Submarine hydrogeology of the Hawaiian archipelagic apron
1. Heat flow patterns north of Oahu and Maro Reef

Robert N. Harris,1,2 Richard P. Von Herzen,3 Marcia K. McNutt,4 
Grant Garven,5 and Kelsey Jordan1,3,6

Abstract. We present two profiles of collocated single-channel seismic reflection and heat flow determinations across the Hawaiian flexural moat: one north of Oahu and the other north of Maro Reef. Seismic reflection data are used as an aid in determining depth to basement and interpreting moat stratigraphy. Moat sediments are locally up to 2 km thick and result mostly from slumps, debris avalanches and turbidites from volcanoes capping the Hawaiian Ridge. Each heat flow profile is ~200 km long and consists of paired thermal gradient and conductivity measurements made at an interval of 1-2 km. The mean heat flow, uncorrected for the cooling effect of sedimentation, along the Oahu profile is 63.9 mW m⁻² (s.d. 11.9 mW m⁻²) and along the Maro Reef profile is 59.8 mW m⁻² (s.d. 5.0 mW m⁻²). Our preferred sedimentation correction model gives a corrected heat flow mean along the Oahu profile of 74.2 mW m⁻² (s.d. 14.6 mW m⁻²) and for the Maro Reef profile of 64.0 mW m⁻² (s.d. 4.9 mW m⁻²). These values are not significantly different despite the fact that oceanic crust at Maro Reef is 19 Myr older than at Oahu. The mean heat flow along the Oahu profile is within one standard deviation of the global mean value for oceanic crust at the same age, while the mean heat flow along the Maro Reef profile differs by greater than one standard deviation for the global mean. However, the mean heat flow along the Maro Reef profile is within the uncertainty of previous heat flow determinations made offshoat at Maro Reef [Von Herzen et al., 1989]. Variability along each profile is significantly greater than the measurement uncertainty (measurement uncertainty is 3 and 2 mW m⁻² for the Oahu and Maro Reef profiles, respectively) and is significantly greater along the Oahu profile than the Maro Reef profile. While this variability could result from fluid flow within the archipelagic apron or underlying crust, simple moat-wide circulation models do not appear to explain the dominant variability in heat flow. Heat flow means and variability of both profiles are inconsistent with simple conductive models of lithospheric reheating.

1. Introduction

The Hawaiian Swell is one of the best studied in the world, and yet its thermal state remains controversial. Midplate swells are regions of anomalously shallow seafloor, up to 2 km high and 500-1000 km in width, and are frequently associated with young volcanic islands. While these characteristics and other proxies for the thermal state of the Hawaiian Swell suggest the presence of anomalously basal heat flux, the increased heat flow predicted by simple models has not been observed [Von Herzen et al., 1989]. Nevertheless, it seems clear that these swells must fundamentally have a thermal origin, whether through thermal rejuvenation [Crough, 1978, 1983; Detrick and Crough, 1978], dynamic uplift [Courtney and White, 1986], thermal expansion of hot plume material at the base of the lithosphere [Crough, 1983, Davies, 1988, Sleep, 1990], compositional buoyancy [Jordan, 1979] or some combination of these mechanisms [Phipps Morgan et al., 1995]. The strongest evidence for the thermal origin of these swells is the large amounts of melt needed to generate the underplated magmatic material [von Brink and Brocher, 1987] and volcanic seamounts that sit atop these swells. Additionally, the bathymetry and subsidence of midplate swells generally follow cooling curves appropriate for seafloor warmer and thinner than indicated by their crustal age [Crough, 1978]. For example, the systematic depth variation of the Hawaiian midplate swell suggests a cooling curve appropriate for seafloor 60 Myr younger than its crustal age of about 90 Ma [Detrick and Crough, 1978; Crough, 1978]. Other evidence for the thermal origin of midplate swells includes gravity and geoid anomalies [Detrick and Crough, 1978; McNutt and Shure, 1986] which indicate a relatively shallow mass compensation characteristic of thinner, warmer lithosphere.

The first Hawaiian heat flow survey (Figure 1) specifically designed to investigate the thermal state of the Hawaiian Swell measured heat flow between the southwestern flexural moat and arch [Detrick et al., 1981, Von Herzen et al., 1982]. These measurements indicate that heat flow increases systematically

---

Copyright 2000 by the American Geophysical Union.

Paper number 2000JB900165.
0148-0227/00/2000JB900165$09.00
Figure 1. Locations of seismic reflection profiles at Oahu, French Frigate Shoals, and Maro Reef (thin lines) and successful heat flow profiles (heavy line). The geologic map of the Hawaiian Ridge is generalized from Moore et al. [1994] and shows the prevalence of mass wasting deposits. Previously collected heat flow data from Von Herzen et al. [1982, 1989] are depicted as circles and triangles, respectively. The multichannel seismic (MCS, thin line) reflection profile of Watts et al. [1985] is also shown. Except for the mid-Pacific Mountain lineation extending to the southwest, the Hawaiian midplate swell is approximated with the 5000-m bathymetric contour.

from Hawaii (52.9 ± 2.3 mW m⁻²) to near Midway (58.8 ± 2.3 mW m⁻²) despite an increase in age of oceanic lithosphere and are consistent with broad-scale reheating of oceanic lithosphere. However, a significant uncertainty existed in the reference heat flow associated with "normal" oceanic seafloor in this area. A second heat flow survey, perpendicular to the axis (Figure 1) was conducted both to establish the off-swell background heat flux and to investigate the variation and magnitude of the anomalous heat flow across the swell [Von Herzen et al., 1989]. This survey was positioned downstream of the Hawaiian hot spot, at Maro Reef, where the magnitude of the heat flow anomaly was predicted to reach a maximum. Simple reheating models [Crough, 1978; Von Herzen et al., 1982] predict a Gaussian-shaped heat flow anomaly coincident with and perpendicular to the swell, with the amplitude primarily dependent on the unperturbed lithospheric thickness and time since reheating. At the location of the cross-swell survey, the amplitude of the anomaly was predicted to reach a peak of 15-20 mW m⁻² and to have values ≥10 mW m⁻² to a distance of ~300 km [Von Herzen et al., 1989].

