On the Thermal Balance of a Mid-Ocean Ridge

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(Received 1971 June 18)

Summary

The northernmost expressions of the East Pacific Rise spreading centre, the Juan de Fuca Ridge and Explorer Trough, have unusual heat-flow distributions. Crestal values are only moderately high, but flank areas that are models of undisturbed sedimentation have mean heat flows near 7 HFU (290 mW m⁻²). The values in these areas retain considerable scatter, but the detailed distribution is close to the opposite of that expected from conductive heat-flow refraction. The high mean values confirm that seafloor spreading occurs by the upwelling of hot (magma temperature) material. The low crestal measurements and the non-refractive scatter in the flank areas strongly suggest that hydrothermal circulation is the dominant heat transfer process in fresh crestal material. Much of the heat loss from the new lithospheric material must occur through open circulation near the ridge crest and is never reflected in measurements of surface conduction.

1. Introduction

The investigation of the heat-flow anomaly at mid-ocean ridges has been an important aim of marine heat-flow measurement since the earliest studies were made (Bullard & Day 1961). A substantial body of data is now available, and has been summarized by several authors (Lee & Uyeda 1965; Langseth, Le Pichon & Ewing 1966; Le Pichon & Langseth 1969). Although the mean heat flow near ridge crests is usually higher than the global average, it is not dramatically so, and the scatter in the measurements is such as to compromise the validity of these means, both statistically and physically. The overall results are not in agreement with the generation of new oceanic crust at the ridges from material at a reasonable temperature (McKenzie 1967), and the recent reinforcement of the sea-floor spreading hypothesis by evidence from other fields (Vine & Matthews 1963; Peterson et al. 1970) requires that the disagreement between the heat flow and the other data be resolved.

It has been pointed out before (Lister 1970a; Sclater et al. 1970) that some environmental control is essential for any measurement whose scatter is sufficient to arouse suspicion of physical cause for the variation. In this study, standard surveying techniques were carefully applied to the area of interest to obtain enough topographic data at least for the topological situation of each heat-flow measurement to be estimated. The influence of thermal refraction cannot be separated from other causes of variation unless the sign and approximate magnitude of the expected correction are known individually for each station. This information has been obtained for almost all of the measurements quoted here, together with data that set an upper limit on the possible magnitude of recent surface disturbances. This last is important because of the suggestion by Sclater, Mudie & Harrison (1970) that the low heat-flow
values often measured in valleys are caused by recent slumps of sediment from nearby
hills or ridges. The assumption that the heat flow is governed solely by conduction
in the crustal materials can therefore be tested directly without making alternative
assumptions about the thermal regime at depth. Only when this test has been made can
the overall thermal balance of a ridge crest be discussed.

The region of the North-east Pacific off the coast of North America is dominated
tectonically by two long active transform faults and the most northerly expressions
of the East Pacific Rise spreading axis. The San Andreas Fault, running from the
Gulf of California to Cape Mendocino, is a right-lateral transform fault joining the
equatorial East Pacific Rise to the ridge segments off the coasts of Oregon and
Washington (see Fig. 1). The magnetic survey of Raff & Mason (1961) suggests that
there are three segments of active spreading: the Gorda Ridge, the Juan de Fuca
Ridge, and Explorer Trough (Vine 1966). They are joined by the Blanco Fracture
Zone, and an ill-defined fault zone between the Juan de Fuca Ridge and Explorer
Trough. The system ends in another long active transform fault, the Queen Charlotte
Fault, and this extends toward the Alaskan end of the Aleutian Trench, completing
the source-to-sink pattern (Atwater 1970).

Of the three spreading segments, the Juan de Fuca Ridge is the longest and the
best defined physiographically. On the western side it has all the characteristics of a
mid-ocean ridge: mountainous topography rising gradually to a crest at the centre
of magnetic symmetry. The eastern side, however, appears dramatically different
on the map, because the ridge has been inundated by the sediment fill of Cascadia
Basin. Heavy continental sediment runoff through the Fraser and Columbia River
systems has been ponded near the coast by the Juan de Fuca Ridge and by high
topography on the Blanco Fracture Zone. Thus the Juan de Fuca Ridge is a unique
example of a classic 'mid-ocean' ridge which has been inundated by terrigenous
sediments almost to the crest on one side, but is sufficiently far from the continental
margin to be structurally integral.

Freedom from structural interference is not a characteristic of Explorer Trough,
however, for it is a small feature amid complex topography and structure just off
the continental rise, north-west of Vancouver Island. The magnetic anomalies
associated with it are poorly defined (Vine 1966), and interpretation is further compl-
icated by the fact that it is a trough not a ridge. However, the lack of sediments in the
topographic lows, relative to the surrounding ocean floor, confirms the magnetic
and now thermal evidence that rifting of the crust has taken place recently (see Fig. 9).
The feature is thus interesting as a spreading centre that has not, as yet, developed
into a full-scale ridge, and will be useful for structural comparisons in the future.

This paper presents field results from two studies: one on the Juan de Fuca Ridge
and one in the Explorer Trough region. The principal investigation is a transect
across the Juan de Fuca Ridge at 47° N, and is shown on the location map (Fig. 1)
as rectangle ' G '. Data for this area have been accumulated over several years from
short cruises of RV Thomas G. Thompson, and include a substantial body of bathy-
metric surveying. The smaller study is within rectangle ' H ', and consists of a heat-
flow profile across the north-eastern extremity of Explorer Trough. This was completed
on Phase VII of Hudson 70, a circumnavigation of the Americas by C.S.S. Hudson.
No detailed surveying was carried out in this area, but the topographic features are
large enough in scale to be delineated adequately by the bathymetric data of the
Pioneer survey (cf. Fig. 1, Raff & Mason 1961).

2. Experimental techniques

Heat-flow measurements were made by examples of both generic types of marine
apparatus. The corer-outrigger instrument has been described in Lister (1970a), and
consists essentially of a 450 × 8 cm gravity corer equipped with a digital bridge
On the thermal balance of a mid-ocean ridge

517

On the thermal balance of a mid-ocean ridge

reducer and three thermistor outriggers. Sediment conductivities were estimated by oven-drying samples from the cores and using the water content versus conductivity nomograph of Ratcliffe (1960). The probe instrument uses the same digital recorder but a sensor arrangement modified to measure thermal conductivity in situ over the whole length of the probe. Geothermal gradient and conductivity are measured sequentially in 30 min on the ocean floor, and the probe dimensions have been made small to permit this (200 x 1 cm). The instrument and the interpretation techniques have been fully described in Lister (1970b). The corrections for initial-disturbance decay and the conductivity values were obtained by graphical methods from a large-scale analogue printout of the digital data. The technique has the advantage of permitting a selection of the data points to be computed, avoiding any disturbances experienced by the instrument and the occasional digitisation errors.

The cores obtained by the corer-outrigger apparatus were split and examined stratigraphically in addition to being used for the water content/ conductivity determinations. Total carbonate assays were made on selected parts of some to assist in distinguishing biogenic from terrigenic material, since the carbonate is exclusively of biogenic origin. The carbonate fraction of the cores was also used to obtain radiometric ages for samples from the uppermost 70 cm by measurement of the carbon-14 decay activity in derived methane (A. W. Fairhall*). Three normal cores were dated to establish the sedimentation rate and pattern, and the technique was then used to examine two disturbed cores that could have included slump or slump-exposed material.

