JOURNAL OF GEOPHYSICAL RESEARCH, VOL. 95, NO. B12, PAGES 19,495–19,516, NOVEMBER 10, 1990

Heat Flow in the State of Washington and Thermal Conditions in the Cascade Range

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Heat flow data for the state of Washington are presented and discussed. The heat flow in the Okanogan Highland averages 75 mW m^{-2}, and the gradient averages 25°C km^{-1}. The heat flow in the Columbia Basin averages 62 mW m^{-2}, and the mean gradient is 41°C km^{-1}. Both of these provinces are interpreted to have a mantle heat flow of about 55–60 mW m^{-2}, the same value as in the Basin and Range province to the south and the intermountain region of Canada to the north. These areas comprise the high heat flow, back arc region of the Cordillera. In the coastal provinces and the western part of the southern Washington Cascade Range the heat flow is below normal and averages 40 mW m^{-2} with an average gradient of 26°C km^{-1}. This low heat flow is related to the absorption of heat by the subducting slab, part of the Juan de Fuca plate, that is beneath the Pacific Northwest. Thus the low heat flow area represents the outer arc part of the subduction zone. Within the volcanic arc, the Cascade Range, the heat flow pattern is complicated. The heat flow is best characterized in the southern Washington Cascade Range. The heat flow there averages 75 mW m^{-2} and the gradient averages 45°C km^{-1}. The heat flow peaks at over 80 mW m^{-2} along the axial region that coincides with the Indian Heaven, Mount Adams, and Goat Rocks centers of Quaternary volcanism. As is the case in northern Oregon and southern British Columbia, the western edge of the region of high heat flow has a half width of 10 km, implying a heat source no deeper than about 10 km. In the northern Washington Cascade Range the data are too sparse to determine the average heat flow. There are two saddles in the heat flow pattern in Washington, along the Columbia River and in central Washington. The origin for the contrasting heat flow may be segmentation of the heat source or some more local effect. The heat flow of the Cascade Range is well characterized in several locations, and the pattern is similar at all localities. The most striking feature is the 10 km half width of the western side of the high heat flow region where it abuts the low heat flow outer arc region. The axial heat flow ranges from about 80 to greater than 100 mW m^{-2}. The midcrustal temperatures are interpreted to range from about 400°C to 800°C. The source of these high temperatures is interpreted to be a long-lived midcrustal zone of magma residence that is characteristic of the Cascade Range regardless of crustal type, rate of volcanism, or composition of volcanism. For example, in southern Washington the region of high heat flow spans the width of the Quaternary zone of volcanism with Mount St. Helens and Mount Adams at the west and east edges, respectively. On the other hand, most of the Oregon stratovolcanoes are near the center of anomalous region.

INTRODUCTION

It has become clear in the last 10 years that the coastal provinces of the Pacific Northwest of the United States represent a more typical subduction zone terrain than was initially thought. The 1980 eruptions of Mount St. Helens emphasized the active volcanism already demonstrated by the fact that many of the Cascade stratovolcanoes have been active in the last 200 years. Detailed seismic studies have identified clear evidence of a subducting slab under Washington [*Weaver and Malone*, 1987; *Weaver and Baker*, 1988; *Ludwin et al.*, 1990]. In addition, evidence has accumulated indicating that large subduction zone earthquakes may occur in the Pacific Northwest, although perhaps on a time scale more extended than those in areas with faster subduction rates [*Atwater*, 1987; *Heaton and Hartzell*, 1987].

The thermal structure of volcanic arcs is of great interest, but regional thermal characteristics of volcanic arcs are difficult to obtain because of the nonconductive heat transfer processes that operate in such areas and the complex nature of the conductive portion of the heat transfer. The complicated nature of heat transfer requires detailed heat flow studies and extensive data coverage. The interest in geothermal energy potential has sparked much study of volcanic arcs. The Cascadia subduction zone is unique in terms of the amount and quality of thermal data available. One surprising result is that the altered volcanic rocks that comprise the bedrock of much of the area are of generally low permeability so that convective heat transfer by groundwater motion does not predominate and a generally conductive regional picture can be detected. The thermal characteristics of several areas in the Cascade volcanic arc in the United States have been described [*Blackwell et al.*, 1982; *Mase et al.*, 1982; *Blackwell and Steele*, 1983]. Recent studies in the Oregon Cascade Range are discussed in a companion paper [*Blackwell et al.*, this issue]. The heat flow data in the Canadian Cascade Range have also been recently described [*Lewis et al.*, 1988]. The first objective of this paper is to present and discuss thermal data in Washington; the second is to discuss in detail new heat flow data in the Washington part of the Cascade Range; the third is to summarize and compare the regional thermal characteristics of the whole

Paper number 90JB01435.
0148-0227/90/90JB-01435$05.00

Cascade Range. The implications of the thermal results for magmatism in the crust will be discussed in detail.

Summary of tectonics. The tectonic setting of the Washington Cascade Range is summarized by several papers in this volume [*Leeman et al.*, this issue; *Wells*, this issue; *Weaver*, 1989; *Sherrod and Smith*, this issue; *Stanley et al.*, this issue], and more references may be found in those papers. The details of the subduction of the Juan de Fuca plate are becoming clearer, and the location of the subducting slab has been mapped even in areas with no active seismicity using seismic tomography and electrical studies [*Rasmussen and Humphreys*, 1988; *Booker and Chave*, 1989, *Wannamaker et al.*, 1989].

The slab is seismically active to a depth of up to 80 km beneath northern California and in the Puget Sound region. In Oregon, the slab does not appear to be seismically active, although because of the lack of detailed station coverage, small-size slab earthquakes may not be recorded. All studies show that the slab dips at moderate angles from the trench to the west edge of the Cascade Range. At that point the dip rapidly increases so that the slab is typically at depths of 100–150 km beneath the volcanic arc.

BACKGROUND OF THERMAL MEASUREMENTS

The earliest geothermal measurement (directed by Van Ostrand) have been tabulated by *Spicer* [1964] and *Guffanti and Nathenson* [1981]. Estimated heat flow values have been calculated for two of these wells (20N/28W-8 and 11N/26E-20CC) based on thermal conductivity values that we measured on surface samples from lithologic units encountered in these holes.

In the early 1960s, R. F. Roy made heat flow measurements in northeastern Washington [*Roy*, 1963; *Roy et al.*, 1968] and measured temperatures in Development Associates Basalt Explorer 1, a deep hydrocarbon exploration well drilled in the Columbia Basin (DABE-1, 21N/31E-10CB). Also in the 1960s *Sass et al.* [1971] made some heat flow measurements and investigations were started by *Blackwell* [1969]. Preliminary results of additional statewide studies were published by *Blackwell* [1974]. Revised heat flow values for these sites are included in this report.

In the 1970s, detailed heat flow studies of the Turtle Lake Quadrangle, northeastern Washington [*Steele*, 1975] and the Indian Heaven area in the southern Washington Cascade Range [*Schuster et al.*, 1978] were published. In the mid-1970s, water chemistry studies indicated the possibility of anomalous heat flow values in the southern Columbia Basin area [*Swanberg and Morgan*, 1979], and in 1978 a reconnaissance temperature logging program was carried out there. The most recent general discussions are a brief summary of Cascade Range heat flow [*Blackwell and Steele*, 1983] and unpublished final reports [*Blackwell et al.*, 1985; *Kososec*, 1984]. Final reports of data collection are by *Korosec et al.* [1983], *Barnett* [1986], and *Barnett and Korosec* [1989]. This paper presents heat flow values for all the data collected between 1970 and 1988. These data consist of accurate, detailed temperature-depth measurements in selected available holes such as water wells and mineral exploration holes ("free holes") and in holes drilled specifically for geothermal studies (in the Cascade Range). Additional temperature-depth logs have been collected by personnel at Washington State University (WSU) over many years [*Stoffel and Widness*, 1983*a*, *b*]. These logs are not incorporated into this report.

DATA PRESENTATION

Geothermal gradient, heat flow, and ancillary information for drill holes in the state of Washington, excluding pre-1974 published data, are summarized in Table 1. Only holes where gradients or heat flow are considered to be of C quality (see below) or better are included in Table 1. Results for the poor quality data are given by *Blackwell et al.* [1985, 1989]. More complete information on the sites in Table 1 are given by *Blackwell et al.* [1989].

Individual holes are located by latitude and longitude and by township and range. Thermal conductivity values are based on measurements of core or cutting samples, or are estimated from lithology (values in parentheses). Terrain corrections were made for all holes where the correction was estimated to exceed 5% of the measured value. The terrain corrections were made by the technique of *Blackwell et al.* [1980] or *Birch* [1950]. Other aspects of the heat flow/geothermal gradient measurement and calculation techniques are discussed by *Roy et al.* [1968] and *Sass et al.* [1971]. A recent summary of hardware techniques for heat flow studies is given by *Blackwell and Spafford* [1987].

Summary maps of the heat flow and geothermal gradient data from Table 1 are shown as Figures 1 and 2, respectively. Only one symbol is shown for each site, and where appropriate, the results from holes close together have been averaged. Both the geothermal gradient and heat flow maps have been contoured. The contouring was done by hand and in areas of sparse control is based on physiographic setting. This procedure is necessary in the northern Washington Cascade Range and coastal provinces because of the scarcity of data there. Another large data gap exists in southeastern Washington. There is only shallow groundwater development along a broad northwest-southeast trending zone through the east central Columbia Basin. With the exception of a few points near the center of this zone, an area 60–70 km wide and 200 km long has no heat flow data available. It must be emphasized that the gradient contour map particularily is an approximation of actual conditions and will not apply to any specific point. In addition, the gradient will change with depth as the thermal conductivity changes, perhaps on a very fine scale. In the following sections, geothermal data from each major physiographic division of the state will be discussed.

COASTAL PROVINCES

Heat flow values from the Puget Lowland, the Coast Ranges (Willapa Hills), and the western part of the southern Washington Cascade Range (see Figure 1) are generally low. There are no data in the Olympic Mountains subprovince. All these areas are underlain by Cenozoic rocks that have accreted to North America in the Cenozoic [*Wells*, 1990; *Stanley et al.*, 1990; etc.]. Most of the rocks in the province are volcanic and volcanoclastic units with abundant feldspar, and thermal conductivity values tend to be below average for continental material. Hence for a given heat flow, geothermal gradients are higher than in a basement terrain. In northwestern Washington one heat flow measurement (34N/1E-2CBB) was obtained in quartz diorite that is

TABLE 1*a*. Thermal and Location Data for the High Heat Flow Sites in Southern Washington Cascade Range

Township/ Range Section	Latitude N, Longitude W	Depth Range, m	Average Thermal Conductivity, W m^{-1} K^{-1}	SE	*N*	Corrected Gradient, °C km^{-1}	Corrected Heat Flow, mW m^{-2}	Quality Rating	Lithology Summary
18N/9E 29 CB	47°00.54′, 121°41.70′	115–140	2.39	0.21	8	19.6	47	B	Miocene pyroxene rhyolite
14N/10E 8DCB	46°42.89′, 121°34.68′	75–115	2.18	0.08	10	33.8	74	B	Eocene volcanics
14N/10E 11AA	46°43.40′, 121°30.70′	80–152	2.50	0.13	6	39.3	95	B	CZ volcanics, sediments
14N/14E 25BCDC	46°40.39′, 121°01.76′	25–150	1.30	0.13	10	67.0	87	B	CR basalt
13N/12E 4AB	46°39.10′, 121°19.80′	60–150	2.02	0.21	8	46.0	93	A	siltstone and mudstone (JR)
13N/12E 2AB	46°39.00′, 121°17.02′	35–150	2.56	0.13	9	42.3	108	A	amphibolite (JR)
13N/11E 2DC	46°38.32′, 121°23.49′	10–148	1.84	0.15	21	45.0	83	A	Eocene volcanics
13N/9E 16BCA	46°37.22′, 121°41.59′	10–152	1.59	0.10	19	43.5	69	A	volcanic sediments
12N/7E 16CC	46°31.37′, 121°56.37′	35–129	1.75	0.08	10	41.6	73	A	
11N/10E 35BC	46°23.40′, 121°32.21′	50–115	2.10	0.13	7	49.0	103	B	CZ sandstone
10N/6E 8DDA	46°21.58′, 122°04.47′	120–270	3.68	0.17	6	18.5	68	C	dacite porphyry
10N/6E 18BDB	46°21.10′, 122°06.40′	205–210	3.85	0.13	6	18.0	69	C	dacite porphyry
10N/10E 15DBA	46°21.10′, 121°32.85′	80–109.7	1.55	0.08	5	44.6	69	B	volcanic sediments and basalt
10N/10E 19ACC	46°20.68′, 121°35.53′	200–305	2.05	0.10	4	55.7	114	B	volcanic sediments
10N/9E 21CAB	46°20.25′, 121°41.90′	25–148.6	1.97	0.10	5	67.8	133	C	Oligocene volcanic sediments
		90–145	1.97	0.10	5	60.2	119	B	
9N/9E 5DA	46°17.79′, 121°42.55′	35–152	1.29	0.06	7	43.0	55	B	Oligocene volcanic agglomerate
9N/5E 18BB	46°15.95′, 122°14.22′	30–122	1.81	0.06	15	20.1	36	B	andesite lahars
9N/9E 34AD	46°13.48′, 121°39.75′	30–152	1.46	0.07	7	59.4	87	B	CZ volcanic and volcanic sediments
8N/8E 8BB	46°11.69′, 121°49.75′	95–152	1.73	0.08	7	27.9	48	B	Oligocene volcanic sediments
7N/8E 2BCD	46°07.50′, 121°46.20′	85–150	1.25	0.03	5	44.5	56	A	vesicular basalt
7N/12E 9CCC	46°06.25′, 121°19.29′	25–141.6	1.34	0.07	5	47.2	63	C	basalt and sediments
		25–75	1.34	0.07	5	43.0	58	C	
7N/9E 17AA	46°05.90′, 121°42.00′	115–150	1.24	0.11	4	53.4	66	A	volcanoclastic
7N/8E 36C	46°02.90′, 121°45.00′	15–25	(1.17)			58.0	67	C	glacial sediment
6N/10E 7ACD	46°01.30′, 121°36.05′	55–150	1.30	0.07	5	18.3	24	B	basalt, sediments
6N/7E 23BAC	45°59.90′, 121°53.60′	100–150	1.23	0.06	6	49.8	61	A	basalt
6N/9E 25ACC	45°58.60′, 121°37.40′	100–150	1.42	0.03	8	52.7	75	A	volcanoclastic
4N/13E 24BC	45°49.40′, 121°07.72′	70–110	1.39	0.13	6	37.1	51	C	CR basalt
4N/7E 21CDB	45°48.70′, 121°57.25′	45–330	1.77	0.08	5	79.4	141	B	Oligocene volcanic sediments
4N/7E 21CD	45°48.63′, 121°57.19′	50–150	1.22	0.13	9	75.4	92	A	volcanoclastic mudflows
4N/9E 27CB	45°48.23′, 121°40.02′	125–152	1.40	0.07	5	330.0	46	B	Oligocene volcanic sediments
3N/5E	45°46.39′, 122°12.18′	25–305	2.80	0.08	5	23.2	65	A	quartz diorite granodiorite
			Average of Seven Holes in Same Section						
3N/10E 10CCB	45°45.36′, 121°32.68′	99–263	(1.59)			17.0	27	C	CR basalt
		199–263	(1.59)			21.5	34	C	
3N/8E 21BDD	45°44.07′, 121°48.24′	5–114	1.60	0.13	13	166.0	265	G	quartz diorite
3N/7E 25DA	45°42.88′, 121°52.90′	59–213	(1.26)			26.8	33	B	Volcanoclastic rocks

