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HEAT FLOW MODELING OF THE MOUNT HOOD VOLCANO, OREGON

By

DAVID D. BLACKWELL
and
JOHN L. STEELE

Department of Geological Sciences

Southern Methodist University

Dallas, Texas 75275

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INTRODUCTION

Most of the currently active volcanoes are associated with volcanic arcs which form over subducting slabs of lithosphere in plate collision areas. Large stratocone volcanoes of generally andesitic composition are typically associated with these active volcanic arcs, although extrusions of basaltic composition are probably more volumetrically important. Along the western Pacific there is an almost continuous string of volcanic arcs associated with subducting slabs of lithosphere. Along the western margin of the North American Continent the subduction is more fragmented than in the western Pacific; however, there are several distinct regions where volcanic arcs are developed. These regions are the Aleutian arc of Alaska, the Cascade Range of the northwestern United States, the trans-Mexico volcanic belt in Mexico, and the Central American volcanic arc. Active subduction is definitely occurring at the first and last of these areas, as evidenced by large numbers of intermediate and deep focus earthquakes generated within the cold sinking block of lithosphere. Along southwestern Mexico there are numerous earthquakes of intermediate focal depth which generally line up to indicate a slab with a very shallow dip.

On the other hand, in the Pacific Northwest, there are no intermediate or deep focus earthquakes, so direct seismic evidence for the existence of a subducting slab does not exist. However, the deepest earthquakes in the conterminous United States (60 km) are associated with the Puget Sound area of northwestern Washington, and so apparently the tectonic situation must be different in some way from the remainder of the western United States. Moreover, based on the evidence of abundant volcanism in the Cascade Range, and the present-day existence of many large stratocone volcanoes which would be rapidly removed by erosion, it is apparent that subduction has been active in the Pacific Northwest within the past few thousand years. Subduction may be continuing, but at a low level or with the block too hot to generate earthquakes.

Although extrusive rocks of basaltic composition often are volumetrically dominant in the volcanic arcs associated with subduction zones, the associated intrusive rocks are dominantly of granitic composition. Thus, it is felt that geothermal systems are more likely to be associated with volcanic rocks of andesitic to rhyolitic composition where the magmas are more viscous and more likely to solidify as intrusive rocks rather than proceed to the surface and be extruded (see Smith and Shaw, 1975). By cooling below the surface, more thermal energy is transferred into the shallow crust and hydrothermal convection systems are more likely to be formed. In fact, epizonal plutons probably cool dominantly, if not completely, by hydrothermal convection rather than conduction.

Around the world, many geothermal systems are associated with andesite volcanoes. In particular, several of the geothermal systems in Japan, such as Matzukawa and Otaki, are associated with stratocone volcanoes. In Central America, the geothermal fields at Ahuachapán, El Sálvador; Mombotambo, Nicaragua; and in Costa Rica are also associated with stratocone volcanoes and are currently producing or will soon produce electrical power. Therefore, the Cascade Range of the Pacific Northwest represents an obvious focus for geothermal exploration. However, in spite of the abundant evidence of voluminous volcanism of variable composition in the Cascade Range, thermal manifestations are much more subdued than those typically associated with the island arcs. The only thermal manifestations which appear major from their surface evidence are associated with Mt. Lassen, in northern California. The remainder of the andesite volcanoes in the Cascade Range display only weak evidence of hydrothermal activity, although most of the volcanoes have vent fumaroles and have been historically active.

The overall thrust of this project was to investigate the geothermal potential of the Mt. Hood volcano as a guide to estimating the geothermal potential of other stratocone volcanoes in the Cascades. The assumption was made, therefore, that Mt. Hood represents a typical stratocone volcano and that the information obtained in the study of Mt. Hood could be transferred to other stratocone volcanoes in the western United States. A location map showing the Cascade stratovolcanoes is given in Figure 1. In view of the generality of the approach, part of the research carried out in the project was a more generally-oriented study of thermal effects of magma chamber systems associated with stratocone volcanoes and of regional heat flow in the Cascades.

The report is essentially divided into three sections. In the first section, some examples of simple cooling models of different shapes of magma chambers that might be associated

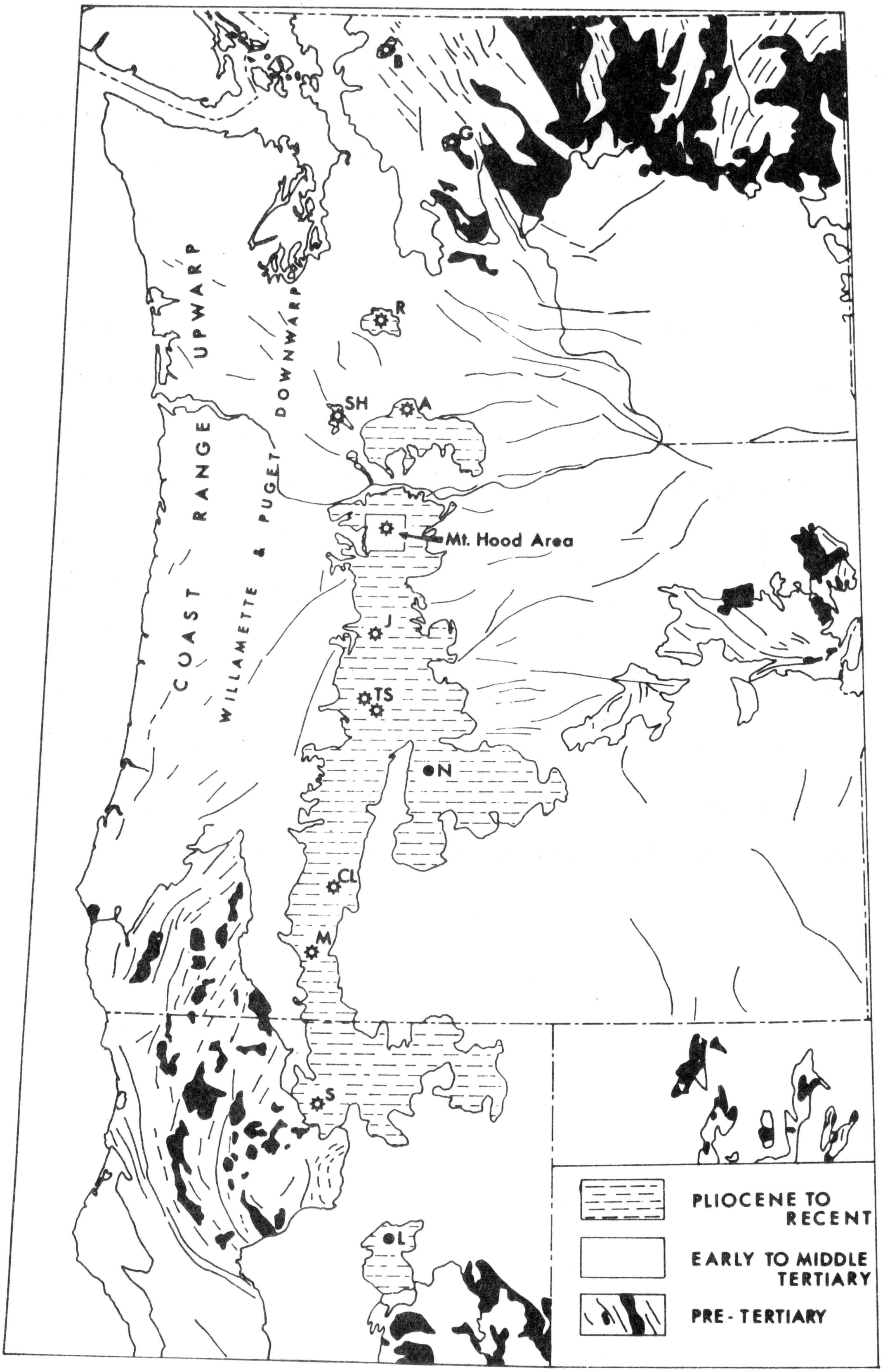


Figure 1. Location map of Cascade stratocone volcanoes (from Wise, 1969).

with stratocone volcanoes are discussed. These simple models are calculated in order to investigate the effects of different geometry on the total heat transfer of the system, rate of cooling, and volcanic recurrence times necessary to keep the crust hot.

In the second section, a description of a detailed study of the Western Cascade-High Cascade boundary in the Western Cascade province is discussed along with the results of heat flow studies along the eastern boundary of the Northern Cascade Range of Oregon. These studies are presented here for completeness and for comparison with the data obtained in the Mt. Hood region.

The final section is a discussion of the heat flow in the vicinity of Mt. Hood, including near-regional exploration holes, and deep exploration holes at Timberline Lodge and at Old Maid Flat at the foot of the Mt. Hood volcano.

THERMAL MODELS OF MAGMA CHAMBERS WITH SPECIFIC APPLICATIONS TO STRATOCONE VOLCANOES

Introduction

A number of different models for the cooling of magma chambers have been discussed in the literature. Most of the models have included discussion of the cooling of a magma emplaced instantaneously at constant temperature, with subsequent cooling from this constant initial temperature (Jaeger, 1963, 1964; Blackwell and Baag, 1974; Lachenbruch and others, 1976). In these models, the magma is assumed to cool conductively once it is in place, and no modifications are included to allow for possible convection in the magma during cooling, permeation of water through the cooling country rock or solidified intrusive rock, or any complexities that might be involved with emplacement of a major magma chamber. In the simplest form, the model predicts that the country rock temperature in contact with the magma chambers never reaches a temperature greater than 0.5 T_m, where T_m is the melt temperature. This model is referred to as the instantaneous model.

Another simple conduction model for magma chamber evolution assumes emplacement of the magma at some time and its subsequent maintenance as a magma chamber over a long period of time. In this case, the rock in contact with the magma will eventually be heated up to the magma temperature. This model, the continuous mode, would be more appropriate for a long-lived magma chamber into which batches of magma were repeatedly emplaced or, for moderate times at least, a very large convecting magma chamber. Of course, at some time such a chamber must eventually begin to cool off, but there is geological evidence that many magma chambers are resupplied by magma and thus can maintain a liquid zone over a long period of time relative to the cooling period of the instantaneous model.

If the magma chamber is maintained for an extreme period of time, the thermal effects eventually reach a steady-state. Ultimately, of course, the magma must start to cool off. Yuhara (1974) has presented some steady-state models of a linear volcano with a plane (vertical) source magma chamber. He illustrates the temperature for such a model for different depths to the plane source.

In general, the geological evidence indicates that most, if not all, magma chambers emplaced in the epizonal region of a crust (1-5 km) actually cool dominantly by convection of heated ground water through the cooling rock in a hydrothermal system. Very seldom is it likely that a large magma chamber would cool conductively. However, analysis of the rate of heat transfer in typical geothermal systems suggests that typical efficiency of convective cooling is not particularly great; the heat transferred by a convection system being on the order of two to ten times the rate of heat loss by conductive cooling alone. So the average heat loss for convective cooling as contrasted with conductive cooling of an active magma chamber might be estimated to be on the order of five.

Basic Model Shapes

Figure 2 shows a topographic cross-section along an

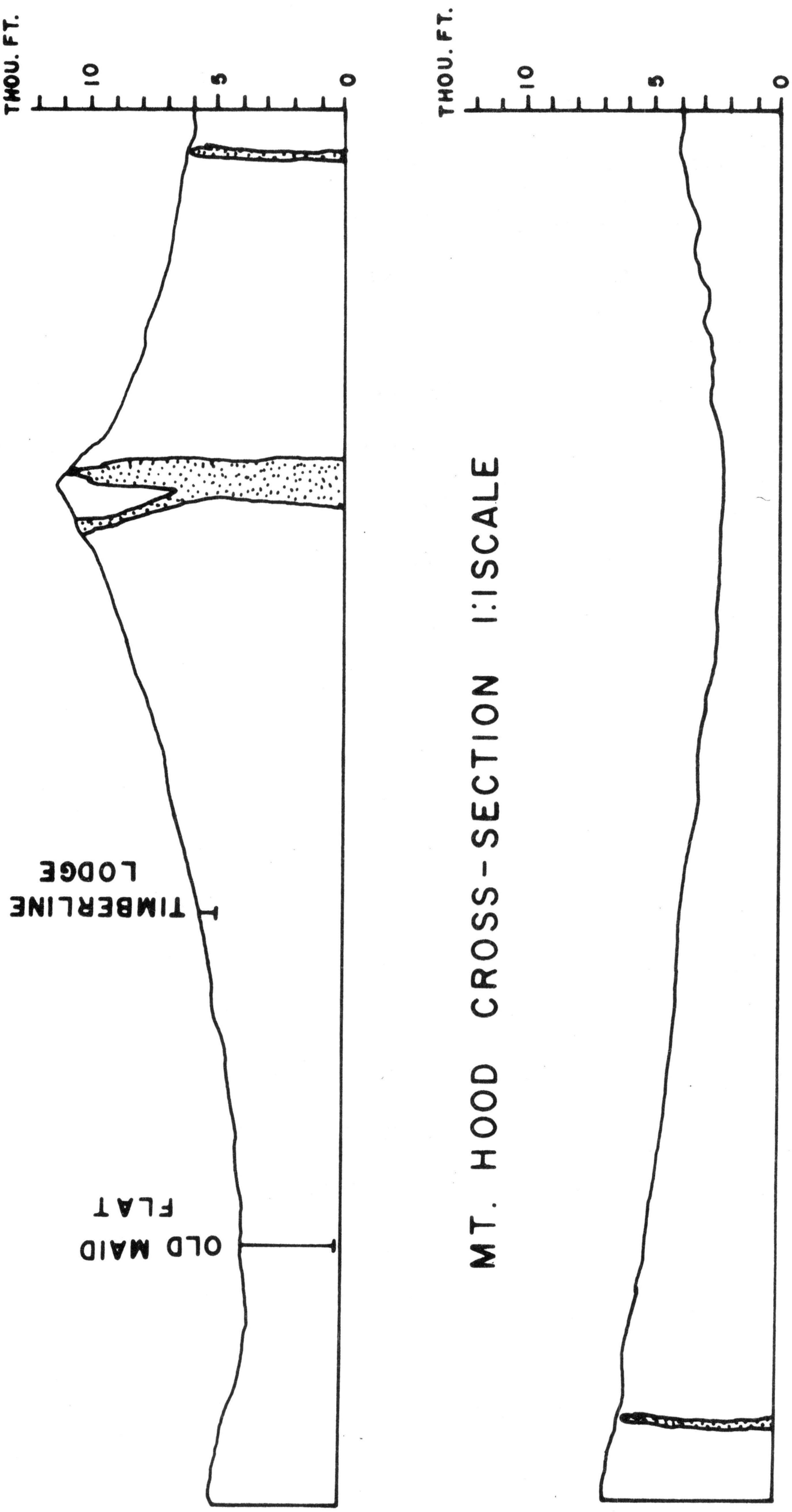


Figure 2. Topographic cross-section of Mt. Hood: east-west profile at a scale of 1:1. Locations and drilled depths of two major heat flow holes are shown.

approximately east-west line through Mt. Hood. This cross-section has equal horizontal and vertical scales. Also shown on the volcano are the locations and scaled depths of two relatively deep exploratory holes which have been drilled for geothermal studies.

Mt. Hood is a typical stratocone volcano. It has a relief of about 1.5 km and a radius of 5 km, giving an approximate surface area of 80 km^2 and a volume of 40 km^3. The geology of the volcano has been discussed by Wise (1968). Most of the volcano was formed in late Pleistocene, and a major eruption which formed most of the south flank of the volcano occurred about 1,600 years ago. The most recent major activity has been dated at about 220 years ago (Crandell and Rubin, 1977). The volcano was in minor eruption between 1848 and 1865, according to newspaper accounts. There is an active fumarolic system near the top of the volcano, but only a "single" thermal spring with several orifices is associated with the volcano (Swim Spring). The spring is along the southern edge of Mt. Hood.

In the discussion of the conductive cooling of magma chambers, two basic shapes will be considered (Figure 3). The first is a parallelepiped with infinite extent in the Y direction, a width equal to 2A, and a thickness equal to L, emplaced at a constant initial temperature (T_o), and at a depth D below the surface. This particular model might apply to a linear magma chamber, such as might exist if several stratocone volcanoes coalesced, or to a dike or sheet-like magma chamber. This sort of model might apply if there is a large regional magma chamber associated with the entire Cascade Range,

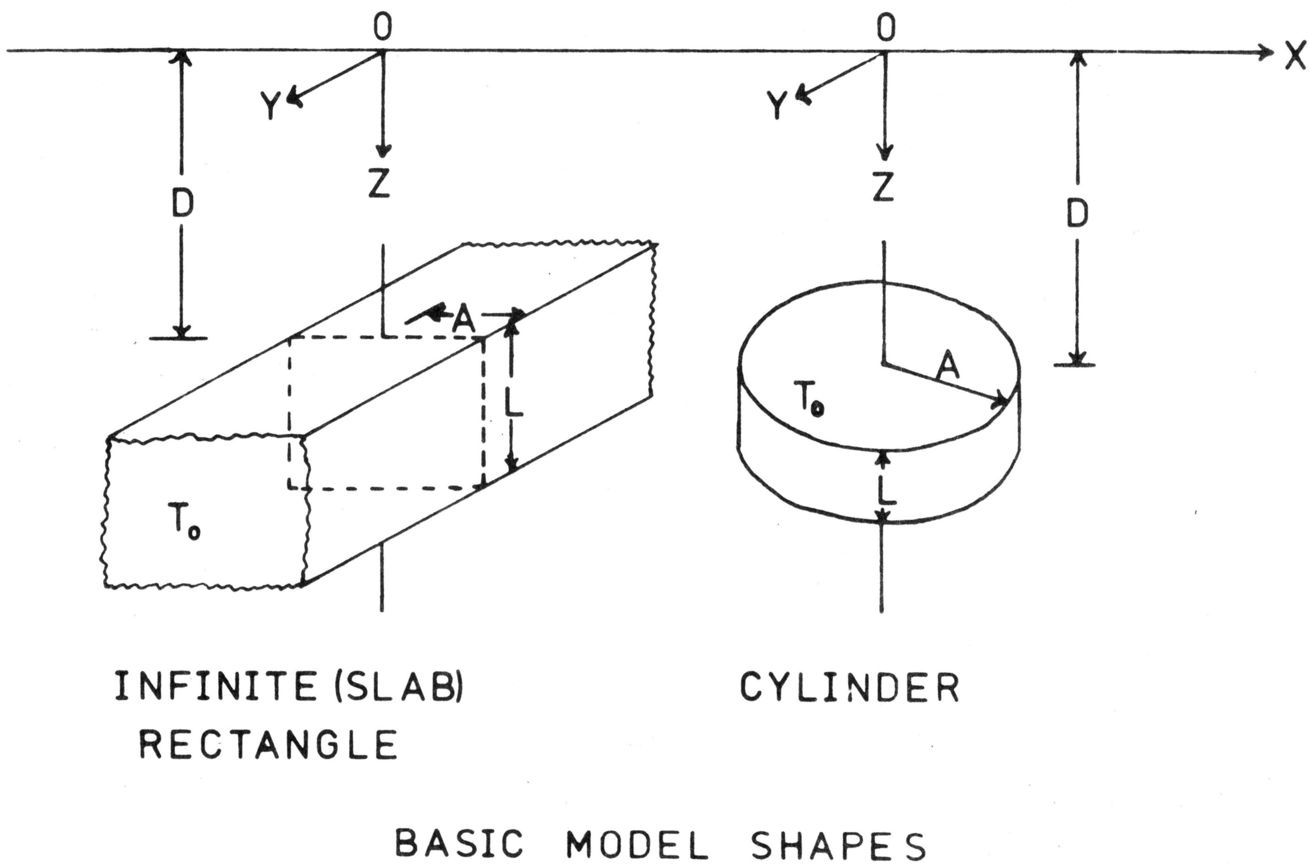


Figure 3. Basic model shapes for two-dimensional magma chamber models. A = half-width or radius; L = thickness; D = depth from surface to top of body; T_o = initial emplacement temperature.

rather than individual magma chambers associated with each stratocone volcano.

