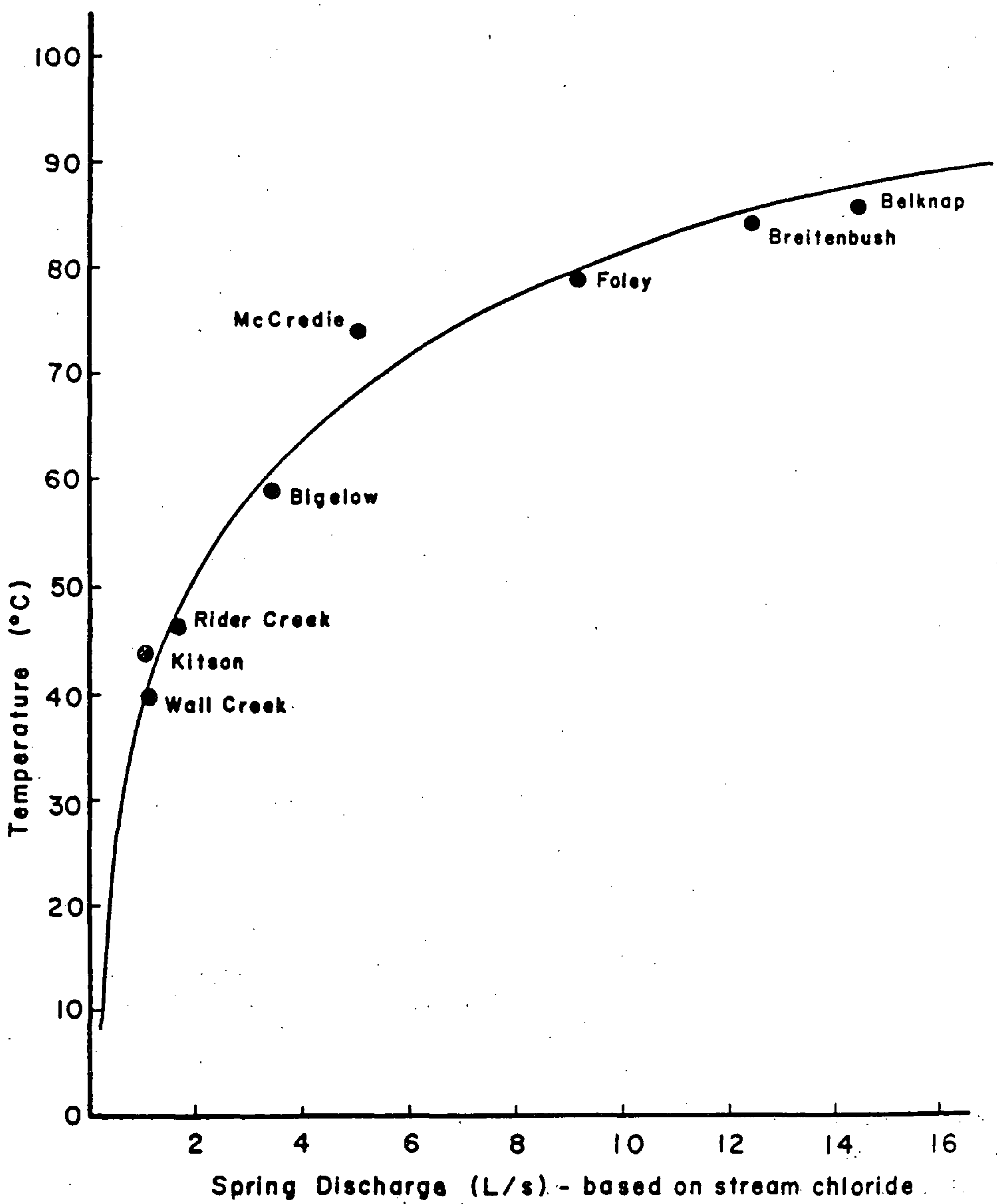

Figure 1.--Measured spring temperature as a function of calculated discharge rate.

TYPES OF HYDROTHERMAL CONVECTION SYSTEMS IN THE IN THE CASCADE RANGE OF CALIFORNIA AND OREGON

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Three general types of hydrothermal convection systems can be delineated within the Cascade Range of California and Oregon. For the purposes of this abstract, they are termed summit-crater systems, caldera systems, and lateral-flow systems. Hydrothermal systems at most major Quaternary volcanic centers within this part of the Cascade Range can be identified with one of these types on the bases of test drilling, geophysical surveys (including heat flow), and the distribution and chemical characteristics of waters from thermal springs and fumaroles.