The general distribution of sediment on the ocean floor in the areas was examined by acoustic reflection sounding. The systems used on the Juan da Fuca Ridge transect emitted short pulses of medium-frequency (150-300 Hz) sound that have a favourable penetration/resolution ratio in deep-ocean sediments. The early profiles were made with several models of a repulsion-type moving coil electromagnetic transducer energized with 11000 J, but the survey east of the ridge crest was made with an attraction-type electromagnetic transducer pulsed by 250 J. The same, somewhat noisy, 15 m line hydrophone was used throughout, and the signal was processed by automatic gain controlled amplifiers equipped with 12 db/octave variable filters. Full-wave variable density recording was used, but the attraction-transducer records were enhanced by a delay-line pulse shape correlator manually set to match the outgoing pulse. The ultimate resolution of both systems was about 10 ms (7.5 m) but the interpretability of the correlated records is better because of the symmetrical printed pulse free of multiples. Except in the flat-lying sediment fill of Cascadia Basion, resolution is limited by the roughness of the bottom rather than by the acoustic system.

The sediment distribution in the Explorer Trough region was estimated by reference to 40 cu in airgun profiles provided by R. L. Chase (University of British Columbia). In particular, the principal heat-flow profile was arranged along airgun profile UBC 70-16-14, that crosses the troughs about 15 km south-west of Paul Revere Ridge. Resolution on this profile was about 60 ms (45 m), as can be judged from Fig. 9.

A bathymetric background is necessary to identify the lateral extent of features observed on transect profiles. In the Explorer Trough region, it was obtained from a map contoured from previous survey data at 9 km line spacing (Mammerick & Taylor 1971). The features in this area appear to have diametral dimensions of about 15 km (Fig. 8), and are therefore adequately delineated. No significant discrepancies were observed between the contour map and echo-soundings on the cruise.

The topography on the western flank of the Juan de Fuca Ridge is, however, far more complex and of finer scale. No adequate map of the area was available at the time the investigation was started, and even the 9-km line spacing of the Pioneer Survey (Mammerick & Taylor 1971) cannot resolve the topographic structure. A bathymetric

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survey was therefore conducted as part of the investigation, to identify topographic
trends near the transect line. Navigation was by Loran-A radiolocation which is good
to ±2 km over most of the area, deteriorating steadily with increasing distance from
the coast. Line crossings were used to improve the topological accuracy of the
contour map by ensuring that track lines were plotted at the correct relative positions.
A standard 12 kHz, 50° beam, echo-sounder gave the depth information, presented
on the 1-s sweep of a precision dry-paper recorder. Two-way travel times were used
in compiling the maps, to emphasize that sound velocity corrections were not made.

The concentration of stations near 47°N 132°W is in an ultra-detailed survey area,
where navigation was assisted by taut-moored radar reflector buoys. Only the outlines
of the detailed map have been transferred onto the figure used in this paper (Fig. 3).
The survey was intended to establish the resolution limits and characteristic response
of standard echo-sounding systems. The information has been valuable in interpreting
the more widely-spaced lines employed in later stages of the study.

Heat-flow stations can be affected by small-scale local topography as well as by the
large-scale features observable on surface echo-sounding records (Sclater et al. 1970).
One method of obtaining information about the local irregularities to be expected
in an area is photography from a camera drifting near the bottom. The method is
severely limited by the loss of visual acuity in bottom water at ranges of more than
10 m, but can produce valuable information about local slope angles, sediment
distribution, and bottom currents. An Edgerton stereo camera cycling every 16 s
was drifted at 5-8 m above bottom in several places on the western flank of the Juan
de Fuca Ridge. On some stations a narrow-band green interference filter was used
to enhance the contrast of distant objects. On the Explorer Trough investigation a
different technique was tried for the first time, and gave valuable information about
the nature of the floor of the main trough. A slow frame-rate cine camera, synchronized
to a pulsed Thallium Iodide arc light, was substituted for the stereo camera. The
film gives almost continuous coverage of a considerable (c. 150 m) section of ocean
floor.

3. Transect of the Juan de Fuca Ridge

A consolidated track chart of the investigations is presented in Fig. 2 to indicate
the extent of bottom coverage. The background for the heat-flow results in Fig. 3 is
bathymetry derived from these tracks and from the more general data available in
earlier studies. The heat-flow measurements are listed by station in Table 1, except for
one measurement from Vema 20 (Paul J. Grim, private communication), whose
relationship to the topography on the map is not known with precision. The first
fifteen entries in Table 1 have been listed in more detail in Lister (1970a), but the
remainder are new. Gradient linearity and the quality of the conductivity measure-
ments were taken into account in establishing the error bands for the heat-flow
measurements.

The distribution of long gravity core samples of the sediment is shown in Fig. 2. There
are wide variations in the appearance and layering of the sediments because both biogenic and
terrigenous materials are deposited on the Juan de Fuca Ridge, sometimes in almost equal proportion. However, most cores exhibit a characteristic
dark surface layer that probably represents postglacial deposition, since carbon-14
radiometric ages below it are greater than 15,000 years (A. W. Fairhall, private
communication). Two important disturbed cores (measurements TT31-3 and TT31-9,
Table 1) were dated radiometrically at more than one depth to make sure that recent
slumping can be ruled out. One, discussed at length in Lister (1970a), is normal
except for the loss of the top section on board ship. The other has older sediment
in the top 20 cm than deeper within it, and dares confirm that a buried dark layer a
25 cm corresponds to normal surface material. It is probable that the corer caused
On the thermal balance of a mid-ocean ridge

FIG. 2. Consolidated track chart of the Juan de Fuca Ridge investigations (not including the detailed survey area near 47° N 132° W). Core samples are shown by filled circles; superscript B means box core of surface material only.

FIG. 3. Bathymetry and heat-flow data across the Juan de Fuca Ridge (axis shown). Map corresponds to area G in Fig. 1. Contour interval is in milliseconds of two-way travel time (× 0.4 to convert to uncorrected fathoms). Heat-flow values are in microcalories per square centimetre per second (HFU).
Table 1