The other basic shape investigated is a cylinder that is considered to be buried at a depth D, to have a radius A, and a thickness L. This model would be more typical of a single magma chamber associated with a volcano, a volcanic pipe, or a very large circular magma chamber which might underlie several adjacent volcanoes. For example, it has been suggested by LaFehr (1965) that a single magma chamber may underlie the Medicine Lake Highland and Mt. Shasta area in California.

Comparing the two geometries, A is the half-width of the slab, or the radius of the cylinder; L is the thickness of the body; D is the depth of the top of the body from the surface; T_o is the initial temperature (assumed constant) of the body. In all the solutions, the medium is assumed to be homogeneous and to have constant thermal conductivity not dependent on temperature. In the solution, the magma is assumed to have the same thermal conductivity as the country rock. Intrinsically, although both are two-dimensional, a slab will have a larger heat content than a cylinder for equal values of A and D, and therefore will tend to cool more slowly.

In addition to the two types of geometry, two different models of magma chamber behaviors were calculated. The first of these is the instantaneous model discussed above, where emplacement of the magma at a given temperature takes place at a certain time and conductive cooling occurs subsequent to this emplacement. For the second set of models, the continuous models, the magma is assumed to be emplaced at a constant temperature, and the boundary of the magma chamber is assumed to remain at this temperature for subsequent time so that the whole half-space is heated by the magma chamber.

In certain of the cases, a plane boundary above the magma chamber has been assumed so that a half-space solution applies. In several of the solutions for the cylinder, the effect of topography has been included so that the effect of the topographic edifice of the volcano on top of the magma chamber can be examined. In these models, the results have been illustrated for certain characteristic locations along the surface such as the apex of the volcano (A), the center of the slope (S), the toe (T), and the plane surface (F), as illustrated in Figure 4.

Plane Surface Instantaneous Magma Chamber Models

The plane surface instantaneous models are the simplest of the solutions and in the rectangular case have been discussed by many authors (Carslaw and Jaeger, 1959; Simmons, 1967; Blackwell and Baag, 1974; Lachenbruch and others, 1976). Some results for a spherical source and for a cylindrical source have been discussed in the literature (Carslaw and Jaeger, 1959; Rititaki, 1959; Blackwell and Baag, 1974). The object of this set of models is to show the relationship between heat flow, depth of burial, and cooling time. All solutions in this section can be calculated analytically from the results given by Carslaw and Jaeger (1959).

Throughout all these models, for consistency, certain shape parameters have been assumed to be constant. These have been

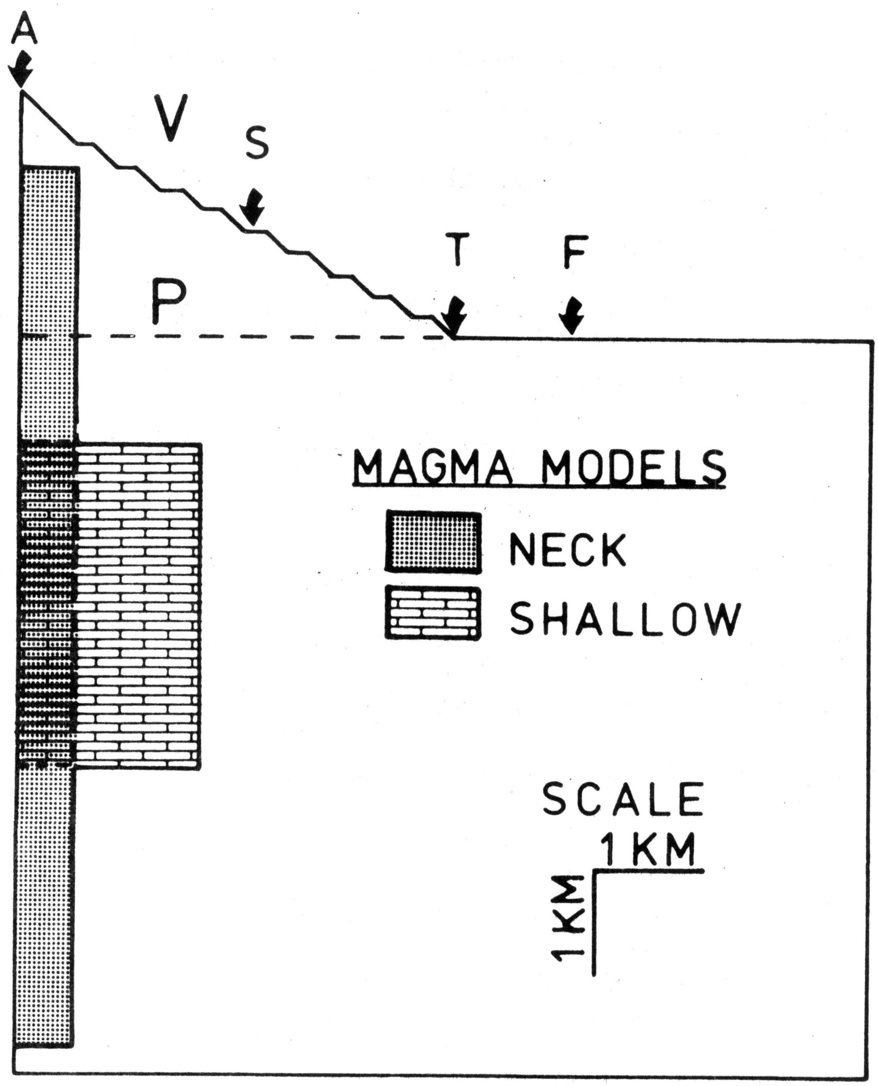


Figure 4. Magma chamber model sizes in relation to surface half-space. P = plane surface half-space boundary; V = volcano half-space boundary. Locations of detailed studies are indicated by A (apex), S (slope), T (toe), and F (flat) of the volcano.

given the dimensions of kilometers, but they can be scaled by the conduction length parameter (Lachenbruch and others, 1976). The assumed values of the parameters are shown in Table 1. Some sample results for the instantaneous models are shown in Figure 5. In all of these models the heat flow at the time of maximum surface heat flow over the apex of the magma chamber is illustrated. Results for both the rectangular and the cylindrical models are shown. The distance is measured from the center line or center point of the body.

The assumed emplacement temperature for these models is 800°C. This temperature is approximately typical of rhyolitic magmas. The actual temperature associated with an andesitic magma might be somewhat higher (1000°C), particularly if the effect of latent heat were included. It is a simple matter to scale the heat flow to any assumed T_o, however, as it is merely a matter of multiplying the heat flow shown on the ordinate by the ratio of the assumed T_o to 800°C.

Some points of interest from these solutions are that in the case of the neck model, where the magma chamber approaches within 100 m of the surface, there is a significant difference between the heat flow associated with the cylindrical neck model and the rectangular neck model. This difference is primarily because of the much larger volume associated with the rectangular chamber. It is also apparent that these anomalies are quite limited in lateral extent. The width of both bodies is assumed to be 500 m and their depth of burial is assumed to be 100 m. The time of maximum heat flow associated with the bodies is approximately 2,000 years in both cases.

TABLE 1

VALUES OF PARAMETERS FOR

THERMAL MODELS OF MAGMA CHAMBERS

		Magma Chamber Type		
		Neck	Shallow	Deep
Radius	(A)	0.25 km	1.5 km	25.0 km
Thickness	(L)	10.0 km	3.0 km	10.0 km
Depth	(D)	0.10 km	1.0 km	10.0 km

Thermal conductivity = 5 mcal/cm-sec-°C

Thermal diffusivity = 0.01 cm^2/sec

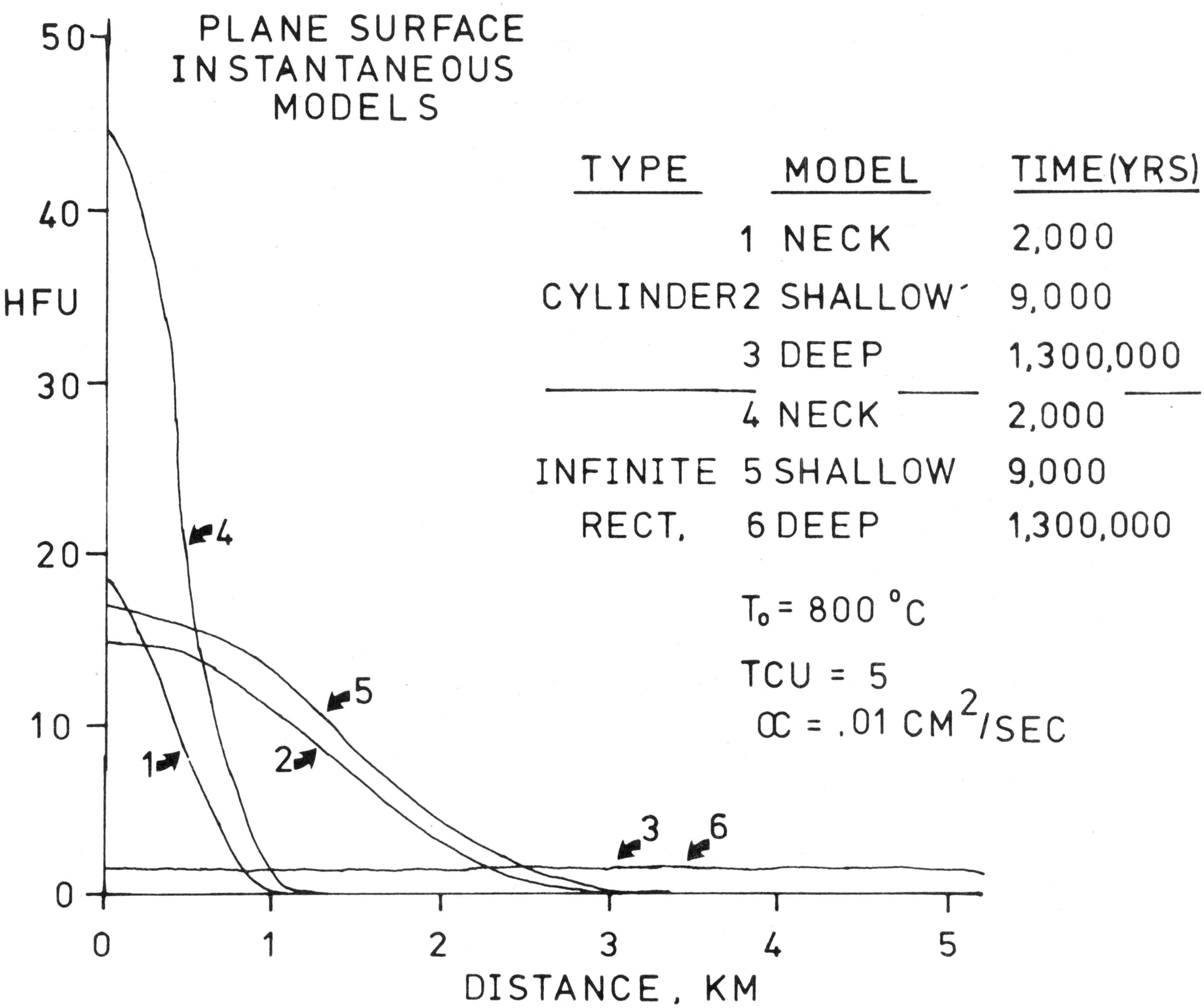


Figure 5. Surface heat flow-versus-distance curves for plane surface instantaneous models. Maximum surface heat flow over the apex of the volcano corresponds to time (yrs.) shown at right of figure; curves are labeled by number of respective models.

As the magma chamber becomes deeper and larger, the difference in heat flow, at the time of maximum heat flow, between the cylinder and the rectangular model chambers becomes less, until in the case of the largest magma chamber modeled (where the depth of burial is 10 km and the half-width, or radius, is 25 km) there is no significant difference between the two anomalies. Also illustrated is the dramatic effect of the depth of emplacement on the maximum heat flow value. Obviously, the cooling times are quite different; the cooling time for the large deep magma chamber is increased by a factor of over 100 compared to the shallow magma chamber, and over 500 compared to the neck model. In general, the cooling times are rather short geologically and, for deep chambers, the increase in heat flow is rather modest. For the rectangular plane surface model, a more complete set of heat flow-versus time curves are given by Lachenbruch and others (1976).

More Realistic Models

In order to obtain solutions for magma chamber models in addition to the simple analytical solutions available, a finite difference program was written for transient-radial heat conduction problems. Using this finite difference model, solutions were obtained that included a model surface shape based on the topographic profile of Mt. Hood (Figure 2 and Figure 4, boundary V). The heat flow effects were calculated at the surface for both instantaneous and continuous models for "neck" and "shallow" magma chambers under the apex of the volcano, of the characteristic dimensions shown in Table 1. A solution was calculated for both cylindrical and rectangular coordinate geometry volcanoes and magma chambers, although the rectangular results are not discussed here. These types of solutions are illustrated schematically in Figure 6 for the "shallow" magma chamber model. The boundary labeled V in Figure 4 was used with the magma chambers placed below V according to the values of parameters given in Table 1.

Because the deepest magma chamber (Table 1) generates a more regional anomaly, rather than a local anomaly, discussion is focused in this and subsequent sections of this chapter on the two smaller magma chamber models calculated. The "shallow" model might correspond to a single magma chamber associated with an individual andesite volcano. The second solution ("neck" model) might be more characteristic of conditions associated with emplacement of a narrow neck or central magma chamber along the conduit of a volcano.

Shallow Magma Chamber

The dimensions of the magma chamber assumed in this discussion are given in Table 1. All models are assumed to be cylindrical, except for model 2 in Figure 6 (the slab model). The two contrasting types of heat conduction solutions, instantaneous and continuous, are shown for the shallow magma chamber model with both the volcano-surface and the plane-surface topography. The results for heat flow-versus-distance at the time of maximum surface heat flow for the models are shown in Figure 6. In Figure 6, the "plane" and "volcano" surface models refer to solutions using a cylindrical magma chamber.

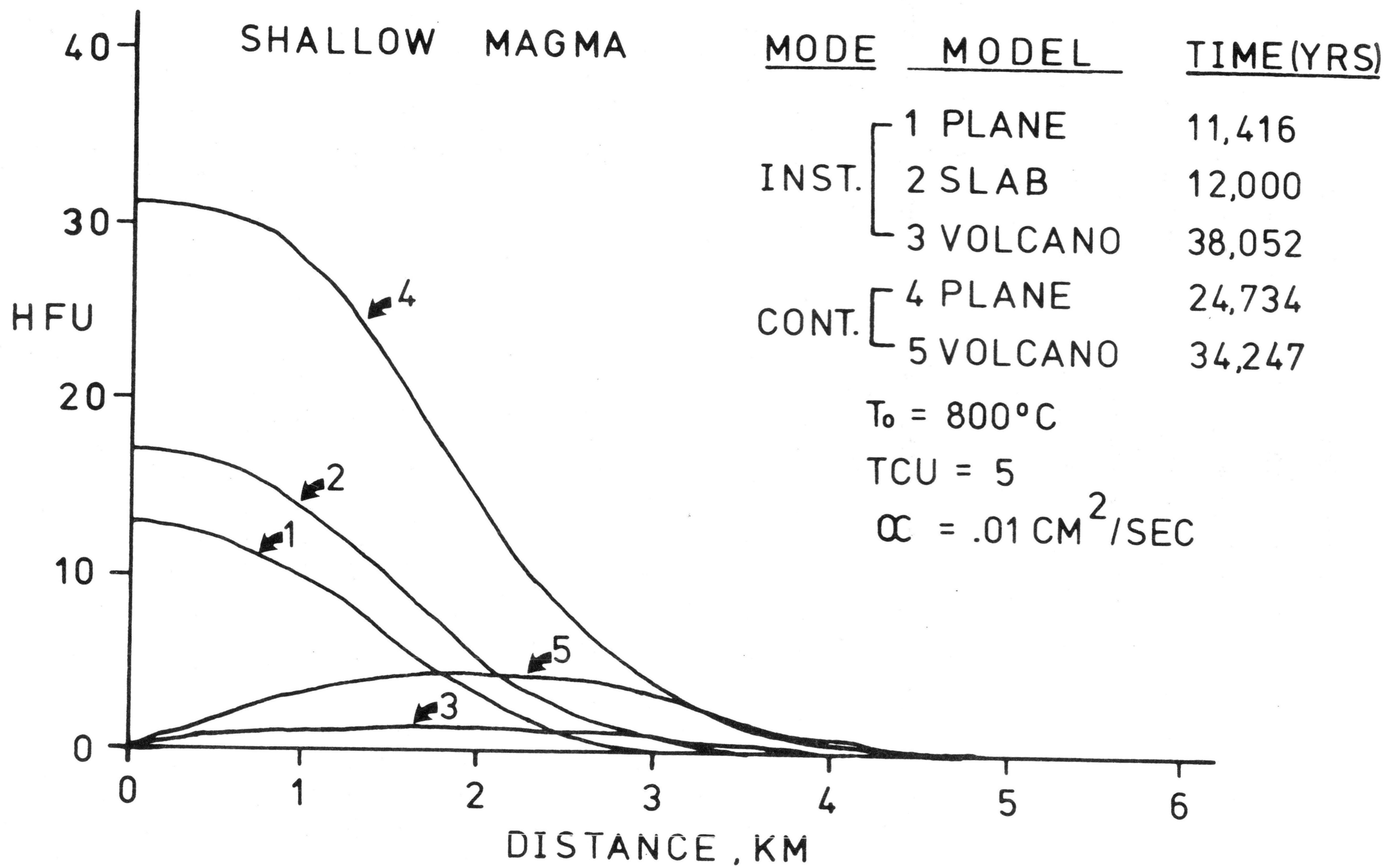


Figure 6. Surface heat flow-versus-distance curves for shallow magma chamber model. Maximum surface heat flow over the apex of the volcano corresponds to time (yrs) shown at right of figure; curves are labeled by number of respective models.