Summit-crater systems occur at Mt. Shasta and Mt. Hood and include surficial discharge of sub-boiling point fumaroles and steam-heated hot springs. Calculations based on empirical gas geothermometers applied to fumarolic samples yield subsurface temperature estimates in excess of 200°C at both areas (C.J. Janik and R.H. Mariner, personal communication, 1985). Similar thermal manifestations have not been detected at other andesite composite cones within the High Cascades of Oregon, possibly because of an absence of hot rocks at shallow depths. Summit-crater systems may include zones where lateral flows of thermal water cool conductively and mix with nonthermal groundwater before discharging in warm springs on the volcano flanks, as for example at Swim Springs south of Mt. Hood.

Caldera systems occur at Newberry Volcano, Medicine Lake Volcano, and Crater Lake. At these locations, ring fractures associated with caldera subsidence may provide conduits for deep fluid circulation. In addition, long-lived silicic volcanism and Holocene eruptions indicate the presence of relatively shallow magmatic heat sources at each area. Test drilling at Newberry has provided temperature measurements and fluid samples that suggest that permeable strata containing water at temperatures near 265°C exist within the volcanic section at depths of 1-2 km. At Crater Lake, heat flow and geochemical investigations indicate the presence of hydrothermal convection to depths of 1-2 km beneath the lake floor and thermal-spring discharge at the lake floor. No clear evidence is yet available that hydrothermal circulation extends outside the caldera structures at these three locations, although attempts to delineate such circulation by test drilling and geophysical surveys are currently planned or in progress, and areas of anamolous soil mercury have been detected along a northeast-trending fissure/fault system on the southeast flank of Newberry Volcano (Priest and others, 1983).

A lateral outflow system occurs at Lassen Park and in the KGRA to the south. This system includes interconnected zones of vapor-dominated conditions and steam discharge beneath the summit region of the Lassen volcanic center and zones of lateral outflow of hot water that discharges in hot springs at lower elevations. Two such outflow zones have been delineated from test drilling and spring chemistry; one is fault-controlled and the other may occur along the contact between rocks from the Lassen volcanic center and underlying rocks from an older andesitic cone-building period. Water in the outflow conduits originates in a central upflow zone in which boiling occurs at temperatures near 240°C and above which a parasitic vapor-dominated zone exists at pressures near 34 bars (Sorey and Ingebritsen, 1984). Similar systems could occur within the Cascade Range in Oregon; necessary conditions include: magmatic heat sources at mid to upper crustal depths, conduits for downward fluid circulation to depths of several kilometers, and permeable strata that allow lateral flow of heated water over distances of 15-25 km. Austin, Breitenbush, Belknap-Foley Hot Springs, and other lower temperature springs along the High Cascades - Western Cascades boundary may represent the discharge points for lateral outflow from High Cascade volcanoes to the east.

Constraints on the type of hydrothermal system(s) associated with these thermal-spring areas in Oregon are provided at present by chemical data from spring waters and thermal measurements in drill holes. Isotopic data from Mariner (this volume) suggest that the systems are recharged on or near the flanks of the High Cascade volcanoes rather than at lower altitudes along the High Cascade - Western Cascade boundary. Chloride flux measurements in streams yield estimates of the total flow of thermal water at Austin Hot Springs (110 L/s), Breitenbush Hot Springs (12 L/s), and Belknap-Foley Hot Springs (33 L/s). Corresponding estimates of reservoir temperatures based on sulfate isotope calculations are near 180°C for Austin and Breitenbush and 150°C for Belknap-Foley. Values of conductive heat flow and temperature gradient beneath the western part of the High Cascades in areas unaffected by vertical groundwater flow are consistently near 100 mWm^{-2} and 65°C/km (Black, 1983). Thus, reservoir temperatures indicated for these spring systems could occur at depths of about 3 km, possibly in marine sediments below the Cascade volcanics. A large component of marine origin in the hot spring waters is also indicated by chloride contents between 400 and 3400 mg/L.

Listed below are estimated convective heat flows, assumed reservoir areas, and conductive heat inputs required to sustain long-lasting regional hydrothermal systems that may be associated with each hot-spring area. The assumed reservoir areas approximate the areas of surface drainage between the springs and the axis of the High Cascades, and as such are probably overestimates of actual reservoir areas at depth.

Spring area	Convective heat flow MW	Reservoir area km^2	Conductive heat input mWm^{-2}
Austin	83	370	220
Breitenbush	9	170	53
Belknap-Foley	21	345	61

If, as implied by the models of Blackwell and others (1982), conductive heating of deeply circulating water occurs over about half the distance between the hot springs and the High Cascade axis, calculated rates of conductive heat input would increase to levels comparable to the estimated regional heat flow for two of the three spring systems. Only for the Austin Hot Springs system are required heat inputs too large to be supplied by regional heat flow, suggesting that for this system a more potent (magmatic) heat source or deeper levels of fluid circulation may be involved. Heat sources for fluid circulation at depth beneath the axis of the High Cascades could be restricted to youthful volcanoes with intermediate to silicic composition that may be underlain by plutonic bodies, like Mt. Hood, Mt. Jefferson, and the South Sister.

