Heat flux measurements in southwestern British Columbia: the thermal consequences of plate tectonics[1]

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Measured heat fluxes from previously published data and 34 additional boreholes outline the terrestrial heat flow field in southern British Columbia. Combined with heat generation representative of the crust at 10 sites in the Intermontane and Omineca belts, the data define a heat flow province with a reduced heat flow of 63 mW m^{-2} and a depth scale of 10 km. Such a linear relationship is not found or expected in the Insular Belt and the western half of the Coast Plutonic Complex where low heat fluxes are interpreted to be the result of recent subduction. The apparent boundary between low and high heat flux is a transition over a distance of 20 km, located in Jervis Inlet 20–40 km seaward of the Pleistocene Garibaldi Volcanic Belt.

The warm, thin crust of the Intermontane and Omenica Crystalline belts is similar to that of areas of the Basin and Range Province where the youngest volcanics are more than 17 Ma in age. Processes 50 Ma ago that completely heated the crust and upper mantle could theoretically produce such high heat fluxes by conductive cooling of the lithosphere. But it is more likely that the asthenosphere flows towards the subduction zone, bringing heat to the base of the lithosphere. Since the reduced heat flow is high but constant, large differences in upper crustal temperatures within this heat flow province at present are caused by large variations in both crustal heat generation and near-surface thermal conductivity. The sharp transition in heat flux near the coast is the result of the combined effects of convective heating of the eastern Coast Plutonic Complex, pronounced differential uplift and erosion across a boundary within the Coast Plutonic Complex, and the subducting oceanic plate.

Les valeurs de flux thermique obtenues à partir de données publiées antérieurement et de 34 nouveaux trous de sondage permettent de circonscrire le champ de flux thermique terrestre dans le sud de la Colombie-Britannique. Ces résultats combinés avec ceux du dégagement de chaleur émanant de la croûte sur 10 emplacements dans les ceintures d'Intermontane et d'Omineca, délimitent une province de flux thermique caractérisée par un flux thermique réduit de 63 mW m^{-2} et à l'échelle de 10 km de profondeur. Il n'existe pas une telle relation linéaire dans la ceinture insulaire et dans la moitié-occidendale du complexe plutonique de Coast où les faibles flux thermiques sont occasionnés par une subduction récente. La limite apparente entre les flux faible et élevé est une zone de transition large de 20 km, localisée dans le bras de mer de Jervis, 20–40 km au large, dans la ceinture volcanique de Garibaldi, d'âge pléistocène.

La croûte chaude et mince des ceintures cristallines d'Intermontane et d'Omenica ressemble à celle des régions de la province de Bassin et Chaîne des Rocheuses où les roches volcaniques les plus jeunes ont un âge dépassant 17 Ma. Il y a 50 Ma, des processus capables de réchauffer complètement la croûte et le manteau supérieur auraient pu créer théoriquement des flux thermiques si élevés grâce à un refroidissement par conduction de la lithosphère. Cependant, l'écoulement de l'asthénosphère vers la zone de subduction accompagné du transfert de la chaleur à la base de la lithosphère offre une explication plus plausible. Vu la réduction considérable du flux thermique et sa constance, les températures fortement différentes dans la croûte supérieure qui existent actuellement à l'intérieur de cette province de flux thermique sont causées par des variations importantes de la production de chaleur crustale et de la conductivité thermique près de la surface. Ce changement de flux thermique en une courte distance sur la côte résulte des effets combinés du transfert de chaleur par convection dans le complexe plutonique de Coast, d'un soulèvement relatif prononcé et de l'érosion traversant une frontière à l'intérieur du complexe plutonique de Coast et de la plaque océanique qui passe dessous.

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Introduction

This paper presents heat flux measurements from 34 boreholes or mine sites from southwestern British Columbia. Combined with published data from 20 other relatively deep boreholes, 17 shallow holes, and oceanic-type measurements in Jervis Inlet, the data for this area allow us to examine the relationship between heat flux and current tectonics. Heat generation has been determined at 16 of these sites where representative samples of the youngest intrusive rocks were available. A separate paper (T. J. Lewis *et al.*, in preparation) is being prepared on extensive heat flux measurements along a single profile through Jervis Inlet into the Garibaldi Volcanic Belt.

Current and recent tectonics appear to govern the heat flux from the deep crust as well as crustal movement. Subduction of the young Juan de Fuca and Explorer plates beneath this area (see Fig. 1) causes the eruption of a characteristic arc volcanic belt (Riddihough and Hyndman 1976), the Pleistocene Garibaldi Volcanic Belt, into the Coast Plutonic Complex. We compare the similarities and differences of the tectonic processes in this area with those of the Cascade volcanic province to the south where the tectonic processes are overprinting very different geological settings.

Geological and geophysical setting

Southwestern British Columbia consists of portions of four main belts or complexes defined by physiographic and geolog-

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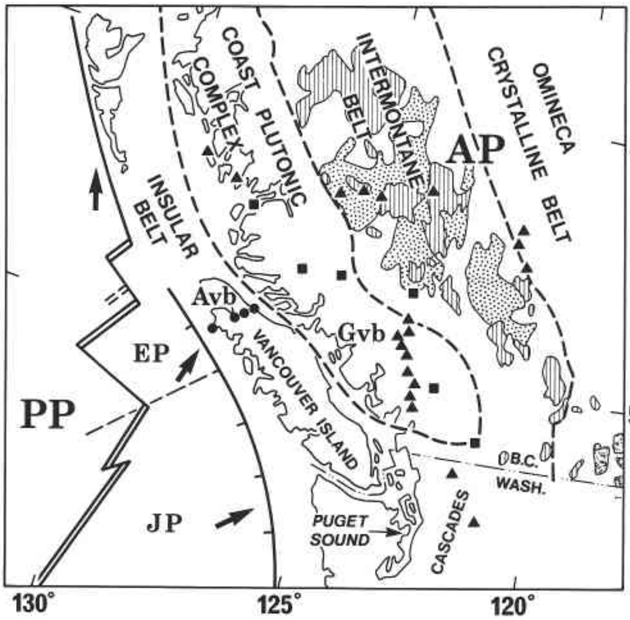

FIG. 1. The young Juan de Fuca Plate (JP) and Explorer Plate (EP) are being subducted under the American Plate (AP) while the Pacific Plate (PP) moves with a slightly convergent velocity with respect to AP. Quaternary volcanism centres, including the Garibaldi Volcanic Belt (Gvb), are shown as triangles; early Tertiary volcanic centres, including the nearby Pemberton Volcanic Belt, are shown as squares. The small, late Tertiary Alert Bay Volcanic Belt (Avb) is shown by solid circles. The Mio-Pliocene Chilcotin volcanics are shown by dotted areas, Eocene volcanics by lined areas, and the Coryell syenites by mottled areas.

ical characteristics. From west to east they are the Insular Belt, the Coast Plutonic Complex, the Intermontane Belt, and the Omineca Crystalline Belt. They have been created by many geologic processes, the most important being the interaction of various oceanic plates with ancestral North America (e.g., Riddihough 1982) and, in the process, strike-slip faulting and the accretion of slices of displaced terranes.