On the other hand, the slab model refers to a solution with a plane surface and an infinite rectangular magma chamber. The times indicated in the figure are the times of maximum surface heat flow associated with the instantaneous magma chamber and the maximum run time for the continuous magma chamber model. The horizontal distance shown in Figure 6 is measured from the apex of the volcano and/or the center of the magma chamber.

Curve 1 (Figure 6) is the maximum heat flow curve for the instantaneous plane magma chamber discussed previously, and curve 4 represents the heat flow for the continuous plane magma chamber at an age of 25,000 years. The continuous magma chamber has essentially reached its equilibrium heat flow in the center part of the anomaly, although the solution has still not reached equilibrium at distances of 2-4 km away from the center line. The maximum heat flow for the continuous (as opposed to the instantaneous) model differs by a factor of slightly over two. Similarly, the continuous plane solution has a much higher heat flow at any distance away from the magma chamber because of the continuous heat input of the replenished or convecting magma chamber. Thus, at any given period of time, the total heat output of the continuous magma chamber will exceed by a minimum of two times the maximum heat output of the instantaneous solution. Over the area of the magma chamber, this heat output might average a maximum of approximately 10 HFU for the instantaneous model, and 20-30 HFU for the continuous model illustrated.

The slab solution is essentially equivalent in dimensions to the instantaneous plane solution, except that the magma chamber is assumed to be infinite in the Y axis direction instead of radially symmetrical. As noted previously, the heat flow is greater for the rectangular model due to the greater volume of magma involved. However, the characteristic cooling time and temperatures are not appreciably different for models 1 and 2 (Figure 6).

Very different results are obtained for the heat flow associated with conductive cooling of a magma chamber beneath a volcanic edifice. In this particular solution (curves 3 and 5, Figure 6) the times of maximum heat flow are much delayed because of the greater average distance between the magma chamber and the surface. Furthermore, the mean heat flow is decreased because of the much larger surface area over which the heat is dissipated in the volcano model, as opposed to the plane surface magma chamber. For these models, the maximum heat flow is associated with the edge of the volcanic edifice. Even for the continuous magma chamber, heat flow values are relatively modest compared to those associated with the cylindrical magma chamber beneath the plane surface. Part of this effect is due to the topographic effects of the volcano, and part is due to the greater effective depth of burial. Of course, if the magma chamber were larger or shallower than that assumed, then higher heat flow values would be associated with the volcanic edifice than with the areas at the toe of the volcano. There are many different solutions that might be developed, and the object is discussion of only a few typical examples.

Heat flow-versus-time curves are shown in Figure 7 for

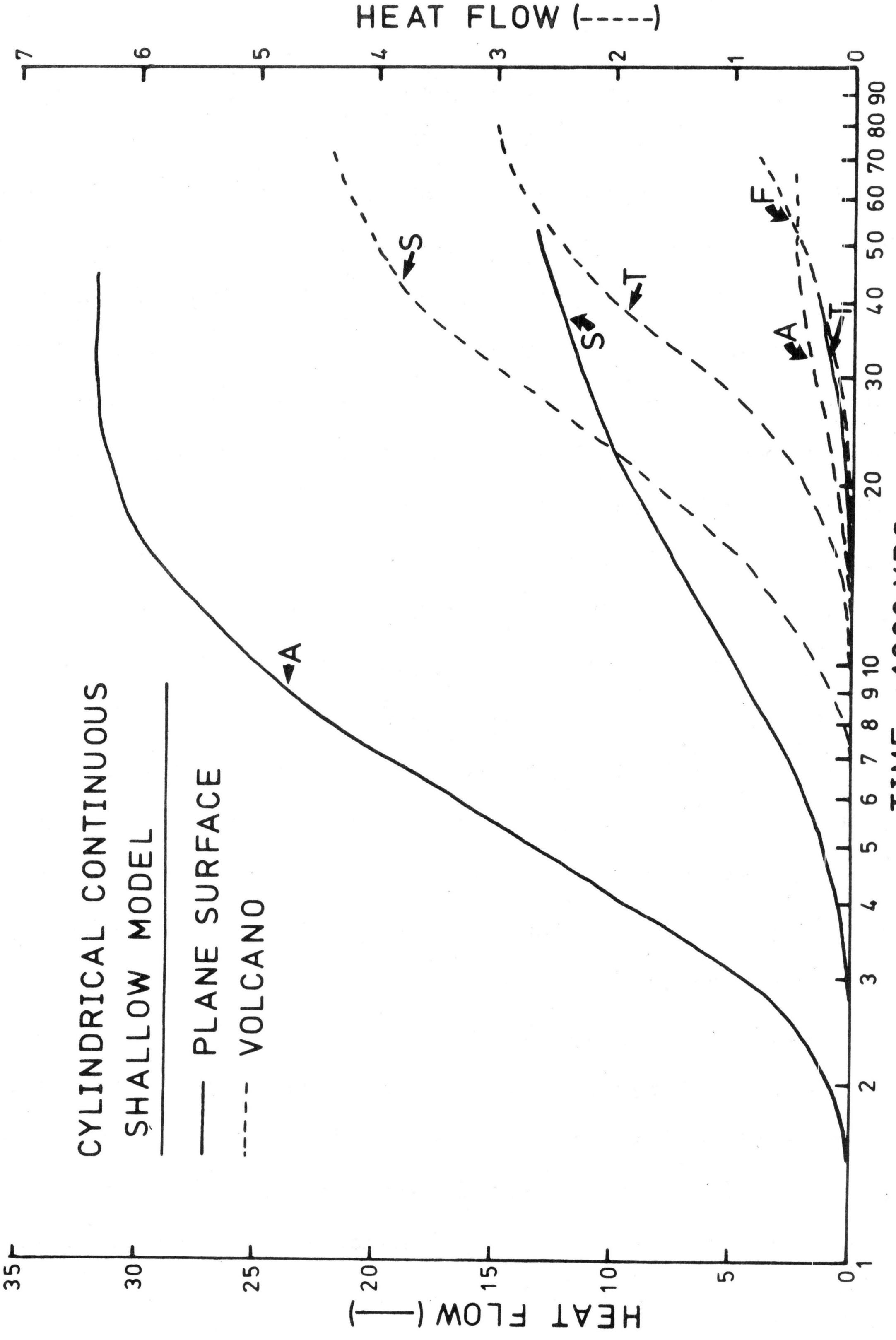


Figure 7. Heat flow-versus-time curves for the shallow cylindrical continuous model. The curves are shown for various positions along the surface of the volcano, keyed to the letters in Figure 4. The scale on the left side of the figure corresponds to the plane surface half-space, and the scale on the right side of the figure corresponds to the volcano surface half-space.

the continuous shallow cylindrical magma chamber model. Heat flow-versus-time curves are shown for various positions along the surface, keyed to the letters in Figure 4. In the case of the instantaneous model, the heat flow builds up rapidly and decays over a longer period of time. The times of maximum heat flow for the particular models described are given in Figure 6. In the case of the cylindrical continuous shallow magma chamber model (Figure 7), the heat flow builds rapidly over the first 10,000-50,000 years of the existence of the chamber to a maximum value (and constant heat flow) after the initial increase. The calculated heat flow continues to rise at larger distances from the volcano for 50,000-100,000 years, but the values are never particularly large.

These models indicate that if the heat transfer is primarily by conduction, then a period of some tens of thousands of years is required to heat up the volcano from a relatively local magma chamber below the volcano. The maximum heat flow values will be observed along the lower slopes of the volcano.

Figure 8 shows isotherms associated with the continuous and the instantaneous shallow cylindrical magma chambers after a period of approximately 38,000 years. The assumed thermal properties are shown in the figure. A comparison is made between the temperatures for the instantaneous cooling solution and the continuous solution. No regional background gradient has been added to these temperature data, so the actual temperature would be slightly higher in proportion to the depth and the background geothermal gradient. After this period of time, temperatures in the volcanic edifice have essentially

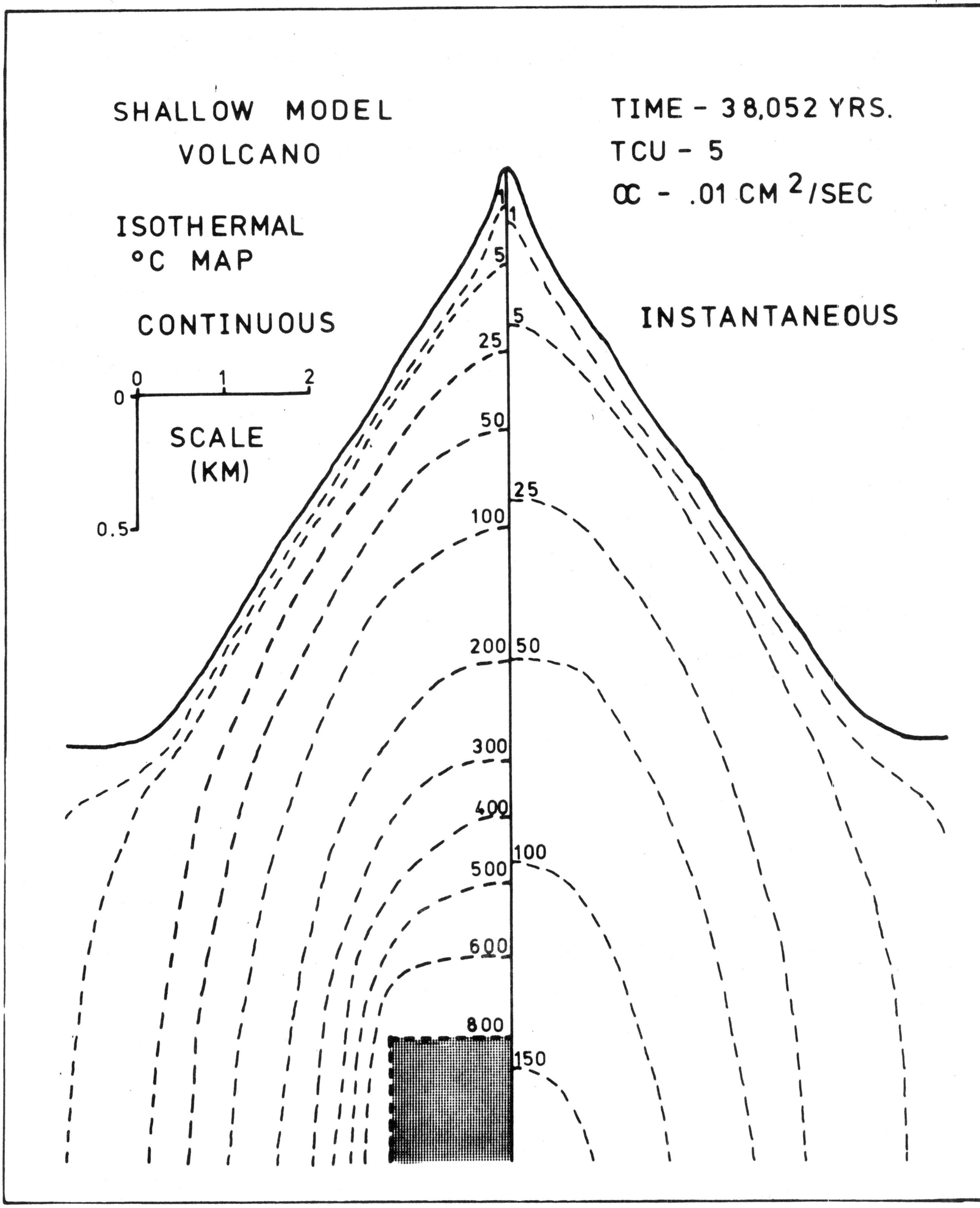


Figure 8. Isothermal cross-section map of the volcanic edifice for the shallow magma chamber model. The continuous model isotherms are on the left side of the volcano apex, and the instantaneous model isotherms are on the right side of the volcano apex. A portion of the shallow magma chamber is indicated by the shaded area at the bottom of the figure.

reached a steady-state for the continuous model; and for the instantaneous model, temperatures are well on their way to complete cooling, as indicated by the low temperatures.

Continuous Neck Magma Chamber Model

The final configuration discussed is the solution for a small neck type of magma chamber associated with a plane surface or with a volcano. The heat flow as a function of time at characteristic positions along the surface is shown in Figure 9. The solid lines are the heat flow values associated with the neck beneath a plane surface, while the dotted lines are the heat flow values associated with emplacement of the neck within a volcanic edifice (note difference in scales).

The spreading effect of the volcano surface is again apparent in the heat flow, as indicated by the factor of five difference in heat flow values associated with the magma chamber beneath the plane surface and the magma chamber beneath the volcano. Of course, the time to reach steady-state is relatively short for the magma chamber within the volcano, because of proximity to the surface. On the other hand, it takes some time for the effect to build up along the slope of the volcano. The heat flow never reaches significant values at the edge of the volcano. The instantaneous cooling of the plane surface and volcano magma chamber is also shown. In both cases, the heat flow anomaly has decayed significantly in a period of 2,000-3,000 years after emplacement. Significant heat flow is not seen along the slopes and flanks of the volcano.

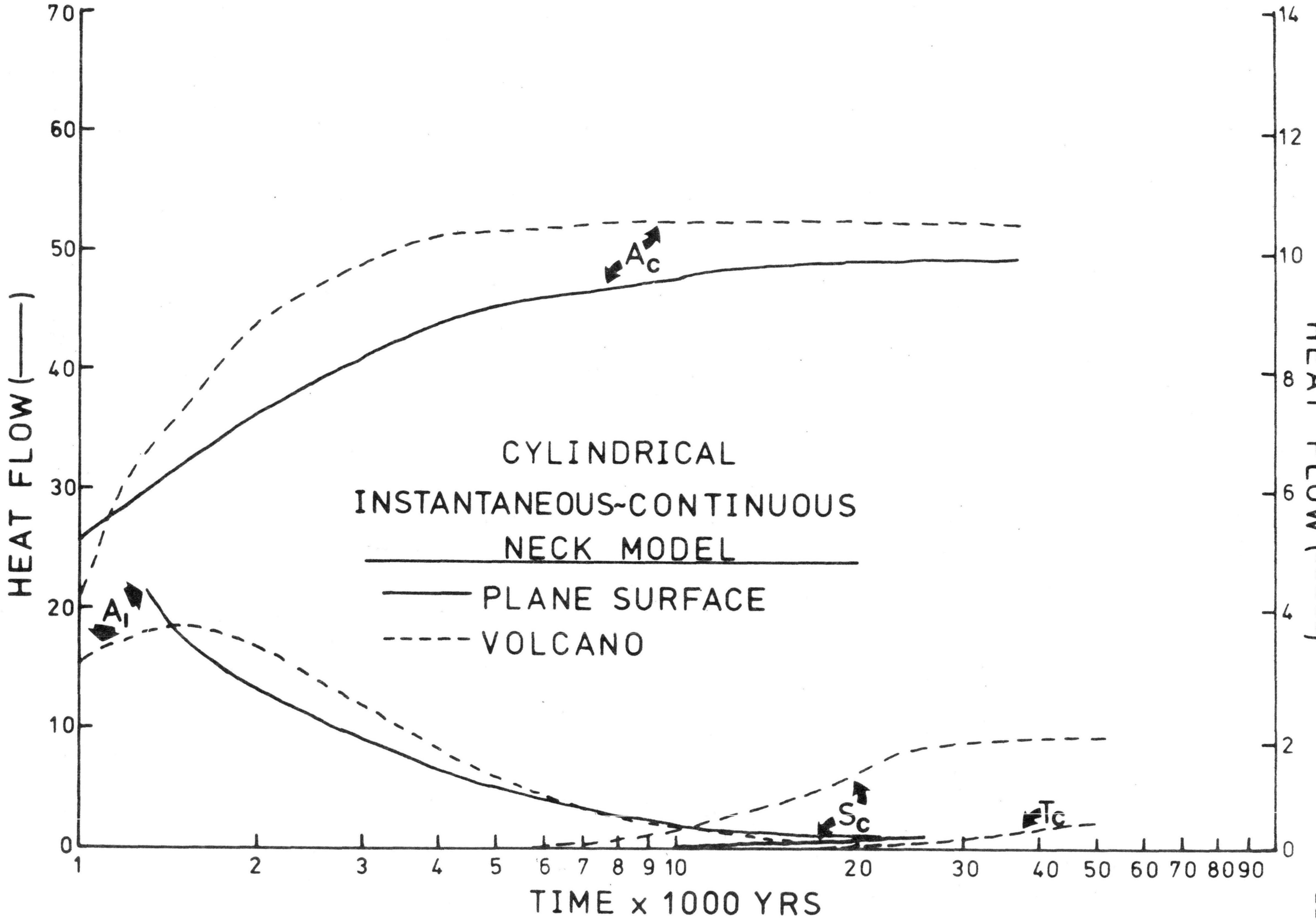


Figure 9. Heat flow-versus-time curves for the cylindrical instantaneous and continuous neck models. The curves are shown for various positions along the surface of the volcano, keyed to the letters in Figure 4. The scale on the left side of the figure corresponds to the plane surface half-space, and the scale on the right side of the figure corresponds to the volcano surface half-space. The subscript letters on the letters labeling the curves indicate the instantaneous (I) or the continuous (C) solutions.

The continuous neck model is reasonable only if the rate of occurrence of volcanism is rapid enough to cycle magma along the neck and thus to maintain a high temperature. Based on these solutions, it would appear that the recurrence of extrusive events would have to be less than 1,000-2,000 years if the axis of the volcano is to be kept at near-magma temperatures.