TABLE 1*a*. (continued)

Township/ Range Section	Latitude N, Longitude W	Depth Range, m	Average Thermal Conductivity, W m^{-1} K^{-1}	SE	*N*	Corrected Gradient, °C km^{-1}	Corrected Heat Flow, mW m^{-2}	Quality Rating	Lithology Summary
			Average of Seven Holes in Same Section (continued)						
3N/11E 29CCA	45°42.76', 121°27.61'	70–100	(1.59)			35.3	56	C	CR basalt
3N/7E 36BDD	45°42.34', 121°51.96'	15–290	(1.46)			32.7	48	C	Oligocene volcanics
3N/11E 34DBB	45°42.12', 121°24.66'	25–100	(1.59)			36.6	58	C	CR basalt
2N/7E 16CCC	45°39.16', 121°57.61'	80–155	(1.46)			82.4	120	G	volcanoclastic rocks
2N/7E 21BAC	45°38.90', 121°57.24'	50–195	(1.46)			129.6	190	G	volcanoclastic rocks
2N/7E 20DCA	45°38.43', 121°58.52'	100.6–179.9	1.40		5	49.3	69	B	volcanoclastic rocks
2N/5E	45°36.75',	30–50	(1.05)			54.2	57	C	basalt and sandy
36DDA	122°07.50'	50–90	(1.59)			45.5	72	C	clay
2N/5E	45°36.72',	30–55	(1.05)			55.9	59	B	clay and basalt
36DDB	122°07.65'	55–105				35.8		B	
		105–129	(1.05)			63.7	(67)	B	

CR, Columbia River; MI, Miocene; CZ, Cenozoic; JR, Jurassic; The section location is subdivided into quarters where a designation 12 ABCD indicates a well in the SE 1/4 of the SW 1/4 of the NW 1/4 of the NE 1/4 of section 12. SE is the standard error of the mean. Values in parentheses are estimated. *N* is the number of thermal conductivity samples. The gradients have all been corrected for topographic effects where necessary. The statistical errors of the heat flow values are usually very small so an estimate of the error is given as a quality rating of A (±5%), B (±10%), C (±25%) and G (geothermal system value, no error implied). More details for each site (including hole name, elevation, and uncorrected gradient) as well as data for sites of less than C quality are given by *Blackwell et al.* [1989].

part of the Middle to Upper Jurassic ophiolite complex [*Brandon et al.*, 1988] exposed on Fidalgo Island.

Throughout much of the Puget Lowland, there is a thick mantle of glacial till and gravel. Gradients are abnormally low within these deposits due to the generally active groundwater flow in these very porous and permeable units. This circulation causes lower temperatures at depth beneath these areas than would be expected based on the average gradient values found in the bedrock units. No data from this geologic setting are listed in Table 1.

Heat flow and geothermal gradient values for these provinces are shown in histogram form (from the data listed in Table 1) in Figure 3. Geothermal gradients range from 10° to 38°C km^{-1}. The higher gradients are usually associated with lower thermal conductivity rocks (such as clays). For example, in drill hole 12N/1W-9CCC gradients range from 18° to 33°C km^{-1} in association with thermal conductivity (lithology) variations.

The average geothermal gradient of 26.0° ± 1.2°C km^{-1} in this area is associated with the volcanic units that make up the bulk of the exposed rocks. Heat flow values are below normal, ranging from as low as 20 mW m^{-2} to as high as 55 mW m^{-2}. The average is 40.1 ± 1.8 mW m^{-2}, which is well below the typical continental average of 60 mW m^{-2}. The only thermal manifestations are warm springs in the Olympic Mountains. No heat flow or temperature gradient data have been obtained in the vicinity of these warm springs.

OKANOGAN HIGHLANDS

The Okanogan Highlands physiographic province occupies the northeastern quarter of the state and represents that part of the larger Northern Rocky Mountains province present in Washington. The geology is complex, reflecting the tectonic history of the Cordillera. The area has extensive exposures of supracrustal rocks of Paleozoic and Mesozoic age and of granitic basement rocks of Mesozoic and very early Cenozoic age. A few areas of Cenozoic volcanic rock exist, such as in the Republic graben and other similar extensional features. These features generally developed from 42 to 55 m.y. ago. The province has been quite stable during the last 30–40 m.y. The topographic relief is high, but the area is not extremely rugged. Unlike areas of similar geology and heat flow in central Idaho, there are no hot springs in the Okanogan Highlands. In addition, no holes in this province have a gradient in excess of 35°C km^{-1}.

Histograms of corrected gradients and heat flow are shown in Figure 4. The average geothermal gradient is 25.3 ± 0.9°C km^{-1}, and the average heat flow is 74.6 ± 2.4 mW m^{-2} (Table 2). The measured gradients for the holes of acceptable quality range from 14°C km^{-1} for hole 32N/40E-4ADD to 35°C km^{-1} in hole 39N/33E-13BCA. This range in gradient predominantly reflects variations in thermal conductivity and not variations in heat flow. In general, gradients of 30°C km^{-1} or higher are associated with Cenozoic volcanic and sedimentary rocks that have lower thermal conductivities than the older rocks. Gradients between 20° and 30°C km^{-1} are associated with granitic rocks that make up a large fraction of the bedrock of the Okanogan Highlands.

There is a small range of reliable heat flow values in the Okanogan Highlands. As discussed by *Blackwell* [1974, 1978], heat flow values are typical of those observed throughout the Cordilleran Thermal Anomaly Zone, which includes the Basin and Range province and the Northern Rocky Mountains in Idaho and Montana. No radioactivity determinations are reported as part of this study, but the correlation of radioactivity and heat flow in this province is described by *Blackwell* [1974]. Lower values of heat flow

TABLE 1b. Thermal and Location Data for the Heat Flow Sites in Washington Coastal Provinces

Township/ Range Section	Latitude N, Longitude W	Depth Range, m	Average Thermal Conductivity, W m^{-1} K^{-1}	SE	N	Corrected Gradient, °C km^{-1}	Corrected Heat Flow, mW m^{-2}	Quality Rating	Lithology Summary
34N/1E 1CBB	48°27.48', 122°38.04'	80–220	2.96	0.04	25	12.6 0.1	37	A	quartz diorite
20N/12W 8	47°14.27', 124°11.47'	304.8–1066.8	(1.30)			27.4	36	B	shale
16N/12W 24DAD	46°51.00', 124°06.00'	60–155	1.48	0.04	22	26.5	38	A	gravel, sand, and clay
13N/4W 7ABA	46°38.02', 123°13.57'	110–128	1.99		1	20.6	41	B	blue clay
13N/3W 35BAB	46°34.49', 123°01.51'	120–135	1.06		1	29.4	47	C	
12N/1W 7AAB	46°32.68', 122°50.81'	50–565	(1.26)			29.0	36	A	shale, sandstone, and siltstone
12N/1W 8DAA	46°32.35', 122°49.46'	110–365 260–578	(1.09) (1.46)	0.04 0.04	59	27.3 21.5	30 31	A A	shale, sandstone, and siltstone
12N/1W 9CCC	46°31.99', 122°48.66'	100–380 710–760	(1.09) 1.77	0.04 0.01	59 5	32.8 18.2	36 32	A A	shale, sandstone, and siltstone
12N/1W* 17BAD	46°31.75', 122°49.48'	90–348	(1.09)	0.04	59	34.8	38	A	shale, sandstone, and siltstone
12N/1W 17ADA	46°31.65', 122°50.07'	90–340 340–690 690–847	(1.09) (1.46) (1.76)	0.04 0.01	59 5	32.7 24.8 22.2	36 36 39	A A A	shale, sandstone, and siltstone
5N/1E 5CDC	45°56.41', 122°42.87'	100–177	(1.72)		1	21.5	37	B	sandstone and shale
5N/1E 23CBA	45°54.12', 122°39.35'	120–246	(1.30)		1	32.1	42	B	sandstone and shale
4N/2E 14BA	45°50.22', 122°31.72	122–192 192–215	(1.05) (1.59)			38.5 25.6	40 41	B B	clay and basalt
2N/3E 21DC	45°38.33', 122°26.42'	30–73	1.34	0.09	9	37.0	49	B	clay, basalt, and gravel
1N/3E 2DB	45°35.92', 122°23.88'	60–150	1.59	0.04	17	31.6	50	A	basalt and claystone

See Table 1a footnotes.
*The thermal conductivity values for the wells in 12N/1W were estimated from *Sass et al.* [1971].

(65–70 mW m^{-2}) are associated with lower values of heat generation. The highest heat flow values, particularly those near the Spokane River in Townships 27N, 37, and 38E (within the 80 mW m^{-2} contour in Figure 1), are associated with granites with heat production values that range up to 4 μW m^{-3} [*Steele*, 1975]. Similar heat flow and radioactivity correlations have been reported for the southern Canadian Cordillera by *Davis and Lewis* [1984] and *Lewis et al.* [1985]. The mantle heat flow in this province is on the order of 60 mW m^{-2}, and the slope on the Q-A plot is about 10 km, similar to other interior provinces in the Cordillera [*Blackwell*, 1978].

COLUMBIA BASIN

Introduction

The Columbia Basin occupies the southeastern third of the state of Washington. Irrigation development of groundwater aquifers that generally occur along flow contacts between basalt layers and in interbedded sedimentary rocks has been widespread. Because of the large number of wells and the importance of the groundwater resources, extensive hydrologic studies have been carried out in various parts of the Columbia Basin. However, the regional setting of the detailed hydrologic studies and the interrelationships of the many aquifers and areas of development are subjects for which there is little information. The complexities and state of knowledge of the aquifer system have been succinctly described by *Lindholm and Vaccaro* [1988]. In the early 1980s, great interest was shown in the area for gas exploration, and several wildcat wells have been drilled with the deepest hole reaching 5340 m [e.g., *Haug and Bilodeau*, 1985], but reliable temperature data for these wells have not been measured. In this case, knowledge of the subsurface temperatures and geothermal gradients is important in evaluating the maturation level of organic material occurring in the rocks beneath the basalt.

The thermal data for the Columbia Basin have been collected by three different organizations: personnel from the U.S. Geological Survey Water Resources Division (USGS-WRD), Tacoma, Washington; personnel from the Department of Hydrology at Washington State University (WSU), Pullman; and personnel from Southern Methodist University (SMU)/Washington Department of Natural Resources. Some aspects of the copious WSU data base have been discussed briefly by *Stoffel and Widness* [1983*a*, *b*] and *Widness* [1983]. Many of the well logs have been presented by *Biggane* [1981, 1983] and *Stoffel and Widness* [1983*b*].