The Insular Belt consists of broadly folded and faulted upper Paleozoic and lower Mesozoic volcanics and sediments. Movement along faults was repetitive over a long time interval. The middle Jurassic Vancouver Island intrusions, principally diorite and quartz diorite, cover large areas compared with small, early Tertiary intrusions ranging from quartz diorite to quartz monzonite. The small Miocene Alert Bay Volcanic Belt crosses northern Vancouver Island.

The Coast Plutonic Complex consists of large volumes of plutonic rocks, mainly quartz diorite and granodiorite, commonly with migmatites, and occasionally with foliation or highly metamorphosed and distorted bedding evident. Older volcanic roof pendants are common. The potassium–argon ages of the plutonic rocks reflect the uplift history, varying from 140 Ma in the west to 40 Ma on the eastern border (Hutchison 1970). Superimposed on these plutonic rocks are the Miocene Pemberton Volcanic Belt and the Pleistocene Garibaldi Volcanic Belt. Rapid uplift and erosion, increased by glacial action, have deeply incised this complex, producing a very rugged topography.

In comparison, the Intermontane Belt has a subdued, upland topography with only a few deep, main valleys. Tectonically it resembles the Basin and Range Province, with small grabens filled with volcaniclastic sediments forming Tertiary basins. In the south these volcanics probably covered the whole Thompson plateau (the Intermontane belt south of 51°N) and were extruded over a period from 48 to 52 Ma ago (Church 1982). In the north, Mio-Pliocene basalts of the Chilcotin Group (Bevier 1983) cover large areas.

The Omineca Crystalline Belt, an uplifted core zone, is characterized by large regions that have been subjected to high-grade metamorphism of mid-Mesozoic age and are surrounded by younger mesozonal intrusions. The Mesozoic intrusions are mostly quartz diorite and granodiorite (more potassic than the Coast Plutonic Complex), but the Coryell syenites and quartz monzonites are typical of the early Tertiary.

Since the various terranes came together to form this region of southwestern British Columbia, interactions between subducting oceanic plates and the North American plate are probably indirectly related to various episodes of faulting, burial, uplift, intrusion, and eruption. A major Eocene thermal event produced the Coryell syenite intrusives, now exposed over large areas of the Omineca Crystalline Belt, and the related Eocene volcanics in the Intermontane Belt. At that time the Coast Plutonic Complex was uplifted (Souther 1977), with maximum uplift rates of 1 km Ma^{-1} (Parrish 1983) for the Cretaceous to Eocene period. The two main periods of eruption of the Chilcotin Group basalts were 2–3 and 6–10 Ma ago, although a few dates are older than the 17 Ma Pemberton Volcanic Belt (Bevier 1983). Post-Miocene uplift along the eastern edge of the Coast Plutonic Complex has tilted Chilcotin basalts up to 15°. The only Quaternary volcanics in the Coast Plutonic Complex, Silverthrone and the Garibaldi Volcanic Belt (Souther 1980), occur south of 52°N where there was rapid uplift (up to 0.75 km Ma^{-1}) from Pliocene to Recent.

Various geophysical studies provide information concerning the crustal structure. Such studies of the continental margin of southwestern Canada define the plate boundaries and relative motions at present and in recent times (Keen and Hyndman 1979). The plate convergence predicted by offshore geophysical data (e.g., Riddihough and Hyndman 1976) has a pronounced effect on the continental crust and upper mantle extending several hundred kilometres inland. The location of presently active spreading centres is indicated by seismicity (Keen and Hyndman 1979). However, there is little seismicity under the Canadian continental crust to indicate a Benioff zone, unlike farther south under Puget Sound (Crosson 1983). A detailed refraction survey (Ellis *et al.* 1983; McMechan and Spence 1983) indicates a Mohorovičić discontinuity dipping to the east under Vancouver Island, where it has been traced to a depth of 37 km. Models to explain observed magnetic anomalies (Coles and Currie 1977) favour a cold, deep crust under the western edge of the Coast Plutonic Complex, with a fairly high magnetite content to a depth of 40 km. Gravity modelling to fit observed profiles extending 500 km northeast from the trench seaward of Vancouver Island (Riddihough 1979: profiles 1 and 2) suggests a subducted ocean crust with a wedge of high-density material overlying a subducting oceanic crust underneath the western Coast Plutonic Complex.

Farther inland, seismic studies show that the low-velocity layer is very close to, if not at, the base of the crust under much of the Intermontane Belt (Wickens 1971; Berry and Forsyth 1975). Magnetic anomalies are relatively subdued over the Intermontane and Omineca Crystalline belts compared with either the Coast Plutonic Complex or the area east of the Rocky Mountain Trench (Haines *et al.* 1971). Geomagnetic depth sounding (e.g., Caner *et al.* 1971) reveals a conductive (low Z) region under the mainland throughout the study area.

TABLE 1. Previously published heat flow values from deep boreholes (>100 m)

Site		Location		Elev.	Heat flow		Heat gen.	Age	Ref.*
		Lat. N	Long. W	(m)	Meas. ($mW\ m^{-2}$)	Corr. ($mW\ m^{-2}$)	($\mu W\ m^{-3}$)	(Ma)	
3	Penticton	49°19.8′	119°37.6′	552	67	79	—	50	*a*
301-2	Meager Creek	50°34.6′	123°27.6′	575	120				*b*
301-3	Meager Creek	50°34.1′	123°28.3′	640	100				*b*
301-4	Meager Creek	50°34.3′	123°29.1′	777	290				*b*
301-5	Meager Creek	50°34.0′	123°30.2′	770	930				*b*
301-6	Meager Creek	50°34.3′	123°30.3′	808	450				*b*
303-1	Lillooet	50°37.8′	123°23.8′	580		132	3.8	7.9	*c*
303-2	Lillooet	50°34.8′	123°16.8′	360		80			*c*
304-1	Cayley	50°07.3′	123°23.6′	170		69			*c*
304-2	Cayley	50°06.4′	123°22.0′	190		89			*c*
310-1	Paynter L.	49°57.5′	119°47.2′	1311	81	87	1.7	(Cretaceous)	*d*
310-2	Trout Cr.	49°34.2′	119°39.1′	389	73	86	2.0	(Cretaceous)	*d*
352-1	White Lake	49°21.5′	119°41.6′	686	74	85	—	50	*e*
352-2	White Lake	49°22.1′	119°46.1′	1183	109	120	—	50	*e*
352-5	White Lake	49°21.9′	119°45.4′	1056	76	87	—	50	*e*
352-6	White Lake	49°22.2′	119°38.8′	655	69	80	—	50	*e*
M6-79D	Meager Creek	50°34.4′	123°29.9′	885	1660	—	2.3	(Tertiary)	*f*
M7-79D	Meager Creek	50°34.1′	123°30.9′	900	2530	—	2.3	(Tertiary)	*f*
M8-79D	Meager Creek	50°34.2′	123°32.3′	875	309	—	2.3	(Tertiary)	*f*
M9-80D	Meager Creek	50°31.2′	123°30.1′	765	120	—	2.3	(Tertiary)	*f*
M11-80D	Meager Creek	50°34.1′	123°28.8′	791	185	—	2.3	(Tertiary)	*f*
M12-80D	Meager Creek	50°32.6′	123°28.7′	792	148	—	2.3	(Tertiary)	*f*
M14-81D	Meager Creek	50°33.5′	123°27.5′	861	146	—	2.3	(Tertiary)	*f*
L1-78D	Lillooet	50°41.1	123°30.8	760	680	—	2.5	(Tertiary)	*f*
L2-80D	Lillooet	50°40.4′	123°32.3′	896	262	—	2.5	(Tertiary)	*f*
L3-80D	Lillooet	50°40.5′	123°31.4′	972	322	—	2.5	(Tertiary)	*f*
L4-81D	Lillooet	50°39.6′	123°32.0′	1097	207	—	2.5	(Tertiary)	*f*
L5-81D	Lillooet	50°40.5′	123°28.8′	774	272	—	2.5	(Tertiary)	*f*
L6-81D	Lillooet	50°37.6′	123°25.0′	535	284	—	2.5	(Tertiary)	*f*
L7-82D	Lillooet	50°38.9′	123°33.5′	1808	227	—	2.5	(Tertiary)	*f*
L8-82D	Lillooet	50°40.5′	123°33.5′	960	176	—	2.5	(Tertiary)	*f*