Heat-flow measurements: Juan de Fuca Transect

<table>
<thead>
<tr>
<th>Station</th>
<th>Latitude (N)</th>
<th>Longitude (W)</th>
<th>Depth (sec, 2-way)</th>
<th>Conductivity ( \times 10^{-3} ) cal cm (^{-2})</th>
<th>Heat flow ( \times 10^{-6} ) cal cm (^{-2}) s (^{-1})</th>
<th>Error band (±)</th>
<th>Heat flow (mW m (^{-2}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>TT3-1</td>
<td>46° 59.5'</td>
<td>131° 48.8'</td>
<td>4.24</td>
<td>2.39</td>
<td>&gt;0.8</td>
<td>0.06</td>
<td>25</td>
</tr>
<tr>
<td>TT3-2</td>
<td>46° 58.0'</td>
<td>131° 57.3'</td>
<td>4.32</td>
<td>1.9</td>
<td>0.8</td>
<td>0.01</td>
<td>10</td>
</tr>
<tr>
<td>TT3-3</td>
<td>46° 58.5'</td>
<td>131° 57.0'</td>
<td>4.89</td>
<td>1.89</td>
<td>0.9</td>
<td>0.06</td>
<td>29</td>
</tr>
<tr>
<td>TT3-4</td>
<td>47° 05.0'</td>
<td>131° 53.5'</td>
<td>4.31</td>
<td>1.93</td>
<td>0.7</td>
<td>0.06</td>
<td>8</td>
</tr>
<tr>
<td>TT3-5</td>
<td>47° 08.4'</td>
<td>132° 19.6'</td>
<td>4.34</td>
<td>2.08</td>
<td>0.2</td>
<td>0.06</td>
<td>8</td>
</tr>
<tr>
<td>TT3-6</td>
<td>47° 09.0'</td>
<td>131° 12.0'</td>
<td>4.09</td>
<td>1.86</td>
<td>0.9</td>
<td>0.04</td>
<td>38</td>
</tr>
<tr>
<td>TT3-7</td>
<td>47° 01.0'</td>
<td>131° 30.0'</td>
<td>3.96</td>
<td>1.96</td>
<td>0.6</td>
<td>0.03</td>
<td>25</td>
</tr>
<tr>
<td>TT3-8</td>
<td>47° 05.5'</td>
<td>129° 21.0'</td>
<td>3.40</td>
<td>2.04</td>
<td>1.4</td>
<td>0.10</td>
<td>168</td>
</tr>
<tr>
<td>TT3-9</td>
<td>47° 05.5'</td>
<td>129° 27.0'</td>
<td>3.61</td>
<td>2.14</td>
<td>2.1</td>
<td>0.10</td>
<td>88</td>
</tr>
<tr>
<td>TT3-10</td>
<td>47° 05.0'</td>
<td>129° 46.0'</td>
<td>3.78</td>
<td>2.09</td>
<td>0.86</td>
<td>0.04</td>
<td>36</td>
</tr>
<tr>
<td>TT3-11</td>
<td>47° 00.0'</td>
<td>130° 00.0'</td>
<td>3.61</td>
<td>1.89</td>
<td>10.9</td>
<td>0.20</td>
<td>456</td>
</tr>
<tr>
<td>TT3-12</td>
<td>47° 10.0'</td>
<td>131° 44.0'</td>
<td>4.42</td>
<td>1.87</td>
<td>0.5</td>
<td>0.03</td>
<td>21</td>
</tr>
<tr>
<td>TT3-13</td>
<td>47° 05.0'</td>
<td>131° 52.5'</td>
<td>4.16</td>
<td>2.18</td>
<td>0.76</td>
<td>0.04</td>
<td>32</td>
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<tr>
<td>TT3-14</td>
<td>47° 02.0'</td>
<td>130° 47.0'</td>
<td>3.75</td>
<td>1.93</td>
<td>9.46</td>
<td>0.20</td>
<td>396</td>
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<tr>
<td>TT3-15</td>
<td>47° 02.0'</td>
<td>130° 21.5'</td>
<td>3.71</td>
<td>2.17</td>
<td>3.4</td>
<td>0.20</td>
<td>142</td>
</tr>
<tr>
<td>TT3-16</td>
<td>47° 01.0'</td>
<td>130° 17.5'</td>
<td>3.82</td>
<td>1.81</td>
<td>3.4</td>
<td>0.20</td>
<td>142</td>
</tr>
<tr>
<td>TT3-17</td>
<td>47° 01.0'</td>
<td>130° 05.0'</td>
<td>3.52</td>
<td>2.39</td>
<td>6.0</td>
<td>0.50</td>
<td>251</td>
</tr>
<tr>
<td>TT3-18</td>
<td>47° 01.0'</td>
<td>129° 45.0'</td>
<td>3.65</td>
<td>2.15</td>
<td>2.6</td>
<td>0.30</td>
<td>109</td>
</tr>
<tr>
<td>TT3-19</td>
<td>47° 03.0'</td>
<td>129° 55.0'</td>
<td>3.57</td>
<td>2.15</td>
<td>5.0</td>
<td>0.10</td>
<td>210</td>
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<tr>
<td>TT3-20</td>
<td>46° 58.5'</td>
<td>129° 06.0'</td>
<td>3.51</td>
<td>2.20</td>
<td>10.5</td>
<td>0.20</td>
<td>440</td>
</tr>
<tr>
<td>TT3-21</td>
<td>46° 59.0'</td>
<td>129° 00.5'</td>
<td>3.56</td>
<td>2.08</td>
<td>4.5</td>
<td>0.30</td>
<td>189</td>
</tr>
<tr>
<td>TT3-22</td>
<td>47° 09.0'</td>
<td>128° 54.5'</td>
<td>3.56</td>
<td>2.21</td>
<td>7.2</td>
<td>0.30</td>
<td>302</td>
</tr>
<tr>
<td>TT3-23</td>
<td>47° 01.0'</td>
<td>128° 46.0'</td>
<td>3.59</td>
<td>2.18</td>
<td>6.7</td>
<td>0.30</td>
<td>281</td>
</tr>
<tr>
<td>TT3-24</td>
<td>47° 00.0'</td>
<td>128° 37.0'</td>
<td>3.65</td>
<td>2.06</td>
<td>9.4</td>
<td>0.90</td>
<td>394</td>
</tr>
<tr>
<td>TT3-25</td>
<td>47° 00.0'</td>
<td>128° 30.0'</td>
<td>3.65</td>
<td>2.89</td>
<td>4.5</td>
<td>+0.60</td>
<td>189</td>
</tr>
</tbody>
</table>

* In situ heated probe measurement (Lister 1970b)
† Assumed
‡ Partial penetration: lowest sensor
On the thermal balance of a mid-ocean ridge

The disturbance by touching the sediment before penetrating fully, but even if a local slump has occurred here, the gradient is unlikely to have been affected. The thermal time constant of 25 cm of sediment of normal diffusivity (0.0025 cm² s⁻¹) is only 0.01 yr, and it is only 1 yr for the whole 200 cm sediment column at this station.

The most striking feature of the heat-flow distribution west of the ridge crest is the zone of sustained high values, between 45 and 75 km (Fig. 3). An acoustic reflection profile across the western boundary of this zone is shown in Fig. 4, where the measured heat flow changes by a factor of 50 over a distance of 25 km. The high value is on a ridge, and the low value is in a valley, which is the reverse of what would be expected from simple heat-flow refraction. The core from the valley station does show evidence of turbidite deposition in the past, but it also has the dark-coloured surface layer indicative of slow deposition in the post-glacial environment. Total sediment thickness is 50 m in a valley 6 km wide: the steady-state refraction correction due to the sediment fill is negligible, and the time constant of the whole layer is 300 yr. There is no evidence of sediment disturbance in the last 13 000 yr, and the good constancy of the heat flow between the long and short probe intervals rules out any significant water temperature change.

The oceanic basement in the middle of the high heat-flow area is remarkably smooth on a small scale, so that the acoustic profiler was able to resolve 30–50 m of sediment draped over the abyssal hills. A section of profile near the two values of 3.4 HFU (Fig. 3) is reproduced in Fig. 5. Notable is the hard basement echo indicative of the absence of surface roughness greater than 10 m. In contrast to this, a section of the same acoustic profile, taken under identical recording conditions, but east of the high heat-flow zone, shows diffuse basement echoes except for short segments of valley floor (Fig. 6). The wider valley contains 25 m of sediment, but the narrower one, leading down to a ponded basin, contains 70 m. This latter valley is probably connected to the Vancouver Island turbidite source, but the 25 m cover is consistent with reasonable quasi-pelagic sedimentation rates and the 1 My magnetic spreading age of the ocean floor at this point (Vine 1966). The basement roughness on the hills is here greater than 25 m, so that bare rock may be expected to outcrop on the hills although they are no larger than those in the high heat-flow zone further west. The gross topographic relief of 0.2 km over a wavelength of 10 km can generate a refractive deviation of only 6 per cent, and the lowest measurement in the area (0.86) is in any case on the floor of a narrow valley (see Fig. 3).

