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Heat Flow Measurements in the Inlets of Southwestern British Columbia

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The geothermal heat flux has been measured at 15 sites in the inlets of southwestern British Columbia by using the ocean probe technique. Corrections have been applied for variations in bottom water temperature, sedimentation, thermal refraction by the sediment prism, topography, warm-rim effect, Pleistocene thermal history, uplift, and erosion. The accuracy of the values is about ± 0.3 μcal cm^{-2} s^{-1} (± 13 mW m^{-2}). The relative accuracy between the values is about half of this range. The data indicate a pattern of low heat flow from the coast inland about 200 km to the heads of the inlets (13 values average 0.9 μcal cm^{-2} s^{-1} or 37 mW m^{-2}). Stations at the heads of two inlets (1.5 μcal cm^{-2} s^{-1} or 63 mW m^{-2}) mark the transition to high heat flow further inland (about 2.0 μcal cm^{-2} s^{-1} or 84 mW m^{-2}). From west to east the heat flow transition coincides with the first occurrence of recent volcanic centers and hot springs, the transition from high to low Bouguer gravity, and the transition to low deep crustal and upper mantle electrical resistivity and to recent regional uplift. The coast low heat flow is explained partly by low crustal radioactive heat production and in part by the heat sink effect of the cold oceanic Juan de Fuca plate being subducted along the continental margin. The high inland heat flow probably is produced by the upward convective transport of heat by magma and thinning of the lithosphere that occurs when the sinking plate reaches a depth where temperatures are sufficient for partial melting. Low mantle temperature and thus high density in both the continental lithosphere and the underlying subducted oceanic lithosphere under the coast zone explain the high Bouguer gravity and thick crust in this region compared with the low gravity and thin crust inland. To the north of about 51°N where at present there is transform fault motion along the continental margin, there probably was subduction prior to about 10 m.y. ago. The gravity and crustal thickness data suggest that the high density in the subducted oceanic lithosphere has disappeared but that upwelling and a hot thin continental lithosphere persist inland of 200 km from the continental edge.

INTRODUCTION

A region of lithospheric plate convergence or subduction ex-
ts along the west coast of North America between triple
oints off northern Vancouver Island and northern California.
o the north and south of this zone there is strike slip trans-
rm fault motion along the Queen Charlotte and San
ndreas fault systems. The subduction can be deduced from
fshore geophysical data on plate motions [e.g., *Morgan,*
'68; *Tobin and Sykes,* 1968; *Silver,* 1969, 1971, 1972; *Atwater,*
'70; *McManus et al.,* 1972; *Tiffin et al.,* 1972; *Srivastava,*
'73; *Barr,* 1974; *Barr and Chase,* 1974] and from recent
shore vulcanism [e.g., *McBirney,* 1968; *Dickinson,* 1970;
uther, 1970]. There is no deep Benioff seismic zone, but deep
rthquakes are not to be expected for a slowly sinking plate
acks et al., 1968] particularly if the ridge that produces the
ate is close to the trench so that the plate has not cooled
mpletely before reaching the trench [*McKenzie,* 1969;
water, 1970]. One moderately deep earthquake has been
orded, and seismic ray tracing studies indicate a dipping
ate beneath the continental margin [*McKenzie and Julian,*
71]. The rate of sinking probably is decreasing and actually
y have stopped very recently [*Crosson,* 1972], as the small
in de Fuca plate is beginning to move more parallel to the
rth American plate and is becoming coupled to it. The
nch extended further to the south in the past, but the mo-
n of the northern limit is uncertain. *Barr and Chase* [1974]
e suggested that within the past 0.7 m.y. the Explorer
eading center has been cut off by an extension of the Queen
arlotte fault to the south as far as the northern end of the
n de Fuca ridge. A knowledge of the motion of the northern triple point is essential for an understanding of the recent tectonic development of the southern British Columbia coast region [e.g., *Stacey,* 1974]. Heat flow data will help to provide this information. Heat flow measurements also are the best way to obtain estimates of crustal and upper mantle temperatures which are essential for interpreting and understanding other geophysical and geological data.

Sinking trench and island arc continental margins are characterized by a band of low heat flow just on the continental side of the trench, with much higher heat flow inland. The low zone may be produced by the cold sinking plate acting as a heat sink, absorbing heat that otherwise would reach the surface, until the plate reaches a depth of several hundred kilometers. The high heat flow inland is more difficult to explain. Probably at depths greater than several hundred kilometers the plate begins to partially melt, and magma upwelling occurs which results in high heat flow inland [*Langseth et al.,* 1966; *McKenzie,* 1969, 1970; *Hasebe et al.,* 1970; *Minear and Toksöz,* 1970]. The pattern is well-established for the west Pacific arcs [e.g., *McKenzie and Sclater,* 1968; *Uyeda and Horai,* 1964; *Hasebe et al.,* 1970]. The pattern also is apparent in the northwestern United States [*Blackwell,* 1969, 1971]. A continuation of the subduction zone heat flow pattern north of 49°N would substantiate the conclusion of a present or recent subduction zone off southwestern British Columbia. How pronounced the low is will depend on the sinking history. Rapid sinking of the cold oceanic plate should produce a marked low zone, while slow sinking will produce a less pronounced low heat flow zone. The commencement or speeding up of subduction will be evident in the surface heat flow only after several million years, and if underthrusting stops, the thermal anomaly will disappear only after some tens of millions of years, the thermal time constant of the sinking lithosphere slab. The persistence of the low heat flow zone associated with subduction has been used to locate old now ex-

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tinct subduction zones, in good agreement with data from surface volcanics [*Blackwell*, 1971; *Lipman et al.*, 1971; *Hyndman*, 1972].

Active subduction involves compression, deformation, and andesitic vulcanism along a coast. Initially there may be subsidence along the margin as the low-temperature coastal zone develops and uplift inland where crustal and upper mantle temperatures become high causing thermal expansion. Recent motions that may be caused by such changes and the pattern of recent volcanic activity have been discussed by *Souther* [1970], *Culbert* [1971, 1972], and *Stacey* [1974].

The use of the ocean probe technique in inlets permits a large number of continental heat flow values to be collected at moderate cost, although the accuracy of individual values will be lower than that for borehole or mine measurements. Quite large corrections are required for the recent thermal history of the bottom water, the Pleistocene thermal history of the area, uplift and erosion, the sediment deposition, the thermal refraction of heat away from the low thermal conductivity sediments, the difference in temperature between the land and the bottom of the inlet, and for the steep topography. Large bottom water temperature fluctuations usually make shallow water heat flow measurements impractical, but the bottom temperature variations in the deep inlets are generally less than 0.5°C, and detailed water temperature records for up to 20 years are available for many of the British Columbia inlets. Large heat flow differences are to be expected in this tectonically active area, so that values with even quite low accuracy can give very useful information. Measurements using the ocean probe technique in lakes and inlets of other areas have had accuracies of about ±20% [*Law et al.*, 1965; *Steinhart and Hart*, 1965; *Lubimova and Shelyagin*, 1966; *Reitzel*, 1966; *Von Herzen and Vacquier*, 1967; *Steinhart et al.*, 1968; *Sclater et al.*, 1970; *Hanel*, 1970; *Rankin and Hyndman*, 1971; *Lee and Henyey*, 1971].

Nineteen measurements have been obtained in this study, 17 in inlets and two in Powell Lake, a former inlet recently cut off from the sea by regional uplift. Four additional measurements were made in Knight Inlet and in Sechelt Inlet (off Jervis Inlet), but the gradients were nonlinear and there were not adequate bottom water temperature data to make satisfactory corrections. The inlet measurements were obtained on two cruises of the research vessel C.S.S. *Vector*, March 6–10 and May 1–5, 1972. The two measurements in Powell Lake were from a chartered 6-m outboard launch. The station locations and depths in the inlets were obtained by ship's radar and depth sounder and should be within ±0.15 km and ±10 m. The depths of the inlets ranged from 217 to 685 m. The locations of the measurements are shown in Figure 1.