Von Herzen et al. [1989] report 150 heat flow measurements at eight sites across the Hawaiian Swell near Maro Reef. Surprisingly, average values of heat flux at each of eight sites across the Hawaiian Swell are remarkably uniform and do not show a systematic variation in heat flow across the swell as predicted by simple lithospheric reheating models. Instead, the mean heat flow (corrected for sedimentation) at each site is within ±10% of the mean value for all sites. Some of the highest values of heat flow are at sites farthest from the axis of the swell, and some of the lowest values are measured in the Hawaiian flexural moat, an archipelago apron mostly consisting of volcaniclastic debris eroded from the volcanoes. While every effort was made to ensure that heat flow determinations were unaffected by superficial processes such as fluid circulation associated with basement outcrops, cooling effects of recent debris flows, or local heating by recent volcanism, it is possible that broad scale fluid circulation in the upper oceanic crust or in the sediments filling the flexural moat might lead to the observed pattern of heat flow across the swell.

The purpose of this research is to investigate the patterns of heat flow across the archipelagic apron in an effort to better understand the distribution of heat flow observed by Von Herzen et al. [1982, 1989] and to explore environmental influences on observed heat flow. The combination of seismic reflection data and finely spaced heat flow measurements provides the best opportunity to characterize environmental controls of heat flow variability [Davis, 1988; Davis et al., 1992; Bonneville et al., 1997]. In August and September of 1997 on board the R/V Roger Revelle, we collected seismic reflection and heat flow measurements along two profiles approximately perpendicular to the trace of the hot spot swell (Figure 1). The Oahu profile is located near the multichannel seismic line of Watts et al. [1985] to take advantage of the deep imaging. The second profile at Maro Reef is coincident with the cross-swell heat flow survey of Von Herzen et al. [1989], which allows comparison with previously collected heat flow data. Each profile is ~200 km long and consists of single-channel seismic reflection, heat flow determinations at 1-2 km spacing, multibeam 3.5-kHz echo soundings, and marine gravity. A third seismic reflection profile, at French Frigate Shoals, was...
surveyed, but the heat flow probe could not penetrate the sediments along this profile. In this paper, the first of two, we describe seismic and heat flow profiles across the archipelagic apron and explore and rule out a number of candidate sources that might account for the observed high-frequency variations in heat flow. Our preferred interpretation is that these variations result from fluid circulation within the archipelagic apron, upper oceanic crust, or both. Numerical models of fluid circulation are explored in the companion paper [Harris et al., this issue] but indicate that simple moat-wide pore fluid circulation does not produce the dominant modes of variability observed in our heat flow profiles.

2. Stratigraphic Setting

The Hawaiian Archipelago consists of a broadly linear chain of islands and seamounts, which load the underlying lithosphere, creating a flexural moat and peripheral bulge. The moat is depressed more than 2 km below the surface of the surrounding seafloor and is partially infilled with volcanoclastic debris from the islands. The prevalence of mass-wasting events as an important mode of deposition has recently been recognized along the Hawaiian Archipelago [Moore et al., 1989, 1994] and other volcanic islands such as the Marquesas [Borsczewski et al., 1992; Filmer et al., 1992; Wolfe et al., 1994], Canaries [Watts and Masson, 1985], and Reunion [Lenat et al., 1989; Voogd et al., 1999].

The stratigraphy of the flexural moat is reported by ten Brink and Watts [1985] based on a multichannel seismic reflection profile northeast of Oahu and by Rees et al. [1989] based on several cross- and along-moat single-channel seismic reflection lines. Both sets of profiles reveal a thick wedge of stratified material. The seismic interpretations of ten Brink and Watts [1985] and Rees et al. [1989] indicate that the moat stratigraphy generally consists of a basal pelagic unit, thick landslide units, and an upper unit consisting of turbidites and pelagic muds. The basal unit drapes and infills the rough upper surface of the oceanic crust and is interpreted as representing the formation of the volcanic loads and structural moat. Landslide units show progressive overstepping toward the flexural arch and are interpreted as the subsurface expression of older buried slumps and debris avalanches. These units are seismicity opaque and are characterized by chaotic diffractions showing little or no coherent structure. The incoherent diffractions suggest the presence of basaltic slump blocks embedded in a matrix of poorly sorted landslide debris. Overlying but also interbedded with the landslide unit is a sequence of highly reflective, continuous horizons that offset the flexural arch [ten Brink and Watts, 1985; Rees et al., 1989]. Rees et al. [1989] interpret the off-lapping unit as representing the distal facies of landslide deposits or turbidites associated with small-scale slope failures. The observed stratigraphic patterns are thought to be controlled by the combined effects of material influx rates, temporal and spatial changes in the distribution of the volcanic load, and the flexural response of the underlying lithosphere.

The anomalously thick sequence of sediments infilling the moat and the morphology of sediment surface, as illuminated by GLORIA side-scan sonar images, indicate that the source for the majority of the sediments is mass wasting of the Hawaiian edifice [Moore et al., 1989, 1994]. Moore et al. [1994] documented two styles of landslides, slumps and debris avalanches, and found that both are pervasive along the Hawaiian Ridge and adjacent seafloor (Figure 1). Although we were able to use the GLORIA data to avoid regions of shallowly buried and apparently recent magma flow, both profiles cross debris avalanches [Moore et al., 1994]. The Oahu profile crosses the northwestern edge of the Nuuauu debris avalanche, one of the largest debris avalanches mapped by Moore et al. [1989]. This debris avalanche is about 230 km long, covers 73,000 km², and extends across the flexural moat and up the flank of the arch. Individual basaltic blocks are as large as tens of kilometers in maximum dimension and apparently have been carried 50 km or more by these events. Moore et al. [1989] report scattered blocks ranging up to 1 km in size at the distal portion of the deposits. While the ages of these events are poorly known, the largest landslides are interpreted to occur late in the period of active shield growth when the volcanoes were close to their maximum size. It appears that slope failure begins early in the history of the individual volcanoes but may continue at reduced rates long after the volcanoes become dormant [Moore et al., 1989]. These landslide deposits are thought to segment the moat into a series of subbasins and to control lateral transport of sediment along the axis of the moat [Moore et al., 1989; Rees et al., 1989].