*References: *a* = Jessop and Judge (1971); *b* = T. J. Lewis and Souther (1978); *c* = J. F. Lewis and Jessop (1981); *d* = Davis and Lewis (1984); *e* = T. J. Lewis (1984); *f* = Reader and Fairbanks (1983).

Previous measurements

Previously published heat fluxes measured in holes deeper than 100 m within the study area are listed in Table 1 and included with new data in Fig. 6. Values near the Garibaldi Volcanic Belt at the Meager, Cayley, and Lillooet sites exhibit a large scatter, as expected in areas where heat is being redistributed locally by convection and groundwater flows. Steele and Blackwell (1982) observed such scatter near Mount Hood. Other values near the Okanagan Valley in the Intermontane Belt indicate a conductive heat flux of 80 mW m^{-2}, a high value as measured in most 50 Ma old continental crust (Jessop *et al*. 1976). Results from several shallow wells (Davis and Lewis 1984) give a fairly uniform heat flux of 83 mW m^{-2} on a 200 km long east–west profile crossing Okanagan Lake. This southern part of the Intermontane Belt is therefore a northern extension of the Cordilleran Thermal Anomaly Zone first described by Blackwell (1969). Compared with the Basin and Range Province, there are relatively fewer basins. In one of these, the White Lake Basin, the heat flux varies up to a maximum of 116 mW m^{-2} as a result of refractive effects and water flows (T. J. Lewis 1984). Blackwell (1983) suggested that the average measured value of heat flux for the Basin and Range Province is biased because of such effects combined with the location of most measurements on ranges.

An abrupt change in heat flux from 40 mW m^{-2} to 80 mW m^{-2} over a distance of 20 km along Jervis Inlet has been measured (Hyndman 1976; T. J. Lewis and Hyndman 1981; T. J. Lewis *et al.* 1983, and in preparation) using oceanic techniques. This transition is located 30 km seaward of the Garibaldi Volcanic Belt.

Heat flux to the east of the Rocky Mountains in the thick sediments of Western Canada Basin is controlled by systematic water flow (Majorowicz and Jessop 1981; Majorowicz *et al.* 1984). Values are high compared with those from other Precambrian platforms or from areas of the same tectonic age (50 mW m^{-2}, Jessop *et al.* 1976). Preliminary results indicate high heat flux in the southeastern Omineca Crystalline Belt, and Judge (1977) favoured deep groundwater circulation as the cause. To the north in the Intermontane Belt (Jessop *et al.* 1984) the heat flux ranges from 63 to 100 mW m^{-2}, and a reduced heat flow of 45 mW m^{-2} is likely. A heat flux of 73 mW m^{-2} (Mathews 1972) was measured at the eastern edge of the northern Coast Plutonic Complex.

The heat flux in the western United States is well defined by many measurements (e.g., Sass *et al.* 1976, 1980; Hyndman 1984). Low values in the Cascade Range of California, less than 30 mW m^{-2}, are attributed to regional circulation of groundwater (Mase *et al.* 1982) within the young volcanic rocks. The much higher than average heat flux of the Basin and Range Province is attributed to extension within the region

(Lachenbruch 1979), whereas the Snake River plain, an arc extending from eastern Oregon to Yellowstone Park, may be the trace of a hot spot. Although a high heat flux is expected here, the measured values in 200 m deep holes are low (Brott *et al.* 1978). In northern Oregon the heat flux changes abruptly from values of 40 mW m^{-2} in the coastal provinces and the older Western Cascade Range to 100 mW m^{-2} in the younger, High Cascade Range (Blackwell *et al.* 1982; Blackwell 1983). This transition zone is also 30 km seaward of a volcanic belt, the Cascades, similar to the relation observed for the transition measured in Jervis Inlet. In southern Washington available data (Blackwell and Steele 1983) indicate a continuation of this transition. However, in northern Washington the transition is not so well defined.

Heat fluxes measured offshore in young oceanic crust (Davis and Lister 1977; Hyndman *et al.* 1978, 1982) generally are high, consistent with the previously described tectonics.

Measurements

The measured parameters for the new heat flux determinations are given in Table 2. Standard methods were used to measure the temperatures in boreholes and the thermal conductivity and heat generation from representative samples. A calibrated bead thermistor, the temperature sensor in our portable temperature logging system, has an absolute accuracy of 0.02°C and a relative accuracy of 0.002 K. Depth is resolved to ±0.3 m, but only over a small depth interval of 20 m is this a cause of significant error. In general the temperature gradient in a borehole is measured to an accuracy of 3% when corrections for dip and cable stretch (T. J. Lewis 1975) are included. The interval over which the gradient was calculated is indicated for each borehole in Table 2.

The thermal conductivity was measured on either cores or chip samples from nearly every borehole. A divided bar was used to measure 1 cm thick water-saturated disks or small cylindrical cells carefully packed with water-saturated chips (Sass *et al.* 1971). The thermal conductivity of the chips in each cell was calculated using the geometric model. Reproducibility of the measurements is within 3% for disks and within 15% for each repacked cell. The number of samples, type of sample, and thermal conductivity are given in Table 2 for each borehole. At best, mean conductivities are estimated to be accurate to 5% for disks and 10% for chips.

The concentrations of the long-lived, naturally occurring radionuclides, uranium, thorium, and potassium, were measured using gamma-ray spectroscopy. The method, described by T. J. Lewis (1974), has since been altered to use a constant sample mass of 330 g and assumes constant absorption. Checks for equilibrium of the uranium series were made using three weak peaks at energies of 186, 144, and 63 keV, corresponding to concentrations of ^{235}U and ^{234}Th. The accuracy of measurement of heat generation, generally better than 10% for values greater than 0.5 μW m^{-3}, is much greater than required, given the natural variation in nearly all rocks over distances of the order of a metre. Table 2 contains the average heat generation where representative samples of the most recent magmatic event could be obtained. Such samples are thought to be most related to the crust. Individual values of heat generation measured on many surface samples representing the crust as well as individual values used in this and previous studies are contained in T. J. Lewis *et al.* (1984) and T. J. Lewis (1976).