MEASUREMENTS

Gradient. The temperature gradients were measured with a 4.5-m-long, 2.8-cm-diameter solid Bullard type probe having four sensors and with a 2.5-m-long, 1.3-cm-diameter solid probe with three sensors. Water temperature and instrument tilt were recorded at most stations. Complete or nearly complete and vertical penetration was obtained at all of the successful stations. However, little or no penetration could be obtained at a number of stations attempted in parts of Bute, Toba, and Knight Inlets where glacially fed streams bring down coarse sediments. These sites are unsuitable for accurate measurements in any case, because of the uncertainties in bottom water temperature variations arising from sediment-laden bottom currents and because of the large and uncertain sedimentation correction.

A small strip chart recorder was used to record the output or balance voltage of a Wheatstone bridge in which a water-temperature thermistor located just above the instrument package is compared to the probe thermistors successively. The accuracy is from ±2 to ±10 m°C (about 2–10% error in gradient) primarily depending on the temperature difference sensitivity used. The zero gradient reference was obtained in the inlets by holding the probe stationary near the bottom where water temperature profiles have shown the temperature gradients to be very small. Powell Lake has a large gradient in the water column near the bottom, and so the zero gradient reference was found in a constant temperature laboratory tank in which the probe and electronics were kept at the temperature of the bottom water. The probes were left in the sediment for from 20 to 40 min, a duration which requires a small extrapolation to equilibrium temperatures following the frictional heating of penetration [*Bullard*, 1954]. At three sites the probe was lowered twice because the initial sensitivity was too low for good accuracy or so high that the deepest sensor was off scale.

Conductivity. Sediment cores were taken at each station for conductivity measurement, in most instances with a plastic-lined 4-m-long, 6-cm-diameter gravity corer. A few 3.5-cm-diameter cores also were used. The core lengths were from 1.5 to 2.5 m, considerably less than the depths of penetration of most of the gradient measurements, but no systematic variation of conductivity with depth was observed. Thermal conductivities were measured at about five places along each core by using the transient needle probe technique [*Von Herzen and Maxwell*, 1959] while the core remained encased in the plastic liner. Most of the measurements were completed within 24 hours of the coring and all within 5 days. The values were corrected by −4% after *Ratcliffe* [1960] for the difference between in situ and measurement temperatures. Fused silica was used as a conductivity reference [*Ratcliffe*, 1959]. Sediments from Saanich Inlet and from Powell Lake were chemically reducing and expelled gas when they reached the surface. Most of the gas will have escaped before the conductivity measurement, but that remaining will have resulted in values that are slightly too low. To some extent this effect is offset by the consolidation and compaction that occur during coring and handling this gelatinous kind of sediment. The maximum estimated error in the harmonic mean conductivity used for each station is ±10%, most values being within ±5%.

HEAT FLOW CORRECTIONS

Six major corrections to the measured thermal gradients are required to give the regional heat flux. They commonly have magnitudes up to 30% and in a few instances range up to 60%. They have been treated separately to obtain the final heat flows. They are not strictly independent, but it is exceedingly difficult to solve the complete problem. By applying the corrections in order from the smallest to the largest dimension scale and from the shortest to the longest characteristic time scale, the error is minimized. The corrections were applied in the order given in the text and in Table 1. The corrected heat flows are given in Table 2.

Recent thermal history of bottom water. The inlets are all very deep and are connected to the sea by shallow sills, so that there are virtually no annual variations in bottom water temperature. However, irregular changes of up to 2.0°C do oc-

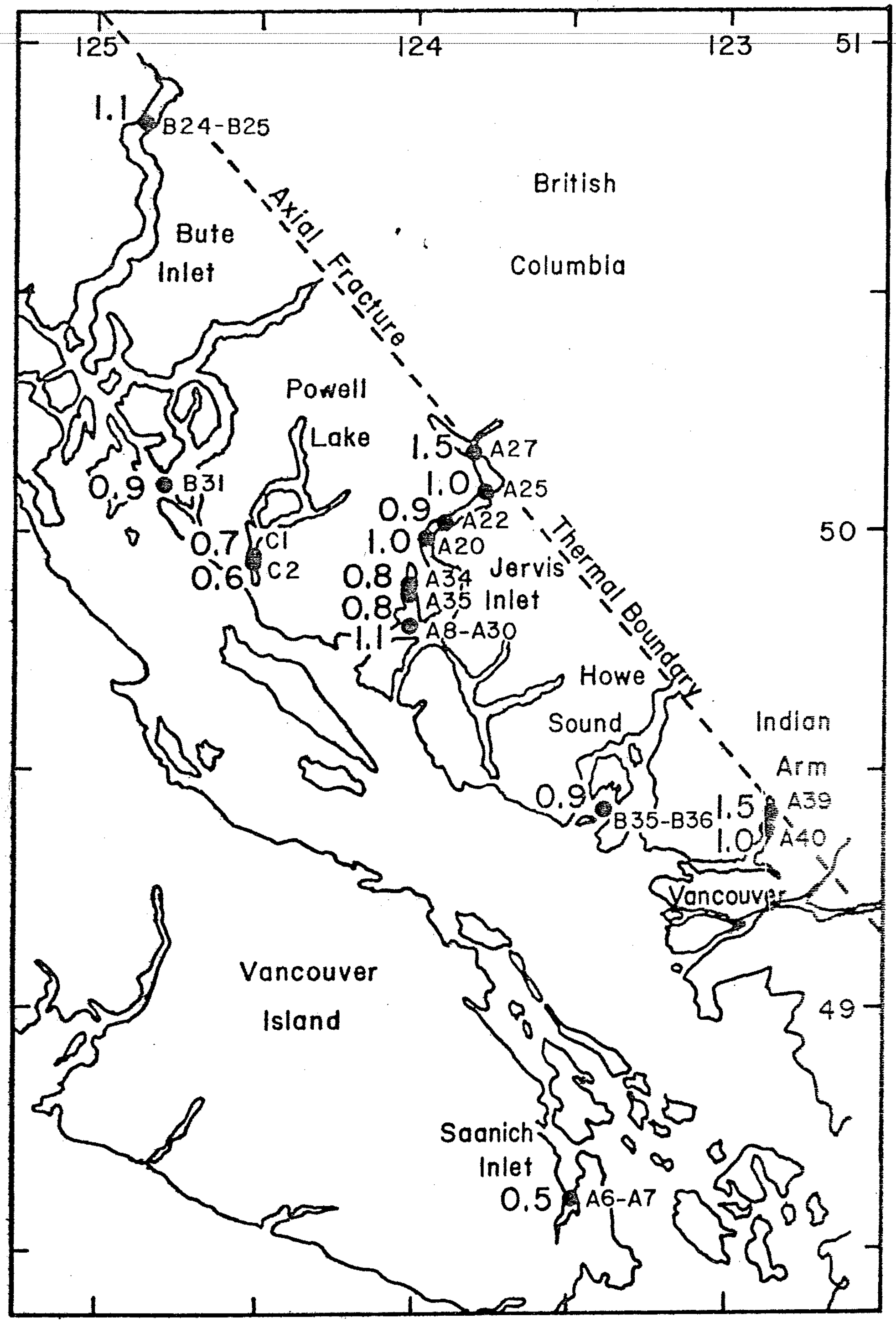

Fig. 1. Heat flow measurements in British Columbia inlets. The line of Recent and Pleistocene volcanic activity and hot springs which coincides with *Culbert*'s [1971] axial fracture is shown marking the transition from low to high heat flow. It also coincides with the transition from high to low gravity and with the heads of most of the large inlets, to suggest recent subsidence to the west and uplift to the east.

…in part from storm-driven surges. At several stations the …sturbance to the gradient exceeds the corrected or regional …ue. There are two sets of data that can be used to correct for …s disturbance: the records of bottom water temperatures …om physical oceanographic measurements and the sediment …perature-depth profiles from the thermal gradient probe. …cords of bottom water temperatures since 1952 are available …r many inlets, acquired by the University of British Columbia Institute of Oceanography and by several government departments [*University of British Columbia*, 1952–1972; *Pickard*, 1961; *Herlinveaux*, 1962; *Gilmartin*, 1962].