Discussion

The magma chamber models herein presented are highly idealized and the purpose of the following discussion is to briefly summarize their applications to the problem of geothermal resources associated with a stratocone volcano. In order to have a commercially attractive geothermal prospect, two things are needed: (1) high temperature, and (2) fuild flow sufficient to allow generation of economic amounts of power. The minimum heat loss that might be associated with an economic geothermal system is somewhere between 5×10^6 and 10×10^6 cal/sec. For a volcano such as Mt. Hood, with a radius of approximately 5 km and a total basal area of approximately 50 km^2, a mean heat flow, or thermal extraction rate, of 10-20 HFU would be required to maintain such a geothermal system. Furthermore, as discussed above, the rates of cooling of magma chambers associated with hydrothermal convection are probably greater by a factor of two to ten times than those associated with conductive cooling. Thus it is obvious that a single batch of magma, which might be approximated by the instantaneous conductive cooling model, would not be sufficient to generate a very long-lived hydrothermal system, and the cooling of a magma chamber associated with a stratocone volcano (and of the same order of dimensions) would take place in less than a few thousand years.

It might appear that the application of the plane models to the stratocone volcano is somewhat questionable. However, one approach to modeling the cooling of a magma chamber associated with a stratocone volcano might be to assume that, for thermal purposes, the actual edifice of the volcano does not exist. Typically, the volcano edifice is composed of an alternating series of ashes and blocky flows, with a large horizontal permeability. Furthermore, these volcanoes are typically large topographic features and thus are subject to relatively high rainfall. Large amounts of meteoric water which are transferred into the subsurface along porous units tend to flow out at the edge of the volcano. Water flow may be so rapid as to prevent any appreciable heating of the volcanic edifice, or to confine heating to areas which have been hydrothermally altered where permeability has been significantly decreased. In a simple model, where the fluid flow is very high, the bottom of the aquifer could be treated as an upper boundary condition of constant temperature in the conductive solution, instead of treating the volcano surface itself as a constant temperature boundary. Therefore, the heat flow for the plane boundary model might approximate the amount of heat input from a magma chamber beneath the volcano into the circulating groundwater system.

It would appear, then, that if significant geothermal systems are associated with andesite volcanoes, then either larger instantaneous magma chambers than those modeled are required, or volcanism must recur over a considerable period of time, so that the continuous solutions give a better estimate of potential heat transfer into the hydrothermal system.

While a neck type of magma chamber is capable of transmitting a large amount of heat into the volcanic edifice, it is relatively short-lived and a recurrence time of a few hundred years would be required for enough heat input from a neck type magma chamber to drive a hydrothermal system of any significance within the edifice of the volcano.

A somewhat more deeply buried magma chamber typical of the shallow magma chamber model could, if replenished over periods of a few thousands of years, easily impart the amount of heat required to drive significant geothermal systems. The presence of the volcano surface, as well as the probable flux of groundwater associated with the topographic features of the volcano, serve to indicate that whatever the shape of a magma chamber beneath a stratocone volcano, the maximum heat loss will probably be located along the lower slope of the volcano. Unless a magma chamber actually penetrates into the volcano itself, it is unlikely that significantly high temperatures will be associated with the upper part of the edifice. The general locations of thermal manifestations associated with stratocone volcanoes are usually near the outer edges of the volcanic carapace. The localization of the manifestations may be controlled by several different mechanisms, such as ring structures, etc. However, it appears that both the hydrologic and the conductive heat loss models are consistent with such a localization, except in cases where the volcano has been active enough to thoroughly heat up its cone.

REGIONAL HEAT FLOW IN THE NORTHERN CASCADE RANGE OF OREGON

Regional Heat Flow Setting

The heat flow associated with the stratocone volcanoes in the Cascade Range can only be understood with respect to the regional background setting of heat flow in and around the Cascade Range itself. For several years DOGAMI and SMU have been involved in a systematic program of exploring the geothermal character of the Northern Cascade Range in Oregon. The studies have included location and logging of free holes in the Willamette Valley-Western Cascade Range areas during 1976; a drilling program involving 15 holes in the Western Cascade and High Cascade Range provinces during 1976; a free hole study of heat flow along the eastern margin of the High Cascade Range and the western portion of the Deschutes-Umatilla Plateau province in 1977; and regional investigation of heat flow around Mt. Hood in 1978. In addition, during 1977-79, a heat flow evaluation of deep holes drilled in the vicinity of, and on, Mt. Hood has been part of the program. In this section regional heat flow in the Northern Oregon Cascade Range, exclusive of the Mt. Hood region, will be discussed. Most of these data were obtained during 1976-77, but have not been summarized in any readily available report. Qualitative results have been discussed by Blackwell and others (1978).

A generalized heat flow map of Oregon based primarily on the data by Blackwell and others (1978), and updated using the data from Table 2, is shown in Figure 10. All of the heat flow

TABLE 2. Thermal and location data in the Northern Cascade Range of Oregon.

Location	N Latitude	W Longitude	Collar Elevation	Depth Interval	Avg. Thermal Conductivity [standard error]	N†	Corrected Gradient	Corrected Heat Flow	Quality
3N/12E-32cd	45°41.8'	121°20.7'	53.20 m	20.0 - 67.5	3.8		46.0	1.7	B
2N/12E- 6bb	45°41.4'	121°22.6'	85.10	10.0 - 80.0	3.8		28.0	1.1	B
2N/12E-16cba	45°39.5'	121°19.8'	380.00	100.0 - 185.0	3.8		34.0	1.3	A
2N/11E-20ab	45°38.8'	121°28.1'	629.30	60.0 - 120.0	2.8		37.0	1.0	B
2N/ 9E-29ad	45°37.8'	121°42.8'	1146.00		4.15 [0.92]	3		**	
2N/12E-30ad	45°37.7'	121°21.4'	457.30	20.0 - 120.0	3.8		39.0	1.5	A
2N/ 1W-32bb	45°37.2'	122°50.6'	282.70	0.0 - 200.0	3.8		25.0-32.0	1.1	B
2N/ 7E-31bd	45°36.8'	121°59.8'	536.70	50.0 - 150.0	3.79 [0.14]	5	42.2	1.6	A
2N/13E-31cd	45°36.7'	121°14.7'	279.70	90.0 - 175.0	3.8		52.0	2.0	B
1N/ 9E- 1ac	45°36.2'	121°38.1'	267.50	0.0 - 175.0	3.8		64.0	2.4	C
1N/ 2E-24da	45°33.3'	122°30.0'	4.60	60.0 - 95.0	4.8		27.0	1.3	B
1N/ 1W-25bc	45°32.6'	122°45.6'	326.80	70.0 - 150.0	3.8		23.0	0.9	B
1N/ 2E-29da	45°32.4'	122°34.9'	62.30	160.0 - 230.0	3.0	6	32.1	1.0	A
1N/ 3E-33ad	45°31.7'	122°26.0'	60.50	130.0 - 330.0	3.0	1	37.3	1.1	A
1N/ 6E-31cd	45°31.2'	122° 7.0'	743.70					**	
1S/10E- 9bc	45°29.8'	121°33.8'	560.00	10.0 - 25.0	3.70 [0.27]	5	256.0	9.5	C
1S/13E-20da	45°28.1'	121°11.8'	354.00	50.0 - 135.0	3.50		41.9	1.4	B
1S/10E-29ca	45°27.1'	121°34.8'	725.40		4.38 [0.20]	5			
2S/15E- 3bb	45°25.9'	120°55.5'	793.40	10.0 - 55.0	3.8		39.0	1.5	B

Location	N Latitude	W Longitude	Collar Elevation	Depth Interval	Avg. Thermal Conductivity [standard error]	N	Corrected Gradient	Corrected Heat Flow	Quality
2S/11E- 6aad	45°25.7'	121°27.7'	1161.00 m	95.0 - 150.0	3.80 [0.15]	5	30.4	1.15	A
2S/ 4E-18dda	45°23.5'	122°21.4'	196.60	50.0 - 225.0	2.8		34.0	0.95	A
2S/ 8E-15cd	45°23.5'	121°48.5'	777.40	1000.0 -1200.0	4.19 [0.14]	21	55.0	2.3	A
2S/ 8E-17cc	45°23.3'	121°51.6'	658.40	100.0 - 200.0	4.04 [0.32]	5	51.2	2.07	A
2S/ 4E-16cd	45°23.4'	122°19.3'	220.40	20.0 - 75.0	3.0		39.0-42.0	1.2	B
2S/ 2E-20bd	45°22.8'	122°35.4'	18.30	45.0 - 75.0	3.8		21.0	0.8	B
2S/ 6E-24ca	45°22.8'	122° 1.1'	359.70	70.0 - 150.0	4.33 [0.21]		42.7	1.85	A
2S/ 7E-34bb	45°21.5'	121°56.1'	442.10	35.0 - 85.0	4.30 [0.80]		47.0	2.0	A
2S/13E-36cd	45°20.9'	121° 8.0'	753.90	10.0 - 45.0	3.8		35.0	1.3	C
3S/ 6E- 3a	45°20.6'	121°55.4'	148.60					**	
3S/11E- 1aa	45°20.6'	121°21.6'	919.90	50.0 - 125.0	3.60 [0.30]	3	43.8	1.58	A
3S/ 9E- 6dd	45°19.9'	121°42.5'	1798.80		4.42 [0.16]	4		**	
3S/ 9E- 7ab	45°19.7'	121°42.6'	1761.70					**	
3S/14E- 7dc	45°19.0'	121° 5.9'	832.40	10.0 - 65.0	3.8		26.0	1.0	B
3S/ 8E-16cd	45°18.4'	121°49.9'	762.20	50.0 - 120.0	5.22 [0.15]	9	24.0	1.25	A
3S/ 8E-24bbd	45°18.1'	121°46.5'	1106.60	10.0 - 60.0	5.20		60.0	3.1	C
3S/8½E-25aa	45°17.2'	121°43.7'	1167.70		4.53 [0.26]	6		**	

Location	N Latitude	W Longitude	Collar Elevation	Depth Interval	Avg. Thermal Conductivity [standard error]	N	Corrected Gradient	Corrected Heat Flow	Quality
3S/ 8E-29ddc	45°16.6'	121°54.5'	722.40 m	80.0 - 140.0	6.32 [0.44]	5	28.2	1.78	C
4S/13E- 1ca	45°14.9'	121° 7.4'	340.50	10.0 - 75.0	2.8		77.0	2.2	B
4S/12E-10dd	45°13.8'	121"16.7'	530.50	20.0 - 90.0	3.8		36.0	1.4	B
4S/ 9E-28dd	45°11.3'	121°40.1'	1036.00					**	
4S/13E-32dc	45°10.5'	121°12.2'	547.20	20.0 - 145.0	3.0		60.0	1.8	B
4S/14E-33cb	45°10.5'	121° 4.1'	313.10	50.0 - 70.0	3.8		68.0	2.6	B
5S/16E-20cb	45° 7.1'	120°50.9'	755.40	10.0 - 80.0	3.8		34.0	1.3	B
5S/12E-31aa	45° 6.0'	121°20.5'	677.90	20.0 - 107.5	3.8		38.0	1.4	C
6S/15E- 5ba	45° 5.0'	120°57.9'	802.60	35.0 - 70.0	3.8		44.0	1.7	B
6S/ 7E- 4dc	45° 4.3'	121°57.6'	686.00					**	
6S/ 1E-13da	45° 2.9'	122°37.2'	326.10	95.0 - 140.0	3.9	1	39.0	1.4	A
6S/14E-13dd	45° 2.6'	120°59.8'	940.90	80.0 - 120.0	3.8		45.0	1.7	B
6S/ 7E-21cd	45° 1.8'	121°57.7'	603.70	10.0 - 40.0	3.52 [0.17]	4	168.1	5.91	C
6S/ 7E-30bb	45° 1.3'	122° 0.5'	512.10	50.0 - 130.0	3.95 [0.19]	9	203.5	8.04	C
6S/ 6E-34cd	44°60.0'	122°3.8'	487.80	10.0 - 50.0	3.91 [0.09]	9	65.7	2.57	A
7S/ 1E-11aca	44°58.8'	122°38.8'	214.00	0.0 -2379.0	3.50		26.0	0.9	B
7S/ 5E-22aa	44°57.1'	122°10.4'	655.50	20.0 - 90.0	3.48 [0.12]	7	66.1	2.3	A
8S/ 1E- 8db	44°53.3'	122°42.5'	303.30	95.0 - 215.0	4.1	1	28.0	1.1	A
8S/ 1E-17da	44°52.2'	122°42.3'	315.50	105.0 - 110.0	3.8		25.0	1.0	B
8S/ 2W-24bc	44°51.7'	122°53.0'	127.10	22.5 - 60.0	3.8		25.0	0.95	B

Location	N Latitude	W Longitude	Collar Elevation	Depth Interval	Avg. Thermal Conductivity [standard error]	N	Corrected Gradient	Corrected Heat Flow	Quality
8S/ 1W-32bb	44°50.3'	122°50.4'	115.80	30.0 - 80.0	3.8	1	34.0	1.3	A
8S/ 5E-31cc	44°49.9'	122°14.8'	705.30	35.0 - 345.0	4.30 [0.80]		28.0	1.2	A
9S/17E- 6cad	44°48.7'	120°44.0'	987.80	45.0 - 120.0	4.5		37.0	1.7	A
9S/ 3E-11ba	44°48.5'	122°24.5'	317.00	47.5 - 85.0	3.2	1	25.4	0.7	A
9S/ 3E-11cb	44°48.1'	122°24.8'	333.80	25.0 - 60.0	3.2	2	24.3	0.8	A
9S/ 6E-23bb	44°47.0'	122° 2.4'	550.00	30.0 - 105.0	3.84 [0.27]	7	54.1	2.08	A
9S/ 7E-21ad	44°46.7'	121°57.1'	725.40	70.0 - 150.0	3.67 [0.23]	12	81.5	2.38	A
9S/ 2E-21da	44°46.2'	122°33.6'	213.40	22.5 - 47.5	3.0	1	41.0	1.2	B
11S/ 1E- 7da	44°37.5'	122°43.3'	158.50	40.0 - 57.5	3.2	1	25.0	0.8	B
11S/ 1W-14dd	44°36.5'	122°45.9'	182.90	30.0 - 125.0	3.2	1	41.0	1.3	A
11S/15E-22cd	44°35.6'	120°55.1'	963.40	605.0 - 820.0	6.5	1	31.0	2.0	A
11S/ 1W-32bbb	44°34.6'	122°50.6'	108.20	0.0 -1345.0	3.8		19.0	0.7	B
12S/ 1W- 4dc	44°33.1'	122°48.7'	135.00	30.0 - 65.0	3.20 [0.30]	3	36.4	1.2	A
13S/ 2W- 3aa	44°28.3'	122°54.4'	317.00	40.0 - 95.0	3.2	1	31.0	1.0	A
13S/ 1W- 8db	44°27.1'	122°49.9'	329.20	15.0 - 195.0	3.8	1	21.0	0.8	A
13S/ 1W-10ca	44°27.1'	122°47.7'	149.40	27.5 - 62.5	3.2	1	23.0	0.75	B
13S/ 2W-18cb	44°26.2'	122°59.1'	378.00	95.0 - 190.0	3.8		26.0	1.0	B
13S/ 1E-20ba	44°25.9'	122°43.0'	402.90	90.0 - 130.0	3.2	1	40.0	1.3	B
13S/ 1E-35ab	44°24.1'	122°38.9'	310.90	90.0 - 150.0	3.2	1	40.1	1.3	A
14S/ 3W-24dc	44°20.0'	122°59.6'	207.30	10.0 - 47.5	3.2	1	25.0	0.8	B
15S/ 6E-11dc	44°16.1'	122° 3.2'	716.50					**	

Location	N Latitude	W Longitude	Collar Elevation	Depth Interval	Avg. Thermal Conductivity [standard error]	N	Corrected Gradient	Corrected Heat Flow	Quality
15S/ 7E-28aa	44°14.8'	121°58.4'	1143.30 m					**	
15S/15E-30ad	44°14.8'	120°58.0'	1002.80	40.0 - 65.0	>3		96.4	>2.9	B
15S/14E-36ac	44°13.2'	120°59.6'	1023.00	20.0 - 75.0	>3		123.5	>3.7	C
16S/ 6E- 2ca	44°12.1'	122° 3.0'	70.10	100.0 - 150.0	4.15 [0.06]	11	81.7	3.39	A
16S/14E-16daa	44°11.2'	121° 2.8'	1024.00	15.0 - 75.0	>3		181.8	>5.5	C
16S/ 4E-14dbb	44°10.1'	122°17.5'	457.20	12.5 - 45.0	4.30 [0.80]	2	38.0	1.6	B
16S/ 6E-27bb	44° 9.1'	122° 4.7'	573.00	30.0 - 150.0	3.75 [0.12]	12	73.8	2.77	A
17S/ 1W-26da	44° 3.7'	122°47.1'	327.70	20.0 - 145.0	3.8		27.3	1.0	A
17S/ 2W-36ca	44° 2.8'	122°52.7'	213.40	40.0 - 105.0	3.2	1	25.0	0.8	A
* 18S/12E- 5bbd	44° 2.8'	121°19.1'	1102.00					**	
18S/ 2W- 4ad	44° 2.2'	122°55.8'	175.30	45.0 - 125.0	3.2	1	33.0	1.1	A
* 18S/11E-25bd	43°59.4'	121°21.4'	1195.00					**	
18S/ 1W-32cc	43°57.3'	122°50.4'	192.10	70.0 - 215.0	3.8		37.0	1.4	A
19S/ 2W- 2ac	43°56.9'	122°53.7'	231.70	25.0 - 120.0	3.2	1	32.0	1.0	A
19S/ 2W-10ad	43°55.9'	122°54.6'	218.00	10.0 - 42.5	3.2		30.6	1.0	A
* 19S/16E-16dc	43°54.9'	120°49.1'	1376.00	25.0 - 300.0	3.6		51.1	1.8	B
* 19S/11E-25ba	43°54.4'	121°21.5'	1373.00					**	
* 20S/14E-25aa	43°48.8'	120°59.4'	1428.00	45.0 - 125.0	<4.4	4	34.4	1.5	B
20S/ 3E-26da	43°48.0'	122°25.0'	719.50	70.0 - 140.0	3.8		38.0	1.4	B
20S/ 3E-26cd	43°47.9'	122°25.2'	707.30	10.0 - 125.0	3.8		29.0	1.1	B
* 21S/17E- 1ad	43°47.0'	120°36.9'	1440.00	40.0 - 90.0	2.4	4	82.5	2.0	B

Location	N Latitude	W Longitude	Collar Elevation	Depth Interval	Avg. Thermal Conductivity [standard error]	N	Corrected Gradient	Corrected Heat Flow	Quality
21S/ 3E-10ad	43°45.6'	122°25.9'	548.80 m	22.5 - 100.0	7.6	1	35.0	1.3	B
* 21S/15E-16ab	43°45.2'	120°56.2'	1476.00	70.0 - 150.0	4.2	6	55.0	2.3	C
* 21S/11E-25bb	43°43.9'	121°21.7'	1515.00	27.5 - 35.0	3.6	2	65.3	2.4	C
21S/ 4E-28ad	43°43.2'	122°20.0'	533.50	10.0 - 150.0	4.54	13	58.0	3.0	A
* 22S/19E- 5cc	43°41.4'	120°28.7'	1450.00	10.0 - 37.5	2.4	1	83.0	2.0	B
22S/ 3E-10cd	43°40.3'	122°27.0'	490.70	20.0 - 90.0	3.76		39.4	1.48	B
22S/ 5E-26bc	43°38.2'	122°11.3'	975.40	30.0 - 150.0	4.72 [0.14]	13	53.0	2.7	A
* 22S/19E-32ad	43°37.6'	120°27.4'	1520.00	42.5 - 47.5	1.60	1	118.0	1.9	B
* 23S/19E- 5b	43°37.0'	120°28.4'	1550.00	70.0 - 150.0	3.90	1	50.9	2.0	A

† N is number of samples used to determine the average thermal conductivity. Where N is blank, value used is average estimate of type rock. Where standard error is blank when a value is given for N, this indicates value used is estimate from composite samples from drill hole.