East of the ridge crest, the heat-flow measurements are high, much more consistently than on the western side. The scatter is still considerable, but no longer...
over an order of magnitude: the highest measurement is only 47 per cent above the
6 station mean of 7.15 HFU. Some, but not all, of the variation can be accounted
for by the basement relief under the sediments, and the effect of the high turbidite
sedimentation rates. The five innermost measurements fall on the acoustic reflection
profile shown in Fig. 7. The value of 10.5, 12 km from the ridge crest, was taken at the
extreme edge of a small sedimented valley, within 100 m of the outcropping rock
slope that itself is 100 m high. The situation of the measurement is akin to a mixture
of the cases treated by Lachenbruch (1968) and Von Herzen & Uyeda (1963). On
the one hand the measurement is within one scale height of what may be a 45° slope,
and on the other, it is in a sediment-filled trough that may not be too far from elliptical
in cross-section. While the solutions referred to may not be added, a rough idea of the
effect of the actual topography may be obtained by multiplying the corrections: the
observed flow should be \times 1.17 because of the slope and \times 0.9 because of the trough,
or only about 5 per cent too high. At the known spreading rate of the ridge (Vine
1966), the valley should be about 0.4 My old, and unless the 40 m of sediment were
deposited recently, the sedimentation correction is only 3--5 per cent (Fig. 9 i in Von
Herzen & Uyeda 1963). Thus the expected net correction for this station is nearly
zero, in so far as it can be estimated.

Another station that is notably out of line is the first on the abyssal plain proper,
4.5 HFU at 18 km from the crest. This is in a substantial sediment-filled trough,
where both the refractive and sedimentation corrections are significant. A minimum
sedimentation correction results if all 200 m of sediment has been deposited when the
valley was formed: +7 per cent. A reasonable maximum corresponds to sedimenta-
tion 0.1 cm yr\(^{-1}\) for the last 200 000 years, or +15 per cent. The ellipsoidal trough
solution, which should be a good approximation for a centrally located measurement,
gives a further +10 per cent. Thus the measurement should be raised to between 5-2
and 6 HFU, not enough to be in line with a decreasing trend from the ridge crest, but
at least changed in the right direction.

The last major out-of-line value is the 9.4 HFU at the extreme left of Fig. 7. There
is some rise in the basement near the measurement, at least 100 m in 5 km, but it is
less pronounced than the buried hill under the nearby 6.7 HFU value. In the case of
a thin high-resistivity layer overlying topography on a medium of lower resistivity,
the Jeffreys (1938) approximation must be modified to a surface thermal resistivity
model. This shows that the effect of buried topography is enhanced by the ratio of the
thermal resistivities minus one, provided the slopes are gentle enough for the basic
Jeffreys approximation to be valid. Even with the \times 1.5 enhancement, the
correction to the measurement (on the sinusoidal ridge approximation) is only
-5 per cent for the visible basement relief; it could be higher if basement deepened
suddenly beyond where the profile terminates (cf. Fig. 2). There is some evidence
that this may be the case, but basement returns on that section of Fig. 7 are not
clear enough to be sure.

The prominent basement ridge under the measurement of 6.7 HFU (Fig. 7) should
be taken into account. As it is flat topped, the inclined plane solutions of Lachenbruch
(1968) may be used, and show that the -7 per cent correction is less than would be the
case for a sinusoidal ridge of similar height. The enhancement factor is applied to the
slope height in this case.

4. Explorer trough study

The pertinent measurements made from the Hudson are listed in Table 2. They are
distributed over the ocean floor on, or south-west of, the prominent fault line associated
with the upwarping of the Paul Revere Ridge (Fig. 8). This feature is capped by
sediment layers that dip toward the north-east and outcrop on the south-west slope
(R. L. Chase, private communication). It is therefore a piece of normal ocean floor
FIG. 5. Reflection profile of evenly sedimented basement within the high heat-flow zone, near 47° N and between 130° 23’ W and 130° 09’ W. Horizontal scale is 15 km long with 5-km divisions.

FIG. 6. Reflection profile near 47° N, from 139° 51’ W to 129° 37’ W, an area of moderate heat flow near the ridge crest. Horizontal scale is 15 km long with 5-km divisions. To the left of the 3.5-s identifier, a small peak reaching 3.35 s does not register on this record, due to strong scattering of the sound.
Fig. 7. Reflection profile east of the ridge crest near 47° N, with heatflow values (in HFU) superimposed. Vertical exaggeration 15 : 1.

Fig. 9. The heat-flow measurements near the Explorer troughs projected onto UBC airgun profile 70-16-14. Profile by courtesy of Dr R. L. Chase, University of British Columbia. The vertical scale is from 2·5 to 5 s in 20 divisions, and the profile is 77 km long, so that the vertical exaggeration is 10 : 1.
that has been upwarped, rather than a linear volcanic pile, and represents a clear boundary to the spreading centre. The first seven measurements form a profile across the two troughs that abut the upwarped scarp, and were sited as close as possible to airgun reflection profile UBC-70–16–14 (R. L. Chase, private communication).

The measurements have been projected onto the profile in Fig. 9, and it is striking that the floors of the troughs are almost bare of sediment although the surrounding topography is well covered: in the case of south-easterly trough, to the very brim. The measurement in the north-westerly trough obtained only partial penetration with the 2-m probe, and a cine camera station was run nearby to determine the structure of the smooth, acoustically-hard floor. It is indeed quite flat, and consists of sediments interrupted every few metres by outcropping, though sediment-covered, boulders. A lava flow covered by 1 m of sediment would fit the appearance precisely. The relative absence of sediments strongly suggests that the floors of both troughs were formed recently, and the distribution of heat-flow measurements is therefore extraordinary. All the measurements on the sedimented basement near the troughs are high, but those in the troughs themselves are not, even though the bench in the middle of Fig. 9 is sediment-covered. The trend is exactly opposite to what would be expected from refraction through the topography, and so refraction corrections have not been made.
### Table 2