3. Seismic Data

We collected three seismic profiles (Figure 1) using an array of Bolt PAR 1500-L airguns towed at an approximate depth of 6 m below the sea surface. Chamber sizes were 29,005 cm³ for the French Frigate Shoal profile and 33,921 cm³ for the Maro Reef and Oahu profiles. The shots were received on an AMG Corp. small-diameter marine seismic streamer with two active sections (French Frigate Shoals deployment) or three active sections (Maro Reef and Oahu deployments). Each active section is 50 m in length and has 42 hydrophones at 1-m spacing. A 50-m stretch section and a 7-m weighted section were in place ahead of the first active section, while another acted as a tail section behind the second active section. The tow lead was ~150 m in length. Thus the setback between the transom to the center of the first active section was ~232 m. The guns were towed between 90 and 110 m behind the ship. Guns were fired every 20 s with a 9-s recording window. This corresponds to an average shot spacing of ~60 m but is variable depending on ship speed. Sampling rate was 1 ms, with a low cut to 10 Hz and a high cut of 150 Hz and a rolloff of 24 dB/ octave. Seismic data were recorded digitally using a system developed by the Hawaii Institute of Geophysics. Digital processing included a debias, a band-pass filter (12 to 45 Hz) and a time-varying gain (×t² or t). Although three channels are available, our interpretation is based only on the near trace. Adding the other two traces did not appear to improve the seismic image.

The Oahu profile is close to the multichannel line reported by ten Brink and Watts [1985], and our interpretation is guided by theirs. We identified five stratigraphic units, each bounded by relatively large-amplitude continuous reflectors (Figure 2). Over the flexural arch, reflectors generally can be followed laterally but often exhibit a discontinuous and hummocky character. In the deep part of the moat nearer the islands, reflectors are discontinuous and are difficult to trace laterally. While the sediment-basement contact is generally identifiable over the flexural arch, it becomes difficult to identify in the deepest
portion of the moat. Our picks in the deepest portion of the moat (distances < 100 km) are based on connecting prominent diffractions. The two-way travel times generally correspond to those reported by ten Brink and Watts [1985] with maxima of ~7.5 s. The top of the bottommost unit is bounded by a relatively continuous high-amplitude reflector. Overlying this unit are three units which we interpret in terms of mass-wasting units based on the characteristics described by both ten Brink and Watts [1985] and Rees et al. [1989]. These characteristics include progressive overstepping toward the arch, seismically opaque pods, and chaotic diffractions. The topmost unit consists of distinct parallel reflectors.

At a distance of ~150 km a region exists showing hummocky discontinuous reflectors, chaotic diffractions, and seismic opaqueness, which might be explained as the presence of basement highs, poorly sorted material from mass wasting events, basement intrusions, or faulting. This position corresponds to the location of the toe of the Ninamu debris ava-

Figure 2. (top) Oahu seismic reflection profile. The profile location is shown in Figure 1. Five stratigraphic units have been identified on the basis of relatively large amplitude continuous reflectors. A line drawing interpretation of the prominent reflectors is referenced to approximate axis of island chain. Subvertical lines are interpreted faults. Distance scale.

Figure 3. (top) Maro Reef seismic reflection profile. The line location is shown in Figure 1. Six stratigraphic units have been identified on the basis of relatively large amplitude continuous reflectors. Note the presence of basement highs where the profile crosses the Murray Fracture Zone. (bottom) A line drawing interpretation of the seismic profile. Distance scale is referenced to approximate axis of island chain.
lanche [Moore et al., 1994] unit as observed on GLORIA sidescan images. Toward the arch, reflectors are generally more coherent.

In general, reflectors are more continuous along the Maro Reef profile (Figure 3) than the Oahu profile. We subdivided the Maro Reef seismic profile into six units based on seismic character and prominent reflectors. The Maro Reef profile crosses the Murray Fracture Zone [American Association of Petroleum Geologists (AAPG), 1984], and consequently, the basement shows a great deal more relief. Two prominent basement highs appear to truncate otherwise continuous reflectors. A relatively continuous high-amplitude reflector characterizes the top of basement, although in the deepest part of the moat this reflector becomes less distinct and is picked by joining prominent diffractions. The top of the basal unit is picked as the last prominent and continuous high-amplitude reflection. Overlying this unit is a series of fairly continuous high-amplitude reflectors. These reflectors bound units containing seismically opaque pods and chaotic diffractions that we interpret as multiple landslide units. The uppermost unit consists of distinct parallel reflectors that are laterally continuous across the entire profile.

To convert from travel time to sediment thickness, we use seismic velocity models obtained from analyses of seismic refraction and sound-buoy data [Brocher and ten Brink, 1987; ten Brink and Brocher, 1988]. Relatively high seismic velocities ($V_p = 3.5 \text{ – } 4.4 \text{ km/s}$) characterize the flexural moat. The highest velocities are closest to the volcanic edifice and decrease toward the flexural arch. These velocities and two-way travel times to basement indicate a maximum sediment thickness within the moat greater than 2 km.

4. Gravity Data

To help estimate sediment thickness and to provide a check on depth to basement within the moat, we collected free-air gravity data using a NAVO Bell BGM-3 gravimeter. The gravity data are referenced to the best fitting spheroid, and high-frequency noise was filtered using a nine-point running mean. Figure 4 shows free-air gravity anomalies along the Oahu and Maro Reef profiles.

We modeled the gravitational effect of the sediment wedge using a simple Bouguer slab approximation. This simple geometry appears adequate given the large width of the moat (200 km) as compared to the depth of the density interfaces (<10 km) and the linearity of the Hawaiian chain. We used a sediment density of 2400 kg m$^{-3}$ and a crustal density of 2900 kg m$^{-3}$. The high sediment density was used because of the relatively high seismic velocities observed in the flexural moat and is similar to that used by Watts et al. [1985] in their analysis of crustal flexure. Without further iteration these densities and sediment thicknesses give reasonably good fits to the data (Figure 4). Root-mean-square misfits are 4 and 5 mGal for the Oahu and Maro Reef profiles, respectively. While an infilling density of 2400 kg m$^{-3}$ may be appropriate for the moat, it may be a bit high for sediments over the arch crest where seismic velocities are somewhat lower [Brocher and ten Brink, 1987; ten Brink and Brocher, 1988]. Both models give maximum misfits of the order of 10 mGal at the distal ends of the profiles and suggest either thinner sediments or a smaller density contrast. However, because reflectors are well imaged here, the misfits at the distal ends of the profiles are likely a result of using too small a density contrast.
5. Heat Flow Determinations