Complexity of Temperature Logs

Blackwell et al. [1985] compiled an extensive data base of temperature logs in an attempt to evaluate the background thermal conditions of the Columbia Basin and look for local

TABLE 1*c*. Thermal and Location Data for the Heat Flow Sites in Okanogan Highlands

Township/ Range Section	Latitude N, Longitude W	Depth Range, m	Average Thermal Conductivity, $W\ m^{-1}\ K^{-1}$	SE	*N*	Corrected Gradient, $°C\ km^{-1}$	Corrected Heat Flow, $mW\ m^{-2}$	Quality Rating	Lithology Summary
40N/27E 6BDC	48°59.74′, 119°29.19′	60–180	3.16	0.08	19	22.3	70	A	greenstone schist
40N/33E 2ACD	48°59.71′, 118°35.91′	75–205	3.17	0.13	22	22.7	72	A	Mesozoic greenstone
40N/33E 2DBB3	48°59.67′, 118°35.95′	80–215	3.15	0.13	20	22.5	71	A	Mesozoic greenstone
39N/33E 14ABB	48°53.11′, 118°36.03′	14–74	(2.30)			29.6	68	C	granite
39N/33E 13BCA	48°52.84′, 118°35.32′	39–89	(2.30)		1	34.5	79	B	granite
37N/26E 8DBC	48°43.01′, 119°35.46′	165–435	3.50	0.13	26	21.5	75	A	biotite granodiorite
37N/26E 24DCA	48°41.15′, 119°29.95′	24–123	(3.18)		1	23.3	74	B	schist and granite
37N/32E 33AAC	48°39.99′, 118°46.38′	150–260	2.41	0.04	17	31.1 0.2	76	A	Cenozoic andesite
36N/20E 19ADC	48°36.48′, 120°23.07′	140–275	2.88	0.08	15	26.7	77	A	Mesozoic metamorphics
36N/20E 19DAB	48°36.35′, 120°23.15′	240–360	3.19	0.29	4	23.0	73	A	Mesozoic metamorphics
34N/26E 23DA	48°25.77′, 119°30.96′	104–184	(1.97)		1	27.0	53	B	granite
33N/31E 14BDA	48°21.75′, 118°52.59′	130–230	3.36	0.13	17	20.3 1.0	68	B	argillite and quartz monzonite
33N/31E 14ACB	48°21.75′, 118°52.35′	205–270	3.11	0.08	9	22.4	69	A	quartz monzonite porphyry
33N/31E 14BDC	48°21.65′, 118°52.78′	120–200	3.36	0.13	18	19.5	66	A	argillite and quartz monozonite
32N/40E 4ADD	48°18.13′, 117°45.64′	49–117	(4.39)		1	14.1	62	B	quartzite
31N/44E 34ADD	48°08.77′, 117°13.27′	95–155	(2.05)		1	27.7	57	B	granite
29N/37E 36DDD	47°58.50′, 118°04.00′	60–380	3.13		25	27.0	85	A	phyllite-argillite
28N/42E 5D	47°56.90′ 117°31.55′	19–144	(2.43)			27.5	66	B	sediment/clay to granite
28N/37E 9DBD	47°56.60′, 118°09.20′	100–133	3.00	0.04	6	25.1	76	B	phyllite-argillite
28N/44E 31ADD	47°52.99′, 117°16.98′	154–264	(2.76)		1	24.9	68	B	sediment/clay to granite
27N/37E 2BBB	47°52.40′, 118°07.40′	90–150	3.31	0.04	6	26.7	87	B	porphyritic quartz monzonite
27N/37E 3AAA	47°52.40′, 118°07.40′	60–100	3.26		6	27.8	91	B	porphyritic quartz monzonite
27N/38E 28BBB	47°49.00′, 118°01.50′	100–145	2.54		2	33.0	84	B	Paleozoic marb-hornfel
26N/44E 12AAC	47°46.10′, 117°10.80′	120–159	(3.26)		1	27.2	89	B	granite
24N*44E 6BC	47°36.30′, 117°17.99′	64–81	(2.38)		1	20.0	48	C	granite
24N/44E 10AD	47°35.38′, 117°13.22′	25–147	(3.43)		1	30.1	103	B	granite

See Table 1*a* footnotes.

thermal anomalies that might indicate higher than average potential for geothermal energy in particular areas. In spite of this extensive data base, however, there are still some uncertainities about the thermal conditions in the Columbia Basin. Most of these uncertainties arise from the fact that all the holes logged were drilled as water wells, and casing and cementing is minimal. Furthermore, there are no holes that have been continuously cored and from which reasonably accurate in situ thermal conductivity estimates can be obtained except on the Hanford reservation [*Sass et al.*, 1971]. Hence we have little idea of possible vertical or lateral variations in thermal conductivity that undoubtedly must occur due to interbedded sediments and the different proportions of vesicular versus nonvesicular basalts.

The most serious problem, however, is the nature of the fluid flow associated with the basalt aquifers. On a local scale, permeability across individual basalt flows is quite low, whereas along contacts between flows permeability is high. The generality of these conditions is indicated by the ubiquitous upward or downward flow between aquifers made possible by the open hole nature of the well completions. Consequently temperature logs for water wells in the

TABLE 1*d*. Thermal and Location Data for the Heat Flow Sites in Columbia Basin

Township/ Range Section	Latitude N, Longitude W	Depth Range, m	Average Thermal Conductivity, W m^{-1} K^{-1}	SE	*N*	Corrected Gradient, °C km^{-1}	Corrected Heat Flow, mW m^{-2}	Quality Rating	Lithology Summary
26N/34E	47°45.87′,	109–199	(1.59)		1	55.4	88	B	CR basalt?
10CAB	118°31.18′								
26N/33E	47°45.20′,	25–129	(1.59)			31.9	51	C	CR basalt
18ACA	118°42.50′	100–129	(1.59)			52.9	84	C	
24N/36E	47°34.70′,	66–265	(1.59)			53.7	85	B	CR basalt
16AAB	118°16.37′								
24N/31E	47°34.55′,	19–149	(1.59)			52.8	84	B	CR basalt
16BCC	118°56.07′	19–204	(1.59)			58.3	93	B	
23N/43E	47°30.41′,	95–125	(1.34)		1	41.6	56	B	CR basalt
8AB	117°23.81′								
22N/20E	47°22.11′,	310–900	2.18	0.21	18	28.4	62	A	rhyolite, arkose,
26CBB	120°18.00′								and sandstone
21N/31E	47°20.00′,	61–1250	1.67	0.21		42.0	70	B	CR basalt
10CB	118°55.00′								
20N/22E	47°14.65′,	49–264	(1.59)		1	32.0	51	C	CR basalt
12ABC	120°00.78′								
19N/17E	47°07.27′,	10–170	(1.38)		1	29.4	41	B	
23CBD	120°40.88′								
18N/26E	47°00.26′,	70–100	(1.59)			49.8	79	C	CR basalt?
35C	119°31.73′								
16N/19E	46°53.37′,	59–333	(1.55)		1	>27.3	>42	C	CR basalt
12DB	120°23.45′	59–154	(1.55)		1	29.4	46	C	
16N/31E	46°52.75′,	10–230	(1.59)			59.1	94	C	CR basalt
15B	118°55.10′								
16N/21E	46°49.63′,	10–100	1.38		4	37.2	51	B	
33CD	120°12.37′	40–90	1.38		4	44.8	62	B	
15N/17E	46°46.90′,	40–22	(1.59)			36.5	58	B	sandstone and clay
23ABC	120°39.95′								
15N/17E	46°45.13′,	75–595	(1.46)			31.6	46	B	clay, sand, gravel
36AAA	120°38.49′								
14N/16E	46°43.77′,	20–110	(1.46)			35.9	53	C	andesite
1CBD	120°46.74′								
14N/44E	46°43.72′,	54–89	(1.59)			22.9	36	C	CR basalt
1BCC	117°13.28′								
14N/18E	46°40.40′,	20–55	(1.46)			39.0	57	C	sand and clay
29DBB	120°36.05′								
13N/19E	46°37.50′,	50–160	(1.59)			32.8	52	B	CR basalt
11DBC	120°24.75′								
13N/29E	46°36.77′,	10–210	(1.59)			48.3	77	C	CR basalt
13DB	119°07.28′								
12N/17E	46°33.06′,	80–120	(1.46)			64.2	94	C	gravel, clay, and
11BBB	120°40.11′								sand
12N/20E	46°31.56′,	100–400	(1.59)			41.5	66	C	CR basalt
16CA	120°19.54′								
12N/22E	46°31.25′,	60–200	(1.59)			33.3	53	B	
13CDC	120°00.71′								
12N/20E	46°28.64′,	10–350	(1.59)			32.3	51	C	sediments and CR
36CD	120°15.78′								basalt
11N/21E	46°27.53′,	85–450	(1.59)			32.9	52	B	CR basalt
7AAC	120°13.50′								
11N/21E	46°26.10′,	130–210	(1.59)			41.0	65	C	sediments and CR
16CDB	120°12.15′								basalt
11N/21E	46°25.84′,	150–330	(1.59)			28.8	44	B	sediments and CR
22AC	120°10.53′								basalt
11N/28E	46°25.68′,	20–280	(1.59)			37.7	60	B	CR basalt
23BB	119°16.62′								
11N/22E	46°25.00′,	50–270	(1.59)			49.4	79	C	CR basalt
30BA	120°07.13′	50–150	(1.59)			43.9	69	B	
11N/22E	46°24.57′,	27–210	(1.59)			39.7	63	B	CR basalt
28CB	120°04.51′								
11N/46E	46°23.49′,	15–405	(1.59)			35.6	56	B	basalt with clay
32BCA	117°04.52′								
11N/45E	46°23.28′,	10–192	(1.59)			32.0	51	B	CR basalt
32DAB	117°11.45′								
10N/41E	46°22.37′,	10–90	(1.59)			27.2	43	C	CR basalt
3DBD	117°39.50′								
10N/28E	46°21.97′,	5–120	(1.59)			34.3	54	C	CR basalt
10ACC	119°17.43′								
10N/28E	46°21.73′,	110–605	(1.59)			41.1	66	B	CR basalt
14ABB	119°16.13′								

TABLE 1*d*. (continued)

Township/ Range Section	Latitude N, Longitude W	Depth Range, m	Average Thermal Conductivity, W m^{-1} K^{-1}	SE	*N*	Corrected Gradient, °C km^{-1}	Corrected Heat Flow, mW m^{-2}	Quality Rating	Lithology Summary
10N/24E 36BD	46°18.70′, 119°45.43′	125–200	(1.59)			34.0	54	C	CR basalt
10N/39E 32BCA	46°18.33′, 117°57.87′	49–229	(1.59)		1	31.6	50	C	CR basalt
9N/32E 13BA	46°16.04′, 118°45.21′	135–210	(1.59)			34.8	55	B	basalt with interbeds
9N/27E 23CA	46°14.82′, 119°24.17′	160–350	(1.59)			43.1	69	B	CR basalt
8N/33E 21D	46°09.31′, 118°40.99′	50–200	(1.59)			37.7	60	B	CR basalt
8N/25E 36AB	46°08.38′, 119°37.78′	5–240	(1.59)			35.0	46	C	
7N/46E 2AA	46°07.99′, 116°59.88′	160–275	(1.59)			36.1	57	C	CR basalt(?)
7N/36E 17CAD	46°05.39′, 118°20.30′	100–225	(1.59)			35.4	56	B	CR basalt
7N/46E 13BA	46°05.25′, 116°59.01′	15–85	(1.59)			40.8	65	C	CR basalt
7N/35E 25AAC	46°03.66′, 118°22.22′	85–260	(1.59)			34.0	54	C	CR basalt
6N/33E 1DBD	46°01.46′, 118°37.30′	60–305	(1.05)			69.7	73	B	cemented gravel/ clay
6N/33E 10AD	46°01.01′, 118°40.65′	20–85	(1.05)			106.8	112	C	cemented gravel
6N/19E 24BC	45°59.52′, 120°22.97′	59–139	(1.59)		1	41.2	65	C	CR basalt
5N/21E 16CAB	45°55.00′, 120°11.49′	74–224	(1.59)			49.2	78	D	CR basalt
		224–394	(1.59)			27.2	44	C	
5N/14E 22BCB	45°54.46′, 121°02.86′	40–137.5	(1.59)		1	31.2	50	B	CR basalt
5N/15E 28DDD	45°53.04′, 120°55.60′	55–145	(1.59)			24.7	39	B	Quaternary basalt
3N/13E 13DA	45°44.66′, 121°06.95′	160–195	(1.59)			39.0	62	C	CR basalt
3N/13E 21AB	45°44.23′, 121°10.94′	65–150	(1.59)			33.5	53	C	CR basalt
3N/21E 19BAB	45°44.12′, 120°13.55′	30–70	(1.59)			53.8	85	C	CR basalt

See Table 1*a* footnotes.

Columbia Basin are typically complex and are difficult to interpret. Recharge to the aquifers is quite distant from the points of development of the aquifers, and flow paths within the aquifer system must be complex and time dependent.

When the holes are drilled, drilling stops when water flow of appropriate quantity is encountered. As a consequence, water wells almost without exception bottom in or just below aquifers (sites of potential up-or-down flow in the hole). Another limitation with many of the temperature logs measured by WSU and USGS-WRD is that logging starts at the

TABLE 1*e*. Thermal and Location Data for the Heat Flow Sites in Northern Washington Cascade Range

Township/ Range Section	Latitude N, Longitude W	Depth Range, m	Average Thermal Conductivity, W m^{-1} K^{-1}	SE	*N*	Corrected Gradient, °C km^{-1}	Corrected Heat Flow, mW m^{-2}	Quality Rating	Lithology Summary
38N/8E 20	48°45.85′, 121°48.30′	10–140	2.33	0.10	5	(309)	720	G	Quaternary volcanics
26N/13E 27BA	47°43.20′, 121°07.21′	40–150	2.68	0.17	7	26.1	70	B	biotite schist
26N/13E 28CD	47°42.66′, 121°08.50	20–100	2.06	0.17	7	48.4	100	B	phyllite, near hot spring
		60–100	2.06	0.17	7	55.6	115	B	
25N/9E 4ADA	47°40.97′, 121°38.92′	120–145	3.03		2	20.0	61	C	greenstone
22N/11E 4BDA	47°25.54′, 121°24.80′	110–145	2.97	0.21	7	14.7	44	B	phyllite

See Table 1*a* footnotes.