* Hull *et al.*, 1977.

** Considered unsuitable for heat flow calculations.

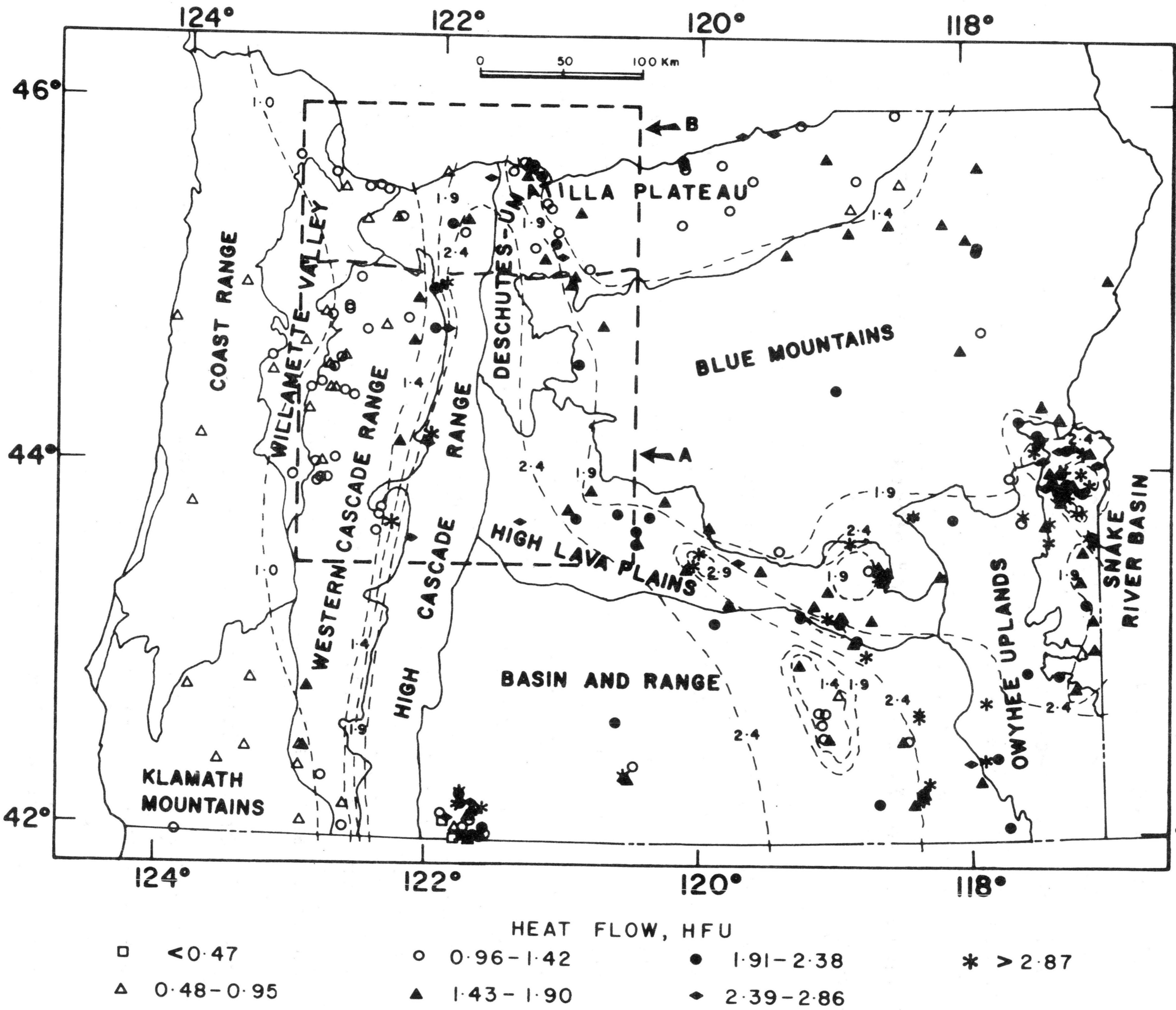


Figure 10. Generalized heat flow map of Oregon.

data discussed in this report are listed in Table 2. Included in this Table are all of the pertinent thermal and location data including latitude and longitude, township and range, elevation, interval of gradient calculation, average thermal conductivity, geothermal gradient and heat flow. All appropriate reductions have been performed to the data set, including terrain corrections. The data values are ranked by quality as discussed by Blackwell and others (1978), and Sass and others (1971).

It is apparent from Figure 10 that the northern part of the High Cascade Range in Oregon represents an area of higher heat flow than the Willamette Valley-Western Cascade Range provinces to the west and the Deschutes-Umatilla-Blue Mountains provinces to the east. The boundary between the High Cascade Range heat flow provinces is well-defined on the basis of the available data. The boundary between the High Cascade Range and the Deschutes-Umatilla-Blue Mountain provinces does not appear to be as well defined, and additional data are needed in order to completely delineate the location of the heat flow boundary and the magnitude of the heat flow contrast.

Northern Oregon Cascade Range Heat Flow

The heat flow data in the vicinity of the northern Cascade Range of Oregon are shown in greater detail in Figure 11. Available heat flow data from southern Washington (Schuster and others, 1978; Blackwell, unpublished) are also shown. Only the data south of Mt. Hood are discussed in the section (Box A, Figure 10).

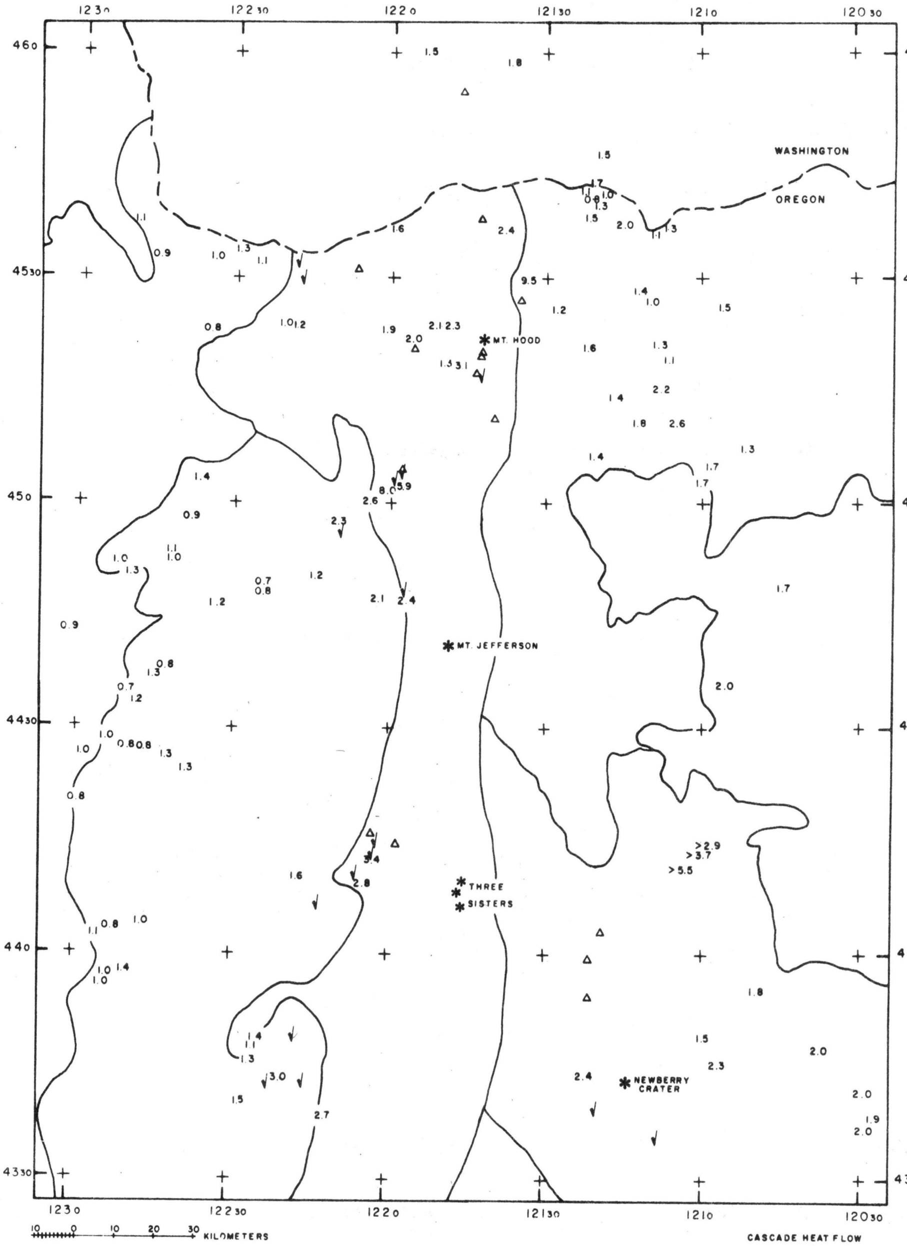


Figure 11. Heat flow map of the Northern Cascade Range area of Oregon (detail of area enclosed by heavy dashed lines in Figure 10). Heat flow values (in HFU) are plotted over their locations, with the decimal point of each value being the actual hole location. Holes drilled for heat flow studies, but considered unsuitable for heat flow calculations, are indicated by open triangles. Wells logged which are associated with regional aquifer disturbances are also shown by open triangles. Physiographic province boundaries are shown by solid lines. Locations of major volcanoes are indicated by the asterisks, and locations of major hot springs are indicated by √s.

The heat flow data in the Western Cascade Range west of the high heat flow boundary are very homogeneous. A histogram of the heat flow data from Figure 11 in the Western Cascade Range and Willamette Valley provinces is shown in Figure 12. Also shown in this figure is a histogram of the heat flow associated with the Western Cascade Range-High Cascade Range boundary south of 45°10'N as well as a histogram of heat flow values along the eastern boundary of the High Cascade Range.

The heat flow data in the Western Cascade Range-Willamette Valley average 1.1 HFU. Heat flow values along the Western Cascade Range-High Cascade Range boundary average 2.5 HFU, while heat flow values along the eastern boundary of the High Cascade Range average approximately 2.0 HFU, and the average has a large standard error. The average gradients associated with these heat flow values are 30°C/km, 60°C/km and 50°C/km respectively.

Typical temperature-depth curves observed in domestic water-supply wells in the Western Cascade Range-Willamette Valley provinces are shown in Figure 13. The holes are generally along major drainages where development is taking place. In general, the heat flow regime is completely conductive within the bedrock units of these two provinces with only occasional evidence of local water circulation and no evidence of regional water flow. The basic-to-intermediate composition volcanic and volcanoclastic rocks typical of these provinces seem to be pervasively altered to clay and zeolite minerals, resulting in relatively impermeable rocks. The staircase temperature-depth patterns, typical of water flow within drill

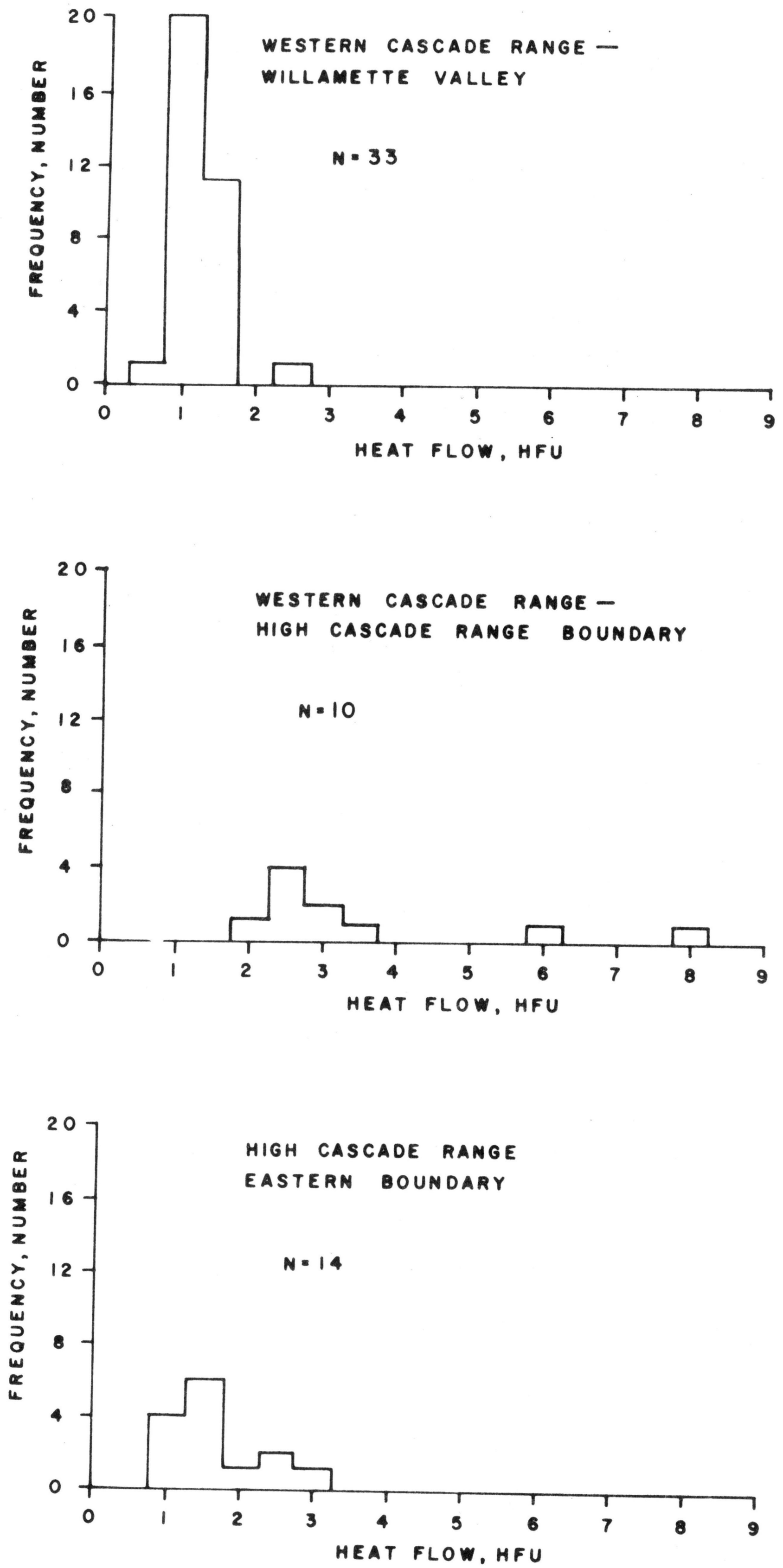


Figure 12. Heat flow histograms for the Northern Cascade Range province boundaries. Values for the histograms are shown in Figure 11.

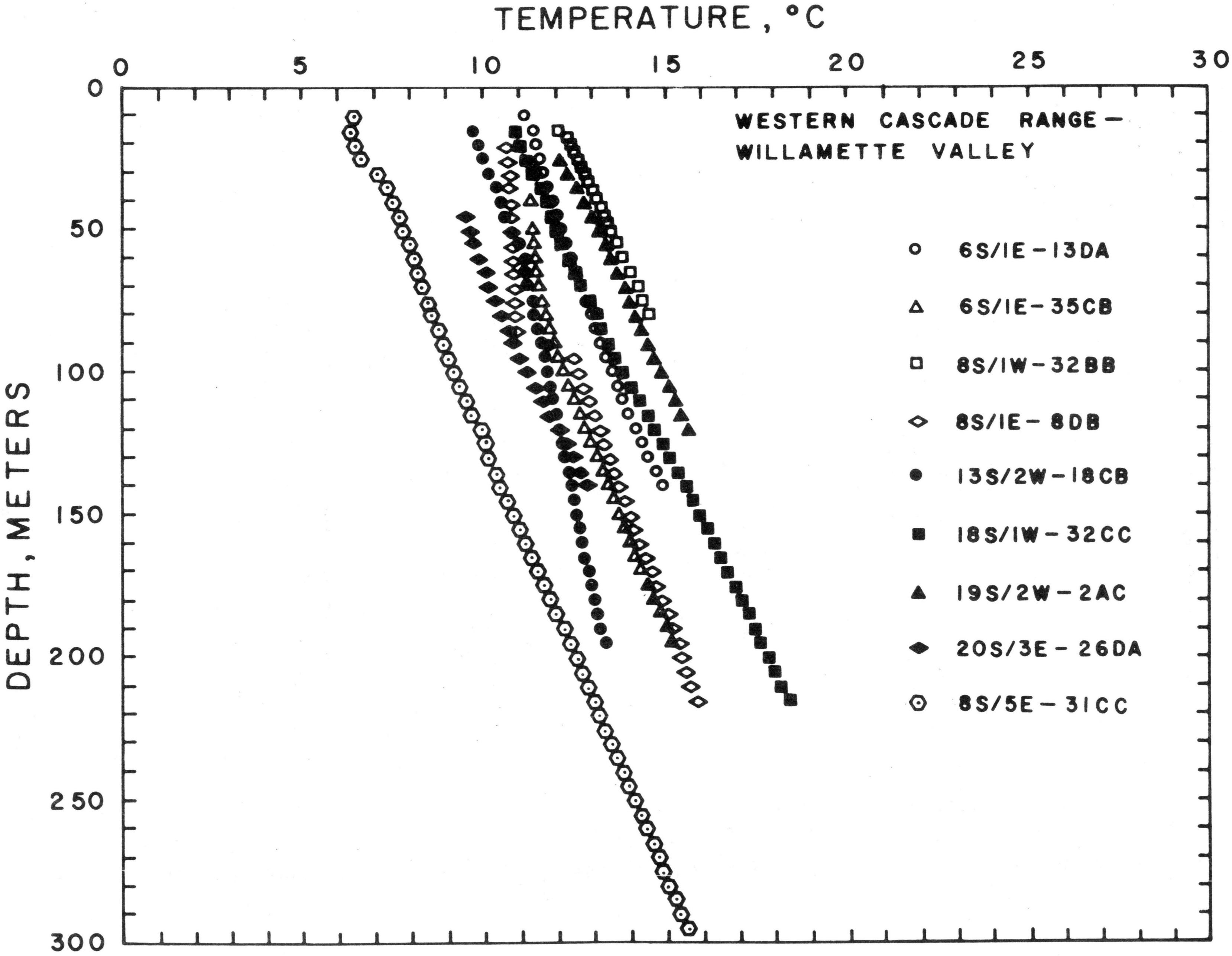


Figure 13. Typical temperature-depth curves for the Western Cascade Range-Willamette Valley province boundary. Locations of holes are given by township/range-section.

holes, which are typically noted in relatively high relief areas and in more brittle acidic rocks are seldom observed in the altered volcanic rocks of the Western Cascade Range-Willamette Valley regions.