**Heat-flow measurement: Explorer Trough area**

<table>
<thead>
<tr>
<th>Station</th>
<th>Latitude (N)</th>
<th>Longitude (W)</th>
<th>Depth (sec, 2-way)</th>
<th>Conductivity ($\times 10^{-3}$ cal cm$^{-1}$)</th>
<th>Heat flow ($\times 10^{-6}$ cal cm$^{-2}$ s$^{-1}$)</th>
<th>Error band (±)</th>
<th>Heat flow (mW m$^{-2}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>H70-4</td>
<td>49° 58'.60'</td>
<td>129° 32'.60'</td>
<td>2.78</td>
<td>2.19</td>
<td>3.8</td>
<td>0.04</td>
<td>159</td>
</tr>
<tr>
<td>H70-5</td>
<td>50° 01'.57'</td>
<td>129° 43'.04'</td>
<td>3.13</td>
<td>2.08</td>
<td>8.4</td>
<td>0.30</td>
<td>352</td>
</tr>
<tr>
<td>H70-6</td>
<td>50° 04'.36'</td>
<td>129° 46'.36'</td>
<td>4.36</td>
<td>1.94</td>
<td>1.3</td>
<td>0.10</td>
<td>54</td>
</tr>
<tr>
<td>H70-7</td>
<td>50° 07'.70'</td>
<td>129° 49'.80'</td>
<td>3.70</td>
<td>1.92</td>
<td>1.9</td>
<td>0.20</td>
<td>80</td>
</tr>
<tr>
<td>H70-8</td>
<td>50° 07'.50'</td>
<td>130° 01'.80'</td>
<td>2.79</td>
<td>2.19</td>
<td>6.6</td>
<td>0.20</td>
<td>276</td>
</tr>
<tr>
<td>H70-9</td>
<td>50° 13'.20'</td>
<td>130° 09'.20'</td>
<td>3.22</td>
<td>2.34</td>
<td>4.7</td>
<td>0.10</td>
<td>197</td>
</tr>
<tr>
<td>H70-10</td>
<td>50° 15'.26'</td>
<td>130° 17'.84'</td>
<td>3.25</td>
<td>1.71‡</td>
<td>1.5</td>
<td>0.30</td>
<td>63</td>
</tr>
<tr>
<td>H70-11</td>
<td>50° 21'.00'</td>
<td>130° 06'.60'</td>
<td>3.53</td>
<td>2.43</td>
<td>16.8</td>
<td>0.40</td>
<td>704</td>
</tr>
<tr>
<td>H70-12</td>
<td>50° 45'.90'</td>
<td>130° 36'.60'</td>
<td>3.28</td>
<td>2.27</td>
<td>4.8</td>
<td>0.10</td>
<td>201</td>
</tr>
<tr>
<td>H70-13</td>
<td>50° 39'.50'</td>
<td>130° 33'.00'</td>
<td>3.27</td>
<td>2.06</td>
<td>4.0</td>
<td>0.06</td>
<td>168</td>
</tr>
</tbody>
</table>

All these measurements were made with the *in situ* conductivity probe.

‡ Incomplete penetration: lowest sensor value.
The remaining three measurements are essentially on the Paul Revere fault line (Fig. 8). One, at the end of the north-westerly trough, is in a small shallow sediment pond at the base of the ridge slope. It is unusually high, even for this area. The other two are in a larger and deeper sediment pond to the north-west. This straddles a possible final spreading rift leading to the Queen Charlotte Fault proper, which essentially follows the continental margin off the Queen Charlotte Islands (Fig. 1). In view of the exceedingly high value farther down the Paul Revere Fault, the two measurements of 4 and 4.8 HFU do not confirm that this trough is the site of further spreading, but neither do they rule it out.

5. Discussion

The heat-flow measurements of this study differ from previous results (cf: Le Pichon & Langseth 1969) primarily in the presence of numerous zones of high values. Near Explorer Trough, both the plateaus have high heat flow (Fig. 9) and so does the fault line associated with the Paul Revere Ridge (Fig. 8). The more extensive transect of the Juan de Fuca Ridge establishes two major high zones: the sedimented flank east of the crest, mean 7.1 HFU, and between 50 and 130 km west of the crest, mean 6.4 HFU (Fig. 3). It is notable that all the areas where the heat flow is high have an even sediment cover thicker than local basement roughness (Figs 5, 6, 7 & 9), though this by itself is not sufficient to ensure a high measured heat flow (Fig. 4).

If an even sediment cover is necessary to measure consistently high heat flow, one must test the possibility that local sediment ponds separated by outcropping rocks can result in substantially depressed heat flow in those ponds. Von Herzen & Uyeda (1963) have shown that an isolated circular pond depresses the heat flow most, and if the conductivity contrast were as high as 3 and the basin were hemispherical, a measurement in it would be 0.4 of the true value. If the rock topography between basins is elevated, the depression is reduced, and the interference between neighbouring basins also reduces the effect. The 0.4 factor therefore represents a maximum for any reasonable configuration of lightly sedimented ocean floor, and cannot be expected to apply on an average basis to randomly located measurements. If the extreme correction factor were applied to all the inner western flank values (mean 2.4) it would bring them up to the same level as the others—yet all the heat-flow measurements can hardly have been made in deep local sediment pockets. The heat-flow contrast shown in Fig. 4 is orders of magnitude larger than the 2.5:1 expected, and moreover the high value is on the ridge where outcropping rocks may be common, not in the valley. The explanation cannot be applied at all to the measurements in the Explorer Troughs themselves: the local ponds would have effects smaller than the gross topographic effects of opposite sign. Therefore, while local refraction is likely to have some depressant effect on heat-flow measurements in rough topography, it cannot explain the gross contrasts observed in this study.

The next question that arises is whether even the highest values are high enough to be consistent with a conductively cooled sea-floor spreading model fed by material at observed lava temperatures (1400°C). The simplest spreading model has been analytically solved by McKenzie (1967), but an even simpler model is adequate for use here: cooling of a half-space from uniform initial temperature, neglecting lateral conduction entirely. The solution is given in Carslaw & Jaeger (1959, p. 59) and the surface gradient is $T_0/\sqrt{(\pi k t)}$, where $T_0$ is the initial temperature, referred to surface zero, $\kappa$ is the diffusivity of the rock material, and $t$ is time. The time axis can be translated into spreading distance, and although the solution is clearly inapplicable close to the ridge crest, where lateral heat flow is important, the approximation is good once the spreading distance substantially exceeds the penetration depth of conductive cooling $2\sqrt{(\kappa t)}$, at which depth $T = 0.84T_0$. The Juan de Fuca spreading velocity is 3 cm yr$^{-1}$ (Vine 1966) and if a diffusivity of 0.007 cm$^2$ s$^{-1}$ may be assumed
for the rocks, the spreading distance is three times the penetration depth at a distance of 30 km from the ridge crest. The reason this is considered adequate is apparent from Fig. 10, where the upper curve is the heat flow predicted by the cooling model: it comes on scale at 30 km. All the heat-flow values of the ridge crest study are also plotted in Fig. 10 by their distance from the ridge crest along the 47th parallel. Only the highest values west of the ridge crest are above the conductive cooling curve; the mean in the high heat-flow zone, at 6.4, is not too different from the mean of the cooling curve in that region, 7.5. However, none of the measured values less than 50 km from the crest reach the curve, and the discrepancy increases as the ridge crest is approached.