There are two major sources of error in obtaining a regional heat flux from measured values in this terrain: refraction of the conducted heat and redistribution of heat by groundwater. Except for the Intermontane Belt, the topography for the entire area is very rugged and groundwater flows are prevalent, supplied by ample precipitation and flowing through rocks of quite basic composition, which do not quickly form clays to seal off fractures. The effects of refraction caused by surface topography and (or) known structures of different thermal conductivity can usually be modelled. The effects of water flows can be divided into two types, dependent on whether the measured temperatures indicate flowing water (Drury *et al.* 1984; T. J. Lewis and Beck 1977), or whether water flows along and changes the temperature of rock boundaries (joint systems, fractures, fault zones, volcanic flow tops, or permeable strata) remote from the borehole. The latter can affect the measured heat flow and yet in a single borehole be impossible to detect, let alone model.

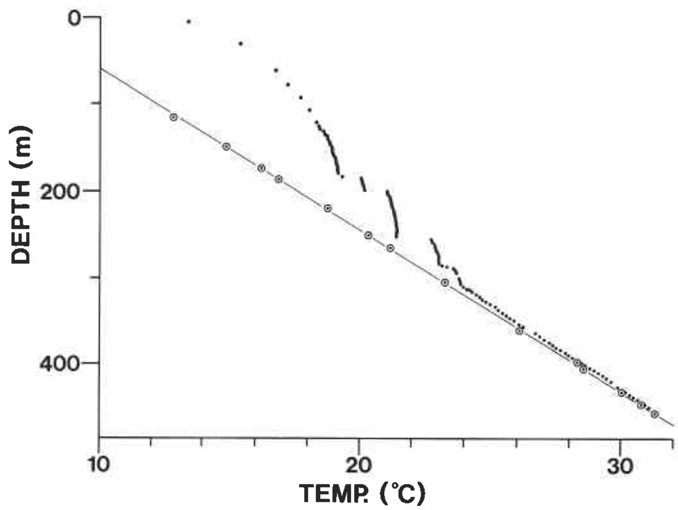

FIG. 2. Bottom-hole temperatures measured during drilling (enclosed points) and a complete temperature log of a borehole near Burrell Creek in the Coryell syenites.

For example, we interpret the temperature log shown in Fig. 2 to indicate that water enters this hole at depths of 184, 200, 256, 284, and 310 m and flows upwards. Bottom-hole temperatures measured during pauses in the drilling verify this interpretation (T. J. Lewis *et al.* 1979). We consider these excellent data, but if we had just the single final temperature log down to a depth of 284 m, then these data would be discarded just as we have discarded other logs. If the hole had been successfully grouted, then the water flows would have been stopped, but we would have no indications of where the permeability occurred and the availability of water.

Except for sites described otherwise, all heat fluxes given in Table 2 were calculated in the following manner. Temperatures were used from the large intervals of depth given in Table 2 in which there appeared to be no direct or indirect influences of water flows. The temperatures were corrected for the effects of topography and lakes or inlets using the methods of Jeffreys (1940) and Lachenbruch (1957), with lapse rates varying between 3 and 10 mK m^{-1}. The temperatures were plotted as a function of the thermal depth (Bullard 1939), and the slope of the best fitting straight line determined the heat flux without glacial correction shown in Table 2. To indicate the size of the topographic correction, the measured heat flux without this correction is also shown.

Corrections for the effects of glacial periods (Jessop 1971) were added and at each site depended on the extrapolated

TABLE 2. Heat flux and associated

Site	Location		Collar elevation (m asl)	Measured interval (m)	Thermal conductivity			
	Lat. N	Long. W			n	m*	(W mK^{-1})	SD
McLease L.	52°31.1′	112°17.4′	1070	190–290	20	c	3.79	0.36
	52°31.2′	112°17.6′	1050	0–230	16	c	3.72	0.26
Mt. Polley	52°38.0′	121°33.3′	1158	80–310	19	c	2.55	0.21
	52°38.1′	121°33.5′	1148	30–225	13	c	2.23	0.27
Bralorne	50°46.′	122°48.′	1212	0–1520	4	d	3.24	†
Coquihalla	49°30.7′	121°17.3′	1208	410–490	20	c	3.12	0.50
Goldstream R.	51°37.7′	118°25.0′	666	60–280	19	d	2.70	0.72
	51°37.7′	118°24.9′	667	100–260	18	d	2.82	0.75
	51°37.8′	118°25.0′	665	125–260	18	d	2.70	0.52
Zenith	49°19.4′	121°55.5′	732	250–438	18	d	2.47	0.37
Bethlehem	50°28.6′	120°58.2′	1207	0–460	25	c	2.89	0.40
	50°28.6′	120°58.2′	1196	0–460	22	c	2.79	0.47
	50°28.5′	120°58.3′	1188	0–430	14	c	3.03	0.48
	50°28.6′	120°58.1′	1199	0–300	28	c	2.80	0.43
	50°28.3′	120°57.9′	1185	0–460	17	c	2.52	0.20
Clearwater	51°31.2′	119°48.5′	1708	50–275	17	c	4.20	0.82
	51°31.1′	119°48.5′	1683	50–250	21	c	4.32	1.18
Myra Cr.	49°34.6′	125°36.6′	392	100–780	90	d	3.35	0.80
	49°34.3′	125°35.2′	312	100–650	67	d	3.50	0.58
Britannia	49°36.7′	123°09.6′	−389†	150	19	c	4.67	1.52
	49°36.7′	123°09.5′	−389†	40	7	c	4.09	0.52
	49°36.6′	123°09.1′	−393†	60	8	c	3.90	0.63
	49°36.8′	123°10.1′	−386†	150	0	—	4.15	—
	49°36.8′	123°10.1′	−386†	100	0	—	4.15	—
Getty	50°39.1′	120°28.5′	765	70–250	18	c	2.76	0.34
Afton	50°39.7′	120°30.6′	669	100–265	17	c	2.36	0.21
Leemac	50°38.2′	120°27.5′	945	85–135	12	c	2.72	0.44
Maggie	50°54.4′	121°24.6′	512	90–200	9	c	2.95	0.42
	50°54.1′	121°24.5′	512	80–170	8	c	2.47	0.39
Owl Cr.	50°25.6′	122°50.2′	1295	35–90	9	c	2.59	0.23
Granby	49°26.2′	118°25.4′	720	50–460	33	d	2.14	0.11
	49°25.1′	118°31.1′	750	50–450	34	d	2.14	0.44
Island Copper	50°36.3′	127°28.6′	244	0–760	40	d	2.80	0.77
Gambier Is.	49°30.7′	123°22.0′	137	160–280	19	c	2.29	0.34

*The conductivity measurements were made on chip samples (c) or disk samples (d).
†See text for explanation of this value.
‡See text for method of calculation.
[ ] Value from samples of the surrounding region.

surface temperature, the geothermal gradient, and the depths of measurements. This correction is very dependent upon these site parameters, as Jessop stressed. For example, the extrapolated surface temperature, 8.8°C, at a borehole within 10 km of and at the same elevation as the Saanichton Experimental Farm, is 3.6 K less than the 1.5 m depth temperature at the farm used by Jessop to calculate approximate corrections. For this borehole site the correction is 58% less than Jessop's generalized correction because of this one parameter.