Sediment temperature-depth profiles, linear to within the estimated measurement accuracy, were obtained at 11 of the 15 inlet stations. The gradient corrections at these stations from the recorded bottom water thermal history were quite small except for the station in Saanich Inlet where the probe penetra-

TABLE 1. Thermal Gradient Corrections

Station	Uncorrected Gradient	Gradient Corrected for Bottom Water Variation	Sedimentation Correction	Thermal Refraction of Sediment Prism	Topography and Warm Rim	Pleistocene Thermal History	Uplift and Erosion	Corrected Gradient
Bute Inlet								
B24-B25		40	+6.8	+10.8	−15.0	+5.7	−9.5	39
Desolation Sound								
B31	54	54*	+5.9	+11.0	−18.0	+9.2	−12.2	50
Powell Lake								
C1	46	46*	+2.8	+32.6	−37.2	+11.7	−11.0	45
C2	44	44*	+2.8	+31.3	−35.7	+11.7	−10.7	43
Jervis Inlet								
A27	77	79	+17.4	+26.4	−40.4	+8.1	−17.8	73
A25	52	64	+4.5	+14.1	−28.2	+8.7	−12.4	51
A22	54	60	+9.0	+19.7	−38.5	+8.5	−11.6	47
A20	53	54	+9.2	+18.0	−30.5	+8.0	−11.6	47
A30-A8	55	61	+14.6	+14.8	−24.2	+9.5	−14.9	61
Hotham Sound								
A34	50	60	+2.4	+17.0	−38.9	+8.7	−9.7	40
A35	52	59	+3.0	+15.9	−31.9	+9.5	−10.9	45
Howe Sound								
B35-B36		36	+6.1	+9.7	−6.7	+7.8	−10.4	43
Indian Arm								
A39		96	+11.5	+30.6	−55.0	+8.3	−18.0	73
A40		71	+7.1	+31.5	−55.7	+8.3	−12.3	50
Saanich Inlet								
A6-A7	43	24	+2.4	+8.8	−9.0	+9.8	−7.1	29

Nonlinear sediment temperature-depth profiles were measured at the stations of Bute Inlet, Howe Sound, and Indian Arm.
*No bottom water temperature variation correction made.

tion was small, and so the gradient linearity was poorly determined. Thus a linear gradient provides a fair assurance that recent bottom water temperature changes can be neglected. There are, of course, thermal histories that would result in a large gradient disturbance and still give a linear gradient.

For all of the inlet stations, with the exception of Desolation Sound, the measured sediment temperatures were corrected by using the thermal history of the bottom water. No bottom water temperature records were available for Desolation Sound but the temperature-depth profile in the sediment linear and the penetration into the sediment was over 6 m, the deepest achieved; so the uncorrected gradient probably reliable. At the other stations the corrected gradient was accepted only if it was linear. Several stations with limited bo

TABLE 2. Heat Flow Data

Station	Location: North Latitude	Location: West Longitude	Elevation, m	Corrected Gradient, m°C m^{-1}	Deepest Sensor, m	Number of Temperature Points	Number of Conductivity Values	Harmonic Mean Conductivity*	Heat Flux†	Estimated Error
Bute Inlet										
B24-B25	50°49.8′	124°53.5′	−360	39	3.5	4	5	2.96	1.1	0.4
Desolation Sound										
B31	50°05.5′	124°48.8′	−527	50	6.0	3	5	1.84	0.9	0.3
Powell Lake										
C1	49°56.3′	124°32.0′	−350	45	2.5	3	5	1.45	0.7	0.4
C2	49°56.1′	124°32.0′	−350	43	2.5	3	5	1.45	0.6	0.4
Jervis Inlet										
A27	50°09.0′	123°51.0′	−360	73	4.0	4	6	2.09	1.5	0.3
A25	50°04.1′	123°48.1′	−540	51	5.0	4	5	1.95	1.0	0.3
A22	50°00.6′	123°56.3′	−547	47	3.5	3	5	2.00	0.9	0.3
A20	49°59.2′	123°59.1′	−561	47	3.0	3	5	2.12	1.0	0.3
A30-A8	49°47.7′	124°02.1′	−685	61	3.5	6	5	1.79	1.1	0.2
Hotham Sound										
A34	49°53.0′	124°03.7′	−505	40	4.0	3	3	1.96	0.8	0.3
A35	49°52.1′	124°02.8′	−560	45	3.5	3	5	1.79	0.8	0.3
Howe Sound										
B35-B36	49°25.0′	123°24.1′	−249	43	4.5	5	6	2.17	0.9	0.4
Indian Arm										
A39	49°24.7′	122°52.2′	−220	73	4.5	4	5	2.04	1.5	0.3
A40	49°23.5′	122°52.5′	−217	50	4.5	4	5	2.05	1.0	0.3
Saanich Inlet										
A6-A7	48°35.5′	123°29.8′	−228	29	2.5	6	6	1.73	0.5	0.3

*Given in mcal cm^{-1} s^{-1} °C^{-1} (0.419 W m^{-1} °C^{-1}).
†Given in μcal cm^{-2} s^{-1} (41.9 mW m^{-2}).

them water records did not meet this condition. Bottom water temperature records are available for locations within 5 km of each heat flow station. The records show quite uniform temperature over much of each inlet; so the error introduced by this distance should be negligible. The temperature changes were assumed to occur as steps half way between the temperature-time points. The theory is given by *Carslaw and Jaeger* [1959, p. 63]. The bottom water temperature data for the inlets that have been used are given in Figure 2, and the uncorrected and corrected sediment temperature-depth profiles in Figure 3. Subjective estimates of the errors in the corrected values from this source range from less than ±5% for the linear 4-m probe measurements such as Jervis Inlet to about ±20% for the very nonlinear sites of Indian Arm and Howe Sound. The statistical errors from the linearity of the corrected gradients are much less.

The bottom water temperature of Powell Lake is strongly stabilized by a relict bottom saltwater layer [*Williams et al.*, 1961; W. H. Mathews, unpublished report, 1962; T. Osborne, personal communication, 1972] so that significant temperature changes should not occur. This conclusion is substantiated by the linear and identical gradients measured by *Reitzel* [1966; unpublished report, 1966] in 1965 and by ourselves in 1972.

Sedimentation. Part of the geothermal flux from the bottom of the inlets goes into establishing a temperature gradient in the sediment which is suddenly or continuously deposited. The sedimentation histories in the inlets are poorly known. *Pickard and Giovando* [1960] estimated a recent rate of 35 cm per 1000 years for lower Jervis Inlet. Seismic profiles in several inlets [*MacDonald*, 1970; *St. John*, 1972] indicate a thin layer of recent sediment up to 50 m thick overlying more compacted sediments up to 250 m thick. The latter will have originated as glacial till and glacial outwash during the Pleistocene ice advances. There appear to be terminal moraines in the upper parts of the inlets from the most recent ice advance (Sumas Stade, 11,000–8000 years ago), and most of the sediment probably was deposited during and immediately following the more extensive earlier ice advances (e.g., Vashon Stade, 25,000–13,500 years ago or Salmon Springs glaciation, prior to 36,000 years ago [*Armstrong et al.*, 1965]). The correction has been computed by assuming rapid deposition of the estimated sediment thickness 20,000 years ago. This time is poorly determined. The slow deposition of recent sediments will have only a small effect. For the stations in the lower parts of the inlets, much of the sediment may have been deposited earlier than 20,000 years ago, and the correction may be overestimated. The correction probably has been underestimated for stations near the heads of the inlets where much of the deposition was more recent.