TABLE 1f. Thermal and Location Data for the Low Heat Flow Sites in Southern Washington Cascade Range

Township/ Range Section	Latitude N, Longitude W	Depth Range, m	Average Thermal Conductivity, $W\ m^{-1}\ K^{-1}$	SE	N	Corrected Gradient, $°C\ km^{-1}$	Corrected Heat Flow, $mW\ m^{-2}$	Quality Rating	Lithology Summary
16N/4E 23DDD	46°51.01′, 122°15.36′	25–86.5	(1.97)		2	10.2	21	C	basalt
15N/4E 14DDA	46°46.78′, 122°15.32′	100–122	1.72	0.17	2	19.8	34	C	basalt
13N/5E 18ABD	46°36.96′, 122°13.63′	10–55	(1.09)			26.9	29	C	Eocene(?) sedimentary
12N/2E 5DCD	46°32.86′, 122°34.73′	125–155	(1.67)			25.4	38	B	Tertiary volcanics
12N/2E 9DAD	46°32.07′, 122°33.17′	60–205	(1.67)			23.0	38	B	basalt
12N/3E 17DCB	46°31.33′, 122°27.54′	90–138.5	(1.67)			20.5	34	B	Eocene(?) basalt
12N/3E 19CBD	46°30.44′, 122°29.23′	65–232.5	(1.67)			27.0	45	C	Eocene-Oligocene volcanics
10N/4E 2DBA	46°22.65′, 122°15.95′	15–141	1.95	0.19	4	36.9	72	B	
9N/2W 1ADB	46°17.72′, 122°52.14′	20–135	(1.59)		1	14.2	23	C	clay, sandstone, and basalt
8N/4E 14DD	46°10.37′, 122°16.10′	50–150	1.85	0.06	17	24.2	45	B	
5N/2E 25DBC	45°53.16′, 122°30.08′	70–156	(1.59)			32.0	51	B	basalt?
5N/3E 28CCC	45°52.96′, 122°27.00′	70–97.5	(2.01)			20.4	41	C	
4N/3E 20ADD	45°49.10′, 122°27.33′	135–220	(1.59)		1	38.0	60	B	Eocene basalt
3N/3E 21DAB	45°43.77′, 122°26.23′	30–103	(1.59)			27.0	42	B	Eocene basalt
3N/3E 21DBD	45°43.60′, 122°26.40′	30–188	(1.59)		1	27.0	42	B	Eocene basalt
2N/3E 12CAD	45°40.03′, 122°22.58′	25–127	(1.59)		1	23.1	37	B	Eocene basalt

See Table 1a footnotes.

water table. Thus, if the well has flow up to, or down from, the water table, the first measured temperature may be much different from in situ rock temperature. Temperature logs of many of the wells consist of a vertical line (isothermal). Given the propensity of fluid to flow within the wells following completion, it becomes virtually impossible to decipher the nature of the thermal and fluid flow conditions existing at a particular location from a single well. Based on local aquifer conditions within the wells, gradients both higher and lower than the "true" value might be encountered. Thus only the overall average gradients can be determined with the present data set.

Temperature logs clearly show differences in aquifer heads and also demonstrate that these aquifer head differences may change with time. Thus it might be expected that local vertical leaks along dikes, in fractures zones, or at erosional gaps in flows could cause local fluid circulation anomalies. However, no convincing evidence has been discovered so far for geothermal anomalies associated with fluid circulation of this sort.

Some of the complexities of temperature-depth curves for a single well are illustrated by reference to DABE-1 (21N/31E-10CB, see Figure 5). This well was drilled as a hydrocarbon exploration test in 1960. It was first logged by R. F. Roy in 1961. Subsequently, the hole had a checkered history that included its being plugged with railroad cross ties for a time by local ranchers because they could hear fluid cascading down the hole. It was opened by the USGS-WRD for hydrologic tests in 1972, at which time it was logged by the WSU group. The hole was then plugged back to a depth of 730 m and is maintained as a water level monitor well by the USGS-WRD. Well logs measured in 1961, 1972, and 1981 are shown in Figure 5. In 1961, water entered the hole at the water table (about 40 m), flowed down the hole to a depth of 700–800 m, and exited into the formation. Below 800 m, the gradient is nearly constant, and the temperature-depth curve is linear except for the bottom part of the log. The bottom isothermal section may have been due to additional intrahole water flow or due to the fact that the cable had hung up (and this fact was not recognized). The hole was logged to the end of the cable at about 1250 m. By 1972 the flow pattern had changed dramatically, and water flow was both up and down from an entry point at about 500 m. The rate of flow down the hole was quite low and occurred between depths of about 500 and 800 m. Artesian (up) flow occurred between 500 and 300 m. The presence of upflow is indicated by temperatures above the interpolated temperature-depth curve from below 900 m and the surface temperature intercept (about 12°C). The curved temperature-depth relationship between 200 and 300 m and the fact that the observed temperature is higher than the interpolated temperature suggest upflow, with the 200–300 m portion of the hole being heated by the upward flowing warm water (from 500 m). The sharp negative-slope part of the curve between 500 and 600 m is very difficult to

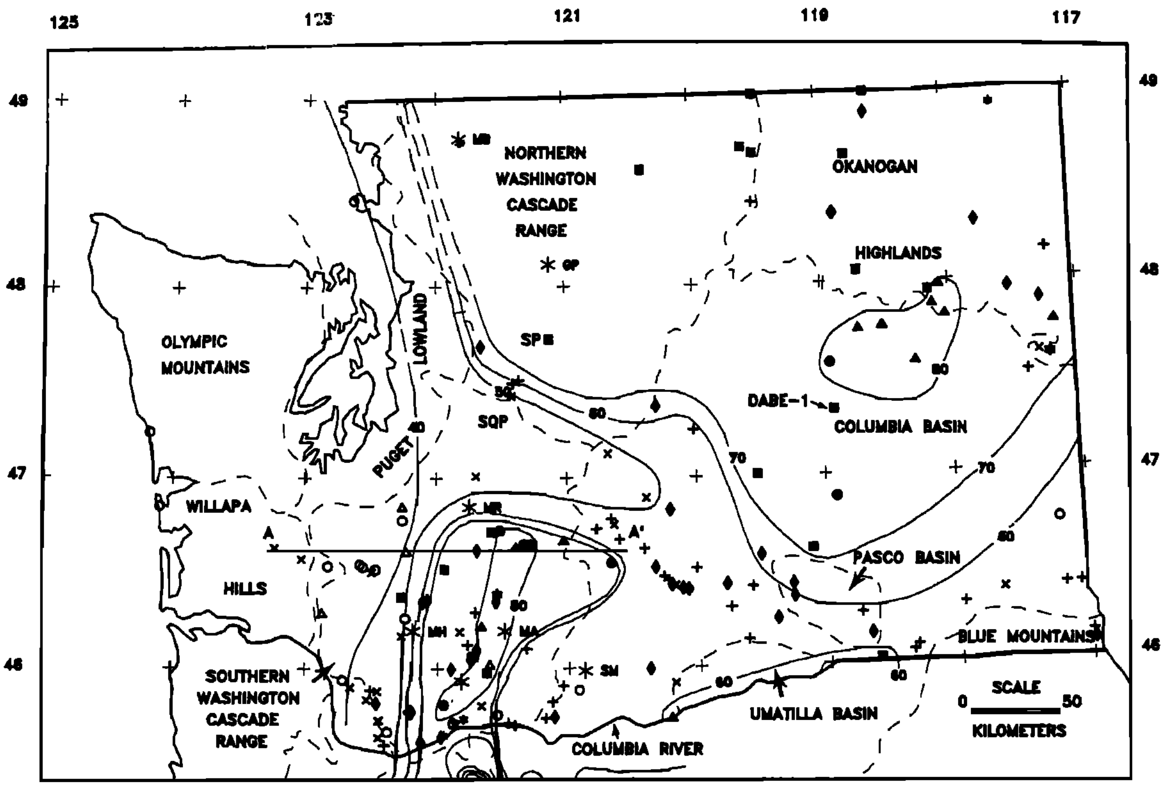

Fig. 1. Heat flow map of Washington. Physiographic provinces are outlined with dashed lines and identified. The Pasco and Umatilla basins are subprovinces of the Columbia Basin. The volcanoes in the Cascade Range (large asterisks with a two-letter identification) are also shown for reference. Locations of Stevens Pass (SP), Snoqualmie Pass (SQP), Simcoe Mountains (SM), and hole DABE-1 are identified. Data of acceptable quality (Table 1) are plotted on this figure. The map is contoured at 10 mW m^{-2} intervals. Heat flow symbols are triangle, 20–30 mW m^{-2}; circle, 30–40 mW m^{-2}; cross, 40–50 mW m^{-2}; pluses, 50–60 mW m^{-2}; solid diamond, 60–70 mW m^{-2}; solid square, 70–80 mW m^{-2}; solid triangle, 80–90 mW m^{-2}; dot, 90–100 mW m^{-2}; star, 100–110 mW m^{-2}; small asterisk, >110 mW m^{-2}. The location of the cross section in Figure 7 is indicated.

explain unless up-or-down flow in an aquifer outside the well bore is included. The temperature data between 100 and 200 m may represent the background gradient unaffected by water flow. Following plugging of the hole and its development as a water level monitor well, the flow pattern in the hole changed again. When the well was logged during 1981, there was downflow through most of the hole, with water entering at the water table and exiting at two locations, between 210 and 300 m and at a depth of approximately 450 m. An aquifer at 300 m was either donating or extracting fluid.

Heat Flow and Geothermal Gradients

Based upon analysis of the temperature logs [*Blackwell et al.*, 1985], the gradient range throughout the Columbia Basin is between 30° and 55°C km^{-1}, with a mean value of 40.6° ± 1.8°C km^{-1}. The mean heat flow, excluding data from the western edge of the province is 62.2 ± 2.0 mW m^{-2}. Histograms of geothermal gradient are given in Figure 6. There is a lack of detailed thermal conductivity information, so it is not clear whether the variation in gradients is associated with thermal conductivity variations within the basalts or between basalt and sedimentary rocks, or whether it reflects genuine changes in heat flow in various locations throughout the plateau. There is some evidence, as presented in Figures 1 and 2, for higher than average gradients on the western and eastern margins of the province. The results of this study suggest that the high predicted SiO_2 heat flow values in the Columbia Basin [*Swanberg and Morgan*, 1979] are related to the high gradients.

There are two interpretations of the heat flow in the Columbia Basin. The first is that the measured heat flow values reflect the conductive heat flow from the crust below the Miocene basalt. This hypothesis is supported by the fact that the heat flow variations are consistent with probable regional variations of crustal radioactivity. Heat flow values are generally 50–60 mW m^{-2} in areas where the crust is more mafic [*Catchings and Mooney*, 1988] and along the southern and western margins of the province where the crust appears to be tonolitic in composition (west of the 0.706 line of initial Sr isotope ratios [*Kistler and Peterman*, 1978]) and most of the exposed granitoids are very low in heat generation [*Swanberg and Blackwell*, 1973]. Similar heat flow values are found in western Idaho where such rocks (of Mesozoic age) crop out at the surface [*Blackwell et al.*, 1990]. On the other hand, along the north central part of the Columbia Basin, where granitic rocks of higher radioactivity occur, heat flow values are 70–93 mW m^{-2}, typical of those found in basement areas of the Okanogan Highlands in northeastern Washington [*Blackwell*, 1974; *Steele*, 1975].

The second major effect which may influence the heat flow is transfer of heat by groundwater flow. The high permeabil-

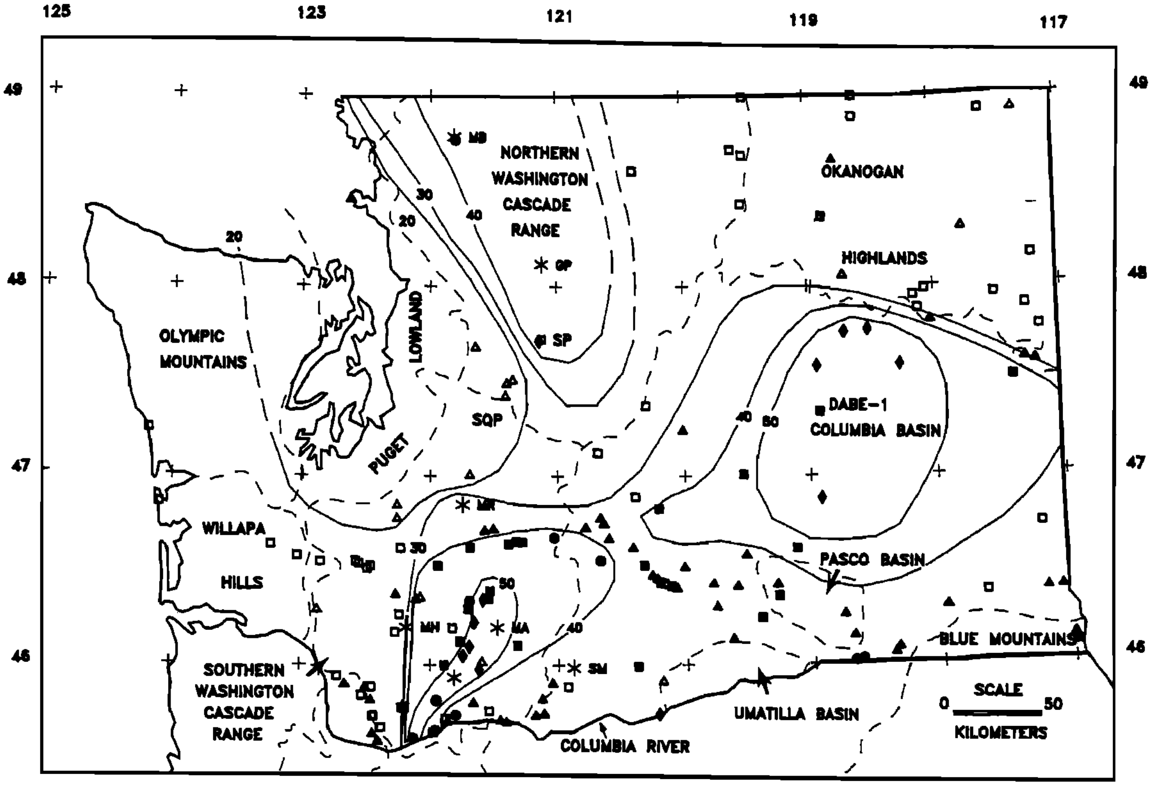

Fig. 2. Geothermal gradient map for Washington (same base as Figure 1). Geothermal gradients from holes or sites of acceptable quality (Table 1) are shown. Generalized gradient contours are shown. Gradient symbols are circle, 0°–10°C km^{-1}, open triangle, 10°–20°C km^{-1}; open square, 20°–30°C km^{-1}; solid triangle, 30°–40°C km^{-1}; solid diamond, 50°–60°C km^{-1}; dot, 60°–70°C km^{-1}.

ity of the basalt flow interbeds and the general lack of coincidence of piezometric levels certainly indicate potential for groundwater flow. Clearly, fluid flow must effect local areas, but in spite of the many holes drilled in the basalt, no anomalously hot wells have been reported. Any flow that exists must be relatively subtle. There are several high values of heat flow along the Snake River in the Umatilla subprovince; these may indicate some flow toward the topographically low areas along the river.