Heat flow measurements were made using the heat flow instrumentation of Woods Hole Oceanographic Institution, consisting of a steel lance 6 m long with six to seven externally mounted thermistor probes. Each thermistor probe is approximately equally spaced along the length of the lance and consists of both a thermistor and a linear heat element. This configuration allows the measurement of both equilibrium temperatures and thermal conductivity at the depth of each probe. The lance was deployed in a multipenetrator "pogo" mode [Von Herzen, 1987]. Each penetration consists of raising the probe up to 1 km off the bottom from the previous penetration, slowly moving (at 1 2 knots) the ship to the next penetration site, and letting the wire angle become nearly vertical before dropping the probe in the sediment. Once the probe is in the bottom it is left undisturbed for 15-20 min, ~6 min for the equilibrium temperature measurement and ~10 min for the thermal conductivity measurement. Power requirements are such that this instrument can be operated for periods > 60 hours. Real-time acoustic monitoring of the measurements allowed monitoring of instrument performance and data quality. Thus the number of successful penetrations is maximized before data quality degrades to the point where the lance needs to be recovered and refurbished.

The geothermal data (comprising temperatures of the thermistor probes, near-bottom water temperature, internal temperature, pressure, tilt, and the resistance of two high-stability reference thermistors) were logged at 1-s intervals during deployment and recorded in solid-state memory. The data, reduced utilizing software described by Bonneville et al. [1993], is summarized for each successful penetration in Tables A1 and A2 (Appendix A1). Each station consists of a single deployment and for our survey typically consisted of 20 30 penetrations at 1.2 km spacing.

6. Thermal Gradient

Although the thermistors do not reach complete equilibrium in the 6 min allocated to this measurement, it is sufficient time to allow extrapolation to equilibrium temperatures with an uncertainty usually better than 1 mK. These uncertainties include the differences in near-bottom water temperatures recorded both before and after penetration. Near-bottom water gradients are small. Spacing between thermistor probes is known to better than 0.5 cm, and the penetration depth of the probe into the sediments is calculated by extrapolating sediment temperature gradients to the reference bottom water temperature and accounting for the tilt of the probe. Penetrations with excessively high tilts (>30°) were discarded. Gradients are then calculated as the best fitting line to the equilibrium temperatures as a function of depth. Data points are inversely weighted by the standard deviation of the equilibrium temperature. In most cases, equilibrium temperatures increased linearly with depth, so the fit of the data to calculated gradient is good and no systematic curvature or breaks in gradient were observed. Examples of the data are shown in Figure 5a. Occa-

---

1Supporting Appendix tables are available on diskette or via Anonymous FTP from koamog.ago.org, directory APEND (Username = anonymous, Password = guest). Diskette may be ordered from American Geophysical Union, 2000 Florida Avenue, N.W., Washington DC 20036 or by phone at 800-966-2481: $15.00. Payment must accompany order.
Figure 6. Oahu (solid circles) and Maro Reef (solid triangles) thermal profiles. (a) Thermal gradient and (b) thermal conductivity determined from in situ measurements and from needle probe measurements; open circles for Oahu and open triangles for Maro Reef.

Additionally, data from thermistors were rejected because of non-penetration or probe leakage. The statistical uncertainty of the thermal gradients is computed using standard formulas [Bevington, 1969] and is a function of the uncertainties of the equilibrium temperatures, tilt and standard deviation of the equilibrium temperatures to the thermal gradient. The majority of gradients are determined from four or more equilibrium temperatures, although a few have only three thermistors (electronic Tables A1 and A2). Thermal gradients for both the Oahu and Maro Reef profiles are shown in Figure 6a.

7. Thermal Conductivity

In situ thermal conductivities were determined at the position of each thermistor probe using a continuous heating technique [Jenske and Von Herzen, 1989]. Applying a known heat input at constant rate along the length of the heater wire and recording the transient rise in sediment temperature allows the thermal conductivity to be computed [Blackwell, 1954; Jaeger, 1956]. At each penetration the reported thermal conductivity is the harmonic mean computed from the thermal conductivity of each thermistor. As with the temperature measurements, conductivity measurements that showed signs of probe leakage or those that did not penetrate were not used in the calculation of the mean conductivity. A sample of the data scatter for four penetrations is shown in Figure 5b. The overall uncertainty of the in situ conductivities is calculated from the errors of the individual measurements and standard deviation of these values from the harmonic mean. In general the uncertainty is dominated by the standard deviation of the data fit to the harmonic mean (Figure 5b). Thermal conductivities for the Maro Reef and Oahu profiles are shown in Figure 6b. A comparison of these two plots indicates the mean thermal conductivity along the Maro Reef profile is ~0.1 W m\(^{-1}\)K\(^{-1}\) greater than along the Oahu profile. Although the sediment cores were not investigated for lithology, the difference in mean thermal conductivity may be a result of different mean carbonate content of the sediments [Matsuda and Von Herzen, 1986]. The geographic variation in thermal conductivity (i.e., increasing downwedge) is consistent with those measured by Von Herzen et al. [1982]. A second point about Figure 6b is that the uncertainty in thermal conductivity is substantially larger along the Oahu profile and is attributed to greater probe to probe variability in computed thermal conductivities.

As a check on in situ thermal conductivities, we collected five gravity cores (two along the Maro profile and three along the Oahu profile, Figures 6b) so that thermal conductivities could be determined using the needle probe technique of Von...
Herzen and Maxwell [1959] Most cores were several meters in length, and measurements were made at 10-cm intervals along the cores. Measurements were performed at ambient pressure and temperature conditions and corrected to seafloor values assuming an increase in conductivity of 0.5% per kilometer of depth and a decrease in conductivity of 0.25% per degree of temperature decrease [Ratcliffe, 1960]. Thermal conductivities as a function of core depth are plotted in Figure 7. In general, variations along the core are small, and the harmonic mean is of these values is a good estimate of the mean conductivity. Although this value is determined from the gravity cores, it is generally somewhat lower than those based on in situ measurements, these values agree to within 5% and are within the uncertainties of the nearest in situ values (Figure 6b).