The simple theory of the sedimentation correction for a material of constant thermal properties has been given by *Von Herzen and Uyeda* [1963], *Jaeger* [1965], and *Lachenbruch and Marshall* [1966]. The situation in the inlets is complicated by the sediment layer overlying rock of quite different thermal properties. Since the thermal time constant of the sediment layer is less than the time since major deposition, the development of the thermal field depends primarily on the diffusivity of the rock (0.01 cm² s⁻¹). An approximation of sufficient accuracy is to assume a diffusivity equal to the rock but sediment thickness equal to that of an equivalent layer of rock, i.e., the same heat capacity per unit area. The heat capacity per unit volume of the sediment is about the same as that of the rock since the low thermal conductivity and low density of the sediment are about balanced by the high heat capacity of the water present). Thus a sediment thickness simply equal to that observed (see next section) was used in the computation. The correction ranges from +4 to +24%. The correction would be similar for continuous deposition of the sediment layer from about 50,000 to 8000 years ago.

Thermal refraction by bottom sediments. There is a major contrast between the low thermal conductivity bottom sediments and the surrounding rock which refracts heat away from the bottoms of the inlets. The mean conductivity of the sediments will be higher than that measured in the cores, since the deeper sediments will be compacted and most of the valley bottom fill is glacial outwash and glacial till from the Pleistocene ice advances. The mean sediment thermal conductivity has been estimated to be 3.0 mcal cm^{-1} s^{-1} $°C^{-1}$ (1.26 W m^{-1} $°C^{-1}$) for all of the sites except for those in Saanich Inlet and Powell Lake. In the latter two locations the measured conductivities are very low, and probably there is much less glacial outwash fill and glacial till than in the other inlets; so a sediment conductivity of 2.0 mcal cm^{-1} s^{-1} $°C^{-1}$ (0.84 W m^{-1} $°C^{-1}$) has been used for them. The bedrock conductivity has been taken to be 6.0 mcal cm^{-1} s^{-1} $°C^{-1}$ (2.51 W m^{-1} $°C^{-1}$) (quartz diorite) for a contrast of a factor of 2 for all of the sites except for Saanich Inlet and Powell Lake where the contrast assumed is a factor of 3. Seismic profiles of the sediment cross sections were available for a few sites [*MacDonald*, 1970; *Toombs*, 1956; *Gucheur and Gross*, 1964], but for most of the stations the sediment prism cross sections have been estimated

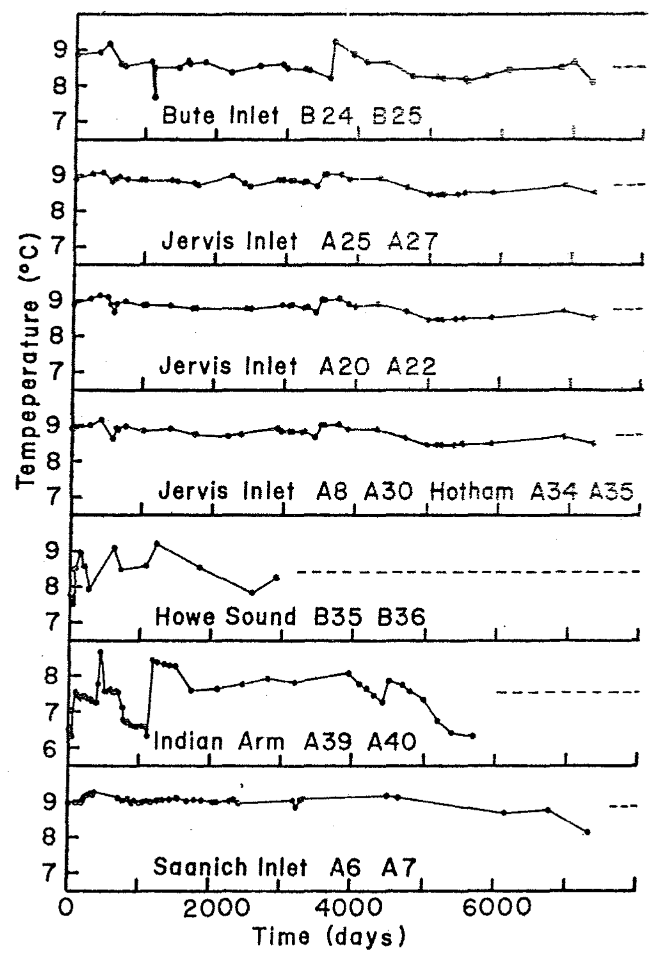

Fig. 2. The bottom water temperature records used for correcting the measured temperature gradients in the bottom sediments. The dashed lines are the mean temperatures used for the times before the recorded water temperature data.

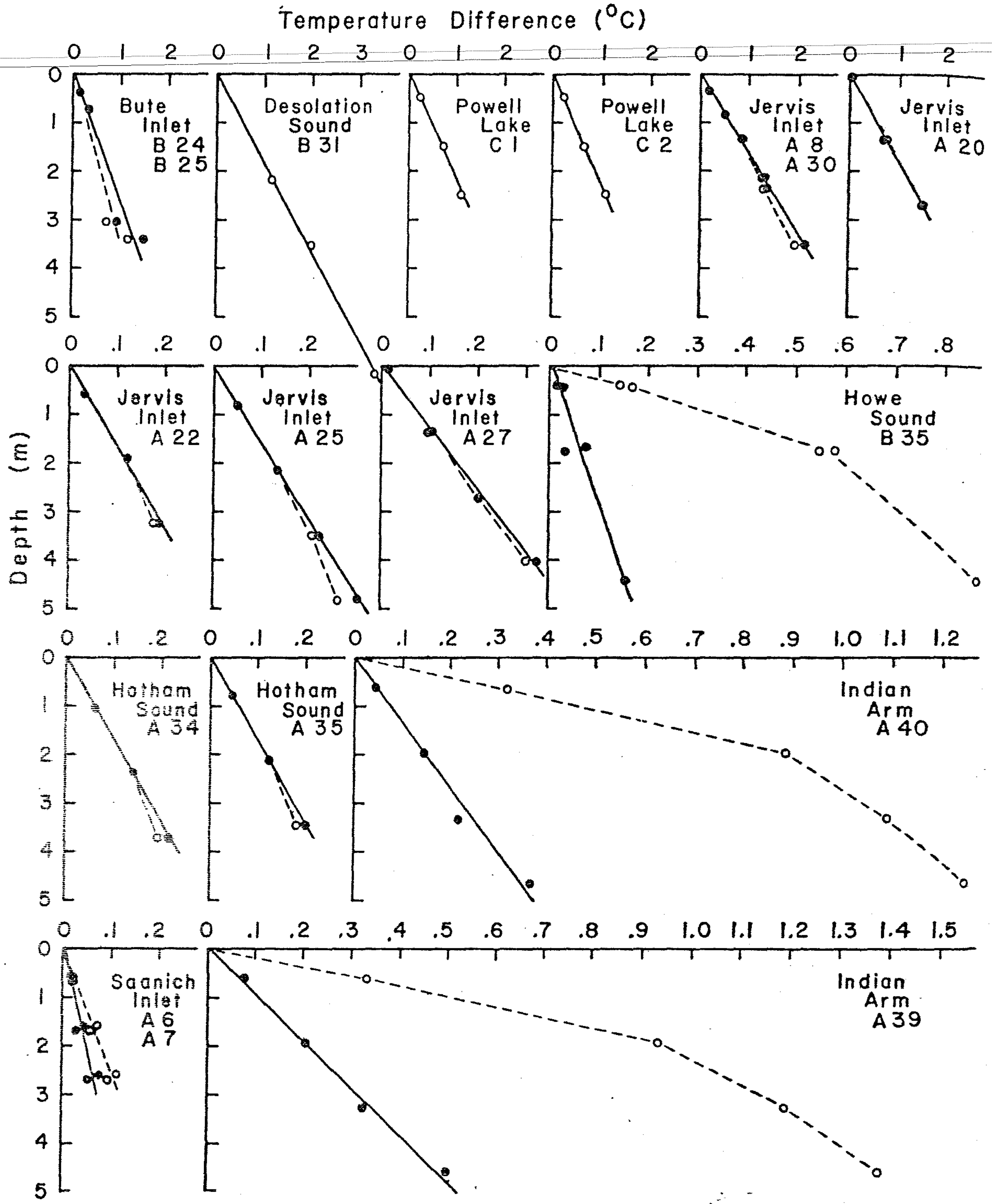

Fig. 3. The sediment temperature-depth profiles for the inlets. The open circles are values uncorrected, and the solid circles are values corrected for variations in bottom water temperature.

by extrapolating the sediment free sides of the inlets. This process is very subjective. A two-dimensional model is sufficient for all of the inlets. *Von Herzen and Uyeda* [1963] and *Lachenbruch and Marshall* [1966] have given the solution for two-dimensional hemiellipsoidal bodies which has been used for the corrections. Numerical solutions could be used to obtain higher accuracy, but the cross sections and station locations are not well enough known to warrant this approach. *Lachenbruch and Marshall* [1966] and *Rankin and Hyndman* [1971] have shown that in the central part of the sediment prism the results are similar for the two approaches. The correction is positive and has a magnitude of from 15 to 65% and tends to balance the negative effects of the topography and the warm rim.