Temperature data that has recently become available from the 2.4 km-deep exploration well drilled in 1981 near Breitenbush Hot Springs provide some needed detail on the nature of the regional hydrothermal systems in the Oregon Cascades. Bottom-hole temperature measurements in this well are consistent with the regional gradient of 65°C/km and heat flow of 100 mWm^{-2}. A heat flow of 380 mWm^{-2} is indicated above a depth of 750 m; this appears to result from lateral flow of water at about 140°C within the Western Cascade volcanic rocks encountered below that depth. This lateral flow could be fed by leakage from the upflow conduit(s) that supply hot water to Breitenbush Hot Springs. The Breitenbush well is located 2 km southeast of the springs, whereas temperature measurements in holes as close as 1 km to some other Cascade hot springs show no anamolous heat flow (Black, 1983).

The probable existence of regional-scale hydrothermal circulation systems within the Oregon Cascades offers promise for successful development of geothermal resources. However, several critical factors must first be evaluated. Confirmation that permeable reservoirs at temperatures near 180°C are present at depths of several kilometers is obviously needed. Although fluid temperatures signficantly greater than 180°C may exist in these systems, such conditions may only occur in close proximity to heat sources along the High Cascade axis. The permeability distributions within these flow systems must also be delineated.

Low values for the average or effective permeability of each flow system between recharge and discharge areas are indicated by the rates of natural throughflow listed above, which are 1 to 2 orders of magnitude lower than throughflow rates in high permeability volcanic environments such as the Taupo graben in New Zealand. Regions of higher permeability may exist within the Cascade systems if throughflow rates are much greater than detected so far or are limited by zones of low permeability in regions of recharge or discharge. Such a possibility can only be evaluated by deep drilling.

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Mount Hood Geophysical Investigations

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A number of geological, geochemical, and geophysical investigations were conducted around Mount Hood, Oregon (Williams et al., 1982) as part of a geochemical resource assessment jointly sponsored by DOE, U.S. Geological Survey, U.S. Department of Agriculture, Forest Service, and the State of Oregon. This particular Pleistocene-Holocene stratovolcano of the High Cascade Range was selected for study for several reasons:

(1) its proximity to a major city and to nearby potential users of geothermal energy for space heating,

(2) reasonably good road access, and

(3) evidence for a thermal source beneath the summit.

Gravity, aeromagnetic, seismic and electromagnetic surveys were carried out by several research groups. We at LBL conducted (a) shallow controlled-source electromagnetic soundings, primarily on the accessible south and west flanks of the cone (Goldstein et al., 1982) and (b) deeper magnetotelluric and magnetotelluric-telluric soundings at station clusters at accessible locations around the cone (Mozley, 1982; Mozley et al., 1985). Our objectives were to determine whether we could discern any electrical resistivity anomalies, and to relate these to other geophysical anomalies and to possible features of geothermal interest. Because we were unable to occupy MT or telluric stations closer than 3 km from the summit, we discounted the possibility of detecting an anomaly from the conduit or a relatively shallow zone beneath the summit. We were extremely interested, however, in learning whether there might be a deeper conductor that could be related to a melt zone.

Thermal Gradient Drilling

Geothermal data from 25 boreholes in a 20-km radius around the summit were collected and analyzed by Steele et al. (1982). They found that the thermal conditions are a complicated function of position around the volcano, which is in part related to rock age and rock type. A regional heat flow value of 80 mW/m^2 was determined, the heat flow increasing towards the apex. Although holes could not be drilled sufficiently close to the apex to detect a neck-type magma chamber, the measured data and thermal modeling have placed certain constraints on the depth and size of the heat source. Steele et al. (1982) conclude that there cannot be a large, shallow subvolcanic melt zone less than 3 km deep and with a radius greater than 2-3 km.

Aeromagnetics

Flanagan and Williams (1982) found a large, normally polarized magnetic high associated with the mountain. Removing the effects of a uniformly magnetized cone (the topographic anomaly), they found evidence in the residual anomaly for smaller, older volcanic cones largely concealed by the later eruptions, one cone under the north flank the other under the south flank. In addition, magnetic lows west and southwest of the mountain could be related to less magnetic Pliocence quartz diorite intrusives.

Subregional Gravity

Couch and Gemperle (1979) found that Mount Hood is superimposed on a gravity low that suggests a north-south oriented graben-like structure. The cone itself is a Bouguer gravity high which, like other discrete highs in the survey area, is believed to be the denser core of the volcanic vent.