Variations in thermal conductivity along the length of the core may reflect systematic variations in thermal conductivity. To determine if these variations in thermal conductivity are systematic, we combined temperature and thermal conductivity to form a plot of temperature versus sum thermal resistance [Bullard, 1939]. In almost all cases the ratio of the gradient was not significantly more linear (as indicated by a chi-square parameter), and the heat flow was insignificantly different from that determined by using the harmonic mean conductivity (<1 mW m⁻²). These results together with the pattern of thermal conductivity variations along the gravity cores indicate that the harmonic mean conductivity is a good estimate of thermal conductivity over the profiles. Heat flow values are computed as the product of the thermal gradient and harmonic mean thermal conductivity. The measurement uncertainty is the square root of the sum of the squares of the statistical uncertainties in the thermal gradient and thermal conductivity.

8. Sedimentation Model

The rapid deposition of cold sediments on the seafloor leads to a transient depression in the surface heat flux until the sediments recover to thermal equilibrium. Because of the large sediment pile and catastrophic nature of sedimentation associated with slumping from the Hawaiian Islands, this correction is significant and has important implications for the interpretation of these profiles. Sediment corrections applied to marine heat flow data typically assume that sediments are deposited at a constant rate and are equilibrated to the bottom water temperature at the time of deposition. Slumps, debris avalanches, and large basaltic blocks embedded within debris flows may violate this assumption. We explore the implications of this assumption when we discuss variability in the heat flow profiles. Because the sedimentation history is complex and not well known, we have calculated a set of models spanning a range of parameter values that includes the general history of sedimentation. We calculate this correction using a one-dimensional algorithm based on the formulation of Hutchinson [1985] as implemented in a finite difference scheme by K. Wang (personal commun., 1997). This formulation includes the thermal effects of sediment compaction, variable sedimentation rates, and different thermal properties between the sediment and basement rocks. This correction is a function of the rate at which sediments thermally equilibrate and is most sensitive to the combination of sediment thickness and its rate of deposition over time.

A simple sedimentation model along the Maro Reef profile was used by Von Herzen et al. [1989] to correct heat flow determinations at their sites 4 and 5 for the effects of sedimentation. We have been guided by this approach generally but
Table 1. Sedimentation Models 2 and 3

<table>
<thead>
<tr>
<th>Age, m.y.</th>
<th>Sediment Type</th>
<th>Thickness, m</th>
</tr>
</thead>
<tbody>
<tr>
<td>0-79</td>
<td>Oahu Profile</td>
<td>P</td>
</tr>
<tr>
<td>79-81</td>
<td>volcaniclastic</td>
<td>(T-P)*0.8</td>
</tr>
<tr>
<td>81-83</td>
<td>pelagic/pelagic</td>
<td>(T-P)*0.2</td>
</tr>
<tr>
<td></td>
<td>Mako Reef Profile</td>
<td></td>
</tr>
<tr>
<td>0-85</td>
<td>pelagic</td>
<td>P</td>
</tr>
<tr>
<td>85-87</td>
<td>volcaniclastic</td>
<td>(T-P)*0.6</td>
</tr>
<tr>
<td>87-102</td>
<td>pelagic/volcaniclastic</td>
<td>(T-P)*0.4</td>
</tr>
</tbody>
</table>

T is total sediment thickness, P is thickness of basal pelagic layer. Model 1 assumes continuous sedimentation.

This model adds a negligible correction to the heat flow profiles (Table 2). For more realistic sediment models we divided the sediments into three units corresponding to the three different modes of sedimentation inferred from the seismic stratigraphy. The basal layer represents slow pelagic sedimentation with a rate of sedimentation assumed constant until the onset of volcanism at 79 and 85 Ma for Oahu and Maro Reef, respectively. The middle layer represents catastrophic sedimentation associated with volcaniclastic debris flows and is assumed deposited during a 2-Myr period of constructive volcanism. The uppermost layer represents a combination of volcaniclastic turbidites and pelagic sedimentation. In our second model we follow Von Herzen et al. [1989] and adjust the pelagic layer to a thickness equivalent to a 500 m surface layer. For sediment sections greater than this thickness we partition the remaining sediment 4/5 volcaniclastic and 1/5 pelagic for the Oahu profile, and 3/5 volcaniclastic and 2/5 pelagic for the Maro Reef profile, respectively. We use a different partition for Maro Reef because of the greater time since constructive volcanism. If, however, the total sediment thickness is ≤500 m, we assume that the sediment section is unaffected by mass wasting and use a constant rate of sedimentation over the whole time span of the sedimentation model, 83 Myr for the Oahu profile and 102 Myr for the Maro Reef profile. This occurs only at the distal end of the profiles near or over the flexural arch. In the third model we adjust the

Figure 8. Theoretical effect of sedimentation on surface heat flow. Three models are shown: steady state sedimentation, model 1 (dashed line); a basal pelagic layer of 500 m, model 2 (dash-dotted line); and a basal pelagic layer of 200 m (solid line). (a) Oahu model corresponding to sediment thickness at penetration 8.01. (b) Maro Reef models corresponding to sediment thickness at penetration 5.01.
Table 2. Corrected Heat Flow and Lateral Gradients Over Profiles

<table>
<thead>
<tr>
<th>Model 1</th>
<th>Model 2</th>
<th>Model 3</th>
</tr>
</thead>
<tbody>
<tr>
<td>( \lambda ) km</td>
<td>( q ) (s.d.), mW m(^{-2})</td>
<td>( m ) (s.d.), mW m(^{-2})</td>
</tr>
<tr>
<td>---------</td>
<td>---------</td>
<td>---------</td>
</tr>
<tr>
<td>Oahu Profile(^a)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>1</td>
<td>65.8 (12.2)</td>
<td>6.7 (1.7)</td>
</tr>
<tr>
<td>2</td>
<td>66.2 (12.5)</td>
<td>6.4 (1.7)</td>
</tr>
<tr>
<td>3</td>
<td>66.5 (12.3)</td>
<td>6.1 (1.7)</td>
</tr>
<tr>
<td>Maro Reef Profile(^b)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>1</td>
<td>61.9 (5.0)</td>
<td>2.6 (1.4)</td>
</tr>
<tr>
<td>2</td>
<td>62.4 (5.0)</td>
<td>2.3 (1.4)</td>
</tr>
<tr>
<td>3</td>
<td>62.6 (5.0)</td>
<td>2.1 (1.4)</td>
</tr>
</tbody>
</table>

Model 1 corresponds to continuous sedimentation, models 2 and 3 are described in Table 1 and in text. \( \lambda \) = compaction constant, \( \phi = \phi_0 \exp (-z/\lambda) \), \( q \) is the mean heat flow, and \( m \) is the best fitting linear trend in heat flow across the profile.