Topography and warm-rim effect. The narrow, deep inlet valleys necessitate large corrections for topography. Heat flows away from topographic highs toward the valley bottoms which involve a shorter thermal path. The mean temperature of the land also is different from that of the inlet bottoms. The land is usually warmer causing heat to flow preferentially in the inlets, and thus the designation 'warm-rim effect' is given. The topography has developed very recently, so that the effect of lack of thermal equilibrium also should be considered.

There are two main approaches to the topographic correction [*Jaeger*, 1965]: (1) approximate solutions to (potentially) exact representations of the topography and (2) exact solutions to analytic approximations of the topography. In the first technique the topography is represented numerically to whatever accuracy is desired or to which information is available. Usually, contour maps are partitioned by a grid of polar coordinates and the effect of each element computed [*Jeffreys*, 1940; *Bullard*, 1940]. *Birch* [1950] showed how the evolution of the topography could be included in the correction. Three-dimensional topography can be treated by this method. The accuracy may be poor for very steep slopes [e.g., *Lachenbruch*, 1969]. In the second technique the topography must be approximated by two-dimensional profiles for which analytical solutions to the thermal field are available, for example, Lees type hills [*Jaeger and Sass*, 1963] or inclined planes [*Lachenbruch*, 1969], or else Laplace's equation may be solved numerically for a two-dimensional profile. The approach of *Jeffreys* [1940] has been used in this study because it can be used on all types of three-dimensional topography and because it can readily incorporate the warm-rim effect.

Topographic contour maps of scale 1:50,000 were available for the areas of most of the inlets and 1:250,000 for the remainder. The inlet bathymetries have been included in the polar grid representations of the topography, using the hydrographic charts and a few of our own and others' sounding profiles. The mean annual bottom water temperatures were estimated from the physical oceanographic data for the segments covering the inlets so that the correction includes the warm-rim effect [e.g., *Johnson and Likens*, 1967]. The ground surface temperatures are from the mean annual air temperatures from the nearest weather stations [*Government of Canada*, 1970] (Figure 4) on the assumptions that the extrapolated ground surface temperatures are 2.0°C higher and that the mean air temperature decreases by 10°C/km of elevation. The latter is the atmospheric lapse rate [*Berry et al.*, 1945] and is substantiated by local weather station data (Figure 4). The uncertainties in the inlet bathymetry for the segments nearest the measurement stations and in the mean annual land temperatures are the major sources of error in this correction. Since the stations are all at the bottoms of valleys, the corrections will be overestimated. Fortunately, the bottoms of the inlets are quite flat; so the inherent error for this method applied to steep topography is not too serious for these stations. Some corrected heat flows may be up to 0.10 μcal cm^{-2} s (4 mW m^{-2}) too low from this source. The total corrections range from 15 to 50%, about half arising from the topography and half from the warm-rim effect.

Pleistocene thermal history. The temperature at the bottom of the inlets will have been significantly lower than it is at present during the periods of ice advances during the Pleistocene. The temperature during the advances will have been between about −0.50°C, the pressure melting point for 700 m of ice [*Chow*, 1964] when the glacial ice was in contact with the floors of the inlets, and about 4°C, the temperature of maximum density of freshwater when the ice was floating. An average value of 2.0°C has been used for the periods of advance and the present average temperature of 8.0°C for the interglacial periods, giving a difference of 6.0°C. The times of advance from *Armstrong et al.* [1965] have been used: Sumas Stade 10,000–11,000 years ago; Vashon Stade 13,500–25,000; Salmon Springs glaciation 36,000 to about 50,000 (assumed). The time constant of the sediment prisms is much less than 8000 years; so the thermal parameters of the underlying rock rather than of the sediment must be used. The rock diffusivity has been taken as 0.01 cm^2 s^{-1}, and the conductivity as 6.0 mcal cm^{-1} s^{-1} °C^{-1} (2.5 W m^{-1} °C^{-1}). The correction increases the heat flow by 0.17 μcal cm^{-2} s^{-1} (7.1 W m^{-2}). The correction is 8.5 m°C m^{-1} for an average sediment conductivity of 2.0 mcal cm^{-1} s^{-1} °C^{-1} (0.84 W m^{-1} °C^{-1}). This is about 15% of the measured values.

Uplift and erosion. The area of the inlets has undergone large uplift and erosion during the past few million years which have the effect of increasing the surface heat flux. In addition to the regional uplift and erosion, there has been glacial scouring to produce or at least deepen the inlets themselves. The uplift and erosion rates are poorly known; so the magnitude of the effect has been estimated for the whole region, and the same correction has been applied to all of the measured heat fluxes.

Jaeger [1965] has outlined the theory for the gradient disturbance arising from uplift and erosion. The effect is equivalent to that for material from below moving upward toward the surface at a speed equal to the erosion rate. Changes in elevation of the surface, for example, if erosion does not keep up with uplift, can be treated as for varying surface temperature, the surface temperature being determined from the atmospheric lapse rate. *Birch* [1950] showed how the effect of uplift and erosion can be incorporated in the *Jeffreys* [1940] treatment of topography, but there is not enough information in the region of the inlets to use this method profitably. Most of the uplift of the present coast range of British Columbia of several kilometers took place in the period since the late Pliocene [e.g., *Roddick*, 1966; *Roddick et al.*, 1967; *Culbert*, 1971] producing dissection of a mid-Tertiary erosion surface. A reasonable approximation to the history is an uplift of 0.5 km per million years for 5 m.y. In the inlet valleys themselves, erosion must have approximately matched uplift. This model requires a −20% correction to the measured gradients for a rock diffusivity of 0.01 cm^2 s^{-1}. The great depths of the inlets suggest considerable glacial scouring during the Pleistocene, but as was discussed by *Culbert* [1971], much of the depth has arisen through very recent block subsidence and the valleys were established much before the Pleistocene glaciation. Thus the effect of more recent glacial scouring can probably be neglected.

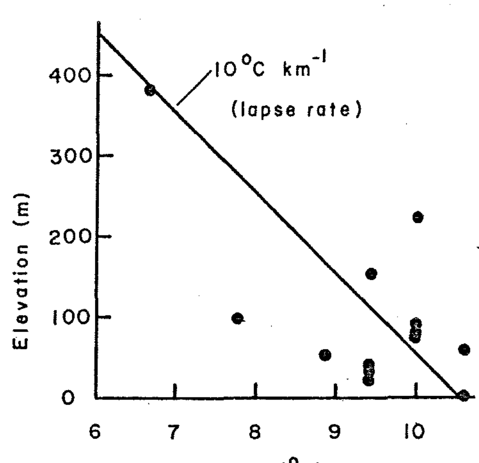

Fig. 4. The mean annual air temperature versus elevation for meteorological stations in southwestern British Columbia.

The corrections for Pleistocene thermal history and for uplift and erosion are both poorly determined and are frequently neglected in heat flow data. The computed values for the inlets are of similar magnitude and opposite sign; so omitting these two corrections will not significantly change the final heat flow value.