Most uncertainties estimated for the heat fluxes in Table 2 are larger than the standard deviations of the straight lines fit to each Bullard plot. The largest standard deviation is 2% for Clearwater. Even though these are very good data, the resolution and accuracy of measurement, especially for thermal conductivity, restrict our claims to a maximum accuracy of 5%. Figure 3 shows the data plotted for the Zenith site as an example. Where sites or boreholes are close together, the average value is given as the accepted value in Table 2.

Site-specific considerations

Some sites require additional consideration, mostly for two reasons: rugged topography, which is not adequately modelled by the Jeffreys approximation, and ill-defined thermal conductivity in complex localized geology, to be expected near parameters for each site

Heat Flux: Measured (mW m^{-2})	Heat Flux: Without glacial correction (mW m^{-2})	Heat Flux: Corrected (mW m^{-2})	Heat Flux: SD (mW m^{-2})	Heat Flux: Accepted value (mW m^{-2})	Reliability (%)	Heat generation: (μW m^{-3})	Heat generation: n	Sample description	Reduced heat flux (mW m^{-2})
67	67	79	1.2	74	5	0.5	3		70
61	60	69	1.0						
69	70	78	0.3	74	5				
58	62	69	0.5						
85	80	85*		85	15				
61	63	64	0.4	64	20				
152	104	110	2.3	116	15	3.7	1	Biotite–hornblende	
111	—	—	—	—				quartz monzonite	84
170	115	122	5.5	—					
46	50	52	0.5	52	5				
77			1.1						
74			1.1						
74	72	78‡	0.7	78	10	0.8	6	Quartz diorite	71
80			0.6						
64			0.4						
98	101	105	2.4	114	5	4.8	1	Biotite granodiorite	73
115	119	123	2.4					and granite	
50	36	38	0.4	40	10	0.6	18		37
56	41	43	0.6						
62	68	68	2.3						
50	56	56	1.1						
78	84	84	0.4	71*	20	1.0	9	Quartz diorite	54
67	73	73	0.4						
68	74	74	1.1						
85	85	93	0.5						
83	84	90	0.5	82	5	0.7	8	Various intrusives	76
55	55	64	1.5						
81	81	89	0.7	82	5	0.9	7	Biotite quartz	74
67	67	75	1.3					diorite porphyry	
44	41	45	1.8	45	20	[0.9]			
111	113	116	0.7	116	5	4.9	19	Syenite	74
112	111	114	1.3	114		4.9†	20	Syenite	72
64	65	69	0.5	69	10	0.9	19	Quartz feldspar porphyry	61
32	31	39	0.4	39	15				

mineral deposits.

Thirteen temperature measurements at Bralorne were made in near-horizontal boreholes from six different levels of the mine beneath extremely rugged surface topography. There is a strong possibility that the heat flow is distorted by contrasting thermal conductivities within the complex geological structure. Samples were not available from all of the formations, so we could not make a thermal conductivity model to represent the structure. However, the vertical temperature gradient over 1500 m is very uniform: 26.2 mK m^{-1} ± 0.9%. An uncertainty in the measurement of 15% is assigned: 10% due to poor conductivity sampling and 5% due to the effect of the structure.

Most of the temperatures in mineral exploration holes on a mountain ridge at the Coquihalla site are directly influenced by water flows. Some of the data from the bottom of a 490 m hole are used to calculate a heat flow of 64 ± 0.4 mW m^{-2}, but a 20% reliability is assigned.

The three boreholes at Goldstream are near the bottom of a deeply incised river valley. Since the valley, with sides rising 1500 m, is linear, the Jeffrey's topographic correction, which tends to underestimate such effects, was replaced by a numerical, two-dimensional calculation, as described by Finckh (1981) and Lee and Henyey (1974). The calculated correction, 51 mW m^{-2}, is nearly twice as much as that from the Jeffreys

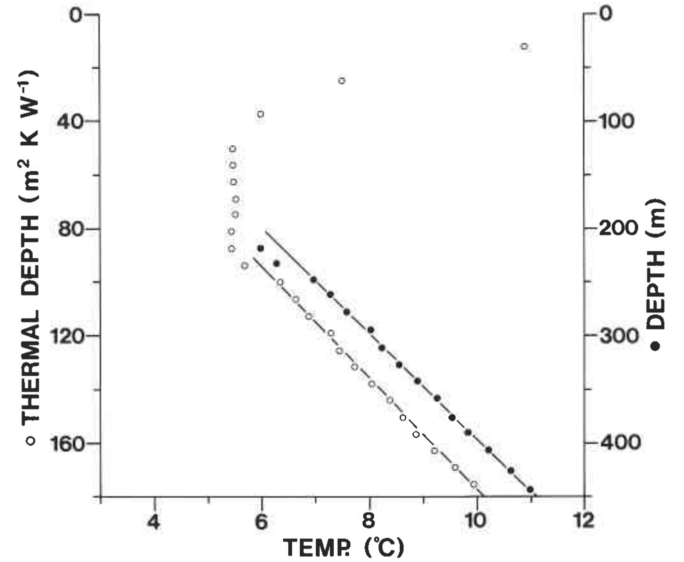

FIG. 3. Measured temperature as a function of depth (solid symbols) and corrected temperatures as a function of thermal depth (open symbols) for the Zenith borehole drilled from a relatively high elevation through volcanic rocks. Water flows are disturbing the upper half of the borehole.

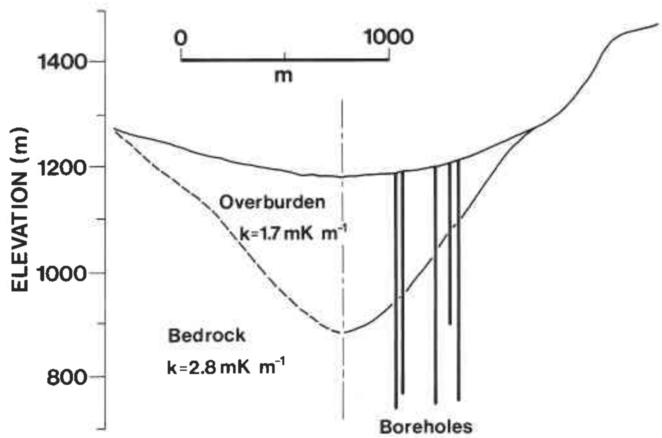

FIG. 4. A model of the Highland Valley used to calculate the combined effects of refraction and topography. Note the exaggerated vertical scale.

approximation. There are indications of water flow in this rugged terrain, and the rocks are steeply dipping metasediments, including phyllites, limestones, and a lower siliceous phyllite (Höy 1983). An uncertainty of 15% is assigned.

At the Bethlehem site several boreholes penetrated up to 300 m of glacial till forming the valley floor before entering basement rock of the Guichon Batholith (Northcote 1969). Nearly all of this valley fill was present before the most recent glacial episode (W. MacMillan, personal communication, 1983). A two-dimensional relaxation model with contrasting thermal conductivities of 1.7 (fill) and 2.8 (bedrock) W mK^{-1} was used to calculate the results of the two competing effects of refraction and topography (see Fig. 4). The thermal conductivity of the overburden was calculated as that of the bedrock multiplied by the ratio of the average measured geothermal gradients. The correction, averaged for five boreholes, was −4%, and an uncertainty of 10% is assigned. The individual corrections tended to produce a uniform heat flux, although the 50 × 50 half-valley grid could not represent small irregularities in the bedrock surface.