\(^a\)Mean uncorrected heat flow 63.9 (11.9) mW m\(^{-2}\), linear trend 7.7 (1.7) mW m\(^{-2}\) /100 km.

\(^b\)Mean uncorrected heat flow 59.8 (5.0) mW m\(^{-2}\), linear trend 3.4 (1.4) mW m\(^{-2}\) /100 km.

9. Heat Flow Profiles

9.1. Oahu Profile

Heat flow values both uncorrected and corrected for the thermal influence of sedimentation are shown in Figure 10. The mean heat flow value along the Oahu profile is 63.9 mW m\(^{-2}\) (s.d. 11.9 mW m\(^{-2}\)) without sediment corrections and 74.2 mW m\(^{-2}\) (s.d. 14.6 mW m\(^{-2}\)) with sediment corrections, based on 101 successful measurements. The measurement uncertainty is 3 mW m\(^{-2}\), indicating that variability about the mean is significant. The uncorrected heat flow values show a slight but discernable trend of lower values toward the island chain. All of the investigated sediment corrections diminish this trend. For our preferred model the mean sediment correction is 10.3 mW m\(^{-2}\), and while the sediment correction becomes larger over the deepest part of the moat, it does not significantly change the pattern of anomalies across the flexural moat and arch.

The greatest variability comes from a series of bimodal single penetration anomalies. At distances of 100 to 130 km (Figure 10) a series of single penetration high values were measured. The amplitude of these peaks is about 30 mW m\(^{-2}\) (50%) above the mean. In the area where the Oahu profile crosses the Niihau debris flow the substrate seems anomalously hard, as there are a number of partial and failed penetrations (denoted by a horizontal bar in Figure 10). This region is also characterized by poor seismic return. Penetrations 10.04 and 10.05 have tilts of 25° and 26°, respectively. Outboard of this region are a series of heat flow lows between about 160 and 180 km with values ~20 mW m\(^{-2}\) (30%) below the mean. In both cases, thermal gradients for these anomalous heat flow values are linear.

9.2. Maro Reef Profile

The mean heat flow value along the Maro Reef profile is 59.8 mW m\(^{-2}\) (s.d. 5.0 mW m\(^{-2}\)) without sediment corrections and 64.0 mW m\(^{-2}\) (s.d. 4.9 mW m\(^{-2}\)) with sediment corrections, based on 83 successful penetrations (Figure 11). The meas-

\[ \phi(z) = \phi_0 \exp \left( -\frac{z}{\lambda} \right) \]
Figure 9. Theoretical effect of sedimentation on surface heat flow along each profile and as a function of compaction constant ($\lambda$). (a) Oahu profile model 2, (b) Oahu profile model 3, (c) Maro Reef profile model 2, and (d) Maro Reef profile model 3.

Measurement uncertainty is 2 mW m$^{-2}$, slightly less than the uncertainty along the Oahu profile because the thermal conductivities, in general, are better determined. Maximum peak to peak variability along the profile is of the order of 20 mW m$^{-2}$. As with the Oahu profile, including the effects of sedimentation removes a small positive trend in the unadjusted heat flow determinations. The mean sediment correction is 4 mW m$^{-2}$ and varies between 7 and 1 mW m$^{-2}$ but does not significantly change the pattern of anomalies across the profile. Except near the beginning and end of the profile (Figure 6), thermal conductivity is relatively constant along the line and heat flow variations reflect variations in thermal gradient.

Significant variations in heat flow along this profile occur at distances of 100-120 km and 150-180 km. Both regions are inboard of prominent basement highs associated with the Murray Fracture Zone. We infer that the strike of these features may be subparallel to the fracture zone lineation. At a distance of 190-200 km, another region of variability exists that corresponds to the flexural arch where small seamounts crop out. The single anomalous high value at 200 km does not seem to be correlated with any clear bathymetric or basement feature.

The 1997 Maro Reef profile lies along the 1980 profile measured by Von Herzen et al. [1989] and overlays their site 4 (triangle, Figure 11). This site is composed of 14 penetra-
Figure 10. Heat flow data, Oahu profile. (a) Heat flow determinations uncorrected (open circles) and corrected for the effects of sedimentation (solid circles). (b) Depth-converted seismic reflection profile. Hatched region shows thickness of debris flow used in sediment correction.

Measurements parallel to our profile, and six penetrations perpendicular to the profile. Heat flow determinations for the 1989 profile are based on nine values of in situ conductivity. The remaining heat flow determinations use a value of 0.96 W m$^{-1}$K$^{-1}$ based on the average value. In the area of overlap thermal conductivity values for the 1997 survey are relatively uniform with a mean of 0.95 W m$^{-1}$K$^{-1}$. The two profiles are offset by ~5 km (Figure 12, inset). A detailed comparison of uncorrected heat flow values for these two data sets shows general agreement. In the interval of overlap the mean uncorrected heat flow and standard deviation from this survey are 57.4 ± 4.7 mW m$^{-2}$, based on 13 points, and from Von Herzen et al. [1989] are 57.8 ± 4.6 mW m$^{-2}$, based on 20 points. However, Figure 12 also emphasizes small-scale (10 km to a few tens of

Figure 11. Heat flow data, Maro Reef profile. (a) Heat flow determinations uncorrected (open circles) and corrected for the effects of sedimentation (solid circles). (b) Depth-converted seismic reflection profile. Hatched region shows thickness of debris flow used in sediment correction.
Figure 12. Comparison of Maro Reef heat flow profile (circles, this study) and individual heat flow penetrations (triangles) reported by Von Herzen et al. [1989]. Inset shows detailed location map. Over the interval of overlap the mean uncorrected heat flow and standard deviation are 57.4 ±4.7 mW m⁻², based on 13 points, and from Von Herzen et al. [1989] are 57.8±4.6 mW m⁻², based on 20 points.

kilometers) variations which may reflect complex three-dimensional influences on heat flow. At a distance of 148 km we made four repeat measurements at the same location to estimate the instrumental variability. The standard deviation of these four heat flow determinations is 0.4 mW m⁻²; thus the variations between penetrations and between surveys are significant. The observation that we can reproduce heat flow at the same site to within 0.4 mW m⁻² implies that the scatter in the 10-km survey box averages [Von Herzen et al., 1989] are real and not measurement error. Indeed, the scatter found in the 10-km survey box [Von Herzen et al., 1989] has the same magnitude as that along the entire profile. The fact that averaging over larger distances with much more data does not reduce the standard deviation implies that these sources of variability are real, and the magnitude implies that it is important.