Individual heat flows and discussion of errors. The heat flux error limits given in Table 2 are subjective estimates based primarily on the magnitude of the corrections and on the accuracy with which they could be performed. The errors in gradient and conductivity measurement are negligible in comparison with those introduced by the corrections. About half of the error limits given are ascribed to systematic errors for all the stations and half to random errors. As indicated above, some values may be up to 0.10 μcal cm^{-2} s^{-1} (4 mW m^{-2}) systematically too high because of a bias in the topographic correction. Thus differences in heat flows between stations of greater than half the error limits (i.e., about 0.2 μcal cm^{-2} s^{-1} or 8 mW m^{-2}) probably are significant.

At the Bute Inlet station B24-B25 the most important source of error is the uncertainty in the correction for bottom water temperature variations, since there were no data for nearly a year before the measurements. The other corrections are quite small and are well determined. The thermal conductivity is unusually high. Since the sediment core could have been obtained up to 200 m from the gradient measurement, the conductivity used could be too high, a reason perhaps explaining why the heat flow at this site is somewhat higher than the average.

The heat flow at the Desolation Sound site B31 is well determined. The major uncertainty arises from the lack of bottom water temperature records in the area.

At Powell Lake the large and poorly determined correction for the refraction of heat away from the low conductivity sediment prism is the main source of error. In addition, the topographic, warm-rim correction is sufficiently large to be seriously overestimated by the *Jeffreys* [1940] technique. The computed flux may be up to 0.10 μcal cm^{-2} s^{-1} (4 mW m^{-2}) too low from this source.

The Jervis Inlet and Hotham Sound sites A27, A25, A22, A20, A30-A8, A34, and A35 all have well-determined heat flows. The lowest heat flows are at the stations with the largest topographic corrections, a suggestion that this correction generally has been slightly overestimated as discussed above.

The heat flow at the Howe Sound station B35-B36 is poorly determined because of the large correction for bottom water temperature variations with limited temperature records. The other corrections are quite small and are well-determined.

At the Indian Arm stations A39 and A40 the bottom water variations are very large, but there are very detailed records. The sediment prism refraction and the topographic, warm-rim corrections also are large but are quite well-determined. The high value of 1.46 μcal cm^{-2} s^{-1} (61 mW m^{-2}) at station A39 near the head of Indian Arm appears to be significantly above that at A40 and the mean of the other stations, but because of the larger error it is not as clearly distinct as the high value at A27 in upper Jervis Inlet.

The Saanich Inlet station A6-A7 requires a considerable correction for bottom water temperature variations which is pronounced only because of the shallow probe penetration. Data are missing for a critical period of time for this correction, and examination of the record suggests that the computed heat flow may be too low as a result. The other corrections are well-determined.

Radioactive Heat Production

The concentrations of uranium, thorium, and potassium in three quartz diorite samples from the region of central Jervis Inlet were measured using gamma spectrometry by G. K. Muecke, Dalhousie University. The results along with those for the U.S. Geological Survey standard G-2 are given in Table 3. The mean heat production is 1.25 × 10^{-13} cal cm^{-3} s^{-1} (HGU) (or 5.23 × 10^{-7} W m^{-3}). The U and Th concentrations are remarkably low for such rocks. Three samples are not sufficient for any firm conclusions, but the results suggest that the low measured heat flows arise in a large part from low crustal heat production. This conclusion is confirmed by more extensive data for which a preliminary report has been given by *Lewis* [1975]. He finds heat production values in the coast range complex ranging from 1.2 HGU (5.0 × 10^{-7} W m^{-3}) for diorites to 3.1 HGU (13 × 10^{-7} W m^{-3}) for monzonites.

Results and Conclusions

Heat flow pattern. The corrected heat flow data for this study extending from 48°30'N to 51°N have an average of 0.96 ± 0.07 μcal cm^{-2} s^{-1} (40 ± 3 mW m^{-2}) (Table 2, Figure 1). There is no significant variation northwest-southeast parallel to the continental margin, but there is an important pattern perpendicular to the coast (Figure 5). The heat flow of southern British Columbia is low from the edge of the continental margin inland for 200 km with high values further to the northeast. The two most northeasterly values of this study (both 1.51 μcal cm^{-2} s^{-1} or 63 mW m^{-2}) are on the transition zone from low to high heat flow. Excluding these values, the mean for the coast low zone is 0.88 μcal cm^{-2} s^{-1} (37 mW m^{-2}). As discussed above, large errors may be introduced by the corrections, and there is a probable bias to too low corrected heat flows. But the average must be less than 1.0 μcal cm^{-2} s^{-1} (42 mW m^{-2}). In Figure 5 the pattern is emphasized by the addition of values from *Blackwell* [1971] in Washington State to the south and one value from *Jessop and Judge* [1971]

TABLE 3. Radioactive Heat Production

Sample	Potassium, %	Uranium, ppm	Thorium, ppm	Heat Production,* 10^{-13} cal cm^{-3} s^{-1}
WC-G1	3.22	0.44	1.60	1.29
WC-G2	2.10	0.81	1.52	1.24
WC-G3	2.14	0.75	1.60	1.23
Average	2.49	0.67	1.57	1.25
Standard G-2	3.77	2.20	25.8	
Recommended	3.74	1.99	25.2	

The source for the standard and recommended values was *Flanagan* [1969].
*Or 4.19 × 10^{-7} W m^{-3}.