Typical temperature-depth curves in the Western Cascade Range-High Cascade Range boundary region are shown in Figure 14. Most of the holes shown in this figure were drilled specifically for heat flow studies and were sited in the general vicinity of existing hot springs which are concentrated along the physiographic boundary between the Western Cascade Range-High Cascade Range provinces.

Detailed cross-sections of the heat flow results are shown in Figure 15 A-B. The cross-sections include corrected geothermal gradients and heat flow. From north to south, clusters of data are along the Clackamas River in the vicinity of Austin Hot Springs, along the Santiam River in the vicinity of Breitenbush Hot Springs, along the McKenzie River in the vicinity of Belknap and Foley Hot Springs, and along the Middle Fork of the Willamette River in the vicinity of McCredie Hot Springs. Each of the cross-sections also includes heat flow values derived from water wells in the western and central parts of the sections. The zero line for distance scale is the mean location of the physiographic boundary between the High Cascade Range and the Western Cascade Range (Figure 11).

Although many of the values plot within the High Cascade Range province (Figure 15 A-B), most of the holes were actually drilled in Western Cascade rocks. The province boundary is quite irregular with the High Cascade rocks generally lying at

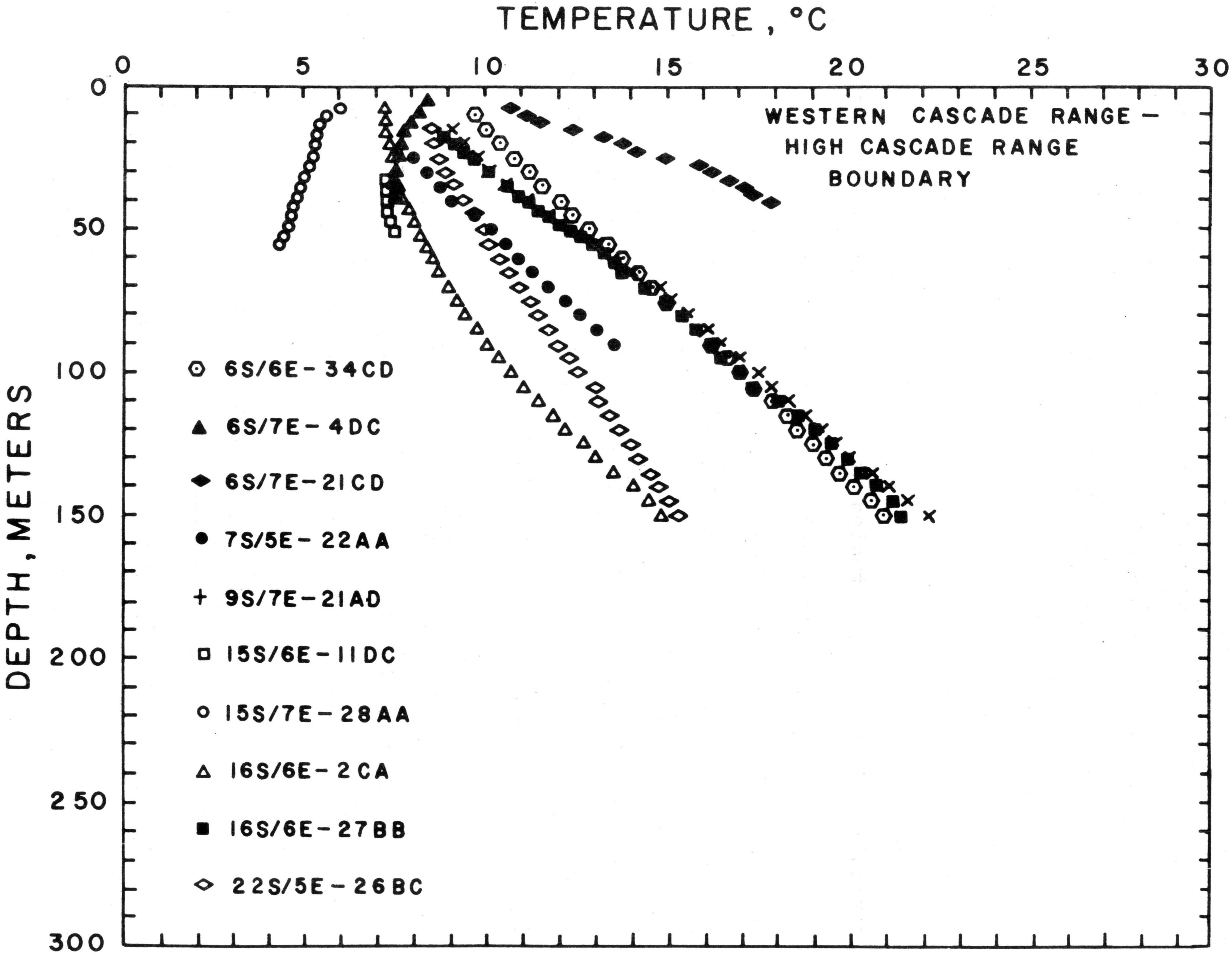
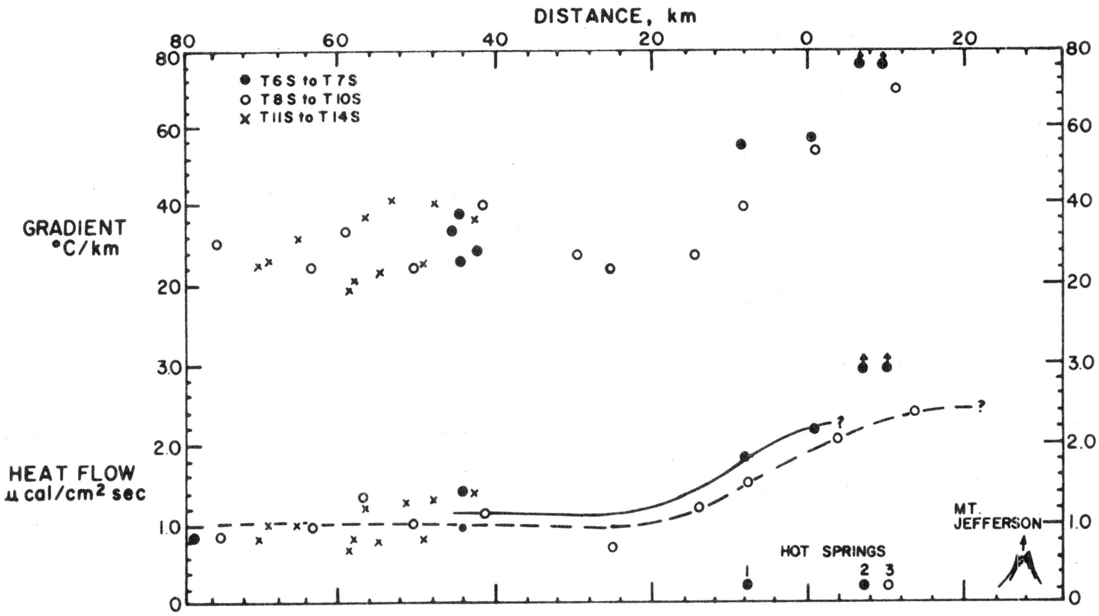


Figure 14. Typical temperature-depth curves for the Western Cascade Range-High Cascade Range province boundary. Locations of holes are given by township/range-section.



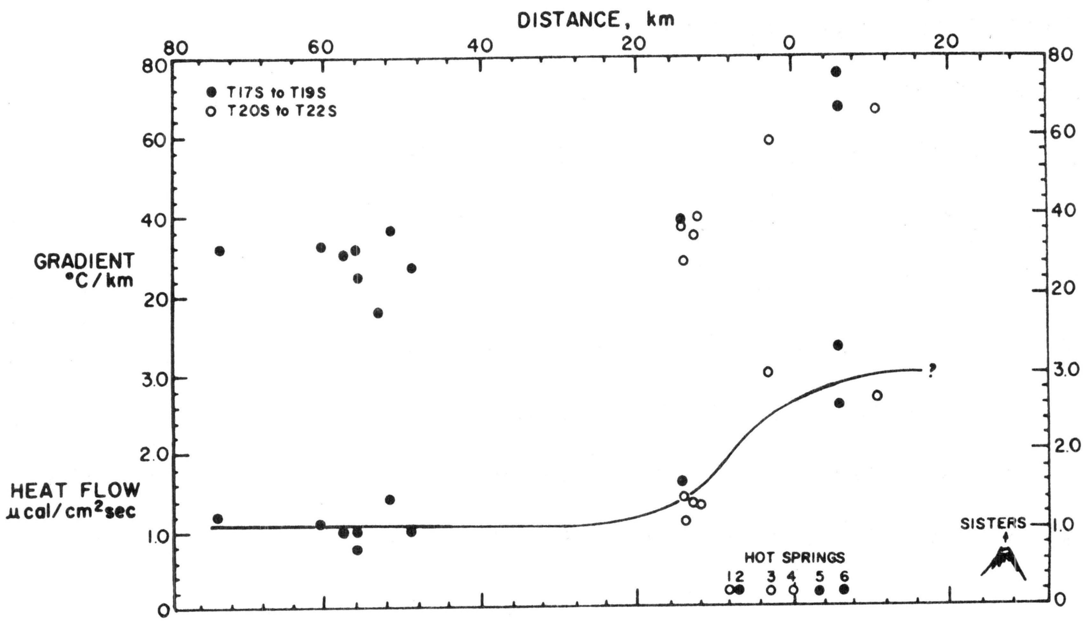


Figure 15 A-B. Heat flow cross-sections of the Willamette Valley-Western Cascade Range-High Cascade Range province boundaries. Heat flow and gradient values are grouped by townships as indicated. The hot springs near Mt. Jefferson (15-A) are (1) Bagby; (2) Austin; and (3) Breitenbush. The hot springs near Three Sisters (15-B) are (1) Cougar Reservoir; (2) Kitson; (3) Foley; (4) McCredie; (5) Wall Creek; and (6) Belknap.

higher elevations than the Western Cascade rocks, so that the latter tend to outcrop along the valleys while the former generally outcrop along the ridges. Three holes in the High Cascade rocks shown in Figure 14 were drilled in inter-canyon flow sequences that descend to lower elevations. All three holes have gradients typical of the downflow or lateral flow regime of an aquifer, although the gradients appear to be regional in the bottom of the drill holes. With these three exceptions, most of the holes have completely conductive gradients.

The data shown on the cross-sections in Figure 15 A-B is discussed in sequence from north to south. The northernmost data set includes the area of Townships 6S and 7S. At the High Cascade Range-Western Cascade Range boundary, the cross-section includes Austin Hot Springs and Bagby Hot Springs along the Clackamas River. A total of eight reliable heat flow values were obtained but the data are relatively widely spaced, one group being in the Willamette Valley and the second group being at the Western Cascade Range-High Cascade Range province boundary. Approximately 10 km west of the physiographic boundary, a heat flow value of 1.8 HFU was noted. The heat flow rises to approximately 2.2 HFU near the physiographic boundary. Two holes within 2 km to the east of Austin Hot Springs display heat flow values in excess of 5 HFU and indicate a relatively large geothermal system along the Clackamas River. This cross-section is the only one along which such high heat flow values, typical of hydrothermal circulation, were located during the drilling phase of the project, even though most of the holes drilled for heat flow values were within 5-10 km of hot springs.

The second set of data was obtained along a cross-section including the areas in Townships 8S and 10S. Eleven reliable heat flow measurements were obtained along this section. This cross-section is primarily along the Santiam River and crosses Breitenbush Hot Springs. The data show an almost constant heat flow of 0.9-1.1 HFU within the Western Cascade Range province. The heat flow rises gradually about 10 km west of the physiographic boundary and attains a value of 2.4 HFU approximately 15 km into the High Cascade Range province and at a distance of about 25 km from the easternmost low heat flow value.

No heat flow data were obtained along the heat flow boundary between Townships 11S and 14S; however, 11 heat flow measurements were obtained in the Western Cascade Range province. These heat flow values average 1 HFU, typical of those observed elsewhere in the province.

A third cross-section lies along the McKenzie River and includes data from Townships 16S to 19S. Holes were drilled just to the west of Belknap Hot Springs and Foley Hot Springs. Ten reliable heat flow measurements were obtained along this cross-section. Approximately 10 km east of the physiographic boundary, a heat flow value of 1.6 HFU was noted. Approximately 5 km east of the physiographic boundary, values of 2.0 HFU and 3.0 HFU were measured near Belknap Hot Springs and Foley Hot Springs, respectively. The distance between the high heat flow and the normal heat flow values is approximately 15 km.

The southernmost cross-section spans the area between Townships 20S and 22S. The holes were drilled in the vicinity of McCredie Hot Springs along the Middle Fork of the Willamette

River. Several free holes were also obtained in this area. These data show that the mean heat flow is 1.2 HFU approximately 8 km west of the physiographic boundary, whereas a value of 3.0 HFU is obtained only 5 km east of the boundary in a drill hole near McCredie Hot Springs. This value drops to 2.7 HFU about 10 km further to the east.

These data document clearly a systematic west-to-east increase in heat flow from 1 HFU to greater than 2.5 HFU over a lateral distance of 10 to 30 km and approximately coinciding with the mean physiographic boundary between the High Cascade Range and the Western Cascade Range provinces.

Interpretation of Heat Flow Transition Zone

All the profiles along the boundary between the Western Cascade Range and High Cascade Range provinces show the same characteristics: an almost constant heat flow with a mean value of 1 HFU in the High Cascade Range, and a mean heat flow of 2.6 HFU in holes drilled in Western Cascade rocks; an exceedingly abrupt transition zone between the two regions of heat flow, not exceeding 30 km in any location. The most remarkable aspect is the uniformity of the transition zone from north to south; although small variations may exist, the data suggest very similar conditions along the whole area under discussion.

In order to interpret the results, heat flow profiles representing several possible transition zones were constructed. These are shown in Figure 16 A-B. The different profiles represent different half-widths for the heat flow transition

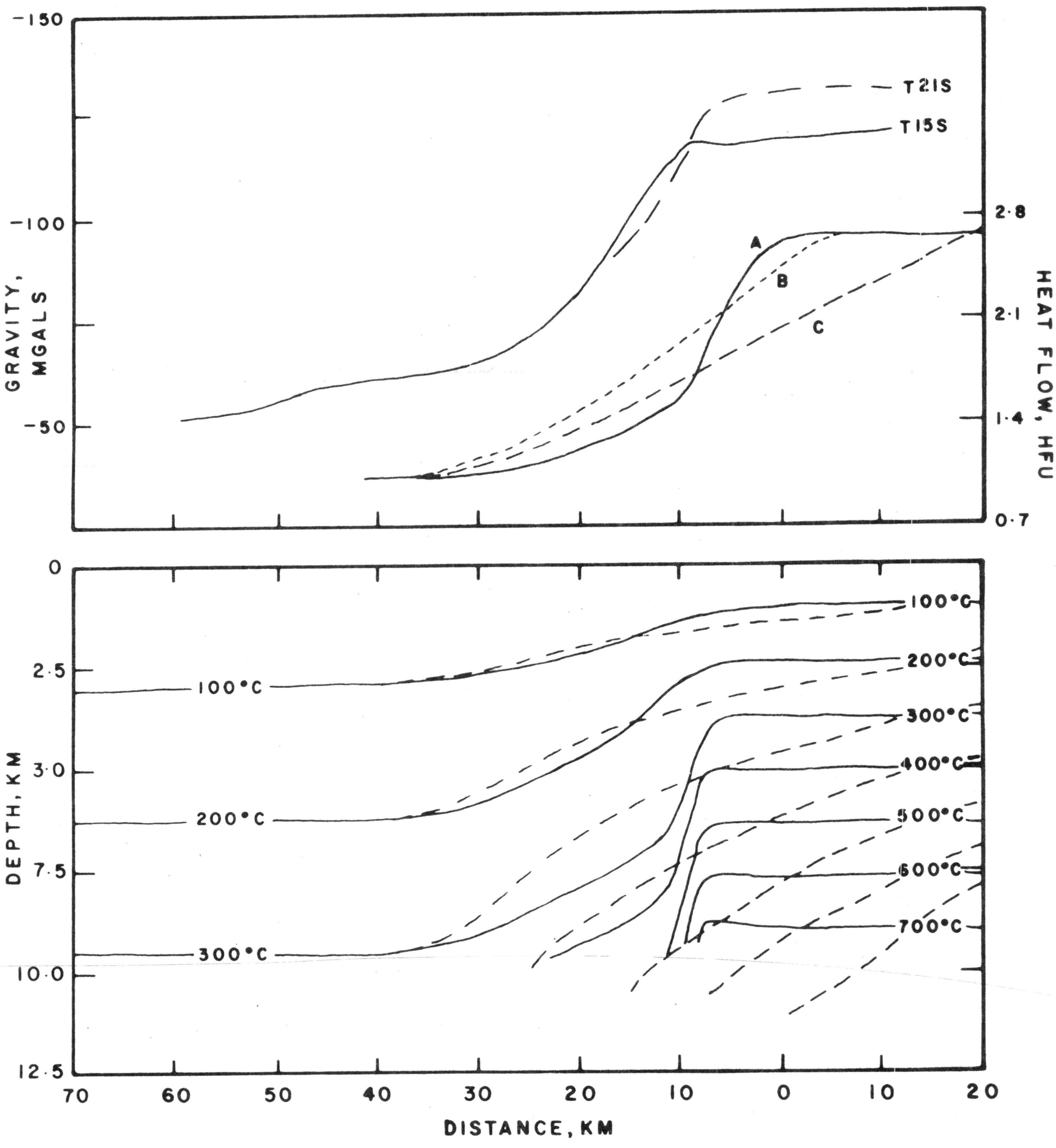


Figure 16 A-B. Heat flow-gravity transition zone models. In 16A, gravity curves are indicated by latitude, and heat flow curves are indicated by A, B, C. The heat flow curves are based on data from Figure 15: curve C, from the 6S-7S profile (15A); curve A, from the 20S-22S profile (15B). In 16B, steady-state isotherms corresponding to curve A are indicated by the solid lines, and those corresponding to curve C are indicated by the dashed lines.

based on the data in the 6S-7S profile (the largest half-width, curve C), the 20S-22S profile (the shortest half-width, curve A), and an intermediate half-width representing an average of the whole data set (see Blackwell and others, 1978).