10. Discussion

Two important facets of these heat flow profiles are the scatter and the mean. Figure 13 shows power density functions for each heat flow profile. Variability along the Oahu profile is significantly greater than along the Maro Reef profile. The sampling frequency for both profiles is ~1 km (1.8 km). The power density function for the Oahu profile shows a clear peak at 5 to 6 km, before dropping off at shorter wavelengths. A well-developed spectral peak exists for wavelengths of 30-70 km. Along the Maro Reef profile, the power density function shows a modest peak at ~15 km before trailing off. At wavelengths of 4 km and less the spectrum is statistically indistinguishable from zero. Each spectral wave band contributes <3 mW m⁻² to the overall variations in heat flow. To highlight areas of heat flow above and below the average, Figure 14 shows a five-point running average for each profile, corresponding to a window length of ~10 km.

The observed variability in these heat flow profiles could be caused by (1) heat from a recent volcanic intrusion, (2) refractive heat conduction, (3) an inaccurate sediment correction, or (4) buoyancy-driven fluid flow. While isolated volcanic intrusions are possible, we do not believe that recent volcanic intrusions are generally the source of variability. Instead, variability along both profiles is greatest over the most. The effects of refractive heat conduction are difficult to conclusively eliminate. Harris et al. [this issue] investigate the effects of heat retraction due to basement relief with a finite element

Figure 13. Spectral analysis of Oahu heat flow data (circles) and Maro Reef heat flow data (triangles). Vertical scale shows magnitude of harmonic coefficient and uncertainty. Spectral smoothing was accomplished through windowing four subsections of the original series.
model. To maximize the thermal conductivity contrast across the sediment basement interface, we set the thermal conductivity of the sediment and basement to 1 and 3 W m⁻¹ K⁻¹, respectively. Because of the thick sedimentary cover over the prominent basement relief, conductive heat flow variations along these profiles are negligible. Heat conduction models along the Maro Reef profile, where variability in basement relief is greatest and sediment cover is relatively thin, indicate that enough sediment is present to attenuate the effects of heat refraction at the surface. If conductivity contrasts in basement had an important refraction effect, then variability along the Maro Reef profile should be greater than along the Oahu profile because of the greater basement relief at Maro Reef. Further, we note that we do not observe the correlation between basement relief and heat flow variability that would be produced by conductive refraction. Additionally, contrasts in thermal conductivity within the debris avalanches, for example, between large basaltic blocks and the matrix in which they are embedded, are not expected to be as large as those used in the model. Similarly, local erosion or deposition of sediments is also difficult to conclusively rule out, although the results of our sedimentation models indicate that it would be difficult to produce the observed variability. Seismic reflection recordings from the uppermost sediments are generally continuous.

Catastrophic sedimentation associated with large landslides of the Hawaiian archipelagic apron introduces perturbations to thermal conditions that could potentially produce some of the observed heat flow variability. The site of greatest variability is along the Oahu profile where it crosses the Nuuanu debris avalanche between 100 and 150 km. The presence of landslide deposits in general and basaltic blocks embedded in the debris flow in particular likely violates the assumption of steady state sedimentation at temperatures equilibrated with bottom water. If these units are deposited at temperatures representative of their preslide conditions (i.e., warmer than bottom water), they might act as transient heat sources. We explore the thermal significance of this effect by using elementary heat conduction theory originally developed to describe temperatures in country rock after being intruded [Jaeger, 1964, pp. 446-449]. This simple transient theory does not incorporate the temperature effects of solidification and therefore only depends on the geometry, initial temperature difference between the source and the surrounding matrix, and thermophysical rock properties. We use a constant temperature upper boundary condition consistent with a constant bottom water condition. Figure 15 shows the transient heat flux into an initially isothermal sheet (representative of a slump or landslide) and above the center of a 1 km³ cube (representative of a basaltic block embedded in a debris flow). Graphs of temperature distributions are given by Jaeger [1964]. In this model the temperature difference between the slide material and the surrounding matrix is 10°C; in one case we assume the material is buried by 100 m of sediment and in another by 10 m of sediment. These are admittedly end-member examples, but they illustrate that while the increase in heat flow is extreme for shallow sources, the transient decays very quickly. Unless the Nuuanu debris avalanche is very young (< 5000 years), this process seems unlikely to be responsible for the heat flow variability.

The cross-swell survey of Von Herzen et al. [1989] indicated the lack of a significant positive heat flow anomaly; the most anomalous points are in the flexural moat closest to the volcanic edifice. This pattern hints at the possibility that fluid circulation might be masking a heat flow anomaly associated
Table 3. Mean heat flow determinations and estimates

<table>
<thead>
<tr>
<th>Line</th>
<th>Uncorrected</th>
<th>Corrected</th>
<th>Uncorrected</th>
<th>Corrected</th>
<th>q''</th>
</tr>
</thead>
<tbody>
<tr>
<td>Oahu</td>
<td>63.9 (11.9)</td>
<td>74.7 (14.6)</td>
<td>---</td>
<td>---</td>
<td>62 (17)</td>
</tr>
<tr>
<td>Maro</td>
<td>59.8 (5.0)</td>
<td>64.0 (4.9)</td>
<td>57.8 (3.4)</td>
<td>59.3 (3.5)</td>
<td>49 (7)</td>
</tr>
</tbody>
</table>