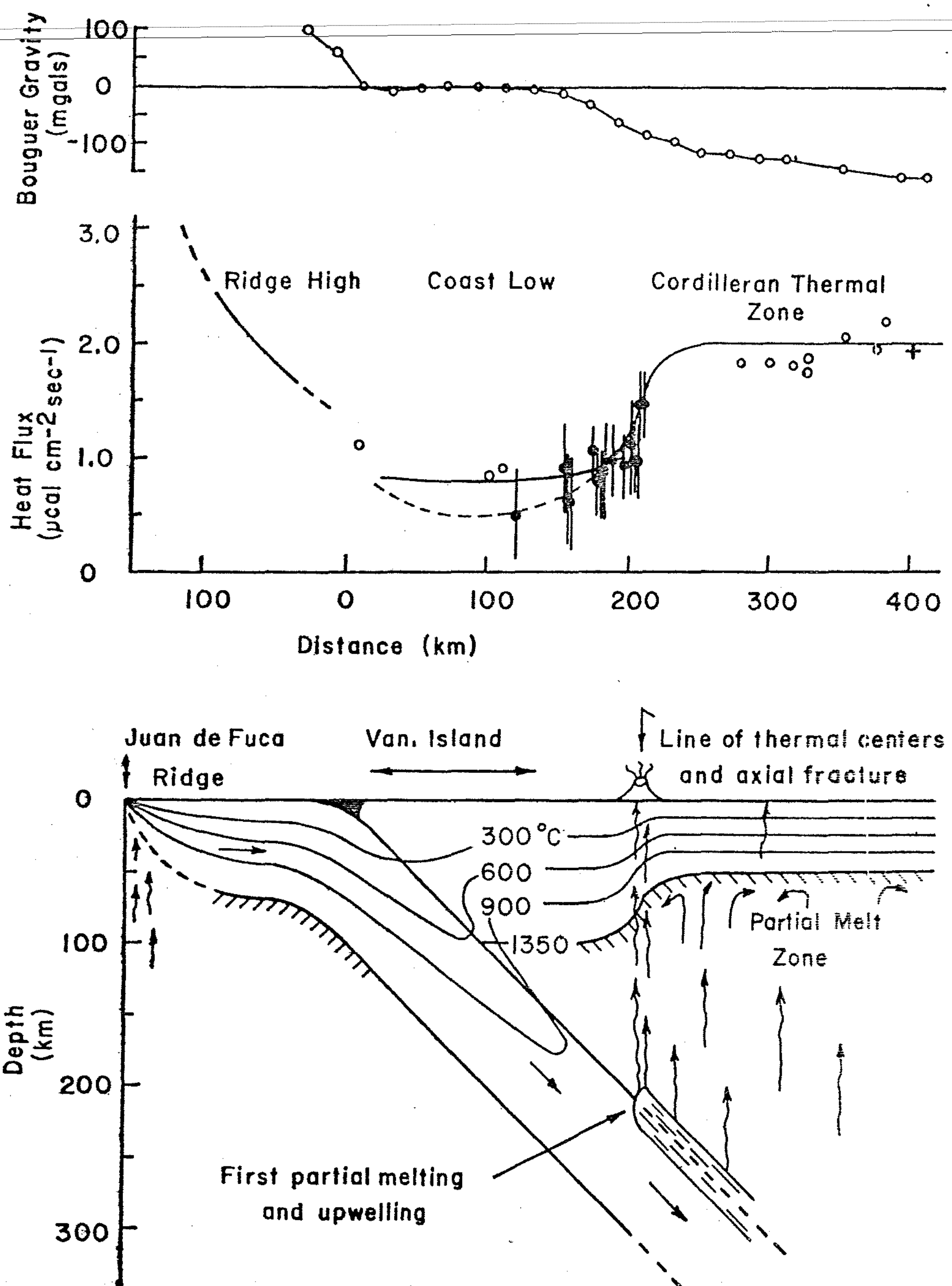

Fig. 5. Inlet heat flow values plotted as a function of distance from the edge of the continental shelf. The open circles are from *Blackwell* [1971] in Washington State to the south and the plus signs from *Jessop and Judge* [1971] in south central British Columbia. Smoothed Bouguer gravity from *Stacey* [1973] and a schematic cross section with illustrative temperatures beneath the area also are shown.

south central British Columbia. Other preliminary values om boreholes in southern British Columbia confirm the low at flow coast zone and the high heat flow of the central and stern Cordillera [*Judge,* 1975].

The transition from low to high heat flow appears to occur a line parallel to the coast running through the head of most the major inlets (Figure 1). *Culbert* [1971] has analyzed the pography of the coast mountains statistically and found that is line marks the transition to much greater uplift inland, nce the termination of the inlets or fjords. He termed the line e axial fracture and pointed out the coincidence of this eleva- on discontinuity with a line of Pleistocene and Recent volcanic centers and hot springs [see also *Bostrom,* 1971]. Pleistocene and Recent volcanic activity is common to the east of this line, but there is virtually none to the west. There also is a pronounced gravity transition at this line which is discussed below. The transition in heat flow occurs over a distance of less than 100 km; so the origin of the contrast must occur at a depth of less than about 100 km. The eastern boundary of the high heat flow zone probably is in the region of the Rocky Mountain trench [e.g., *Berry et al.,* 1971]. The gravity data also suggest this boundary.

Heat flow and heat production. The very limited radioactive heat production data of this study and more extensive preliminary data of *Lewis* [1975] along with the data for Washington of *Blackwell* [1971] suggest that the crustal contribution to the observed heat flow in the coastal low zone is very small. Blackwell computes a reduced heat flow (approximately the heat flux from the mantle) of 0.8 μcal cm^{-2} s^{-1} (33 mW m^{-2}) which is similar to the average heat flow from the inlets. It also is comparable with the reduced heat flow for old stable continental areas [*Roy et al.*, 1968]. The low heat flows of 0.5–0.6 μcal cm^{-2} s^{-1} (21–27 mW m^{-2}) at Powell Lake and Saanich Inlet are poorly determined but may represent the central part of the low zone. A reduced heat flow of about 0.5 is then required, which is similar to that for the Sierra Nevada Mountains [*Lachenbruch*, 1968]. The low reduced heat flows require a heat sink arising from the cold subducted lithosphere.

The heat flow and radioactive production data for the high heat flow region of southern British Columbia [*Judge*, 1975] and Washington State [*Blackwell*, 1971] indicate a reduced or mantle heat flow of about 1.4 μcal cm^{-2} s^{-1} (59 mW m^{-2}), comparable with that for the Basin and Range Province. This high value requires a major heat source in addition to those present under the stable continental regions, which probably comes from frictional heating, partial melting, and upwelling from the subducted slab. The normal lithospheric plate thickness of about 100 km must be thinned to about 50 km (Figure 5) [e.g., *Roy et al.*, 1972; *Sclater and Francheteau*, 1970]. The heat flow–heat production relationship for the study area is compared with the relations for the three main heat flow provinces defined for North America by *Roy et al.* [1968] in Figure 6.

Correlation of heat flow with other geophysical data. A smoothed average Bouguer gravity profile across southern British Columbia between 49° and 51°N from *Stacey* [1973] is shown in Figure 5. The high Bouguer gravity offshore arises from the Bouguer correction over oceanic areas and is not pertinent to this study. Inland for about 200 km from the edge of the continental shelf (western zone) the Bouguer gravity is close to zero. Further inland there is a rapid decrease to about −120 mGal over the central Cordillera (eastern zone). The

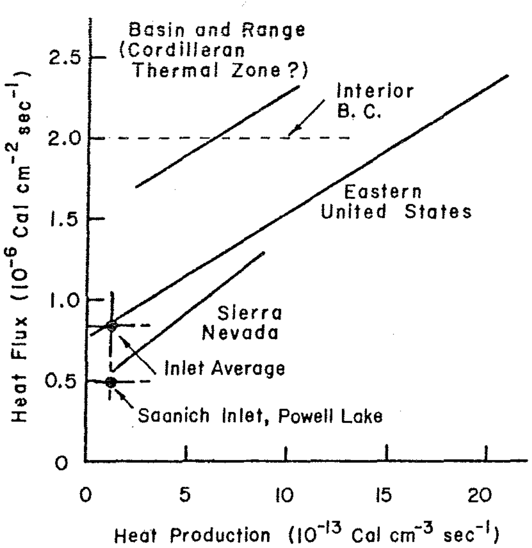

Fig. 6. Heat flow–heat production relations for North America, showing the position of the new data. Note that the coast low reduced heat flow is comparable to old stable continental areas. The very low Saanich Inlet and Powell Lake values require a heat sink, but these values have very low accuracy. The inland data require a heat source with respect to the stable continent.

transition in gravity coincides with the proposed thermal boundary. Thus there is a clear difference in gravity between the high and low heat flow areas. There are many density models that will fit the gravity data, particularly with the complication of a cold high density slab that may exist under the area. But as *Stacey* [1973] has shown, the seismic as well as gravity restraints can easily be satisfied only by having low mantle density under the eastern zone. The lower densities undoubtedly arise from higher mantle and crustal temperatures. If this correlation between gravity and heat flow is accepted, the gravity data show that the low heat flow is indeed a band parallel to the coast.

The crustal thickness under south central British Columbia is about 30 km [*White et al.*, 1968], while it may exceed 50 km under Vancouver Island [*White and Savage*, 1965; *Tseng*, 1968]. The difference is emphasized by the higher mean elevation in the eastern zone. Such a difference can be maintained only by much lower mantle (or possibly crustal) densities under the eastern zone (unless there are very large deviations from isostatic equilibrium maintained by the dynamic forces of subduction). High temperature under the eastern zone explains part of the difference. The formation of the high-temperature zone will have thinned the crust by uplift and erosion. The process is seen in the extensive deep granitic basement exposed in the coast crystalline belt (eastern zone) compared with that in the Vancouver Island or insular volcanic belt.

A difference in mean temperature of 300°C to a depth of 100 km (e.g., Figure 5) will result in a gravity difference of about 200 mGal, approximately that observed. The temperature difference also will give about the correct difference in mean elevation. However, the suggested difference in crustal thickness (very thick under Vancouver Island) should result in a similar gravity difference of opposite sign. Thus in southern British Columbia, additional anomalously high density material must exist under the western zone in the cold, dense subducted oceanic lithosphere. Models for subduction beneath South America by *Grow and Bowin* [1975] indicate up to 160 mGal positive gravity contribution from the cold, subducted oceanic lithosphere in the first 200 km inland from the trench falling to half, 300 km inland. The density anomaly may be less in the more slowly sinking lithosphere beneath British Columbia, but it still must be significant.