Based on these heat flow transition zone profiles, temperature-depth cross-sections were calculated for two extreme cases of transition zones represented by curves A and C. The temperature values are based on a modification of the continuation of thermal data method discussed by Brott (1976). The more gradual transition zone has most of the heat flow difference occurring over a distance of 30 km, while the other shows a much sharper transition zone (over a distance of approximately 15 km). The isotherms were constructed for a steady-state and a transient model. In both models, homogeneous thermal conductivity was assumed. Because the geologic evidence indicates that greater than 6 m.y. BP the Western Cascade Range was the locus of intrusive activity, and because the heat flow now observed is low, there is strong evidence for a major temporal change in heat flow. The temperature sections were calculated assuming steady-state conditions, and assuming a uniform heat flow for the whole region of 2.5 HFU up to 6 m.y. ago with subsequent imposition of a constant strength heat sink beneath the Western Cascade Range which has resulted in the low heat flow values now observed. Over the scale of the area involved here, however, the steady-state and 6 m.y. temperatures do not differ significantly and so only the steady-state results are illustrated.

Both models imply very high temperature at relatively shallow depth beneath the High Cascade Range and extending approximately 10 km into the Western Cascade Range. Any one of the isotherms shown could satisfy the heat flow anomaly so the source does <u>not</u> necessarily have to reach or exceed 700°C; however, the source does have to be relatively shallow and relatively intense. Because of the uniformity of the heat flow data from north to south, it seems unlikely that the boundary can be simply related to hydrology and therefore must be related to some regional crustal effect.

As discussed elsewhere, attempts to investigate the extension of the heat flow pattern further to the east toward the axis of the High Cascade Range have been unsuccessful. Volcanism has been most continuous during the Quaternary along the axis of the High Cascade Range and the cover of young volcanic rocks, with concomitant horizontal water circulation, effectively prevented successful heat flow determinations in this area at depth of 150 m or less.

Also shown in Figure 16A are two Bouguer gravity cross-sections, one at 44°15'N latitude and one at 43°45'N latitude. These profiles have been constructed from data discussed by Couch and Baker (1977). The profile for 44°15'N was extended west of the area based on the regional change in Bouguer areal gravity associated with (but opposite in sign to) the heat flow data. This gravity change is of major magnitude (over 50 milligals) and of relatively short half-width (10-15 km). The short half-width implies a crustal source for the density contrast which results in the gravity change. The coincidence of this gravity anomaly with the heat flow anomaly is additional evidence that the heat flow data are related to regional crustal effects and not to upper crustal groundwater circulation.

Because of the close correspondence of the two sets of data, an interpretation of the gravity data was attempted based on the continuation model of crustal temperatures presented in Figure 16. In an attempt to model the gravity anomaly, a density contrast corresponding to the expansion of the rocks, based on the calculated temperatures, was assumed. These density contrast models did not generate a large enough gravity anomaly. Of course, in a complicated mountainous terrain with young volcanic rocks, density variations may be related to different factors. For example, in the High Cascade Range, shallow volcanic rocks may have a much lower density than the 2.67 g/cm^3 assumed in the reduction, whereas the density of the low-porosity rocks of the Western Cascade Range may be closer to the Bouguer reduction density.

The overall result of the interpretation is that a regional magma chamber or area of thermally anomalous crust exists under the High Cascade Range of Oregon. The width of the thermal anomaly is somewhat larger than the apparent width of the zone of volcanism and it extends at least 10 km west of the physiographic boundary of the province. Regional heat flow and gravity anomalies are associated with this disrupted crustal zone.

Most of the hot springs in the Cascade Range are located near the boundary of this region of high heat flow. This location may be related to a number of different effects such as the hydrologic conditions; location of faults, or fractures, along the boundary which focus the circulation of water;

outcrop of some horizontal unit that transmits water from higher elevations to the east; the location of permeable fractured acidic rocks at depth; and/or the possible location of a slightly more effective heat source.

Heat Flow Along the Eastern Boundary of the High Cascade Range

The heat flow data relating to the eastern boundary of the High Cascade Range geothermal anomaly are considerably less detailed and less consistent than those associated with the western boundary. The heat flow values in this area are shown in Figures 10 and 11. Typical temperature-depth curves are shown in Figure 17.

Around Bend, most of the holes are isothermal or show very irregular gradients to the maximum depth reached in each hole. To the north, generally conductive temperature-depth curves are obtained for holes in the older rocks immediately to the east of the High Cascade lavas. One hole (11S/15E-22cd) which was logged to a depth of over 800 m gives a heat flow of 2.0 HFU. A number of values are available further to the north. They are somewhat variable, ranging from average values for the Columbia Plateau (1.5 $\pm$ 0.3 HFU) to values greater than 2.0 HFU. On the basis of the data available at this time, no clear pattern has emerged and in fact, there may be considerable variations in the regional anomalies along the northeastern margin of the High Cascade Range.

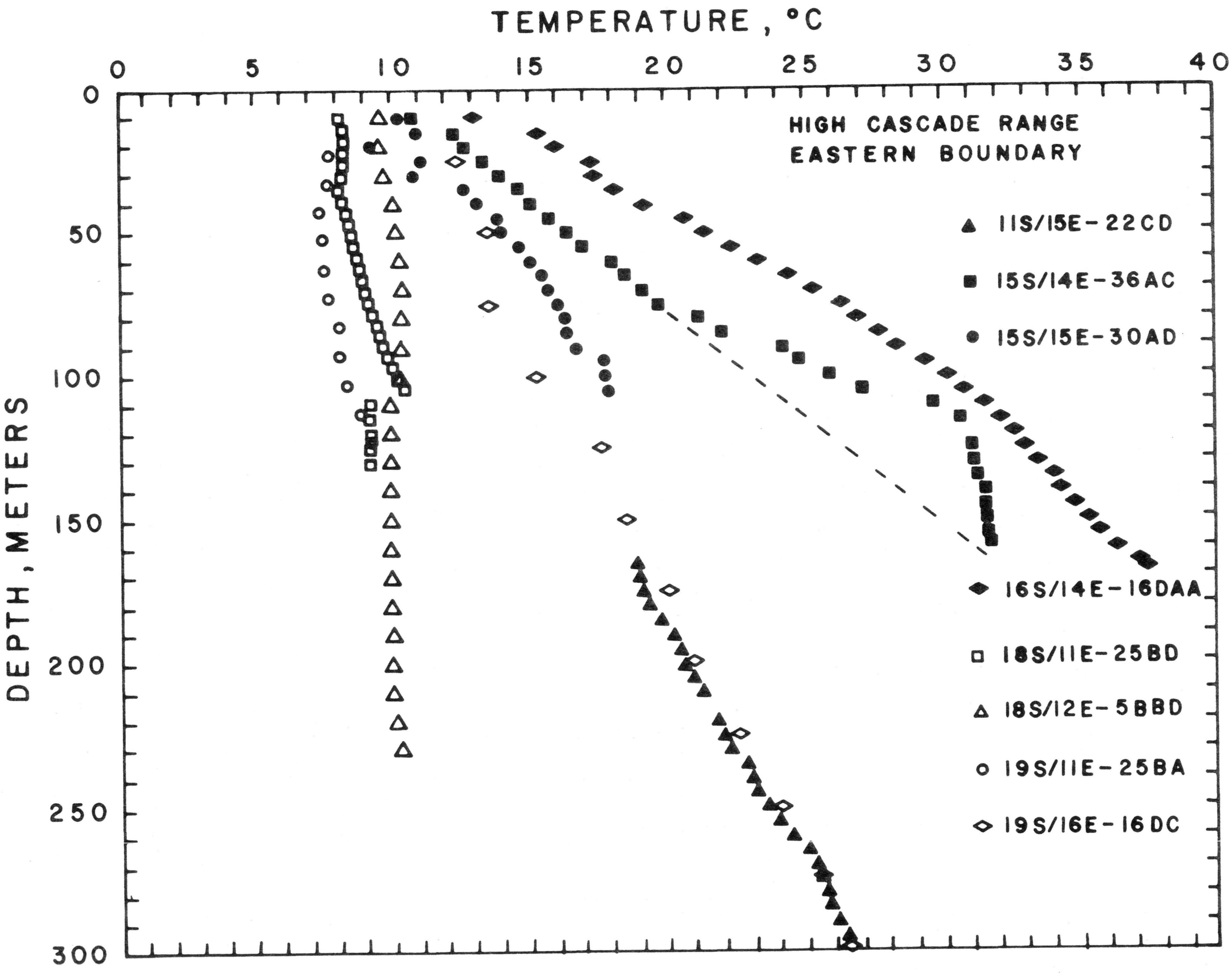


Figure 17. Typical temperature-depth curves for the eastern High Cascade Range province boundary. Locations of the holes are given by township/range-section.

GEOTHERMAL STUDIES ASSOCIATED WITH MT. HOOD

The main emphasis of this project was the analysis or theoretical study of magma chambers possibly associated with andesite volcanoes and a heat flow study of the Mt. Hood region. As part of the heat flow study of the Mt. Hood region, several holes were located during 1976 and 1977 in the Willamette Valley, along the eastern border of the High Cascade Range, and west of the Hood River Valley. In 1976, three holes were drilled specifically by DOGAMI for heat flow, one near Timberline Lodge, one near the southern margin of Mt. Hood, and one in the fine-grained plutonic rocks approximately 10 km southwest of Mt. Hood.

In 1977-78, two deep holes were drilled, one on the slope of Mt. Hood near Timberline Lodge, and one at Old Maid Flat along the Sandy River. In addition, eleven 150 m holes were drilled by DOGAMI for heat flow studies in the near-region of Mt. Hood in the fall of 1978. Temperature-depth curves for most of these holes are shown in Figure 18. The locations of all these drill holes and heat flow values (where appropriate) are shown in Figure 11.

Mt. Hood Regional Heat Flow Studies

The same dichotomy in heat flow regime is observed in the vicinity of Mt. Hood that is seen in holes drilled to the south. Drill holes that encounter rocks characteristic of the Western Cascade Range province, generally display conductive geothermal gradients. However, when the rocks drilled are of Pliocene or Pleistocene age, characteristic of the High Cascade

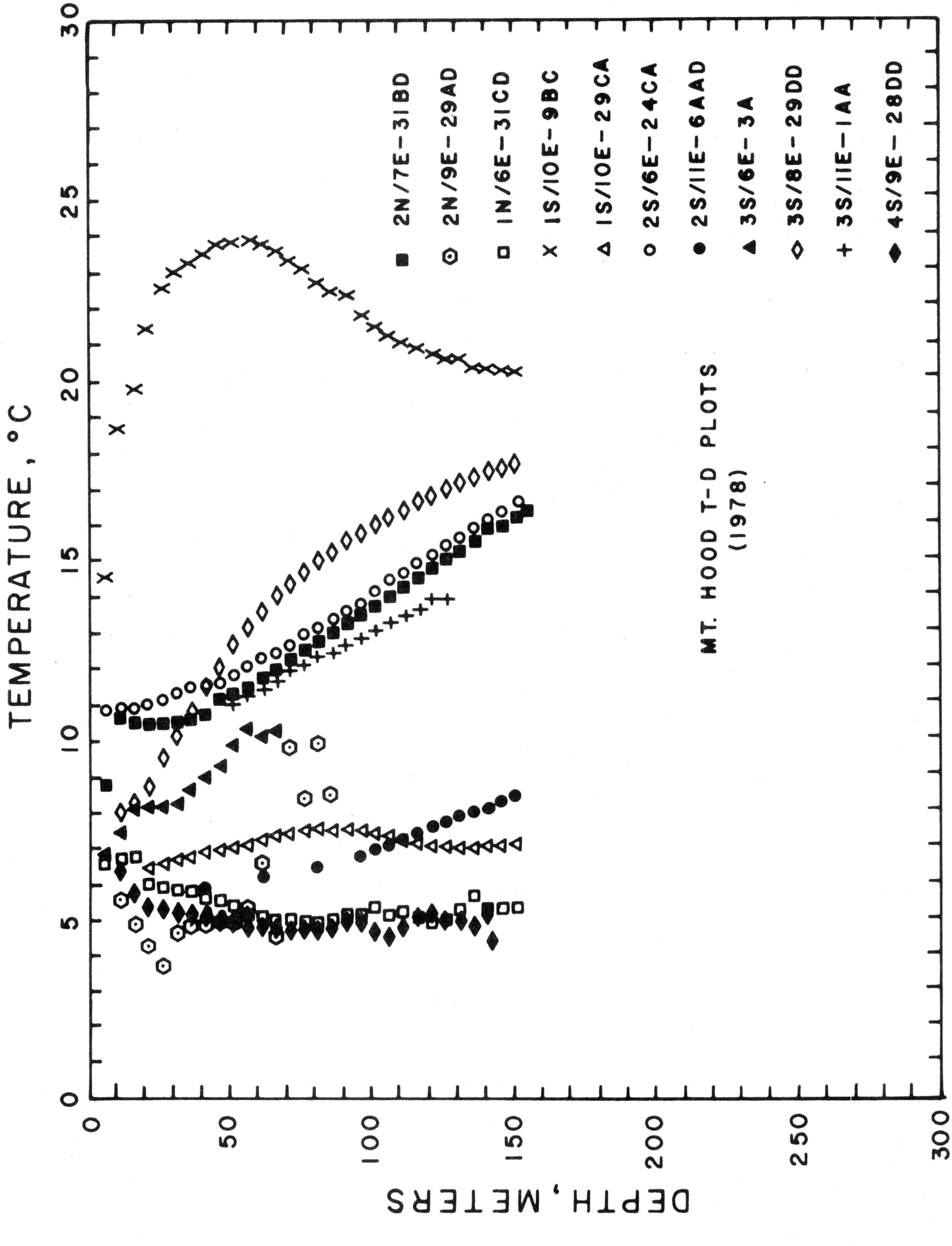


Figure 18. Temperature-depth curves for holes drilled by DOGAMI in 1978 in the Mt. Hood area. Locations of the holes are given by township/range-section.

Range, the holes do not yield useful geothermal gradient data. Most of the 150 m holes in these latter rocks gave isothermal temperature-depth curves.

The heat flow data are shown in Figure 19 projected onto a cross-section of the Cascade Range in the vicinity of Mt. Hood. The major transition zone observed to the south appears to exist in a somewhat subdued form at the latitude of Mt. Hood. Regional values of heat flow away from the volcanic edifice appear to be on the order of 1.5-1.8 HFU, as compared to values in excess of 2.5 HFU in locations to the south.

Only in the Old Maid Flat exploratory hole (2S/8E-15cd), at the toe of the volcano, is a value comparable to those to the south observed. Similarly, a gravity anomaly related to the heat flow transition also appears to be somewhat subdued compared to that further south (see Figure 16). The gravity data currently available are much less detailed than those used in the south. More detailed studies soon to be available (Couch, personal communication, 1979) will possibly allow better analysis of the relationship between gravity data and heat flow data at the latitude of Mt. Hood.

The continuation temperature model using these data shows lower temperatures (except in the immediate vicinity of Mt. Hood) than those shown in Figure 16. This model (not shown herein) is largely hypothetical due to the paucity of data in the immediate vicinity of the volcano and the large scatter of values, especially on the east side of the volcano.