Based on the data available at this time, we conclude that conductive heat transfer dominates on a province wide scale, although in local areas convection might be significant. The surface heat flow pattern reflects the radioactivity of the crust below the basalt and the back arc regional setting. Because the heat flow from below the upper crustal radioactivity layer is interpreted to be similar for the province, even though the surface heat flow varies from northeast to southwest, the Columbia Basin is a single heat flow province in the sense defined by *Roy et al.* [1972].

WASHINGTON CASCADE RANGE

Northern Washington Cascade Range, Snoqualmie Pass Area, Columbia River

Within the state of Washington there are differences in the geologic nature of the Cascade Range. South of approximately 47°20′, the basement of the Washington Cascade Range is predominantly composed of early to middle Cenozoic volcanic rocks [*Walsh et al.*, 1987; *Smith*, 1989]. The major Quaternary stratovolcanoes are Mount Adams, Mount St. Helens, and Mount Rainier. The two principal late Cenozoic basalt centers are the Quaternary Indian Heaven area in the south central part of the range and the Pliocene Simcoe volcanic field in the southeastern part of the range. North of Mount Rainier, the bedrock geology is predominantly pre- to mid-Cenozoic crystalline rocks, including both granitic intrusive rocks and metamorphic rocks. The only late Cenozoic volcanic rocks are the andesite-dacite volcanoes, Glacier Peak and Mount Baker.

The Washington Cascade Range is the province of greatest geothermal activity in the state, with active or dormant volcanic centers, and several hot springs. There are only a few heat flow sites in the northern Washington Cascade Range, and only three of them are far enough west to be near the areas of late Cenozoic volcanism. Two of the three are associated with hot spring areas [*Korosec*, 1983]. At Stevens Pass (marked SP on Figure 1; sites 26N/13E in Table 1) one of the heat flow values is anomalously high and is obviously affected by the hot spring circulation, whereas the other heat flow value is typical of those seen in the southern Washington Cascade Range. The most northerly hole in Washington is on the slopes of Mount Baker near Baker Hot Springs. It is also affected by the hot spring circulation and does not give a typical background value for that area [*Korosec*, 1983]. The sparse results suggest that the regional heat flow may be similar to that found in the British Columbia Cascade Range (about 80–90 mW m^{-2} [*Lewis et al.*, 1988]).

Unlike the situation in Oregon, where the high heat flow zone has been identified for a north-south distance of at least 200 km, the high heat flow zone along the Cascade axis in

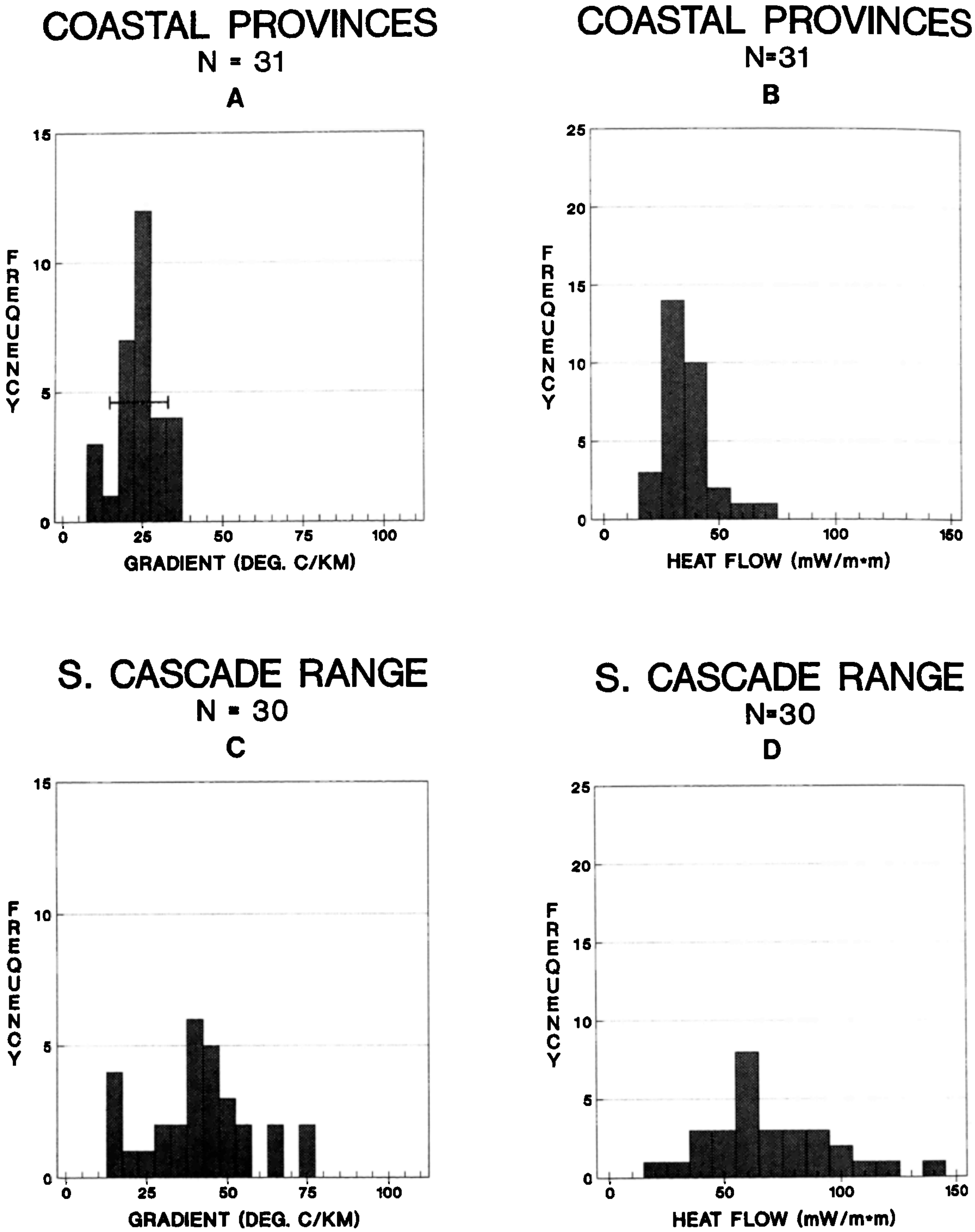

Fig. 3. Histograms of geothermal gradient and heat flow from Table 1. (*a*) and (*b*) For the coastal provinces. Provinces included are the Puget Lowlands, Olympic Mountains, Willapa Hills, and the Cascade Range west of the high heat flow zone. Bar on gradient plot shows range in gradient in hole 12N/1W-9CCC associated with lithologic variations. (*c*) and (*d*) For the high heat flow region in the southern Washington Cascades.

Washington seems to be segmented by areas of normal heat flow around Snoqualmie Pass (SQP on Figure 1) and the Columbia River. For many years the only heat flow values in the Washington Cascade Range were near Snoqualmie Pass [*Blackwell*, 1969]. Heat flow values there are approximately normal for continents, and there appeared to be no significant thermal anomaly in the Cascade Range. These data stood uncontested until drilling for heat flow measurements began in the late 1970s [*Blackwell et al.*, 1982; *Schuster et al.*, 1978].

Three new measurements are low/normal and confirm the values obtained by *Blackwell* [1969]. One hole is northwest of the Snoqualmie Pass area at a mineral exploration location, and two holes (22N/11E-4BDA and 23N/11E-33CAB) were drilled east of the pass as part of this project. Finally, the apparent south edge of the region of normal heat flow is

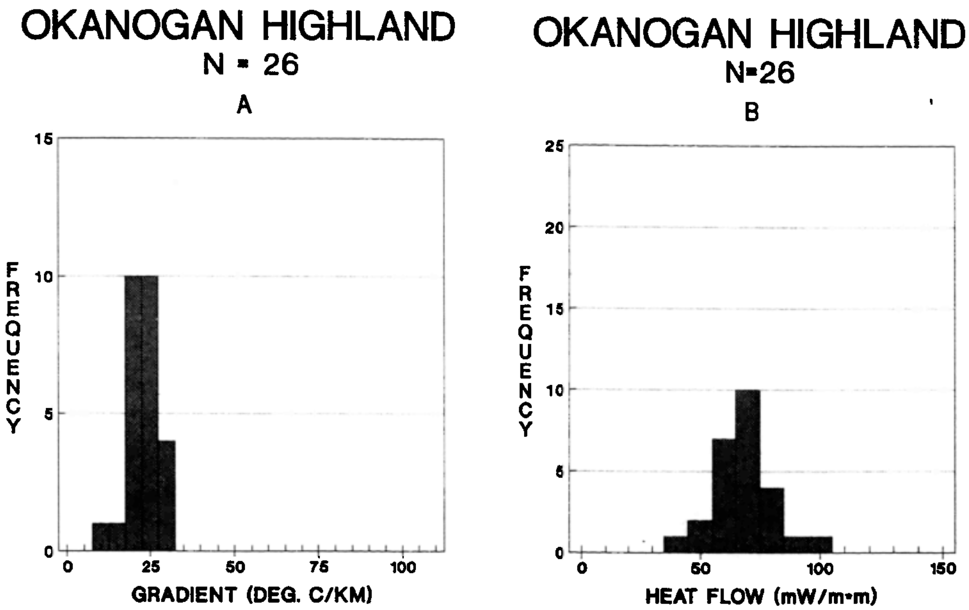

Fig. 4. Histograms of (*a*) geothermal gradient and (*b*) heat flow from Table 1 for the Okanogan Highlands.

poorly defined by a single site north of Mount Rainier. The reason for the lower values observed in the Snoqualmie Pass vicinity is not obvious. The values could be local anomalies due to water circulation, although this explanation seems unlikely since the holes are scattered over a large area and are in a variety of topographic settings. A discontinuous regional heat source at a depth is an alternative explanation for these variations in heat flow. The area contains a structurally complex zone of transition between the southern and northern Washington Cascade Range areas and is cut by several strong lineaments, but there are no late Cenozoic volcanics in the area.

Another anomalous zone of heat flow along the Cascade Range is associated with the Columbia River. East of 122° (excluding holes drilled near the Bonneville and Trout Creek hot springs, marked with a G quality in Table 1), most heat flow values along the Columbia River are less than 60–70 mW m^{-2}, including those at the south edge of the Simcoe volcanic field in Washington and north of the Mount Hood area in Oregon [*Blackwell et al.*, 1982, this issue]. The area is most clearly defined on the gradient maps (Figures 2 and 9). Temperature-depth curves from holes along the Columbia River north and northeast of Mount Hood generally have low gradients and are often curved, possibly indicating regional downflow of water or recent climatic warming. Temperature-depth curves from holes in the Quaternary volcanics often are very complicated and in some cases show evidence of transient lateral flow of warm water. Because no deep holes have been drilled along the Columbia River, or in the Simcoe volcanic field, the observed heat flow pattern may need some revision when such data are obtained. Geothermal conditions need to be tested by drilling

TABLE 2. Summary of Gradient and Heat Flow for Various Areas

	Gradient, °C km^{-1}			Heat Flow, mW m^{-2}		
	Average	SE	*N*	Average	SE	*N*
Coastal provinces (all low values)	26.0	1.2	31	40.1	1.8	31
Southern Washington Cascade Range (high heat flow region)	44.5	3.0	30	75.4	4.4	30
(Columbia River)	36.9	4.8	9	53.6	4.9	9
Okanogan Highlands	25.3	0.9	26	74.5	2.4	26
Columbia Basin (all)	40.6	1.8	55	62.2	2.0	55
Columbia Basin (western margin)	32.8	2.3	7	51.0	1.7	5

SE is the standard error and *N* is the number of data points averaged. Quality A, B, and C data are averaged, but the averages do not change significantly if only A and B quality data are used.

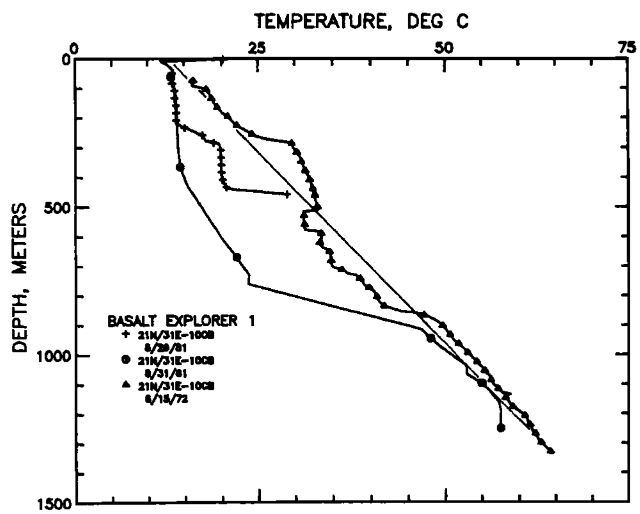

Fig. 5. Temperature-depth curves for Development Associates Basalt Explorer 1. Three logs measured over a 20-year period are illustrated. Logs were obtained by R. F. Roy (1961), personnel from Washington State University (1972), and SMU (1981). Every fifth point on each log is plotted by a symbol.