Detailed Gravity

A detailed gravity survey revealed a dense body, presumably an intrusive, with a diameter of 5 km beneath the summit (Williams and Keith, 1982). The body appears to neck down to a diameter of 1 km with depth.

Seismic

Results of a teleseismic P study indicated no velocity anomaly that one might expect if there were a large, shallow melt zone beneath the volcano (Weaver et al., 1982). The absence of a P-wave delay anomaly at Mount Hood and beneath Cascade volcanoes, in general, (Iyer et al., 1982) has been attributed to the possible small size of these melts (< 5 km in diameter) and the very complex vertical and lateral variations in velocity that makes identification difficult.

On the basis of a refraction survey, Kohler et al. (1982) found high seismic velocities on the west side of the volcano, wraping around to the south. The high velocities seem to correlate with the magnetic low and an electrically resistive zone believed due to a broad area of Pliocene intrusives, only partially exposed by erosion.

Electromagnetics

Several discrete conductive features were discerned after one-, two-, and three-dimensional modeling was done on the CSEM and MT data. Although the modeling did not produce a totally satisfying electrical model, the discrete anomalies could be assembled into a general composite model that seems to be consistent with other geophysical results.

Shallow Conductors

Shallow conductors at depths of about 500 m were mapped on the east and south flanks. These conductors are believed due to the radial flow of meteoric water (mainly snowmelt) in permeable volcanic units. The waters are believed to be heated by hot rocks in the summit area (Wollenberg et al., 1979). The EM results agree with information from the hole at the base of the Pucci chairlift on the south flank.

Conduit Conductor

Only weak evidence could be found for a neck-type conductor beneath the summit. Due to station locations, the presence or absence of a narrow conductor did not seem to affect the 2-D interpretation. However, some of the MT parameters in the 1 to 5 Hz bandwidth suggested a shallow conductor beneath the summit.

Resistive Anomalies

Large areas of resistive rocks, extending from near the surface to 10^{+} km, were indicated on the south-southeast flank and beneath much of the western region of the volcano. There is a relatively strong correlation between the resistor, high velocities from a time-term analysis of refraction seismic data (Kohler et al., 1982), the aeromagnetics (Flanagan and Williams, 1982), and a few quartz diorite outcrops. The resistive nature of the rocks on the west flank and the occurrence of older intrusives was also confirmed by results from the Old Maid Flat well OMF-7A (Blackwell et al., 1982).

Deep Conductors

An elongate conductor (1 ohm·m) striking N20°W was discerned at a depth of 12 km and with a depth extent of 10 km. The conductor is coincident with the trend of the High Cascades and it may be a zone of partial melt. Although this explanation has not been confirmed seismically,it is a reasonable one based on the corrected conductive thermal gradient of ~65°C/km measured in a number of holes around the volcano (Steele et al., 1982). That this conductor was resolved at all by MT came as something of a pleasant surprise in view of the limited spatial and frequency window available. However, the dimensions of the conductor could not be resolved, and this uncertainty admits another resistivity model, one with an additional deeper conductor at ~50 km.

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CHARACTERIZATION OF GEOTHERMAL SYSTEMS IN THE CASCADE RANGE FROM GEOELECTRICAL STUDIES

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Several types of geoelectrical surveys have been made by the U.S. Geological Survey in the Cascades volcanic terrains. A list of the methods used, the areas studied, and the names of the investigators follows:

(a) Magnetotellurics (MT) regional studies over the entire Cascade range, by W. D. Stanley.

(b) Audiomagnetotellurics (AMT), near Lassen National Park and Mt. Hood, by D. B. Hoover.

(c) Telluric profiling, in Medicine Lake volcano and near Lassen National Park, by K. Christopherson and D. B. Hoover.

(d) Schlumberger soundings, in Medicine Lake volcano by A. A. R. Zohdy and R. J. Bisdorf, and in Newberry by R. J. Bisdorf, and south of Lassen by R. J. Bisdorf and A. A. R. Zohdy.

(e) Time domain electromagnetic soundings, in Medicine Lake volcano, by W. L. Anderson and F. C. Frischknecht, and in Newberry by D. V. Fitterman.

(f) Frequency domain electromagnetic soundings, in Medicine Lake volcano by W. L. Anderson and F. C. Frischknecht.

(g) Airborne electromagnetics (DIGHEM), in Medicine Lake volcano and near Lassen National Park, by D. Frazier and D. B. Hoover.

(h) Self potential (SP), in Mt. Hood, by D. B. Hoover.