Two deep mineral exploration boreholes at Myra Creek penetrate cherty volcanic tuffs and breccias of the Sicker Group. High-resolution temperature logs reveal slight offsets of 0.1 K in vertical distances of 15 m, which cannot all be accounted for by changes in the thermal conductivity. One of the holes was deepened from 700 m to 800 m, and a second temperature log revealed that water was flowing up the hole from below 780 m. Similar small water flows along discrete horizons in old volcanics have been observed previously (T. J. Lewis and Beck 1977). It is possible to account for the direct effects of water flow within the boreholes, and this results in the values appearing in Table 2. The deeply incised valley has walls with average slopes of 7 in 10. The Jeffrey's approximation for topographic correction, 25%, was not of adequate accuracy. A two-dimensional model across the valley, as described previously, gave a correction of between 15 and 60% along the length of the deepest hole, similar to a corresponding curvature in the measured temperature gradient.

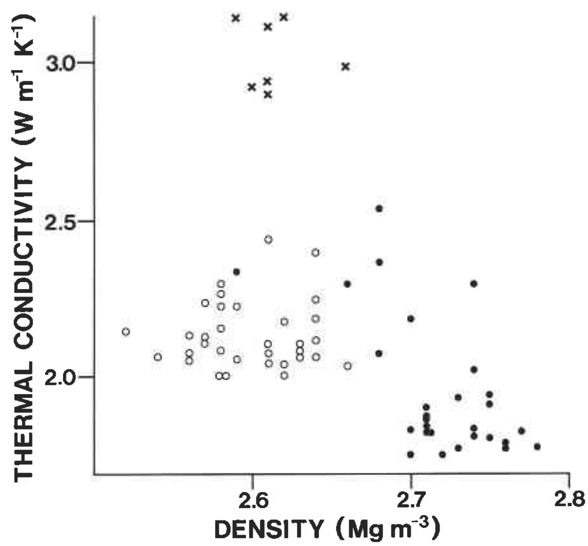

FIG. 5. Measured density and thermal conductivity of core samples from a hole near Burrell Creek penetrating Coryell syenites (open circles) and from a hole near the Granby River penetrating both fine-grained Coryell syenites (×) and a roof pendant of the Nelson Batholith (solid circles).

The difference in the two measured heat flow values and the large variations in thermal conductivity combined with the large topographic correction lead to a low reliability rating (10%) from such a large set of otherwise high-quality data.

The collars of boreholes measured at the Britannia site are deep below the surface (up to 1860 m) in a volcanic pendant intruded by several plutons. In the region of mineralization, located in a geologically complex shear zone, the amount of quartz and silicification of metasediments varies greatly, causing large variations in the thermal conductivity. Samples for the determination of conductivity were not available for two boreholes for which average values from the three other holes were used. The calculated corrections to the heat flux for glacial effects are negligible at the depths of the boreholes. However, the topography is very rugged, as the mountains rise 1700 m from the fjord. A two-dimensional model of the region was used to calculate a topographic correction of +8%. The reliability of the heat flux at this site representing the region is decreased by a possible large variation in measured heat flux

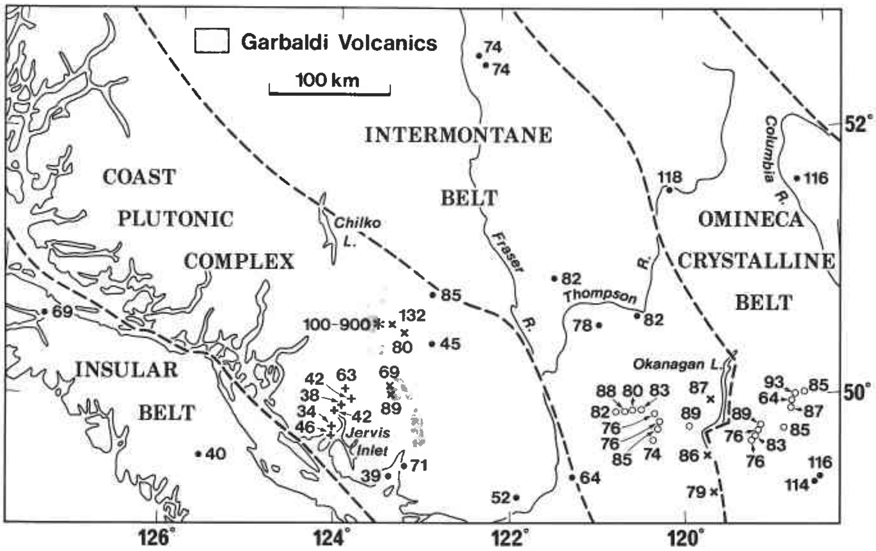

FIG. 6. Heat flux in southern British Columbia. Previously published values are shown as open circles (shallow holes), crosses (deeper holes), pluses (oceanic-type measurements in Jervis Inlet), and an asterisk (Meager Mountain geothermal area). The Garibaldi Volcanic Belt is the stippled area.

caused by this local geological structure.

Our interpretation of representative samples for heat generation can be illustrated on the high-quality data from two boreholes drilled 7 km apart into the Coryell syenites in the Granby River area. Both holes encountered water flows, but series of bottom-hole temperatures measured at the end of regular breaks in the drilling (see Fig. 2) define two linear temperature gradients of 55 mK m^{-1} (T. J. Lewis *et al.* 1979). Unfortunately the second hole intersected the boundary between the Coryell syenite and a roof pendant of Nelson Batholith several times. For calculating the heat flux, values of thermal conductivity from all core samples, shown in Fig. 5, were used. However, the heat generation value for the second hole is taken to be the average of those measured from the younger Coryell syenites in borehole cores and does not include the lower values of the older Nelson Batholith (T. J. Lewis 1976). The differences in these rocks are illustrated in Fig. 5. Note that the youngest magmatic body of any significant size is used to represent the crust.

In a deep hole near Island Copper's open pit on Vancouver Island the measured heat flux varies systematically, from 59 in an upper section of metasediments to 73 mW m^{-2} in a lower section of quartz feldspar porphyry. Since the variation appears to be related to refraction, the heat flow is taken to be 64 mW m^{-2} from the best straight line fit to all of the data below 100 m depth on the Bullard plot. The reliability is set at 10%. A low heat generation was measured on 19 samples of the quartz feldspar porphyry. The heat generation (T. J. Lewis *et al.* 1984) of other intrusives in this region is low as well.

The borehole on Gambier Island in Howe Sound is collared in mineralized andesite surrounding a small porphyry plug. Results from a second hole disturbed by water flow also indicate a low heat flow. However, a low reliability is assigned, since local variations in thermal conductivity are large.