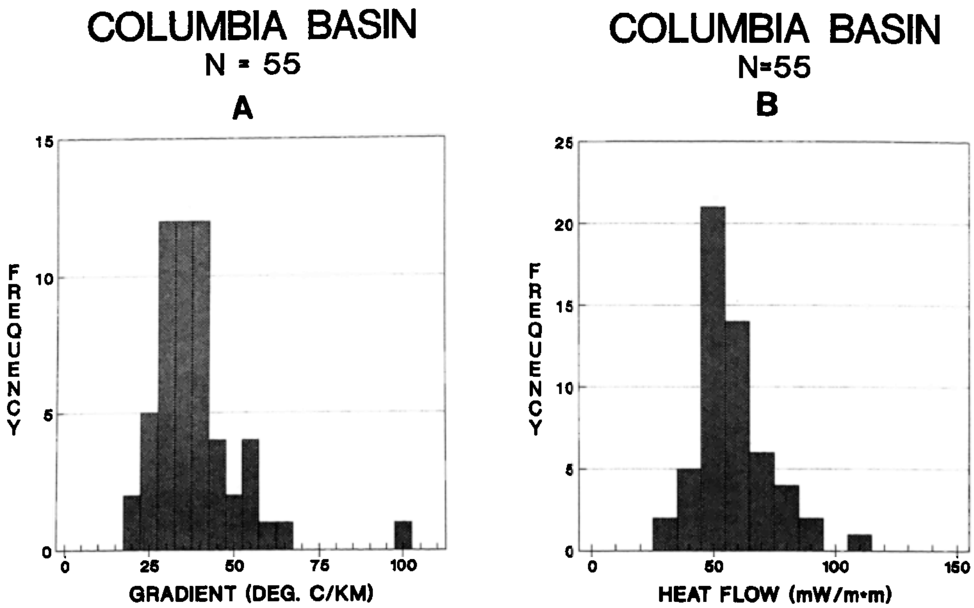

Fig. 6. Histograms of (*a*) geothermal gradient and (*b*) heat flow from Table 1 for the Columbia Basin.

specifically for heat flow/geothermal gradient studies and to depths greater than 150 m to unequivocally determine the heat flow along the Columbia River.

In the Oregon and southern Washington Cascade Range most of the rocks are volcanic or volcanoclastic and have similar thermal conductivities. In the northern Washington Cascade Range and in the Okanogan Highlands, however, there is quite a wide range of conductivity, from very high values in quartzite and dolomite (e.g., hole 23N/11E-4DD in Snoqualmie Pass) to low values in volcanic rock. Consequently, gradients in the northern Washington Cascade Range, even without variations in heat flow and fluid flow effects, can be expected to be much more variable than in the southern Washington Cascade Range. The fractured granitic and metamorphic rocks in the northern Washington Cascade Range may be more brittle and more capable of holding open fractures, and thus be more susceptible to fluid circulation. Indeed the geothermal field at Meager Mountain in British Columbia [*Fairbank et al.*, 1981], where temperatures in excess of 200°C have been obtained from holes drilled in fractured granitic rocks, indicates that there is potential for deep fluid circulation in the geological environment existing in the northern Washington Cascade Range.

Southern Washington Cascade Range

In 1981 a transect of holes was drilled just south of Mount Rainier. The results of the heat flow measurements along this corridor, including data obtained from water wells to the east and west of the drilled holes, are shown in the cross section in Figure 7 (the location is shown as AA′ on Figure 1). A pattern similar to that observed in Oregon [*Blackwell et al.*, 1982, this issue] was obtained, although with a reduced amplitude. Heat flow values west of approximately 122°W are below the continental average of 60 mW m^{-2}, whereas heat flow values between 122° and 120°45′W are 70–108 mW m^{-2}, and gradients range between 34° and 67°C km^{-1}. East of approximately 120°45′, heat flow values decrease to less than 60 mW m^{-2}, and geothermal gradients range from 30° to 45°C km^{-1}. These results are illustrated in Figures 1, 2, and 7. The change from the coastal heat flow of 30–50 mW m^{-2} to high heat flow is well constrained along the profile and occurs over a distance of less than 20 km. Thus the half width of the change (10 km) is similar to that in Oregon [*Blackwell et al.*, 1982, this issue] and in British Columbia [*Lewis et al.*, 1988]. The change occurs within the southern Washington Cascade Range physiographic province. The eastern boundary of the high heat flow zone is not as well located as the western boundary and is of lower relief than the western boundary (10–20 mW m^{-2} instead of 30–40 mW m^{-2}) because it adjoins the high heat flow back arc region.

The heat flow and gradient data for the high heat flow region of the southern Washington Cascade Range are summarized in histograms in Figures 3*c* and 3*d* from the data in Table 1. All data in the region including the data collected as recently as 1988 [*Barnett and Korosec*, 1989] are included. There is a wide range in heat flow and gradient,

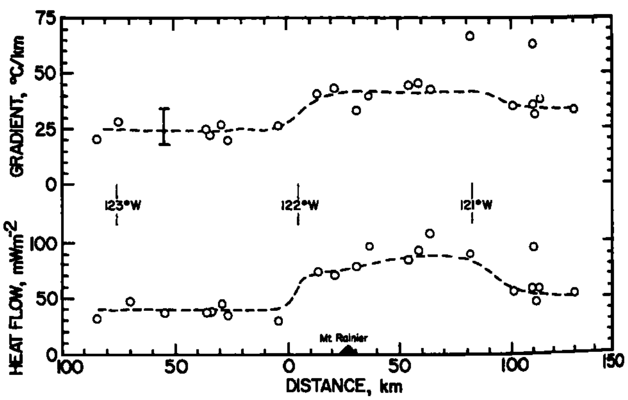

Fig. 7. Heat flow cross section for the southern Cascade Range. Data are plotted as a function of distance along an west to east line south of Mount Rainier (see Figure 1 for location). The range of gradient in hole 12N/1W-9CCC associated with lithologic variations is shown by the vertical line at 50 km on the profile.

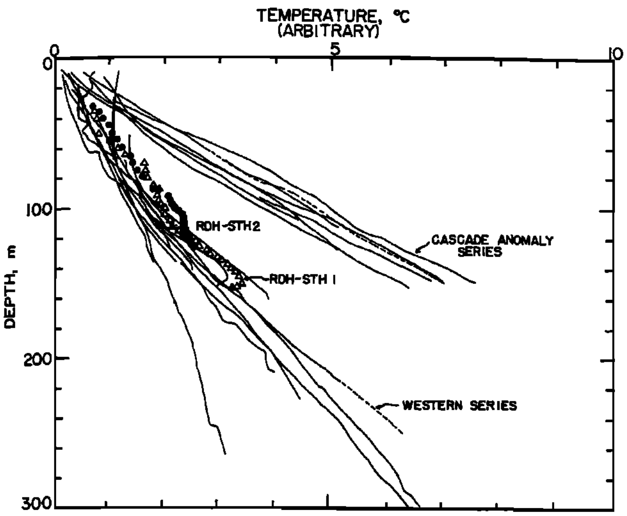

Fig. 8. Temperature-depth curves for the Southern Washington Cascade Range and Coastal provinces. A constant temperature was subtracted from each curve to give a 0°C surface temperature. Data for many of the holes in the low and high heat flow regions are illustrated. The two holes west of Mount St. Helens are clearly associated with the low heat flow area.

indicating that the thermal field in this region is complicated. The average gradient and heat flow for the area are 44.5° ± 3.0°C km^{-1} and 75.4 ± 4.4 mW m^{-2}, respectively.

The major contrast in heat flow and geothermal gradient between the coastal provinces and the Washington Cascade Range is clearly illustrated by the temperature-depth curves from the two regimes in the southern Washington Cascade Range shown in Figure 8. Most of the holes that are in the "Cascade Anomaly" are shown, as are about half of the wells in the low heat flow region to the west.

The difference between the two sets of data is clear. Temperature-depth curves from two holes immediately west of Mount St. Helens are superimposed on Figure 8 and clearly indicate their association with the low heat flow set of holes. These holes were drilled in 1979, and hole 9N/5E-18BB (STH-2) was buried by the debris flow associated with the collapse of the north side of the volcano on May 18, 1980. A drilling project on the east side of Mount St. Helens planned for 1980 was not carried out.

The thermal data in the southern Washington Cascade Range are of sufficient density to warrant detailed interpretation. Because the thermal conductivity does not vary a great deal in the volcanic rocks (see Table 1), a gradient contour map (Figure 9*a*) was prepared and used for interpretation. Alternative interpretations of the details of the pattern are possible given the sparsity of the data, but the main features of the sharp eastern boundary and the northeast/southwest trending axial high are well constrained. The interpretation includes part of the Simcoe volcanic field in the region of high heat flow although there are no heat flow measurements that confirm this inclusion. Also the western boundary is assumed to be oriented north-south. The hand-contoured gradients were digitized on a 15 km by 15 km grid and input into an inversion procedure developed by *Brott et al.* [1981] [see also *Blackwell et al.*, 1982].

The downward continuation technique is based on the placing of a series of point heat sources at a specified depth. Uniform thermal conductivity is assumed in the calculation. This assumption is approximately valid although the values may increase somewhat with depth. The implications of the conductivity assumption are discussed in more detail by *Blackwell et al.* [1982] in their description of the temperatures in the Oregon Cascade Range. Because the downward continuation process is inherently unstable, the calculated temperatures tend to become variable (noisy) as the source is approached. Because we do not know the actual source configuration and the heat flow anomaly is partly interpreted, there is no "correct" answer, and the two temperature distributions are merely indications of the actual temperatures at the depths shown.

Two different models were calculated, one with the sources at a depth of 15 km and one with the sources at a depth of 12.5 km. Calculated temperatures at a depth of 5 km for the source at 15 km are shown in Figure 9*b*. For the reason described above the calculated temperatures are more variable in Figure 9*d* with the sources at 12.5 km than in Figure 9*c* with the sources at 15 km. The maximum calculated temperatures are higher in the second case because the 10 km depth is closer to the plane of sources. Both of the models were calculated assuming steady state conditions, a reasonable assumption if thermal conditions have not changed drastically for 2 to 5 m.y.

The results are consistent with temperatures of 400°–500°C at 10 km over a region of about 100 by 100 km in the southern Washington Cascade Range. The highest temperatures (500°–600+°C) are aligned in a NNE-SSW direction in the center of the anomalous area. The positions of Mount St. Helens (west edge) and Mount Rainier (north edge; although the data on which the position of the northward decrease in heat flow are based are weak) are peripheral to the high-temperature zone, and the area of highest temperature is aligned through the Indian Heaven and Goat Rocks centers of late Cenozoic volcanism. Mount Adams may lie along the apex of the anomaly. This broad area, and especially the Goat Rocks area, is a major center of Quaternary volcanism [*Swanson and Clayton*, 1983; *Smith*, 1989]. Thus the area of highest crustal temperatures appears to be axial to the volcanic arc, the background temperature in the arc is high, and as to the south in Oregon, there is a sharp edge to the west side of the high heat flow region (passing through the site of Mount St. Helens). The locations of Quaternary volcanic centers may occur anywhere within this area, so each volcano is not only an isolated spot of short lived magmatic activity but is also part of a crustal scale thermal anomaly that has developed as magmatism has proceeded.

Stanley et al. [1987, this issue] have described extensive magnetotelluric measurements of crustal resistivity in western Oregon and Washington. Two of the conductance contours from their studies are plotted with the 5-km temperatures in Figure 9*b*. There is not a one-to-one match between the interpreted contours of highest temperature and conductance, but there are no thermal data that constrain the northwest corner of the thermal anomaly where the two interpretations diverge. High crustal temperatures do exist over about 3/4 of the area of the conductance anomaly. *Stanley et al.* [this issue] conclude that lithology (i.e., a deep sedimentary package (an underthrust wedge)) is primarily responsible for the conductance anomaly.

In contrast, the low resistivity (high conductance) zone in the Oregon Cascade Range that is closely associated spacially with the high heat flow is interpreted to be due to