Two crater volcanoes, namely Medicine Lake and Newberry, were studied more extensively by most of these methods. The geoelectric sections in these two areas are similar and may be described from top to bottom by four geoelectric units: (a) A very high resistivity unit of 15,000 to 150,000 ohm-m and of few tens of meters in thickness. This unit is comprised of loose pumice and dry volcanic rocks. (b) A medium-high resistivity unit of 500 to 5,000 ohm-m with thickness of 500 m to 1 km. This unit probably corresponds to Quaternary volcanic rocks that may be saturated in part with fresh, cold water. (c) A low resistivity unit of 5 to 30 ohm meters with a thickness of 500 meters to greater than 1 km. This unit probably represents altered volcanic rocks possibly saturated with circulating hot water. It represents the shallow geothermal resource target. (d) A high resistivity geoelectric basement of > 500 ohm-m which is several kilometers in thickness. The shallowest depths to its top are about 1.5 to 2 km. It is believed to represent Tertiary intrusive rocks.

The exploration of these areas by the direct current resistivity method (Schlumberger soundings) proved to be a challenging task because of very high contact resistances, near absence of long straight roads, and off the road dense forest and fallen trees. Special formulas were derived for computing the proper geometric factors, digitizing the road maps, and correcting the field measurements. The mathematical solution to the winding road problem has opened the door to investigating many other areas with little concern to the presence of straight roads of sufficient length.

In the Medicine Lake area, a DIGHEM airborne electromagnetic survey was made to study the areal distribution of electrical resistivity at shallow depths of the order of few meters or tens of meters. The results showed three low resistivity anomalies (10 to 100 ohm-m), possibly representing near surface hydrothermal alterations. The anomalies are near the northwest corner of Medicine Lake itself, southeast of the Hot Spot, and near Telephone Flat (slightly north of Bullseye Lake). The Schlumberger, time domain, frequency domain, and MT soundings showed that an extensive low resistivity layer of about 5 to 30 ohm-meters underlies a high resistivity cover, and that significant low resistivity anomalies are centered near Bullseye Lake and northwest of Medicine Lake. The results of the 51 Schlumberger soundings are illustrated by two cross sections and by four maps of interpreted-true resistivity at depths of 250 m, 500 m, 750 m, and 1,000 m.

At Newberry, an east-west geoelectric section constructed from the interpreation of Schlumberger soundings shows a low resistivity zone that corresponds to the increase in thermal gradient observed in a USGS test well which encountered 256°C water at a depth of 930 m. Another low resistivity zone about 600 m deep is present just to the west of the caldera boundary. A map of interpreted-true resistivity at a depth of 750 m shows that this western low resistivity zone has an areal extent of about 16 km^2. This zone is interesting enough that at least one company has drilled it. Several time domain electromagnetic soundings also were made in the Newberry volcano area.

Regional east-west magnetotelluric cross sections show that the axis of the Quaternary volcanoes of the High Cascades, to the west of Medicine Lake and Newberry, coincides with an apparent structural trough. The sections also show that, beneath Medicine Lake and Newberry, there are high resistivity rocks believed to consist of multiple small intrusions forming structural highs underneath the low resistivity zones. These structural highs (which were detected by only very few of the Schlumberger and time domain soundings) at a depth of about 2 km or less could be considered as fundamental geothermal targets if they consist of multiple young intrusive units.

DEPARTMENT OF ENERGY DRILLING IN THE CASCADE RANGE

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This year the U.S. Department of Energy, Geothermal and Hydropower Technologies Division (GHTD) will share with the geothermal exploration industry the cost of drilling several thermal-gradient holes to depths of 1 km or more in the Cascade Range. In return for its cost share, GHTD will receive copies of the drilling and geophysical logs, samples of cuttings and cores, and a period of access to the holes for further geophysical measurements. Participation in this prospect drilling will allow GHTD to evaluate the effectiveness of intermediate-depth thermal-gradient holes in locating geothermal resources in areas like the Cascades, where percolating surface water masks the thermal gradient at shallow depths. A total funding of $1 million is available this year for the GHTD share of the drilling.

Several factors were considered in the planning for this project. Shallow holes in the older volcanic rocks of the Western Cascades provided good measurements of the conductive thermal gradient, but in the High Cascades thermal measurements in shallow holes were dominated by cold, moving ground water. Drilling at Mount Hood and Newberry Caldera indicated that thermal-gradient holes should be at least 1 km deep to penetrate the cold ground water and to reach conductive gradients. Much deeper holes would be required to penetrate the geothermal systems thought to exist at these two sites, but the drilling of intermediate-depth holes seemed to offer an attractive research tool. Since thermal-gradient holes can be drilled at a small diameter, they are much less expensive than production wells. Sharing costs with industry is an attractive way of doubling the available experimental data from thermal-gradient sites.

Because of cost sharing, data from this project will be limited to geothermal prospects in the region of the Cascades and Newberry and Medicine Lake Calderas. Industry drilling is confined to those areas where geothermal development is permitted, and it is anticipated that the drilling will be on land with existing geothermal leases. An important aspect of this project from the industry viewpoint is that a reasonable chance of development exists if a resource is indicated.