In distinct contrast to the crust under Vancouver Island the crust is only about 30 km thick to the north under the coast insular belt of central and northern British Columbia [*Forsyth et al.*, 1974; *Johnson et al.*, 1972]. With such a thin crust the coast zone of high gravity in this region can be produced solely by a cold, thick (100 km) high-density continental lithosphere without the presence of a high-density subducted oceanic lithosphere. A hot thin continental lithosphere still is required under the eastern low-gravity zone. Probably north of about 51°N, where there is transform motion along the coast at present, subduction terminated several tens of millions of years ago. The density anomaly in the subducted oceanic lithosphere has disappeared, but upwelling and a thin high temperature lithosphere persist under the inland zone. *Monger et al.* [1972] suggest that such subduction has taken place.

Earthquakes in southern British Columbia are largely limited to the low heat flow zone along the coast [e.g., *Milne et al.*, 1970]. Low crustal temperatures possibly facilitate brittle fracture. In the high-temperature eastern zone, plastic deformation will occur at very shallow depths. There also may be a concentration of earthquakes in the region of the heat flow transition where differential vertical motions are to be expected.

High upper mantle temperatures and a thin lithosphere under the eastern zone are substantiated by seismic studies [e.g., Berry and Forsyth, 1975; Stevens, 1971; low Pn velocities, *White et al.*, 1968], by high electrical conductivity in the depth range of 30–70 km [e.g., Caner, 1970; *Cochrane and Hyndman*, 1970], and by glacial rebound studies [*Fulton and Walcott*, 1975].

Extensions of the observed heat flow pattern. The pattern of a 200-km-wide band of low heat flow extending inland of the continental edge continues to the south into Washington State as shown by several borehole measurements [*Blackwell*, 1969; *Blackwell*, 1971; *Roy et al.*, 1972]. It probably extends south at least to the triple point marking the southern limit of the Juan de Fuca plate off northern California. Further to the south, the low heat flow of the Sierra Mountains and the Peninsular Ranges near the coast of southern California [e.g., *Roy et al.*, 1972] also may represent the extension of this low zone caused by the thermal sink of the subducted oceanic plate. However, in those regions the subducted slab recently has been cut off; so the low heat flow remains only because of the long thermal time constant of the lithosphere.

As discussed by *Berry et al.* [1971], the high heat flow inland in southern British Columbia is the northern extension of the Cordilleran thermal zone of *Blackwell* [1969], which includes the Columbia Plateau, the northern Rocky Mountains of the United States, and the Basin and Range geological province.

The termination of present or very recent subduction to the north of about 51°N [e.g., *Stacey*, 1974] suggests that the pattern of coast low heat flow and inland high heat flow should no longer be present. But as discussed above, the pattern probably persists from earlier subduction. The gravity and crustal thickness in the coastal zone can be satisfied without the cold subducted oceanic slab that is required to the south of 51°N. However, the crustal thickness and low Bouguer gravity inland of 200 km from the continental edge in the northern region [*Government of Canada*, 1974] (Figure 7) requires a thin continental lithosphere and a high heat flow comparable with that found in southern British Columbia.

Origin and development of the heat flow pattern. The low heat flow for 200 km inland of the British Columbia continental margin and the high heat flow inland probably occur in response to present (or recent) subduction of the oceanic Juan de Fuca plate under the continent. Island arc subduction zones such as bound the western Pacific Ocean have been studied in some detail, and the thermal structure in those areas should not differ greatly from that under a continental subduction zone. They are characterized by a band of low heat flow just on the continental side of the trench and very high heat flow inland [e.g., *Vacquier et al.*, 1967; *McKenzie and Sclater*, 1968; *Hasebe et al.*, 1970]. The low zone can be produced readily by the cold sinking slab acting as a heat sink absorbing heat that otherwise would reach the surface. Endothermic dehydration chemical reactions also could be important in the upper part of the sinking slab [*Uyeda and Horai*, 1964]. In fact, a reduced heat flow of 0.8 μcal cm^{-2} s^{-1} (33 mW m^{-2}) can be produced by a normal, stable 100-km-thick lithosphere. *Minear and Toksöz* [1970] have computed the surface heat flow distribution for a number of theoretical subduction models and found that for simple conduction only, the low zone can be reproduced (although somewhat too narrow) but not the inland high zone. *Hasebe et al.* [1970] [see also *Blackwell*, 1971; *Oxburgh and Turcotte*, 1970] show that the inland high heat flow requires first a mechanism of deep heat generation which is probably viscous shear heating. Such heating must be negligible above about 100-km depth, which is reasonable. At shallow depth, shearing should be along a narrow zone perhaps lubricated by hydrated rocks such as serpentinite. Second, a mechanism for the upward transport of the heat is required. They suggest that at about the depth where viscous heating becomes important, the temperatures become sufficient for partial melting to commence in the sinking slab. Heat then is carried upward convectively by magma. This process is in agreement with the presence of a sharp boundary on the trench side of the volcanic arc. The lithosphere over the upwelling thins until most of the increased heat flux can be transported through it conductively to the surface without the plate temperature exceeding the solidus (about 50 km for a mantle heat flow of 1.4 μcal cm^{-2} s^{-1} or 59 mW m^{-2}). Most of the upwelling magma from the sinking slab then reaches only the base of the lithosphere. Surface volcanism occurs by some penetration to the surface along fractures and lines of weakness. A simple schematic diagram of the thermal structure under southern British Columbia is shown in Figure 5.

The volcanic arc in southwestern British Columbia is not well-defined because of the slow sinking rate or recent termination of sinking and because the lithosphere is very thin such that volcanos will not build up to great height [*Vogt*, 1974]. The line of large volcanos to the south in the United States probably reflects a thicker lithosphere.

Dickinson [1973] has shown that the location of the volcanic arc (or first volcanism) moves inland with time at a rate of about 1 km per million years probably because the sinking slab gradually cools the surrounding asthenosphere, so that first melting occurs at greater and greater depths. Thus the coast

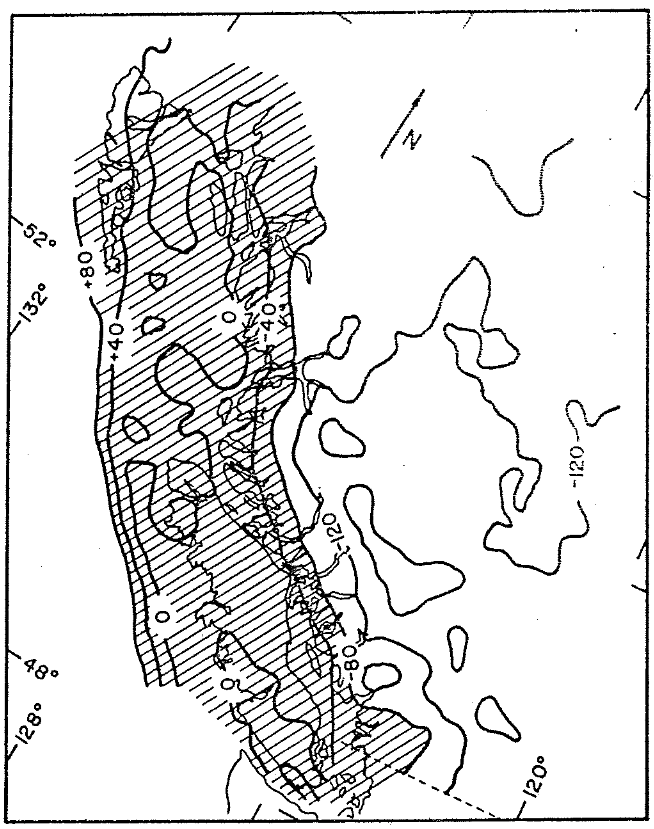

Fig. 7. Simplified Bouguer gravity map of southwestern British Columbia [after *Government of Canada*, 1974] showing band of high gravity (hatched lines between −80 and +80 mGal) that probably is associated with low heat flow.

low heat flow zone is slowly widening to the east. The thermal contraction of the lithosphere resulting from the declining temperatures explains the recent subsidence and development of the mainland inlet or fjord zone. The same process probably produced the fjords of the west coast of South America. The present volcanic arc of southwestern British Columbia is probably represented by the axial fracture line and line of recent volcanic and hot spring activity of *Culbert* [1971] (Figure 1) which is in agreement with the gravity data.

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