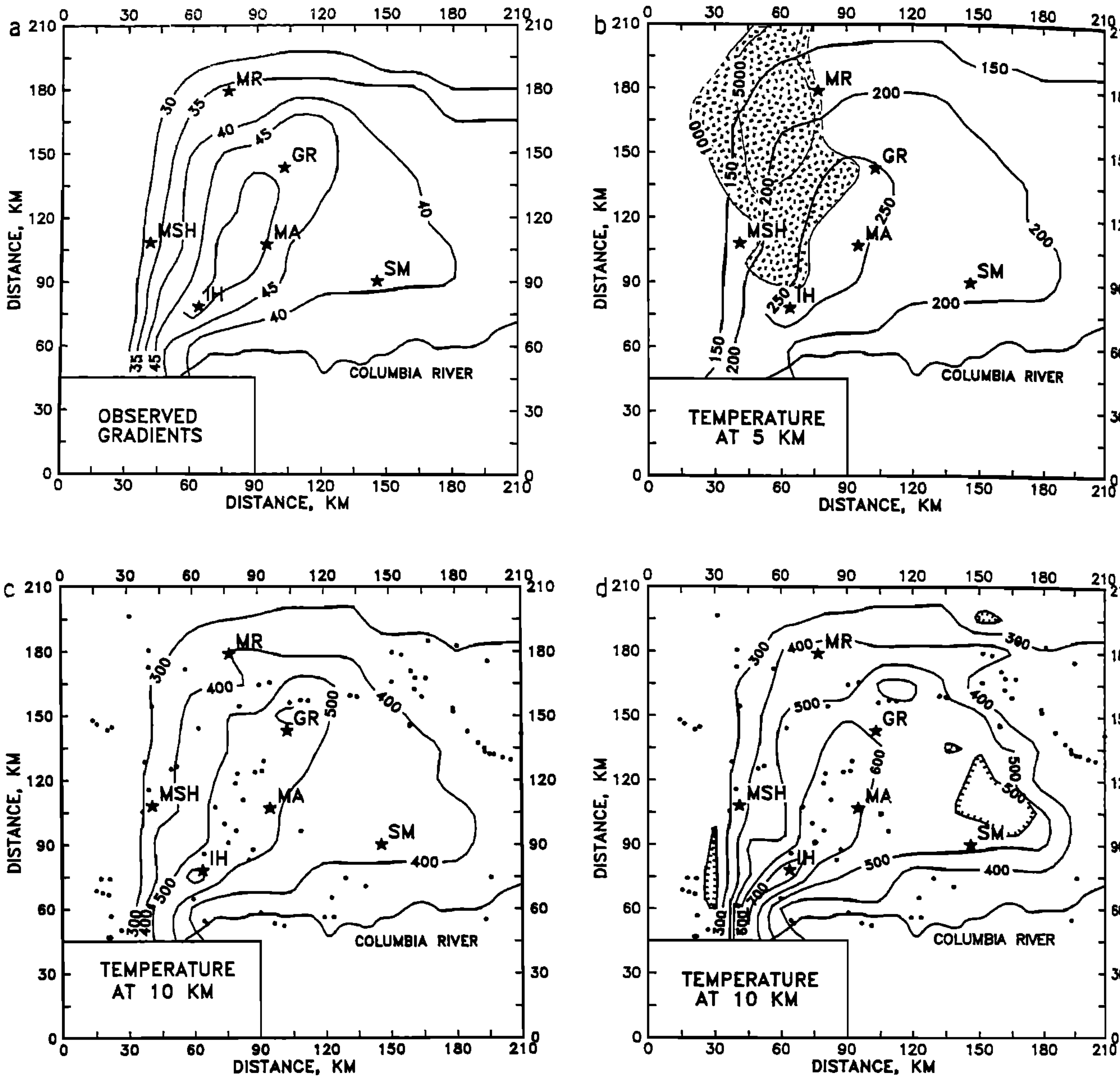

Fig. 9. Interpretive maps of gradient and temperature for the southern Washington Cascade Range. Heat flow sites are shown on Figures 9*c* and 9*d* reference. (*a*) Observed geothermal gradient maps. (*b*) Downward continued temperatures at 5 km for a source plane at 15 km. The shaded region contains the 1000 and 5000 S conductance contours from *Stanley et al.* [this issue]. (*c*) Downward continued temperature at 10 km for a source plane at 15 km. (*d*) Downward continued temperatures at 10 km for a source plane at 12.5 km.

"metamorphically produced fluids and partial melt in a region where ductile behavior occurs at the top of the 6.4–6.5 km/s velocity layer" [*Stanley et al.*, this issue]. The question of the validity of the extrapolation of the heat flow data from the shallow holes to temperatures at midcrustal levels in Oregon has been discussed in detail by *Blackwell et al.* [this issue]. Because most of the arguments apply to this area in a similar fashion, the discussion will not be repeated here. So assuming that the interpreted location of the heat flow anomaly is correct and that the temperatures calculated are reasonable, at the depth of the sedimentary package the temperatures are quite high, and metamorphism might modify the electrical properties of the sediment. The rather close geographic relationship of high electrical conductivity and high midcrustal temperature in northern Oregon and in southern Washington suggests that temperature, or temperature related effects, rather than rock compositional effects may be dominant in causing both of the low resistivity zones.

Discussion

Summary of Washington Heat Flow Distribution

The average heat flow and geothermal gradient values for the various provinces discussed in this report are summarized in Table 2. The thermal patterns in Washington are similar to those discussed by *Blackwell* [1978], except that there is now a significant heat flow data base in and west of the Cascade Range. The thermal provinces include the low heat flow coastal region west of the Washington Cascade Range heat flow anomaly, where heat flow averages 40 mW m^{-2}, the southern Washington Cascade Range heat flow anomaly where heat flow averages 75 mW m^{-2}, and the Okanogan Highlands where heat flow also averages 75 mW m^{-2}.

Average heat flow in the Columbia Basin is 62 mW m^{-2}, significantly lower than in the Okanogan Highlands. This difference in heat flow may be associated with differences in heat loss from the mantle or differences in the heat production from radioactivity within the crust. The crust beneath the part of the Columbia Basin where the heat flow is lowest appears to be unusually mafic, based on seismic refraction data [*Catchings and Mooney*, 1988], so the heat flow from the mantle within this province may be essentially the same as the heat flow at the surface. The average heat flow is nearly equal to the intercept value (60 mW m^{-2}) on the linear heat flow–heat production line (heat flow at an upper crustal radioactivity of zero; see *Blackwell* [1971*b*] and *Roy et al.* [1972] for discussions of the heat flow–heat production relationship) for the Okanogan Highlands.

The pattern of heat flow in the southern Washington Cascade Range is relatively well defined, although many details remain to be determined, particularly along the north, east, and southeast boundaries of the thermal anomaly. Much of the variation may be due to lateral changes in the strength or depth to the heat source that causes the high heat flow. However, some of the variation may be due to fluid flow conditions, or other local situations.

The geothermal gradient pattern for Washington is somewhat different than the heat flow pattern; for example, there is almost a factor of 2 difference in the average gradient in the southern Washington Cascade Range and in the Okanogan Highlands even though the heat flow values are similar. This difference arises because the thermal conductivity of the volcanic rocks in the southern Washington Cascade Range is lower. So in spite of the similar heat flow, temperatures at 10 km beneath the southern Washington Cascade Range may be up to twice as high as those beneath the Okanogan Highlands. High gradients are also observed in the Columbia Basin. These high gradients are partially related to the fact that the basalts have a low thermal conductivity. The average gradients are slightly lower than the average in the southern Washington Cascade Range. The geothermal gradient beneath the Columbia Basin will decrease with depth when the rocks beneath the basalt are encountered because the sedimentary/metamorphic/igneous rocks that make up the basement will have higher thermal conductivities than the volcanic rocks. Hence at a depth of 10 km the temperatures beneath the Columbia Basin are probably closer to those in the Okanogan Highlands than to those in the southern Washington Cascade Range.

The origin for the low heat flow west of the Cascade Range is associated with the subduction of the Juan de Fuca oceanic plate beneath the western part of North America [*Blackwell*, 1971*a*, *b*, 1978; *Blackwell et al.*, 1982]. As the oceanic block sinks beneath the continental plate, the upper part of the oceanic plate absorbs heat from the continental block, causing the low heat flow. This zone of abnormally low heat flow has been identified from British Columbia to Baja California [*Roy et al.*, 1972; *Hyndman*, 1976; *Lewis et al.*, 1988; *Blackwell*, 1978; *Blackwell et al.*, 1990]. Even though subduction is no longer present along much of this length, the long thermal decay time of the low heat flow preserves evidence of subduction for tens of millions of years. Thus the continuity of the low heat flow areas dramatically confirms the validity of the hypothesis that subduction was continuous along the west coast of North America in the early and middle Cenozoic [*Blackwell*, 1971*a*; *Roy et al.*, 1972]. The intermediate focus earthquakes beneath the Puget Sound region are contemporary manifestations of this subduction activity.

It is unfortunate that there are not many data for the transition between the Okanogan Highlands/Columbia Basin and the Washington Cascade Range. Based on the results from this study and data from Canada [*Lewis et al.*, 1985, 1988], heat flow values in the two provinces are similar. However, the details of heat loss in these two provinces must be quite different. The heat flow in the Washington Cascade Range is related to a heat source at an apparent depth of about 10 km. In contrast, the high heat flow in the Okanogan Highlands (and the provinces to the north and south, the Cordilleran Thermal Anomaly Zone of *Blackwell* [1969]) is related to back arc processes with thermal source regions in the upper mantle and lower crust that are associated with past and present thermotectonic events affecting the whole Cordilleran Mountain chain. The back arc thermal anomaly involves a much larger area and owes its origin to the interplay of several mechanisms such as extension, magmatism, and thinning lithosphere [*Roy et al.*, 1972; *Lachenbruch and Sass*, 1978; *Blackwell et al.*, 1990], whereas in the volcanic arc shallow crustal intrusion dominates.

Cascade Range Heat Flow

Extensive studies of geothermal gradient and heat flow have been carried out in the Cascade Range in Oregon. Major results include the location of a sharp eastward increase from lower than normal heat flow to much higher than normal heat flow within the Western Cascade Range (the mid-Cenozoic volcanic arc) and mapping of the western boundary of the high heat flow zone for half the length of Oregon. Average heat flow values in the high heat flow zone are 101 mW m^{-2} and average gradients are 64°C km^{-1} [*Blackwell et al.*, 1982, this issue]. As part of the Oregon studies, extensive drilling was carried out near the Mount Hood volcano (25–50 km south of the Columbia River). *Steele et al.* [1982] concluded that the background heat flow value in the High Cascade Range decreases from 100 mW m^{-2} to 70–80 mW m^{-2} between Mount Jefferson and Mount Hood.

The transition from high to low heat flow along the west side of the Cascade Range over a distance of approximately 20 km is uniquely sharp for a regional thermal boundary. The half width of 10 km indicates a depth of the source of about 10 km. Heat flow measurements in British Columbia indicate a similar pattern of low heat flow along the coast and high heat flow inland, where the Cascade Range heat flow values range from 65 to 90 mW m^{-2}. The transition from low to high heat flow occurs along a line of gravity gradients, anomalous uplift rates, and faulting [*Hyndman*, 1976; *Lewis et al.*, 1985, 1988].

Contours of heat flow for the entire length of the Cascade Range are shown in Figure 10, and heat flow sites are shown

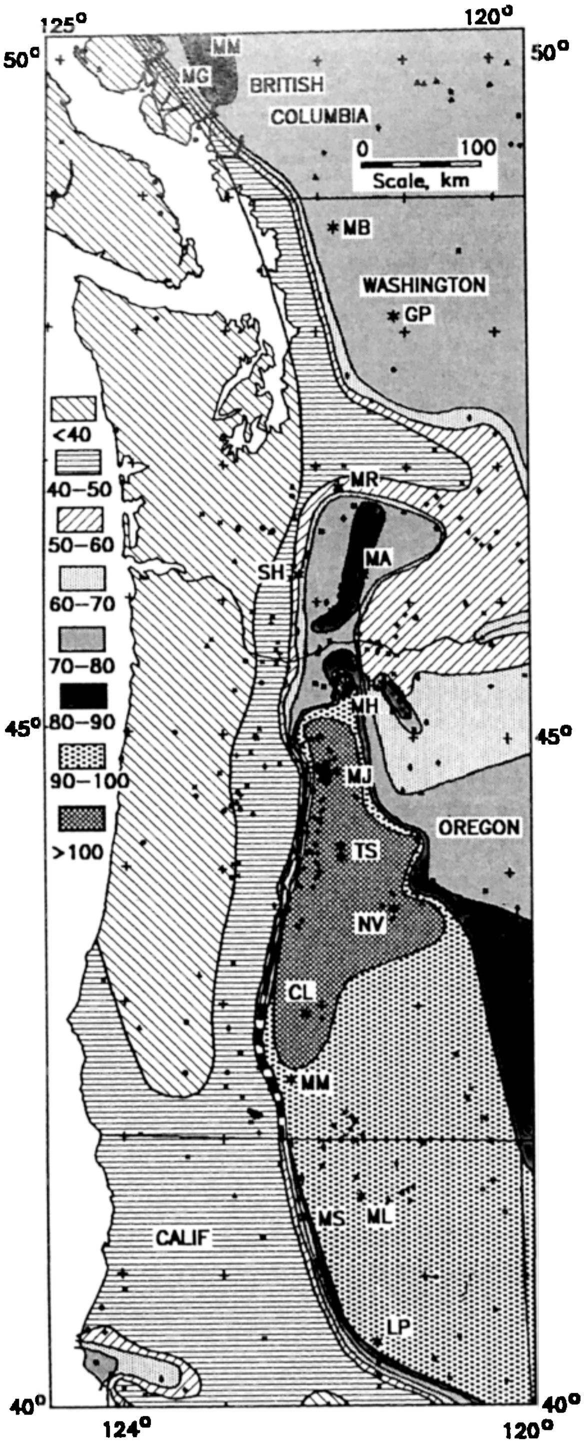

Fig. 10. Summary heat flow map of the Cascade Range. Data points and major volcanic centers are shown. Data points are shown for location only. Data from this paper, *Blackwell et al.* [this issue], *Lewis et al.* [1988], and *Mase et al.* [1982]. Heat flow contours in mW m^{-2}; large asterisks are major Cascade Range volcanoes (identified by two letter abbreviations). The volcanoes are, from north to south, MM, Meager Mountain; MG, Mount Garibaldi; MB, Mount Baker; GP, Glacier Peak; MR, Mount Rainier; SH, Mount St. Helens; MA, Mount Adams; MH, Mount Hood; MJ, Mount Jefferson; TS, Three Sisters; NV, Newberry volcano; CL, Crater Lake; MM, Mount McLoughlin; MS, Mount Shasta; ML, Medicine Lake; LP, Lassen Peak.

to illustrate the data density in the various areas. The most serious data gaps are in northern Washington, in southern Oregon, and in northern California. The contours outline a thermal segmentation of the Cascade Range. This segmentation has been discussed in brief by *Blackwell and Steele* [1983], and other categories of segmentation of the Cascades Range have been discussed by *Guffanti and Weaver* [1988]. Major troughs in the Cascade thermal anomaly are at the Columbia River and in central Washington. The origin of these troughs is not clear; they may be due to gaps in magma production along the arc over the last 5 m.y. or so. However, there are hot springs in both gaps [*Mariner et al.*, this issue]. Deeper drill holes in both areas are needed to resolve the true crustal heat flow in the area.