In addition to sharing the drilling costs, GHTD will support research in and around these gradient holes. Stable temperature measurements will be made during the period of access, and thermal conductivities will be measured in cores from each hole to provide calculations of heat flow. This project requires the drilling of holes to 1 km or deeper, but the method of drilling is up to the bidder. If core holes are drilled, we will have a complete lithologic record that can be compared to surface surveys and provide constraints on geologic models. If larger-diamater holes are drilled, it may be possible to run several geophysical logs to aid in the intertpretation of surface surveys. It is possible that thermal water will be intercepted in one or more holes, and if hole stability and environmental safety can be maintained there are plans to obtain water samples for geochemical analysis. We would appreciate your suggestions as to other research or exploration techniques that would enhance the data from this drilling project.

POSSIBLE SITES FOR SCIENTIFIC DRILLING IN THE CASCADE RANGE

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INTRODUCTION

A great deal has been learned about the Cascade Range from surface surveys and shallow temperature-gradient drilling. Nevertheless, a number of problems have stubbornly resisted solution. Among the most important of these are:

1. Mapping and characterization of hydrothermal systems in both the Western Cascades and the High Cascades.

2. Determination of background conductive heat flow in the High Cascade Range.

3. Mapping of the pre-Quaternary structure and stratigraphy of the High Cascades.

4. Accurate modeling of the source of the regional heat flow anomaly, including energy and mass balance calculations relative to rates of volcanism and subduction.

Drilling of intermediate-depth and deep drill holes is crucial to solution of these and related problems. Only drill holes can provide the necessary fluid and rock samples and down-hole geophysical measurements to place quantitative constraints on possible geologic and hydrologic models.

WHERE SHOULD DRILLING BE FOCUSED?

Given that drilling should occur in the Cascades, the problem then is to decide which areas have the greatest potential for resolving the above problems. Most of the problems are related to the thermal regime of the Cascades, so picking sites can, to a first approximation, be accomplished utilizing a geothermal exploration strategy. The geothermal exploration industry has for decades utilized a phased approach to drilling programs. Surface geological, geochemical, and geophysical surveys are normally followed by shallow drilling of temperature-gradient holes which in turn are followed by drilling of deeper wells to test for hydrothermal fluids. Tables 1 and 2 are qualitative summaries of where various parts of the Cascade Range stand with respect to resource potential, access, and completion of temperature-gradient drilling. These tables take into account only publicly available data. The evaluation categories in the tables could be separated into a number of subcategories corresponding to the various exploration phases, but this would not materially affect the final ranking of sites. The point of this exercise is that there are only a few areas which are at present suitable targets for scientific drilling programs.

TABLE 1. Qualitative summary of the quality of the data base, access, and geothermal resource potential at possible sites for research drilling in the Cascades (0 = Poor; 1 = Moderately Poor; 2 = Moderate; 3 = Moderately Good; 4 = Good; 5 = Very Good). Soil geochemical surveys in the public domain are only available for Newberry and Medicine Lake volcanoes. Data from wells drilled to depths in excess of 600 m are only available for the Mount Hood area, the Breitenbush area, and Newberry Volcano.

Site	Geologic Coverage	Drilling (Heat Flow)	Geophysics (Non-heat flow)	Access	Resource Potential
			WASHINGTON		
Mt. Baker	3	0	3	3	4
Mt. Rainier	3	0	2	1	4
Glacier Peak	2	0	2	1	4
Mt. Adams	5	0	4	1	4
Mt. St. Helens	5	0	4	1	4
N. & Cent. Wash. between volcanoes	2	0	2	2	0
Southern Wash. between volcanoes	2	2	3	4	2
			OREGON		
Mt. Hood	4	2	3	1	4
Adjacent to Mt. Hood	3	4	3	3	2
Mt. Jefferson area	4	0	3	0	4
Breitenbush area	5	5	3	5	4
Santiam Pass area	5	2	3	4	4
North Sister area	5	0	3	0	4
Middle Sister area	5	0	3	0	4
McKenzie Pass	5	1	3	1	4
South Sister	5	0	3	0	5
Century Drive area	5	0	3	2	5
Newberry Volcano	5	2	5	3	5
Willamette Pass area	4	2	3	4	3
Crater Lake	5	1	3	1	5
Adjacent to Crater Lake	3	0	3	4	4
Oregon passes S. of Crater Lake	3	0	3	4	3
			CALIFORNIA		
Mt. Shasta	5	0	5	1	4
Adjacent to Mt. Shasta	5	3	5	4	3
Medicine Lake	5	3	5	5	5
Mt. Lassen	5	2	4	0	4
Between Calif. stratocones	3	1	3	4	3

Table 2. Ranking of sites based on total scores from Table 1. Parentheses indicate site has poor or moderately poor accessibility. Brackets indicate site has poor resource potential.