Figure 6 shows the heat flux determined throughout the study area from boreholes, as well as published values from Jervis Inlet using oceanic techniques (Hyndman 1976). The heat flux is high and variable near the Garibaldi Volcanic Belt, as expected and as observed near other Cascade volcanoes (e.g., Steele and Blackwell 1982). Otherwise the heat flux appears to be generally higher inland compared with values on the Pacific side of a boundary somewhere under the Coast Plutonic Complex. More recent measurements along Jervis Inlet further delineate a transition occurring over a distance of 20 km (T. J. Lewis *et al.*, in preparation).

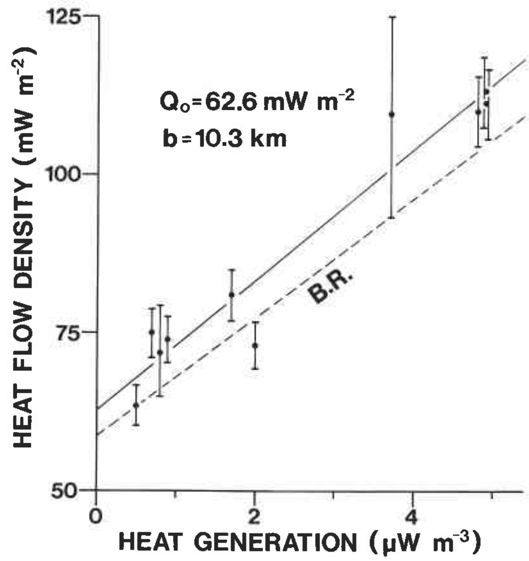

FIG. 7. Heat flux uncorrected for glacial effects and heat generation for sites in the Intermontane and Omineca Crystalline belts.

Interpretation

Roy *et al.* (1968) showed that there exists a linear relation between heat flux and crustal heat generation for large areas of the crust called heat flow provinces. Figure 7 shows the best line fitted to such heat flux and heat generation data from the Intermontane and Omineca Crystalline belts. The distribution of heat generation within the southern Insular Belt and Coast Plutonic Complex is well known (T. J. Lewis 1976; T. J. Lewis *et al.* 1984), and the average values are 1.1 and 0.8 μW m^{-3}, respectively. We could associate a heat generation value with each heat flux within the Insular Belt and Coast Plutonic

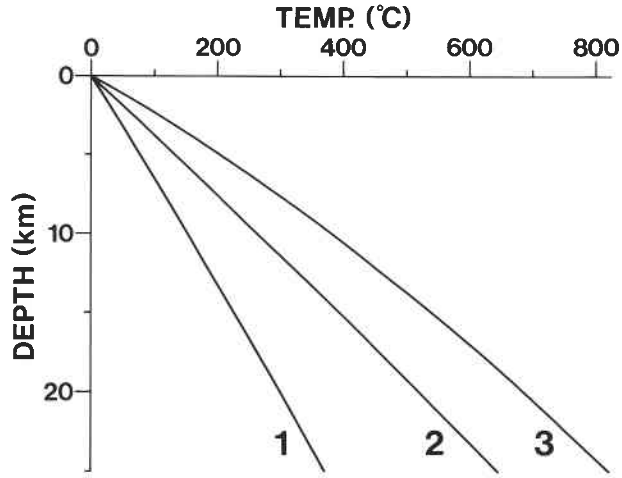

FIG. 8. Crustal temperatures in the Insular Belt and western Coast Plutonic Complex (line 1) and in the Intermontane and Omenica Crystalline belts of southern British Columbia for surface heat generations of 0.5 (line 2) and 5.0 (line 3) $\mu W\ m^{-3}$. Vertical conductive heat flow and a constant thermal conductivity of 2.5 $W\ m^{-1}\ K^{-1}$ were assumed; using a conductivity that decreases with temperature will produce even higher temperatures.

Complex. However, all of these fluxes are less than Q_0 (63 mW m^{-2}) and thus would plot below the line in Fig. 7, except for those measured in the Garibaldi Volcanic Belt.

Intermontane and Omenica Crystalline belts

The Intermontane and Omineca Crystalline belts are parts of one heat flow province with parameters $Q_0 = 63$ mW m^{-2} and $b = 10.3$ km. If glacial corrections are included, the parameters become 69 mW m^{-2} and 10 km, respectively. Data representative of the crust from the Intermontane Belt north of the study area (sites 1, 2, and 5 of Jessop *et al.* 1984) plot below this line. Although Basin and Range data from the younger areas do not all fit such a linear relation (e.g., Blackwell 1978, 1983; Lachenbruch and Sass 1978), for the sake of comparison the original line representing such data is also shown in Fig. 7 as B.R. Note that this line does fit data from areas in the Basin and Range Province having ages of their last major volcanism between 17 and 55 Ma (Blackwell 1978), and these areas may be part of the same heat flow province that includes the Intermontane and Omenica Crystalline belts of southern British Columbia as well as the Washington–Montana area. Such a high reduced heat flow causes high crustal temperatures. The geophysical and geological data previously reviewed are also evidence for a thin, warm crust in the Intermontane and Omineca Crystalline belts.

Within a heat flow province, sections of crust with differing heat production will have different crustal temperatures. Values of heat production in the Intermontane and Omenica Crystalline belts vary by an order of magnitude. The temperatures at depth are a function of the heat flux, thermal conductivity, and heat generation. Figure 8 shows the expected maximum and minimum crustal temperatures in the "crystalline" areas of this heat flow province. For example, where the surface heat generation is 5.0 $\mu W\ m^{-3}$ the temperature at 20 km is likely to be 160 K higher than where the surface heat generation is only 0.5 $\mu W\ m^{-3}$. Beneath the few basins filled with volcaniclastic sediments, temperatures will be higher as a result of the lower thermal conductivity of the sediments (T. J. Lewis 1984). A recent magnetotelluric survey indicates low electrical conductivities rising to a minimum depth of 7–8 km in the basement beneath the White Lake basin.

High heat fluxes and reduced heat flow have been previously investigated for the Basin and Range Province (e.g., Blackwell 1978; Lachenbruch and Sass 1978). To account for the observed values, all models must include convection within the lithosphere during major episodes of extension. Evidence for present-day tectonic extension and magmatism is plentiful in younger regions of the Basin and Range. In the Intermontane and Omineca Crystalline belts of southern British Columbia evidence is limited to a major thermal event heating much of the crust 50 Ma ago and a major volcanic event in the northern part of this area 6–10 Ma ago.

Blackwell (1983) argued that the heat flux data from the Basin and Range Province do not represent the province, since all of the values come from the ranges where refraction and possibly other processes increase the heat flux relative to the valleys. Although the data set from southern British Columbia is beset by corrections, there are very few volcaniclastic filled basins, and White Lake basin is the only basin from which we have data. The variation within this basin has been studied (T. J. Lewis 1984), and the value accepted to represent the basin, the lowest measured value, is comparable to values from nearby areas.

If a warm, thin crust and lithosphere cooled entirely by conduction, then a conductive model could be used to predict the heat flux as a function of time (e.g., Blackwell 1978). This assumes that the crust and lithosphere were completely heated by some process such as large extension and "dyke" intrusion (Lachenbruch and Sass 1978). Conductive cooling of the crust such as that modelled by Royden *et al.* (1980) could probably account for observed heat fluxes at times greater than 20 Ma after the thermal event. The higher heat flux predicted for younger crust may not be observed everywhere if the upper crust is cooled more quickly initially by shallow convective systems.