The near constant half width (10 km) of the west side of the thermal high is an interpretation, but it is documented in three widely separated areas: British Columbia, southern Washington, and northern Oregon. This half width may characterize the depth to a thermal anomaly caused by a zone of shallow crustal magma residence since it does not appear to consistently coincide with any major, crustal, seismic velocity boundary (and by inference any lithologic boundary). A superimposed thermal peak of about 10+ mW m^{-2} coincides with the axis of greatest volcanic production in the southern Washington Cascade Range, and a similar thermal peak has been documented at Mount Hood [*Steele et al.*, 1982]. Other such thermal concentrations have not been resolved probably due to data limitations rather than the lack of such features.

Blackwell et al. [1982] calculated the ratio of intrusive to extrusive volume in central Oregon by assuming that the excess heat flow above the background represented the heat lost by the intrusive process. The calculations were based on volumes of extrusives estimated by *White and McBirney* [1978]. New calculations of intrusive rate along strike for the whole range are shown in Figure 11. These results are based on new along-strike estimates of volcanic extrusion volume for the last 2 m.y. presented by *Sherrod and Smith* [this issue], also shown in Figure 11. These estimates appear to be conservative. *Priest* [this issue] estimates 20–30 km^3 (2 $m.y.)^{-1}$ km^{-1} for central Oregon (about 85% of which is basalt) and M. A. Korosec (unpublished data, 1989) estimates an additional 10 km^3 (2 $m.y.)^{-1}$ for part of southern Washington. The observed heat flow along the arc and the assumed heat flow background are also shown. In calculating the rates of intrusion it was assumed that the 100 mW m^{-2} heat flow observed along the western side of the High Cascade Range in Oregon represents the average heat loss all across the arc. If the axial heat flow is higher (and it probably is), the volume estimates would be proportionally increased.

The ratios of intrusion to extrusion shown in Figure 11 are higher than those estimated by *Blackwell et al.* [1982] because the recent detailed geologic studies mentioned above give somewhat lower volumes of extrusion over the last 2 m.y. than *Blackwell et al.* [1982] inferred from the *McBirney and White* [1978] data. Typical ratios range from approximately 5 or 10–1 in central Oregon to perhaps 50–100 to 1 in British Columbia. This difference might be related to the crustal differences, or perhaps the extrusive volumes in British Columbia are underestimated because of the high rates of glacial erosion there. Furthermore, the estimates for

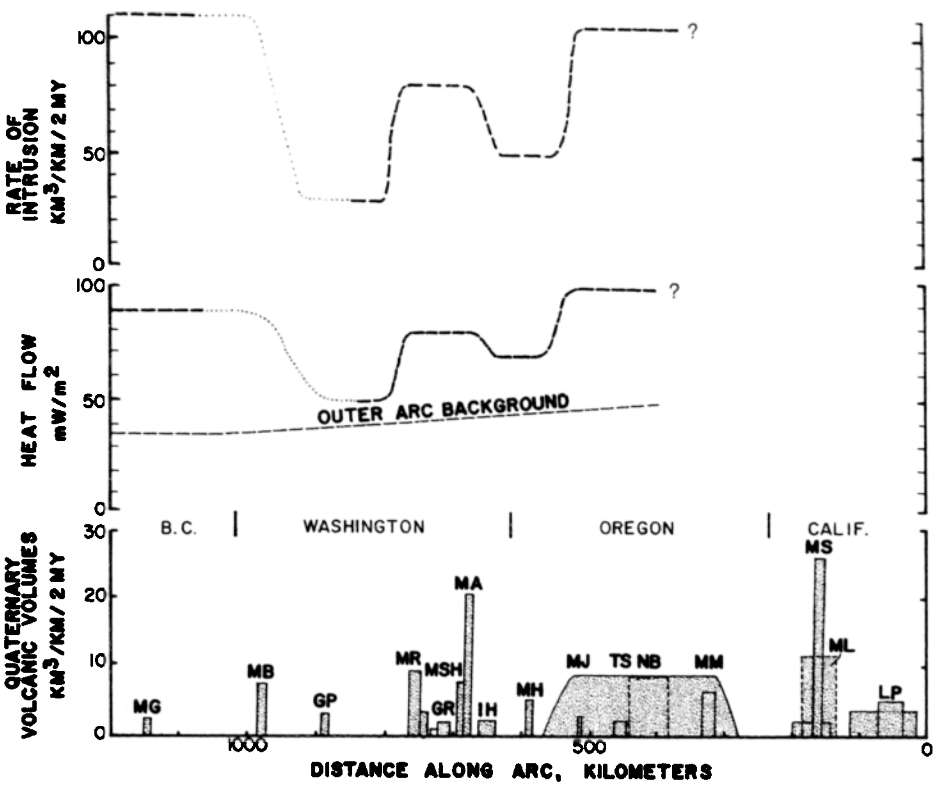

Fig. 11. Volumes of volcanic rocks for the last 2 m.y. [*Sherrod and Smith*, this issue], heat flow, and inferred intrusive volumes along the Cascade Range.

British Columbia relate almost exclusively to silicic volcanic rocks.

There are two major conditions that would modify the calculated amounts of intrusion. First, because the heat flow data have no depth-to-source resolution except along the west side of the anomaly, high lower crustal heat flow along the center and east side of the Cascade Range could decrease the amount of shallow intrusion required below that calculated in Figure 11. It might be argued that the appropriate background to use in the calculation is a high heat flow typical of the Cordilleran Thermal Anomaly Zone or of a region subject to a flux of lower crustal mafic intrusions. In this case the heat flow background could range from 20 to 40 mW m^{-2} higher at various latitudes along the Cascade Range, and the rate of upper crustal intrusion would be decreased. In British Columbia the rate of intrusion would be proportionally most affected and might drop as low as 30–40 km^3 km^{-1} (2 m.y.)$^{-1}$ from the value of 110 km^3 km^{-1} (2 m.y.)$^{-1}$ shown in Figure 11. Along the west boundary of the Cascade thermal anomaly, the 10-km half width requires the highest horizontal thermal gradients to be in the mid-crust, so the outer arc heat flow is the correct background. Second, the heat flow may be higher along the axis of the Cascade Range, in Oregon especially, than assumed for the purposes of the calculation. In this case the intrusion rate would be underestimated, and so the real rate might be higher than the values shown in Figure 11.

The similarity of the heat flow in the British Columbia and Oregon Cascade Range (about 90 mW m^{-2} compared to just over 100 mW m^{-2}) is not reflected in the crustal temperatures. At a depth corresponding to the half width and thus near the source (10 km), the temperatures may be 400°–500°C in British Columbia and 600°–800°C in Oregon. In this case (assuming a constant background) the rate of intrusion holds the heat flow constant, but the temperature varies. The higher thermal conductivity of the upper crust in British Columbia more efficently transmits the intrusive heat to the surface. Consequently, crustal type may play a role in the thermal regime of a volcanic arc.

The temperatures are high enough along most of the Cascade Range that metamorphism should be occurring at the present time. Temperatures in the range of 300° to 600+°C at 8–12 km are well within the range of low-pressure metamorphism of rocks to greenschist to granulite facies. Thus the area may be a natural laboratory for study of a crustal region undergoing metamorphic processes. *Barton and Hanson* [1989] [also *Hanson and Barton*, 1989] have recently discussed the thermal requirements for low-pressure metamorphism. They inferred that temperatures in volcanic arcs are not high enough to form low-pressure metamorphism for much of the time. In the Cascade Range, however, it appears that temperatures are sustained at high enough levels to cause metamorphism for long periods of time (several millions of years).

The ratio of intrusion to extrusion is probably a characteristic property of volcanic arcs, but the relationship has been difficult to quantify. It has been assumed for volcanic terrains in general that lower ratios might be associated with more mafic volcanism whereas higher ratios might be associated with more silicic volcanism. This model was the basis for the use of volcanic type in evaluating the geothermal resource potential of the United States [*Smith and Shaw*, 1975]. The data presented here represent the first heat loss measurements that can be related to magmatic processes and that document quantitively this hypothesis. While there are a number of uncertainties associated with the values calculated in this paper, the rates and their geographic variation represent a new set of parameters to help in understanding the igneous processes in volcanic arcs.

Cascade Range "Magma Chamber"

The conclusion by *Blackwell et al.* [1982] that the Cascade Range is underlain by a broad upper crustal heat source (depth of 8–12 km) that is a magma staging region is not widely accepted. The favored spot for a zone of regional crustal residence, melting, assimilation, and fractional crystalization in volcanic arcs is at the crust/mantle interface. For example, *Leeman et al.* [this issue] describe the evolution of Mount St. Helens dacite by open-system crustal anatexis at the crust mantle boundary. In discussing magma evolution in volcanic arcs in general, *Hildreth and Moorbath* [1988, p. 485] conclude that the

> data lead us to envisage deep beneath each large magmatic center a zone of melting, assimilation, storage, and homogenization (MASH), in the lowermost crust or mantle-crust transition, where basaltic magmas that ascend from the mantle wedge become neutrally buoyant, induce local melting, assimilate and mix extensively, and either crystallize completely or fractionate to the degree necessary to re-establish buoyant ascent. Magmas ascending from such zones may range from evolved basalts to dacites, but all will have acquired a base-level isotopic and trace-element signature characteristic of that particular MASH zone.

The requirements of such a zone to satisfy the geochemical data do not obviate the need for shallower zones of residence, interaction and sourcing, particularly for large volumes of very silicic melts to be generated and erupted. There is no question that such large-scale upper crustal zones existed in some settings in the past. For example, a large system of this sort existed in Idaho during the Eocene [*Criss and Taylor*, 1983], when a zone about 50 km wide and several hundred kilometers long was the site of shallow level granitic intrusions probably overlain by massive caldera systems.

On the North Island of New Zealand the Taupo graben represents the site of a massive, upper crustal, silicic, magma chamber in a volcanic arc setting [*Wilson et al.*, 1984]. In fact, andesite stratovolcanos typical of volcanic arcs are located immediately south of the area of silicic ash flow volcanism. In the Japanese volcanic arcs there is very little basalt erupted and the volcanics are dominately andesite and dacite in composition. The spacing of volcanoes is about 50 km in contrast to the 75-km spacing of the Cascade Range. Perhaps the midcrustal residence zone is an even more continuous zone of crust at melting or near melting temperatures than is the case in the Cascade Range.

Finally, during the Devonian a situation that resulted in a thermal pattern that is in many ways similar to the one proposed for the Cascade Range existed in Maine. In central Maine there is a large region of low-pressure/high-temperature metamorphism grading into a large region of contact metamorphism along strike to the southwest [*Holdaway et al.*, 1988]. Gravity evidence suggests that the low-pressure regional metamorphism is associated with regional development of granite sills at an inferred emplacement depth of 8–12 km, whereas the contact metamorphism is associated with isolated, deeper, more equant, granite bodies. Thermal models are consistent with this origin for the metamorphism [*Yoreo et al.*, 1989].

The surprising consistency of the half width of the heat flow transition in Oregon, southern Washington, and southern British Columbia requires a consistency in heat source depth that is not affected by volume rate of extrusion, average composition of volcanics, crustal type, or stress state. As pointed out by *Blackwell et al.* [this issue], the Curie point depths in Oregon are consistent with the downward continuation of surface heat flow to depths of at least 5–10 km, where temperatures are 300°–500°C. The gravity data in the central Oregon Cascade Range are also consistent with a midcrustal region of low density, although because of the complex regional field and upper crustal density contrasts the depth to this region cannot be well constrained. In Washington and British Columbia [*Lewis et al.*, 1988] there are gravity gradients corresponding to the heat flow transition, although the amplitude of the gravity anomaly that can be isolated is less than that in the Oregon Cascade Range [*Blackwell et al.*, this issue].

These results and interpretations, taken together, lead to the following model. Under most of the Cascade Range where volcanism has occurred over the last 2–5 m.y., a zone of magma interception, storage, and crystallization exists in the depth range of 8–15 km. The horizontal extent of this zone is significantly larger than individual centers, being at least 60 km wide in places, but over time, volcanic vents and major centers of volcanism map out most of the extent of the zone. The correspondence is discussed by *Blackwell et al.* [this issue], who compared the maps of volcanic vents discussed by *Guffanti and Weaver* [1988] to the heat flow anomaly in central Oregon. The temperature in this zone is a function of the rate of intrusion, which may or may not be reflected in the extrusion rate, and crustal type. The absence of primitive basalts at the surface probably means that they stall and fractionate within the crust. The high rate of silicic volcanism in certain segments of many arcs may result from the fact that none of the basalt makes it through the crust and so is able to transfer its heat completely to the crust. An extreme example of this situation is Yellowstone, where volumes of basaltic magma similar to those found in Hawaii are being emplaced in the crust with the result that huge volumes of silicic magma are produced.

In Oregon that time-averaged temperature is in the 600°C range, and at least isolated pockets of melt probably exist. The overall effect may be enough to result in a density contrast compared to the crust outside the Cascade Range. The rate of activity is high enough that the size of the zone is independent of individual volcanic centers. In Washington and probably in northern California and southern British Columbia, the rate of intrusion is much less than in central Oregon, and the temperature in the residence zone is 300°–500°C. In this case the peak thermal anomalies are more closely related to long-lived centers of intrusion. The lower temperatures make the geophysical signitures of this heat source region more subtle than in central Oregon. In no case is a continuous layer of magma envisioned to exist over the whole region at one time.

Acknowledgments. The work in northeastern Washington was supported by NSF grant 11351 to SMU. Studies in the Columbia Basin, Washington Cascade Range, and western Washington were supported by DOE contract ID112307-1 to SMU and by DOE contract DE-A07-79ET27014 and DOE grant DE-GF07-88ID12740 to the Washington Division of Geology and Earth Resources. J. Eric Schuster was involved in much of the work and assisted in the field studies and SMU/WGER coordination. Reviews by L. J. P. Muffler and W. Hildreth were helpful in improving the manuscript, and W. Hildreth pointed out that the background heat flow in the Cascade Range might be higher than the outer arc values.

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(Received August 30, 1989;
revised May 18, 1990;
accepted May 28, 1990.)