The lower curve in Fig. 10 has been obtained from another simplistic model. In this case it is assumed that a linear gradient corresponding to a surface heat flow of 10 HFU is somehow established by cooling of the rocks as they are emplaced. With the figures used (\(K = 0.007 \text{ cal cm}^{-1} \text{ s}^{-1} \text{ °C}^{-1}\), \(T_o = 1400 \text{ °C}\)), the penetration of this cooling is 10 km. The model was chosen because it gives a monotonically decreasing surface heat flow and a simple form for it: \(q_0 \text{ erf} d/(\sqrt{\pi t})\), where \(q_0\) is the initial heat flow and \(d\) is the penetration depth. The solution was obtained from Carslaw & Jaeger (1959, p. 59), and it again ignores lateral conduction, but in this case it is a good approximation throughout. It is apparent from Fig. 10 that this model cannot be distinguished from purely conductive cooling in the western flank area of high heat flow, but it comes much closer to agreeing with the values on the eastern flank. The negligible effect of drastically different crestal conditions on the flank heat flow is important: not only do flank measurements give no insight into dyke injection or other ridge spreading mechanisms, but the explanation for depressed flank heat-flow values must be sought in locally operating mechanisms of non-conductive heat loss. A comparison between the theoretical and measured heat flow at 3 My on this
On the thermal balance of a mid-ocean ridge

7 HFU, and the normalized means for the East Pacific Rise, 2.6 HFU, and the Mid-Atlantic Ridge, 1.9 HFU (Le Pichon & Langseth 1969) shows that there is something to be explained.

At some distance from the ridge crest in any convecting model the upper boundary reaches a state that corresponds closely to a lithospheric layer cooling slowly by conduction. The total heat output between the crest and such a point on the flank must be equal to the heat lost by the lithospheric material in cooling from its initial state. If the temperature distribution at this flank point is similar to that of a half-space cooling from an initial uniform temperature, the total heat output can be obtained simply by integrating the surface heat flux of the one-dimensional model through time. This is only a fair approximation for the more imaginative convective models, but, since the real situation is complicated by processes of partial melting, magma percolation and differentiation, a more elaborate calculation is hardly justified. The interestingly simple result is that the mean heat flow between the ridge axis and a point on the flank is just twice the heat flow at that point computed from the one-dimensional cooling model. For example, in Fig. 10 the mean heat for flow the 100 km (full-width) central zone is nearly 16 HFU, and, so long as volcanic activity is confined to this zone, and cooling is conductive, the result is not affected by the detailed spreading model. In fact, the real heat loss should be substantially greater than this, since a proportion of the uppermost 10 km of material comes up molten, and the latent heat of freezing must be added to the integrated specific heat. The addition could be considerable: the latent heat of melting for dry Forsterite can be estimated from the Clausius–Clapeyron equation and data in Clark (1966); it is 800 cal cm$^{-3}$.

The thermal balance of the ridge crest zone is thus incompatible with the conductive cooling of material upwelling at a reasonable magmatizing temperature, and the observed level of the heat-flow measurements. The presence of high values in rough agreement with the conductively expected heat flow on the western flank suggests that the discrepancy is caused by failure of the assumption of conductive cooling at the crest rather than a more general failure in the sea-floor spreading hypothesis. It is then not necessary to postulate unusual local heat sources for an area of oceanic

FIG. 11. A portion of the Raff & Mason (1961) magnetic survey, with an addition near 47° N 132° W, reproduced to cover the same area as Fig. 3. Positive anomalies are contoured as full lines, negative as dashed lines; contour interval 100y.
crust that is a classic example of undisturbed sedimentation (Fig. 5). It may be argued that the abrupt decrease in heat flow beyond 131° W (Fig. 3) is not in agreement with this idea, but a consideration of the magnetic evidence makes it even more plausible. Fig. 11 is a reproduction of a portion of the Raff & Mason (1961) magnetic anomaly map corresponding to the area of this investigation (Fig. 3). The abrupt decrease in heat flow to more usual flank values occurs west of 131° W. This is the point at which the long anomaly trends begin to diverge from parallelism with the ridge crest: at 131° 30' W the broad negative lineation is almost NS in strike, and the three positive lineations immediately east of it at the northern edge of the map have been compressed into a broad positive band at 47° N. The special conditions of ocean floor-spreading that produced the smooth basement of the high heat-flow zone cannot be expected to extend into the disturbed region probably indicative of the birth of the Juan de Fuca spreading epoch. Note that the well surveyed and sampled area near 47° N 132° W lies on true NS lineations of an earlier spreading era, so that the well established low heat flow could be due to a greater age of the lithosphere there.

6. Hydro-thermal convection

Once it is established that a substantial part of the surface heat loss at the ridge crest is non-conductive, no great imagination is needed to select the most likely mechanism: hydrothermal circulation in the oceanic crust. Iceland is a part of the Mid-Atlantic Ridge, and it is laced with hot springs and geysers. Other tectonically active regions of the world contain large-scale geothermal areas: New Zealand, Italy, California, to name a few. The mechanisms involved in these areas have been discussed at length by Elder (1965) and others; the important question is whether a hypothesis of hydrothermal circulation can help to explain other aspects of the ridge heat-flow observations beside the gross heat loss. The purpose of the following discussion is to show that it does explain both the valley effect in rough topography and the relatively uniform heat flow observed where basement topography has been buried by sediments.

If hydrothermal circulation is to occur, the convective driving forces must exceed the viscous drag of water percolation through the rock. Although both the thermal expansion and the fluidity of water will be decreased somewhat by the high ambient pore pressure beneath the ocean, the principal unknown is the permeability to be expected of the oceanic crustal rocks. Borehole tests in the New Zealand geothermal area (Elder 1965) suggest bulk permeabilities of 0.01 Darcy (0.987 × 10⁻⁸ cm²), but it is possible that the rarity of major geothermal areas is due more to lack of permeability than to lack of heat sources in other regions. If the oceanic basaltic layer is assumed to have a permeability k of 10⁻⁴ Darcy, an effective circulation depth h of 7 km, a diffusivity $\kappa_m$ of 7 × 10⁻³ cm² s⁻¹, and a temperature difference $\Delta T$ of 1400°C, while the percolating water has a thermal expansion coefficient $\alpha$ of 8 × 10⁻⁴/°C and kinematic viscosity $\nu$ of 1.5 × 10⁻³ cm² s⁻¹, corresponding to water at 200°C (Elder 1965); then the Rayleigh stability number $\sigma = k g \Delta T h / \kappa_m \nu$ is approximately 80. It has been shown (Lapwood 1948) that convection can occur for Rayleigh numbers greater than 40 if the pattern is forced by uneven heating from below. Very conservative values have been assumed for the permeability and for the expansion and fluidity of water, since the latter increase with temperature in the range (<300°C) where measurements have been made (Elder 1965). Therefore convective overturn of the pore water in the crustal layer near a ridge crest is highly likely, since the requirement of uneven heating is almost certainly met.

On the flanks of a ridge, both the thermal drive and the forcing effect of uneven heating must decrease as the lithosphere becomes older, and eventually convection will cease. The situation is complicated by the unknown effect of weathering and hydrous alteration of the rock on its permeability, and the natural variability in the
The case where convection ceases as soon as cooling has penetrated to 10 km has been treated above, and the lower heat-flow curve of Fig. 10 describes the subsequent conductive cooling. The heat flow predicted by this curve is greater than the mean of most ridge flank values until the lithosphere is about 30 My old (Le Pichon & Langseth 1969), but although the means are thereafter in rough agreement, there is still more scatter in the measurements than can be expected from conductive refraction. Only in the older ocean basins does the heat flow reach the uniformity purely conductive flow should have (cf. Reitzel 1963). This is, perhaps, to be expected: low permeability regions will cease to convect while comparatively young, but where the permeability is high, say 0.01 Darcy as in geothermal areas, the flow can continue until the thermal drive drops below 140°C over 10 km, a gradient corresponding to less than normal basin heat flow in the absence of convection.