However, conductive cooling of the lithosphere over a period of 50 Ma would require significant subsidence over the entire area. This is not generally believed to have happened, and in some areas there have been significant amounts of erosion between 50 and 15 Ma ago (Ewing 1981). More likely a mechanism to explain the present high heat flux is the flow of the asthenosphere towards the subducting plate where it then sinks downwards with the subducted plate. Davis and Lewis (1984) first suggested this model for this area, and such a model may apply also to the Basin and Range Province. The present high heat flux in the study area is not associated with present-day extension, and to be caused by past extension requires very large rates of extension for which there is no evidence.

Insular Belt and Coast Plutonic Complex

Most of the Insular Belt and western side of the Coast Plutonic Complex are areas of low heat flux, as expected above a subduction zone (Riddihough and Hyndman 1976). The lower crustal temperatures to be expected from conductive heat flow are shown in Fig. 8. Exceptions are found to the north in the Queen Charlotte Basin (north of Vancouver Island), where uplift, rifting, and extension took place 17 Ma ago (Yorath and Hyndman 1983), and near the late Tertiary Alert Bay Volcanic Belt on northern Vancouver Island.

The sharp eastward transition from 40–50 mW m^{-2} to 80–100 mW m^{-2} in the Coast Plutonic Complex occurs over a distance of 20 km about 30 km seaward of the Plio-Pleistocene Garibaldi Volcanic Belt. Any possible model to fit the observations must explain this large transition over a distance of 20 km. If heat flows by conduction at the surface (and it is only the conductive component that we have measured) then the "source" of the extra heat (or "sink" of the missing heat) must lie within 10 km of the surface. In Oregon an abrupt transition in heat flux also occurs, similar in both magnitude and position relative to the youngest volcanics (Blackwell *et al.* 1982; Blackwell 1983). Hot springs are located along the western boundary of the transition in Oregon, as they are along the heads of the inlets in southern British Columbia. The young volcanic centres in the Oregon Cascades, located along faults farther inland from the transition, occupy a graben whose western bounding faults (Priest 1983) nearly coincide with the heat flux transition. Connard *et al.* (1979) detected a shallow Curie isotherm, showing that the high surficial heat flow is the result of a warm upper crust.

There are at least two different possible causes of such heat flux transitions: convective heat transfer upward to a depth of approximately 10 km, and differential uplift and erosion of the crust. The first process may have associated with it partial melting, or intermittent magma chambers, or long lasting magma chambers. Blackwell *et al.* (1982) favoured magma at a depth of 10 km as the source of heat under the young Oregon Cascades. The existence of a low-density material is consistent with the negative Bouguer anomalies both in Oregon and along Jervis Inlet and with the graben structure that formed in Oregon when large amounts of lava were extruded. The convective transfer of heat up through the crust to at least within 10 km of the surface is a complementary model, with the preliminary heating of the crust being a prerequisite for magma to make it up to the surface. A similar model, based on geochemical evidence, has been proposed by Tatsumi *et al.* (1983) for a region in Japan.

In the Coast Plutonic Complex large amounts of differential uplift and erosion have occurred, and it is necessary to calculate their effect on the heat flux. Using fission track data, Parrish (1983) calculated the total uplift over the last 10 Ma as 2.7 km near Bute Inlet and 4.3 km only 25 km inland. If the geothermal gradient is 20 mK m^{-1}, less than the value he used, then the amounts of calculated uplift are increased by 25%. The calculated difference in surficial heat flux between the two areas is 10%, assuming the cause to be constant rates of uplift and erosion (Benfield 1949). The calculated difference is only 30% between the area of maximum uplift and an area of no uplift. If either the rate of uplift and erosion or the period of time were increased by a factor of two, this difference would become 40–60%. Although the effects of uplift and erosion are not enough to explain the magnitude of the transition, they have occurred, and we estimate an effect on the heat flux across the transition of 25%, compared with the observed change of 100%. The existence of a major fault zone for southern British Columbia near the heads of the inlets was rejected by Parrish (1983) who claimed no geological or physiological evidence exists for it even though Culbert (1971) originally suggested otherwise.

In southern British Columbia it is not certain whether the heat flux transition is an important crustal boundary. The depth to the Moho tends to decrease from the Intermontane Belt westward into the Coast Plutonic Complex (Berry and Forsyth 1975). The discontinuity in crustal structure observed by Berry and Forsyth is at or very near the abrupt transition in heat flux. In northern Washington more heat flux data are needed to define any pattern of heat flux. However, present data (Blackwell and Steele 1983) indicate that there is an area extending north through the state of Washington just into southern British Columbia where there is no such abrupt heat flux transition. The transition is not well defined south of Jervis Inlet, and values in the southeastern Coast Plutonic Complex at Owl Creek and Zenith are low. The heat flow density at Owl Creek is of low reliability, and it is possible that water flows deep within the volcanics of the rugged terrain at Zenith indirectly affected the measured heat flux. The Benioff zone under Washington (Crosson 1983) might indicate the presence of a brittle, cool crust, and shallow seismicity outlines a subsiding basin under Puget Sound. Rogers (1983) suggested that the same stress regime exists at the northern and southern boundaries of the area.

Summary

The heat flux in southern British Columbia is low in the Insular Belt and seaward side of the Coast Plutonic Complex. In the Omineca Crystalline and Intermontane belts there is a linear relation between crustal heat production and the observed heat flux with a reduced heat flow of 63 mW m^{-2} and a characteristic depth of 10 km. This region resembles the older (<17 Ma) Basin and Range: the thin, warm crust could be a result of the entire crust being heated 50 Ma years ago. This is an extension of the Cordilleran Thermal Anomaly Zone defined by Blackwell (1969).

A transition in heat flow from 40 mW m^{-2} to 80 mW m^{-2} occurs over a distance of 20 km (also T. J. Lewis *et al.*, in preparation) approximately 30 km seaward of the Pleistocene Garibaldi Volcanic Belt in the Coast Plutonic Complex. Relating this to the general tectonic pattern, subduction of a brittle oceanic crust is taking place under Washington, whereas to the north and south, where heat flux transitions are located, subduction is occurring, with a few exceptions (Rogers 1983), aseismically. In these zones the sharp transitions in heat flow are located where the largest gradients in the gravity field occur. Blackwell *et al.* (1982) found that the gravity anomaly caused by a crustal thermal anomaly was not large enough to explain the gravity observations and proposed a zone of hot, low-density (partially molten) material in the upper crust in Oregon.

In southern British Columbia the geological evidence for large amounts of uplift in the past 10 Ma is overwhelming, but such uplift causes only part of the large differences in observed heat flux. A combination of differential uplift and convective heat transfer is responsible for the transition. The heating of the upper crust is probably a necessary step before magma can stay hot enough during ascent to reach the surface to form fault-controlled centres of eruption. Redistribution or concentration of the heat locally along fault zones may account for the small half width of the anomaly. Convergence under Oregon and British Columbia could be accommodated by a partially molten lower crust and uplift.

The nature of the boundary or boundaries between regions of differing uplift is not known. No seismicity occurs at present, except for deep earthquakes under Texada Island (Rogers 1983). The heat flux transition may be the first indication of an approximate location and cause.

Acknowledgments

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