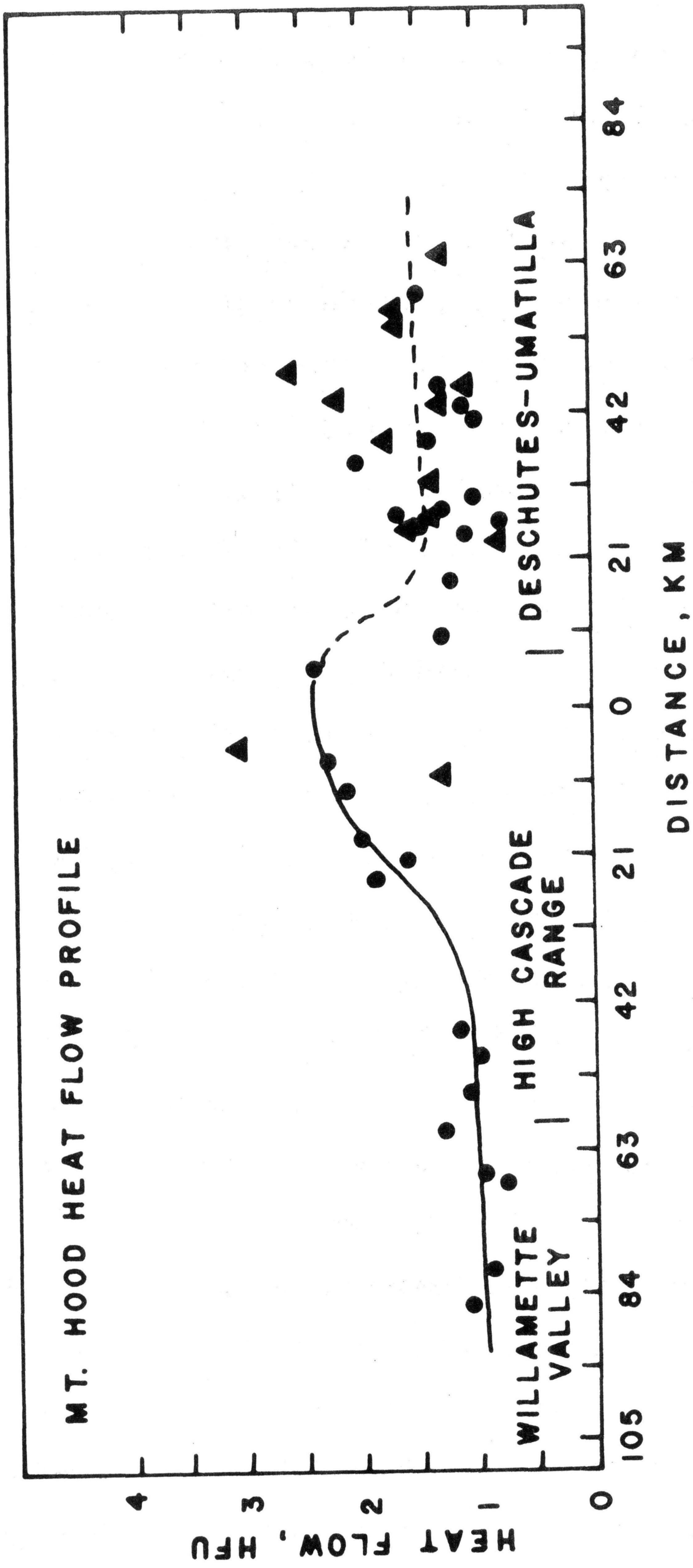


Figure 19. Heat flow profile of Mt. Hood. Baseline reference is through the profile of Mt. Hood. Heat flow values occurring south of Mt. Hood down to Latitude 45°N are represented by solid triangles, and heat flow values occurring north of Mt. Hood up to the Oregon-Washington border are represented by solid circles. All heat flow values are from Figure 11.

Old Maid Flat Holes #1/2 and #3

A major part of the project involved geothermal analysis by SMU and DOGAMI of a 1220.4 m deep geothermal test well drilled in the immediate vicinity of the Mt. Hood volcano. The hole is located at 2S/8E-15cd, along Old Maid Flat in the Sandy River valley, approximately 5 km from the apex of Mt. Hood, but almost 2 km lower in elevation.

The hole was initially drilled in the winter of 1977 to a total depth of 480 m, and designated Old Maid Flat #1. The drilling was done in two phases. A set of near-equilibrium temperature measurements was made during a pause in the drilling, for the purpose of setting casing, when the total depth was 230 m. This log (11/15/77) is shown in Figure 20. The gradient shows a disturbance in the uppermost 75 m of the hole. Below that, the gradient is linear, with a mean value of 67°C/km.

Upon completion of the hole (480 m), the temperature log (12/20/77) shows essentially equilibrium temperatures after completion of the second phase of the drilling. The log shows a somewhat higher temperature at the depths measured in the first log, with a uniform gradient of 67°C/km between 100-300 m. Below 300 m, the gradient decreases and then increases again with a value of approximately 65°C/km observed between 430-475 m. The absolute temperature values between 70-200 m in the 11/15/77 log and between 435-475 m in the 12/20/77 log are compared in Figure 20. At the time of the 12/20/77 log, there was a flow of about 10 ℓ/min from the collar of the drill hole. The nature of the temperature-depth curves indicates that a

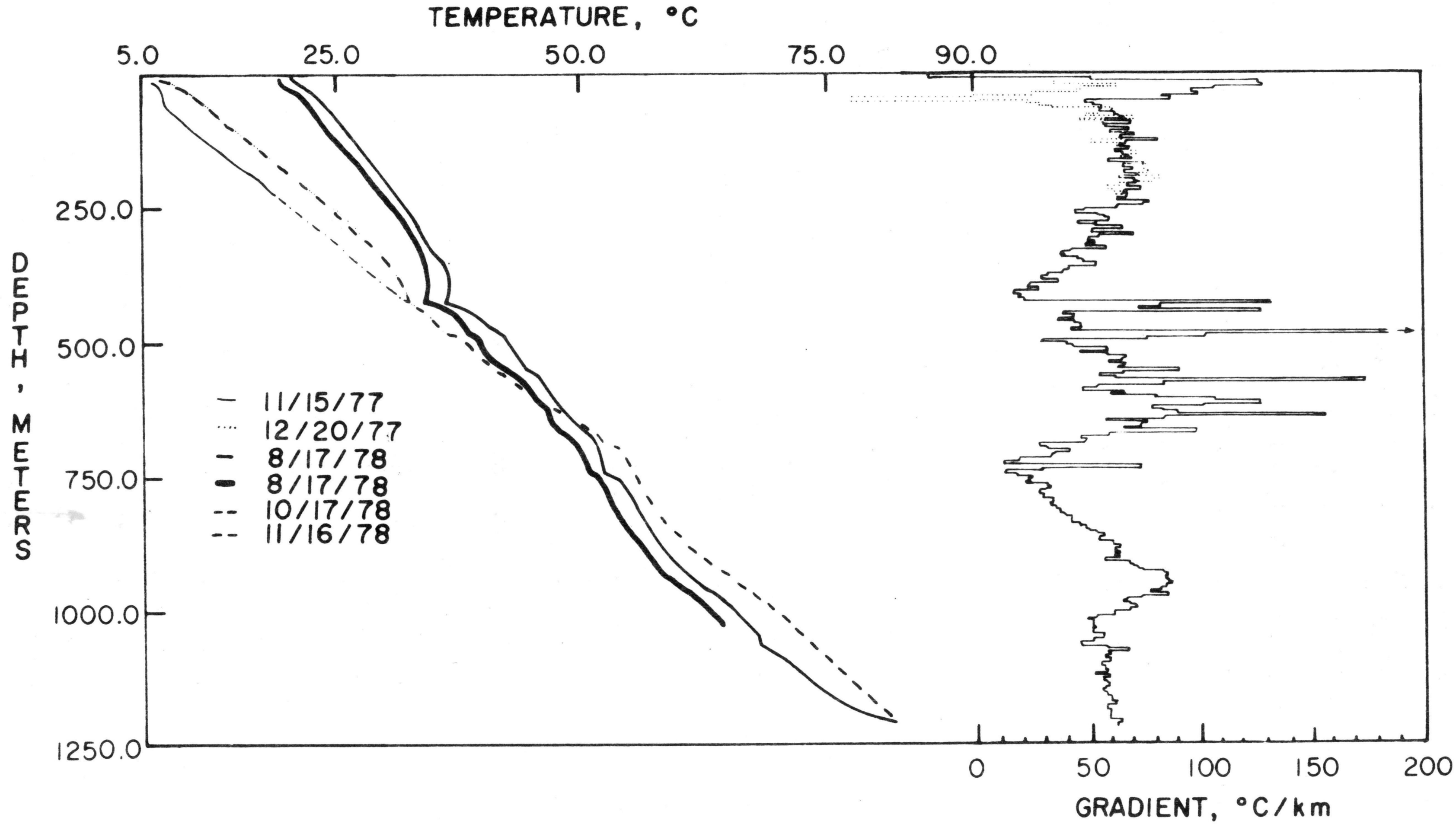


Figure 20. Temperature-depth curves for the Old Maid Flat Hole #1/2. Different dates of logging are represented by the symbols indicated. The bar graph at the right side of the figure is from the 11/16/78 data.

fracture zone with artesian fluid pressure was encountered at a depth of 430 m. At the time of the second logging, water was moving slowly up the borehole from that depth and out at the collar. The temperature of the water column is offset by about 4°C by this upflow in the borehole; thus, undisturbed temperatures in the borehole would agree approximately with the 11/15/77 log.

During the summer of 1978, the hole was deepened to a total depth of 1220.4 m and designated Old Maid Flat #2. Immediately after completion of the drilling, two temperature logs were made on 8/17/78. The lighter line on Figure 20 represents measurements by a commercial well-logging firm, while the heavier line represents measurements made with SMU's truck-mounted logging system, with a total depth capability of 1035 m. The shapes of the temperature-depth curves are essentially the same, but are offset by approximately 2°C. This could be due either to a calibration difference between the two sets of gear, or to differences in the actual temperature of the hole at the times of logging; however, the offset is probably due to a combination of causes.

The rapid return to equilibrium of temperatures in the artesian zone (around 435 m) is illustrated in the logs made immediately after the drilling. After completion, a 2" diameter observation pipe was set into the hole, but no attempt was made to grout the tubing as the hole is to be used to obtain fluid samples. Therefore, natural inter-formation flow in the annulus is not prevented by the completion technique.

After stabilization, the hole was logged on 10/17/78 by

SMU, to a total depth of 1035 m, and on 11/16/78 by USGS, to a total depth of 1214 m (John Sass, personal communication, 1978). Only the 11/16/78 USGS log is shown in detail, as the 10/17/78 SMU log is not significantly different. The temperature-depth curves observed at equilibrium are quite irregular. Even after the theoretical time needed for recovery of the temperatures, gradients in the middle part of the hole (between 300-1000 m) remained highly irregular, with significant changes which cannot be attributed to lithology. These changes are represented by the bar graph on the right side of Figure 20.

The only reasonable explanation for these gradients is a very large amount of borehole fluid communication. In fact, the only part of the hole which appears unaffected by intra-borehole fluid communication is that below 1,100 m. In this bottom portion of the hole, two distinct gradients are observed: a gradient of 52°C/km between 1010-1070 m, and a gradient of 61°C/km between 1170 m-total depth. The gradients in the central part of the hole appear to be due to fluid flow between fracture zones or flow contacts. There appears to be no instance of water flow from a single fracture zone all the way up or down the borehole, but merely localized effects. Each of the spikes of high gradient (Figure 20) corresponds to a fracture zone or flow contact. It appears that upflow is generally seen in the upper part of the hole, whereas downflow is typical of the bottom part of the hole.

If the hole had been grouted, this fluid flow would have been eliminated and a much simpler conductive gradient would be seen. Details of geothermal gradient cannot be deciphered from the complexities of intra-borehole fluid flow. However, it is clear that the mean gradient over the entire borehole is relatively well-established, even if the detailed variations are not.

Thermal conductivity measurements were made on cuttings from between 580-1219 m. The measurements were made at intervals of approximately 30 m. Porosity values for the various depths were estimated from a neutron log. Most porosity values averaged between 1-10 percent, and these values were taken into account when calculating in situ conductivity. The mean thermal conductivity for all samples is 4.12 $\pm$ 0.14 mcal/cm-sec-°C. There are no systematic depth variations based on the bulk conductivity; however, if porosity is taken into account it would appear that slightly lower values of thermal conductivity (approximately 3.8 mcal/cm-sec-°C) occur between approximately 850-1000 m, with higher values (approximately 4.5 mcal/cm-sec-°C) above and below that zone.

The break in gradient at 1070 m, indicated in the temperature-depth curves, does not correspond with a change in thermal conductivity; however, on the commercial temperature log of 8/17/78 a "kink" in the curve was observed at that point, indicating the location of a possible fracture zone. Possibly there is some flow in the hole around this point, with the disturbed conditions only observed below 1070 m.

The mean gradient for the entire hole is 65°C/km, while that for the most undisturbed portion (below 1070 m) is 60°C/km. The gradients for the two apparently undisturbed sections of the upper part of the borehole (25-200 m and 430-475 m) are

60°C/km and 65°C/km, respectively. Taking all these data into consideration, the mean terrain-corrected heat flow for the Old Maid Flat hole is 2.4 HFU. The significance of this value with respect to the surrounding values has been discussed in the preceding section on regional heat flow in the vicinity of Mt. Hood.

Another hole, drilled to completion on 12/13/78 by Northwest Geothermal Corp., to a total depth of 400 m, was designated Old Maid Flat #3. This hole (2S/8E-17cc) is located about 3 km southwest of Old Maid Flat #1/2. The general lithology encountered in the hole is: recent mudflow debris, 0-35 m; pyroclastics with volcanic debris, 35-305 m; and an alternating sequence of Columbia River basalt and andesitic volcanic rock down to 400 m.

The first temperature-depth measurement by DOGAMI (Figure 21) was made on 12/5/78, when the hole was at a total depth of 152 m. The linear portion of the curve, from 75 m to total depth, appears undisturbed and yields a thermal gradient of 50°C/km. The disturbed upper portion of the temperature curve is probably due to fluid motion and drilling effects in the mudflow and upper pyroclastic sequence of rocks.

The hole was temperature-logged again by DOGAMI on 12/21/78 after being drilled to completion (Figure 21). The gap in the curve from 215-325 m is due to lost paperwork. However, from hand-plotted field results, the curve in this interval indicates a fluid disturbance originating around 275 m and affecting the entire missing interval. Excluding the data gap, the upper portion of the 12/21/78 log agrees in general with

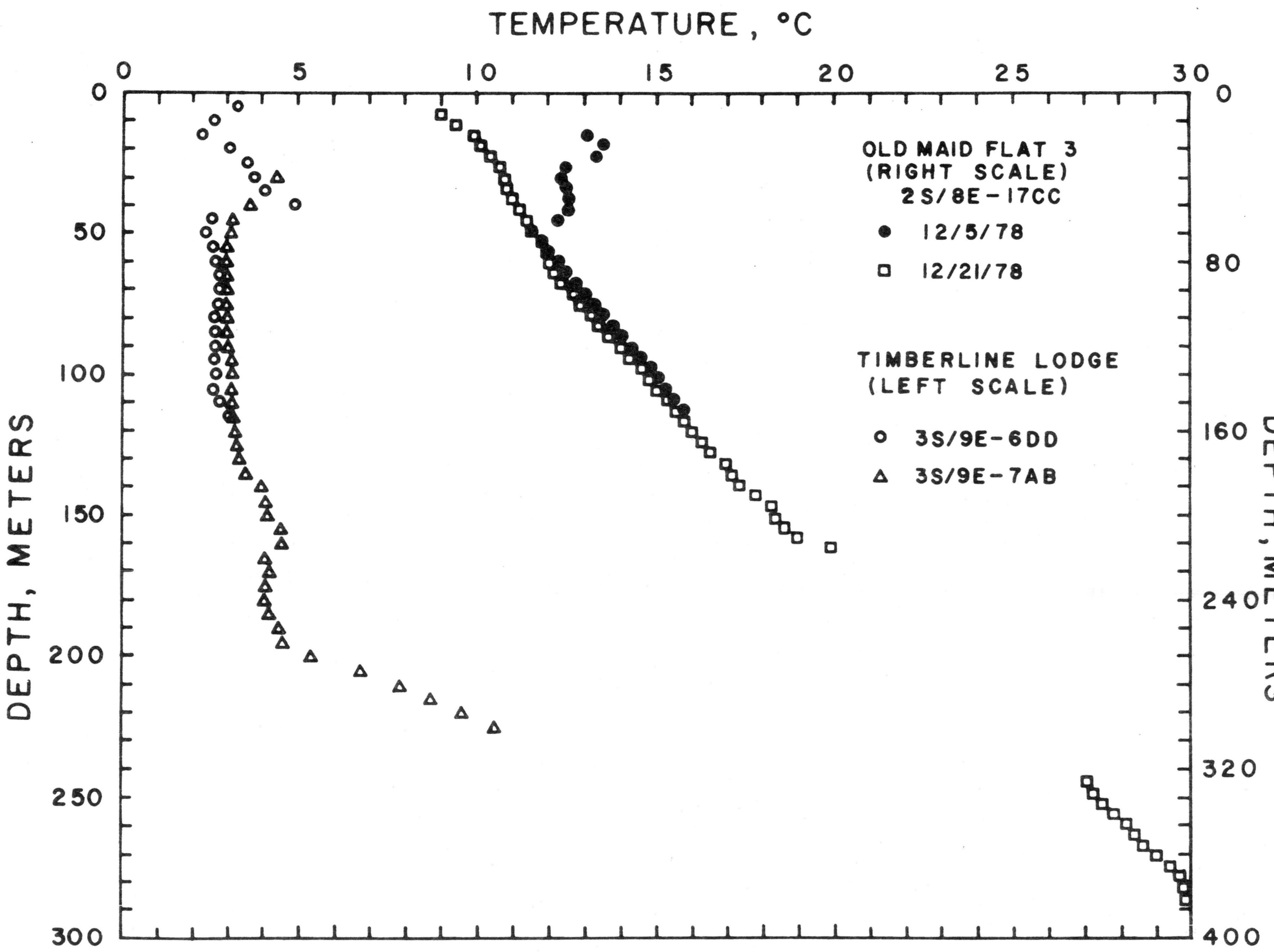


Figure 21. Temperature-depth curves for the Old Maid Flat hole #3 and the Timberline Lodge holes #1 and #2. The scale on the right side of the figure is for the Old Maid Flat hole #3 curve, and the scale on the left side of the figure is for the Timberline Lodge holes #1 and #2 data.

the 12/5/78 log; the lower portion of the 12/21/78 log gives a thermal gradient of 63°C/km.

Based on five measurements of the pyroclastic sequence of rocks, a thermal conductivity value of 4.04 ± 0.31 mcal/cm-sec-°C was obtained. A terrain-corrected thermal gradient of 51.2°C/km, with the above thermal conductivity, yields a corrected heat flow value of 2.1 HFU.

Timberline Lodge Drill Holes #1 and #2

On 9/8/76, a hole was completed to a total depth of 152 m on the carapace of Mt. Hood, about 0.2 km east of the Timberline Lodge Ski Resort. Timberline Lodge #1 (3S/9E-6dd) encountered mostly andesitic-type rocks, ranging from scoriaceous to basaltic andesite with fracture zones at 43 m, 91 m, and possibly at 122 m. The fracture at 43 m had an aquifer flow of about 19 ℓ/m.

The hole was temperature-logged on 9/8/76, 9/13/76, and 9/14/76. The 9/14/76 temperature curve (Figure 21) shows the effect of the aquifer at 43 m. The rest of the curve is essentially isothermal and not suitable for heat flow studies.

Two years later, another hole was drilled about 0.5 km south of the Lodge. Timberline #2 (3S/9E-7ab) encountered the same type of rocks as Timberline #1, to a total depth of 421 m. The resulting 12/13/78 temperature log (Figure 21) only reached 225 m, due to caving and drilling problems in the hole. Again, this hole shows fluid flow problems similar to Timberline Lodge #1, and over a similar depth interval. The linear gradient in the last 30 m of the hole appears to occur below the aquifer zone, but has not yet approached the expected undisturbed gradient. The hole is currently considered unsuitable for heat flow studies.

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