At the ridge crest the hydrothermal circulation must be open, since an essentially infinite reservoir of fluid abuts the free surface of the crest. The free boundary should slightly reduce the Rayleigh number required for the onset of flow, but its most

![Diagram](https://example.com/diagram.png)

**FIG. 12.** (a) Pattern of convection to be expected near a ridge crest, where the permeable layer is open to the ocean. (b) Convective pattern forced by topography in an area of permeable crust bounded by a thin impermeable blanket of sediment. (c) The case of permeable crust with a rough boundary, buried by flat-lying sediments. The varying thermal resistance of the sediment blanket overpowers the direct effect of the topography.
important effect is on the thermal gradient at the boundary. Elder (1965) presents the results of one experiment on convection in a permeable layer heated along an infinite strip of width 2l below it. The heat output of the system is sharply concentrated over a band of width 0.8l, outside of which there is almost no heat output at all. The importance of this can be gauged when the effect of topography on the convective pattern is considered. Ridges and seamounts in the permeable material will act as chimneys for the hot flow, while valleys, representing the deepest penetration of cold bulk seawater, are natural sites for the ingoing return flow. The type of circulation to be expected is shown in Fig. 12(a) for two-dimensional topography, and in a perfect, or uniform percolation, system, zero heat flow should observed in the down-draught zones.

The plausibility of a circulation system for removing the initial heat of the fresh crust can be tested by a basic thermal output calculation. Suppose that at a ridge spreading at 3 cm yr\(^{-1}\) (one side), the top 10 km of lithosphere is cooled an average of 700°C, releasing 700 cal cm\(^{-3}\). The continuous thermal output of 100 m of crest is 1.4 × 10\(^6\) cal s\(^{-1}\) (or 6 MW), and if the circulation mechanism results in a hot spring at about 50°C, it must run at 28 l s\(^{-1}\), equivalent to a 20-cm diameter pipe discharging at 1 m s\(^{-1}\). This would be a large spring, but by no means exceptional, and it could surface anywhere over the central 1 km of the ridge crest. If the heat were dispersed into a cross-ridge flow of 1 cm s\(^{-1}\), and 100 m thick, the temperature rise of the ocean water would be about 0.014°C. Since the thermal gradient in the near-bottom water observed on lowering the heat probe is commonly 0.01°C/30 m, it would not be easy to detect the heating in the topographically disturbed flow near a ridge crest. Only a careful areal study of absolute temperature at a fixed depth, on the lines of Chung et al. (1969), would be able to test for the overall thermal output of the ridge crest.

The second test of the hydrothermal circulation hypothesis is the distribution of measured heat flow over topography bounded by an impermeable layer, such as the high heat-flow region on the western flank of the Juan de Fuca Ridge. There are two reasons why this area can be considered bounded by an impermeable layer of sediment: the mean heat flow is close to what it should be for conductive cooling (Fig. 10), and the pelagic sediment, besides being of low inherent permeability (Appendix), can readily plug the discrete vents of the hard rock structure at the interface. If the heat flow in the region were purely conductive, the even sediment cover over gently undulating basement (Fig. 5) would require the measurements to have a low scatter, and to be higher in the valleys and on the flanks of the ridges or hills than on the crests. Although the sample in Fig. 3 is a small one, the tendency is clear: the highest values are on top of hills or ridges, and lower values are observed on the flanks or in valleys. The high value of 10-9 HFU at 130°06'W, discussed at length in Lister (1970a), is a striking case.

It is situated on the brow of a hill, a place where conductive refraction would generate a minimum of surface heat flux, yet the measurement is nearly twice as high as one made nearby in a small valley or hollow. The values measured near Explorer Trough (Fig. 9) are an even more telling example of heat-flow correlation with topography opposite to that expected from conductive refraction. As the area is complicated by an apparent split in the spreading centre, and thus has an ill-defined recent history, it would be premature to attempt a detailed interpretation. The effect of an undulating impermeable boundary on hydrothermal convection in a permeable layer is illustrated diagrammatically in Fig 12(b). Valleys or surface depressions tend to depress the isotherms beneath them and therefore are places of preferential separation for the cold columns. By elimination, hills and ridges will become the sites for the hot rising columns, and should thus show the highest conductive heat flow through the impermeable sediment blanket. The forcing effect of the topography can be visualised by considering the stability of a cooled boundary layer on an inclined plane: the denser boundary layer will tend to slide down the plane until separation
occurs. The ability of the surface topography to control the cell geometry will depend on the ratio between the inverse wavenumber of the most pronounced topography and the cell size, and the relative effect of variations in the lower boundary of the permeable layer. If the permeability decreases gradually with depth, only large variations in the height of the lower boundary will affect the convection pattern, but such large variations could occur. Variations in the permeability of the crustal material are also likely to distort the cells, so high heat flow should by no means be 100 per cent correlated with the tops of hills or ridges. The evidence certainly suggests a positive correlation between high heat flow and topographic elevations, as opposed to the negative correlation that would be expected from conductive refraction.

The third test of the convection hypothesis is the prediction of the heat-flow variation on a flat sediment plain overlying topography on the basement. The contrast in thermal conductivity between moderately compacted sediment and basaltic rock is about 1:2.5, so that, in the absence of convection, the isotherms in the basement will be distorted by -1.5 times the topography, provided that the Jeffreys (1938) approximation is valid. The coldest areas of the upper boundary of the permeable material are now the highest topographically, and the sense of any convective flow should be reversed from its original configuration before the sediments were laid down (Fig. 12(c)). The geometrical effect aiding the motion and separation of a cold boundary layer is no longer operative, and the forcing effect of the buried topography is probably somewhat weaker than that of similar topography coated with an impermeable skin (see second test, above).

Now, the hottest water is in contact with the thickest sediment blanket, and the coolest water occurs where the sediment blanket is thinnest. Whether the highest heat flow will be observed over basement lows or highs depends on whether the temperature difference between the hot and cold columns of the cell is greater or less than the basal temperature difference between the sediment columns for a uniform heat flow equal to the regional average. In general, when the sediment thickness is much less than 0.4 times the convection penetration into the crust, and the Nusselt number is not too high, the highest heat flows should be observed over troughs in the basement. If convection is vigorous, and the temperature difference across the cell is a small fraction of that required for equal conductive flow, it is possible for the heat flow to be quite uniform even over rough buried basement. Conversely, if the basement relief is too low to force the cell pattern, and the Nusselt number is low, there will be 'hot spots' over the rising columns, uncorrelated with the topography.

The four Cascadia Plain stations in Fig. 7, and the one further east (Fig. 10), bear out the prediction in a general way. The mean heat flow for the region is in fair agreement with the lower cooling curve in Fig. 10: this corresponds to strong initial convection cooling the crest to considerable depth, followed by a resumption of conductive flow. Surface sealing and a reduction in the vigour of convection would produce similar results to true bulk conduction, and there are enough variable parameters to produce a range of models with reasonable mean heat flows. The chief discrepancy from prediction is the low value at the edge of the plain. It is within 3 km of outcropping basement, and it is not too implausible to suggest that the low value is due to forcing of an eddy under the plain by vigorous open cells near the ridge crest. The odd high (9.4 HFU) value over no particular basement disturbance corresponds well with the idea of an upwelling hot spot over a cell not correlated with topography.