Score	Site
22	Breitenbush Hot Springs area, Medicine Lake Volcano
20	Newberry Volcano,
19	Areas adjacent to Mt. Shasta
18	Santiam Pass area
16	Willamette Pass area
15	Areas adjacent to Mt Hood, Century Drive area, (Crater Lake), (Mt. Shasta), (Mt. Lassen)
14	(Mt. Adams), (Mt. St. Helens), (Mt. Hood), (McKenzie Pass), areas adjacent to Crater Lake, areas between northern California stratocones south of Mt. Shasta
13	Mt. Baker, (South Sister), Oregon passes south of Crater Lake, southern Washington Cascades between major volcanoes
12	(North Sister area), (Middle Sister area)
11	(Mt. Jefferson area)
10	(Mt. Rainier)
9	(Glacier Peak)
6	{Northern and central Washington between major volcanoes}

Areas which stand out as good potential drilling sites are:

1. Newberry and Medicine Lake volcanoes
2. The Breitenbush-Santiam Pass transect, central Oregon
3. Areas adjacent to Mt. Shasta, northern California
4. Willamette Pass-Century Drive transect, central Oregon
5. Areas adjacent to Mt. Hood, northern Oregon

The Newberry and Medicine Lake volcanoes, major silicic volcanic centers with large data bases, scored high in the ranking. It is now well known that these volcanoes have a high potential for shallow, high-temperature hydrothermal systems, as shown by the USGS Newberry 2 well (Sammel, 1981). However, the writer is not convinced that further drilling in these areas will tell us much about the High Cascade Range to the west, particularly areas characterized by andesitic to basaltic volcanism. What is needed is an east-west profile of the heat flow in the High Cascade Range and one or more wells deep enough to sample potential hydrothermal systems. Ultimately, it would be of immense scientific and economic interest to penetrate to 7-10 km into the source region for the Cascade heat flow anomaly (Blackwell and others, 1982).

The Breitenbush area, considering that it is in the volcanically inactive Western Cascade Range, ranks very high relative to other sites (Table 2). The abundance of geological, geophysical, and drilling data and the presence of a known, potentially exploitable hydrothermal system caused the Breitenbush area to score high. Sunedco Energy and Development released much of their data for the Breitenbush area including data from their 2.5 km well (Well No. 58-28), the deepest well in the U.S. Cascades. This well intersected a major hydrothermal aquifer at temperatures in excess of 136^{o} C (Waibel, 1985), and it can be reentered for deepening and additional scientific studies (Edward Western, personal communication). Data from numerous shallow temperature-gradient holes are also available for the Breitenbush area. The temperature-gradient data indicate that the aquifer in Well No. 58-28 dips to the east where it might be intercepted at higher temperature (Waibel, 1985).

Well No. 58-28 is also close to the western margin of a regional graben structure inferred from gravity studies of Couch and others (1982) and Couch and Foote (1983). Referred to as the "Cascade graben" (Couch and Foote, 1983), this structure apparently extends from Crater Lake to just north of Mount Jefferson (Couch and others, 1982). The structure is inferred to have experienced active displacements since the mid- to late-Miocene (Couch and Foote, 1983). Deepening the Sunedco Well would test the hypothesis that this structure has a down-to-the east displacement in excess of 3 km (Couch and Foote, 1983).

Deepening the Sunedco well to about 4 km will allow examination of the pre-Cascade crust (Couch, unpublished calculated gravity cross section). This will help to resolve whether this part of the Cascade arc is built on oceanic crust, as postulated by Hamilton and Meyers (1966) and others.

Drilling in the Breitenbush area will help to resolve many questions about the nature of the Western Cascade structure and hydrothermal convection but will not elucidate the structure and thermal regime of the High Cascades. The proximity of the Santiam Pass area to the Breitenbush site, its accessibility, and excellent geological coverage (Taylor, 1967; Taylor, unpublished 1:24,000-scale mapping) make the Santiam Pass area a prime target for a profile across the High Cascades. Santiam Pass is also one of the few easily accessible places in the High Cascade Range where the axis of volcanism can be located with a great deal of confidence. It thus affords an unique opportunity to test the hypothesis that the current axis of mafic to intermediate volcanism may have high conductive heat flow and a high potential for hydrothermal resources. In addition, the area lies in a large Pliocene graben which formed after eruption of voluminous late Miocene to early Pliocene ash flows and lavas from the High Cascades (Taylor, 1980). Drilling could provide quantitative constraints on the amount of displacement on this graben, and the nature of the pre-graben volcanic arc (now buried by Quaternary volcanoes). Drilling would also help to test the hypothesis that the regional residual gravity high under the Three Sisters-Mt. Jefferson segment is caused by a large intrusive complex (R. Couch, 1985, personal communication).