7. Conclusions

The unusual heat-flow distribution across the Juan de Fuca Ridge at 47° N, and, to a lesser extent, the heat-flow distribution near Explorer Trough, can be interpreted to agree with a simple model of sea-floor spreading. The crucial requirement is that
the rock material at the ridge crest be immediately cooled to a depth of 5–10 km by open hydrothermal circulation. If the permeable crustal layer is subsequently sealed off by sedimentation, then the measured heat flow increases to values compatible with conductive cooling models. On the western flank of the Juan de Fuca Ridge, an area of unusually smooth basement appears to have been adequately sealed off by hemipelagic sedimentation; on the eastern side the encroaching turbidities have completely buried the basement topography. In both these areas, the mean heat flow is near 7 HFU, and is consistent with either conductive flow after cooling by hydrothermal circulation, or continued circulation beneath an impermeable boundary. The *distribution* of the measured values over the topography is more consistent in all areas with continued hydrothermal circulation, forced in part by the topography, than with conductive refraction. The continuation of hydrothermal flow in inadequately sealed areas is also necessary to explain the low values commonly found on ridge flanks (cf. Le Pichon & Langseth 1969).

The hypothesis can be further tested by examining the relationship between measured heat flow and topography in areas where topological information is complete for large-scale features, and the distribution of sediments is known. Eventually, the ridge crest hot springs should be found (and possibly tapped for geothermal energy), but it has been shown that the hot water will be hard to detect in an open ocean with significant bottom current flow. The principal value of the data presented in this paper is to confirm the long-held suspicion that the thermal balance of a ridge crest is dominated by circulatory cooling rather than by conductive processes.

**Acknowledgments**

This paper summarizes the work of several years, supported by Grants GA–577, GA–1640 and GA–27947 of the National Science Foundation, and Contract Nonr–477(37), Project NR 083012 of the Office of Naval Research, both of the United States of America. The work of C.S.S. *Hudson* was possible through the generosity of the Department of Energy, Mines, and Resources of Canada, at the Bedford Institute of Oceanography, and through a travel grant made by the Department of Oceanography, University of Washington. I wish to thank Dr R. L. Chase, of the University of British Columbia, for making available one of his seismic profiles, Dr A. W. Fairhall, of the University of Washington, for running carbon–14 dates on critical core samples, and Mr E. E. Davis for his invaluable assistance, both on the *Hudson* and subsequently with the data reduction.

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**Appendix I**

The permeability $k$ of a representative sample of deep-sea sediment has been estimated from the transient loading curve of Mohole Phase I, sample EM8-10, published by Hamilton (1964). Axial loading was applied to a circular disk of sediment by a pair of porous plugs, and the pore fluid was squeezed out into these plugs. During the 'primary consolidation' phase (9 min), the rate of shrinkage of the sample should be controlled by the permeability of the sediment, the permeability of the interface between sediment and plug, and the permeability of the plug itself. It will
be assumed that the permeability of the plug is much higher than that of the sediment, that the interface resistance is negligible compared to the bulk sediment resistance after the first few seconds of loading, and also that the dimensional changes are small enough to be ignored in setting up the flow equations. Since the total primary consolidation in the case considered is 4 per cent, a simple constant-dimension treatment is adequate.

The actual equilibrium response of the sediment fabric to loading is non-linear (Fig. 13(a), points), and varies from sample to sample. The effect of this can be estimated by considering the transient response for two cases: the fully linear compaction OP, and the sudden compaction model OYP where the sediment fabric shows no strength until the final level of compaction is reached. The first case corresponds to a diffusion equation solution, while the second results in a sharp state transition boundary that moves away from the free surface at a steadily decreasing rate.

![Graph of steady-state compaction](https://academic.oup.com/gji/article-abstract/26/5/515/571317)

![Graph of transient primary compaction](https://academic.oup.com/gji/article-abstract/26/5/515/571317)

Fig. 13. (a) Steady-state compaction of Mohole sample EM-10 versus increasing pressure, after Hamilton (1964). (b) Transient primary compaction of the same sample with an applied pressure of 0.7 bars, plotted against square root of time. Replotted from Hamilton (1964, Fig. 9).
The linear model

If a compression state \( \Theta \) is defined by \( \Theta = (V_0 - V)/(V_0 - V_p) = (V_0 - V)/\Delta V \), where \( V_0 = \) initial volume, \( V_p = \) final volume, \( V \) is the actual volume, and \( \Delta V \) is the final volume change, then the pore pressure at any point \( P = P_0(1 - \Theta) \), where \( P_0 \) is the loading pressure. Since it is the volume change that provides the flow, a one-dimensional diffusion equation \( d\Theta/dt = [kP_0/\rho v \Delta V] d^2\Theta/dx^2 \) results for the axial loading case, where \( x \) is axial distance, \( \rho \) is density and \( v \) is kinematic viscosity of the pore fluid. If the sample is long enough, the initial compaction will follow the solution for the semi-infinite solid \( 1 - \Theta = \text{erf} \left[ x/(\sqrt{K t}) \right] \), where \( K' = kP_0/\rho v \Delta V \), at both ends, so that \( dy/dt = [2(d\Theta/dx)_{x=0}] P_0 k/\rho v \) and the displacement

\[
y = -\left(4kP_0 t^4\right)/(\rho v \sqrt{\pi \kappa'}) = -\left(4/\sqrt{\pi} \right) [kP_0 \Delta V/\rho v]^{\frac{3}{2}} t^\frac{3}{2}
\]

The moving transition model

In this model the compacted boundary layer has a linear pressure gradient across it \( P_0/x \), and the velocity of the sharp transition is given by \( dx/dt = (k/\rho v) (P_0/x) (1/\Delta V) \) since the flow is provided by collapse of more material. The single boundary solution for this is \( x = \sqrt{\left((2kP_0 t)/(\rho v \Delta V)\right)} \) and therefore \( y = 2\Delta V x = 2\left[\sqrt{\left((2kP_0 \Delta V/\rho v)\right)}\right] t^{\frac{3}{2}} \).

It is interesting that the two models both give a \( y/t^{\frac{3}{2}} \) curve for the transient displacement in the compaction machine, and the value of \( k \) that would be calculated for the extreme non-linear model is \( 2/\pi \) of that for the linear model. The actual compaction data have been picked off Fig. 9 in Hamilton (1964), and are plotted in Fig. 13(b). The points lie close to a \( y/t^{\frac{3}{2}} \) compaction line of 0.0054 cm s^{-1}, and using the values \( P_0 = 0.7 \) bars, \( \Delta V = 0.042, \rho v = 0.01 \) poise, the values of \( k \) that result are

\[
\begin{align*}
k_{\text{linear}} & = 1.95 \times 10^{-12} \text{ cm}^2 \\
k_{\text{transition}} & = 1.24 \times 10^{-12} \text{ cm}^2.
\end{align*}
\]

The points in Fig. 13(a) represent equilibrium rather than primary compaction, but there is no reason to suspect that the primary compaction response would be more non-linear than the equilibrium response. Since the \( k \) calculated for the extreme OYP path differs little from \( k_{\text{linear}} \), it is safe to take the latter as a good approximation to the steady-state permeability of the sample, some \( 2 \times 10^{-12} \text{ cm}^2 \) or \( 2 \times 10^{-4} \) Darcy. Note that this value is for a sediment sample consisting of 60 per cent water by weight: closer packed material may have up to 10 times less permeability.