A comprehensive study of the Breitenbush area and the adjacent Santiam Pass area is the most logical and cost-effective first step in a scientific drilling program. This would examine both of the major graben structures, the mid- to late-Miocene Cascade Graben and the Pliocene graben in the High Cascades; it also would take advantage of a wealth of previous studies, including the opportunity to reenter the deepest well in the U.S. Cascades. In future years other east-west transects should be studied to help extrapolate the Santiam-Breitenbush data to the north and south.

Drilling in the areas adjacent to Mount Shasta would be of great value both in terms of geothermal assessment and volcanologic research. Drilling would capitalize on the detailed geophysical and geological data base which has been developed for this area by the USGS. Completing a transect to Medicine Lake would also be of interest to study the transition zone between the two areas.

The Century Drive area adjacent to the South Sister is unique in that it has the only Holocene rhyodacite volcanic centers in the High Cascade Range which are also accessible by a major highway. The area is part of a long-lived silicic highland of regional extent (Taylor, 1980). It would be an ideal place to establish whether hydrothermal resources associated with silicic volcanism are present in the main High Cascade Range. Heat flow and geophysical profiles across the full width of the Willamette Pass-Century Drive area would be of interest, especially in comparison with profiles at Breitenbush-Santiam Pass where mafic and intermediate volcanism prevails. Extending the studies to Newberry Volcano would allow investigation of the relationship between the silicic highland at Century drive and silicic volcanism at Newberry.

Areas adjacent to Mount Hood scored well in the ranking because of the large available data base. However, extensive drilling in the area has not been successful in locating indications of major high-temperature hydrothermal systems. According to some interpretations of the geophysical surveys (Goldstein, personal communication), the drill holes may be too shallow and not in the best locations to intercept the hydrothermal systems. In any case, the gravity data indicate that, unlike many of the large High Cascade volcanoes to the south, Mount Hood may not lie above a large, shallow intrusive complex (R. Couch, 1985, personal communication), although Williams and Finn (1983) concluded that a shallow high density intrusion could be present. South of Mount Hood the regional background heat flow increases to values in excess of 100 mW/m^2 (Blackwell and others, 1982), which would be additive to any local heat flow associated with shallow intrusions. This factor would increase the likelihood of large hydrothermal systems in the Cascades south of Mount Hood even if shallow plutons are locally present in both areas.

A transect of detailed temperature-gradient drilling and surface surveys across the southern Washington Cascades would be valuable. This would help to improve the meager heat flow data base in Washington and allow comparisons between this area of somewhat lower regional heat flow (Blackwell and Steele, 1983) and lower rates of Quaternary volcanism to areas with higher rates in Oregon and northern California. This is important for developing a comprehensive model for the Cascades as a whole.

CONCLUSIONS

The Santiam Pass-Breitenbush area is the best site for the first phase of a scientific drilling program in the Cascade Range. This area has a known hydrothermal convection system, a high rate of volcanism, high background heat flow, and a large existing data base for siting drill holes. It is recommended that Sunedco Well No. 58-28 near Breitenbush Hot Springs be reentered, tested, and deepened to about 4 km. Drilling of an additional well to about 2 km depth east of the Sunedco site might cross the east-dipping hydrothermal aquifer at higher temperatures (Waibel, 1985, personal communication). An east-west transect of at least three, and preferably four, 1 km or deeper slim holes should be drilled across the Cascades at the latitude of Santiam pass to delineate the temperatures and heat flow below the zone of rapidly circulating cold ground water. A 2.7 km or deeper well should be drilled near the axis of Quaternary volcanism to test for potential hydrothermal fluids and to explore the nature of the pre-Quaternary stratigraphy. The drilling should be coupled with a program of detailed surface geophysical surveys and extensive radiometric dating of volcanic units to help site the wells and to allow maximum extrapolation of the drilling data. Stratigraphic thicknesses determined from mapping and drilling should be combined with the lateral extent and age of volcanic units to calculate rates of volcanism. This should then be compared to contemporaneous rates of subduction, heat flow, hydrothermal circulation, and crustal deformation to examine causal relationships.

Drilling programs similar in scope to the Breitenbush-Santiam study should be accomplished in the southern Washington Cascades, the Century Drive-Willamette Pass area, and the Mount Shasta area in future years in order to develop a comprehensive hydrologic and geologic model of the Cascade Range. It would also be useful to link up east-west transects across the Cascades to similar transects across the Newberry and Medicine Lake volcanoes to examine the interrelationships between these silicic volcanic centers and the main Cascade arc. Ultimately, drilling to depths of 7-10 km will be necessary to investigate deep magmatic and metamorphic processes operative under the area of the regional heat flow anomaly associated with the Cascade volcanic arc.

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