\*Mean heat flow for values uncorrected and corrected for the thermal effects of sedimentation, using model 3 (Table 2) and a compaction factor of 2 km.
\*Offswell values are based on stations 1 and 8 from the cross-swell survey of Von Herzen et al. [1989] for the Maro Reef profile. Values are uncorrected and corrected for the thermal effects of sedimentation.
\*Values of global mean oceanic heat flow binned in 2-Myr intervals [Stein and Stein, 1993]. Values are based on crustal ages of 83 and 102 Ma for Oahu and Maro Reef, respectively.

with lithospheric reheating. Harris et al. [this issue] investigate the thermal signature of simple moat wide flow models designed to explore this hypothesis. In these models, lateral thermal gradients associated with the edifice provide buoyancy to drive flow from the flexural arch toward the volcanic edifice. A common characteristic of these basin-scale models is that flow is strongest in the position of maximum horizontal temperature gradient and can result in a pronounced but broad heat flow low. Resulting heat flow profiles show strong trends with decreasing heat flow islandward that could potentially hide a heat flow anomaly associated with lithospheric reheating. However, since the mean heat flow is nearly the same both over and off the archipelagic apron for both profiles (Figures 10 and 11), there is probably no thermally significant large-scale circulation (unless we have errors in the sediment correction and/or this correction compensates for large-scale fluid flow effects). The greater local variability over the apron at Oahu suggests fluid circulation at these scales.

Along the Oahu profile a series of heat flow highs is located in the position where the Oahu profile crosses the Nanuau debris flow, and just off profile, SeaBeam data and side echoes indicate a series of basaltic outcrops. We interpret these outcrops as basaltic blocks carried into position by debris avalanches. These blocks may produce permeability pathways capable of hosting significant fluid flow. At ~175 km a broad heat flow low (Figure 14) may indicate an area of recharge. It is interesting to note that the standard deviation of the five-point running average is greater over heat flow highs than heat flow lows and may be indicative of the diffusive nature of recharge and the more discrete nature of discharge.

Heat flow variability along the Maro Reef profile is considerably less than along the Oahu profile. However, the rough basement relief along the Maro Reef profile provides greater potential to image fluid flow in the basement. Along the Maro Reef profile between ~120 and 160 km, two prominent basement highs and the intervening low appear modestly correlated with heat flow (Figure 11). If strong local convection occurred within the basement, it could produce a positive correlation between basement relief and heat flow. However, the heat flow peak at ~115 km is offset (~10 km) toward the volcanic edifice. One mechanism for displacing the heat flow high is through lateral fluid flow within the archipelagic apron. Over the prominent basement high at a distance of 120 km, the sediments are ~600 m thick. Assuming heat is transported vertically through the apron by conduction, a thermal length calculation indicates that it would take ~2800 years for a thermal anomaly to reach the surface. Over the same time interval the heat anomaly needs to be transported ~10 km toward the edifice. This requires extremely high Darcy velocities of the order meters per year. Thus, while there may be fluid flow in the upper oceanic crust, we do not observe a strong correlation between basement relief and heat flow variability as is observed at young sedimented ridge flank settings [e.g., Davis et al., 1989]. Instead, our preferred interpretation is that the fluid flow within the archipelagic apron and possibly in the upper oceanic crust is responsible for the variability in heat flow.

The relatively high frequency of heat flow variability along the Oahu profile suggests that if fluid flow is present, it may lie within the archipelagic sediments. Further, the greater variability along the Oahu profile suggests that fluid flow may be stronger along the Oahu profile, possibly because the Oahu edifice is larger than that at Maro Reef and the profile at Oahu skirts alongside the Nanuau debris flow, which may provide greater permeability capable of hosting fluid flow. The edifice and mass wasting at Oahu is geologically younger than at Maro Reef, and this aspect may contribute to greater variability, either through enhanced permeability or thermal transients. A greater accumulation of pelagic sediments along the Maro Reef profile may also diminish the thermal effects of fluid flow by scaling zones of recharge and/or discharge.

Within the scatter the mean values for these heat flow profiles (both uncorrected and corrected for the effects of sedimentation) are not significantly different from each other (Table 3). Compared to global mean values for crust of the same age [Stein and Stein, 1993], the heat flow mean at Oahu is within one standard deviation of the global mean, while the heat flow mean at Maro Reef is slightly greater than one standard deviation of the global mean. However, offswell values at Maro Reef [Von Herzen et al., 1989] provide a local reference for evaluating anomalous heat flow. These offswell values are within the uncertainty of the mean heat flow for Maro Reef and therefore do not support an interpretation of anomalous heat flow over the Hawaiian Swell at Maro Reef. These observations are somewhat inconsistent with simple models of conductive reheating of the oceanic crust. Von Herzen et al. [1989] reviewed models proposed to explain the origin of the Hawaiian Swell and the critical observations that these models
need to explain. The rapid uplift, predictable subsidence, and apparent compensation of the swell at depths within or below the lower part of thermal plate are most easily explained in terms of lithospheric reheating [Detrick and Crough, 1978]. However, this model predicts a significant heat flow anomaly downstream of the hot spot and, when coupled with other geophysical data [Crough, 1978; Detrick and Crough, 1978; McNutt, 1987], suggests that its maximum should be near the position of Maro Reef. In the absence of a thermal anomaly, Von Herzen et al. [1989] preferred dynamic support of the Hawaiian Swell by plume-related convection as discussed by Courtney and White [1986]. In this model the topographic swell is supported by a combination of dynamic convective uplift and the thermal expansion of the overlying conductive lid. Robinson and Parsons [1988] applied this model to the Hawaiian Swell and predicted heat flow anomalies of 9-10 mW m$^{-2}$ in their best fitting model. These anomalies are larger than observed. To reconcile this difference of 9-10 mW m$^{-2}$, Von Herzen et al. [1989] argued that the latent heat of fusion could consume enough energy and buffer maximum temperatures so that no heat flow anomaly would be detected at the surface. While this model explains many of the observations and reconciles the presence of a bathymetric anomaly with the lack of a heat flow anomaly, the calculated subsidence rate for the Hawaiian Swell is smaller than observed.

Another explanation for the lack of a heat flow anomaly consistent with reheating models is that heat flow values within the archipelagic apron may reflect shallow disturbances to conductive heat flow. The most likely candidate source is fluid flow. Because fluid flow has the potential to obscure basal heat flow patterns associated with the Hawaiian hot spot, this interpretation does not bode well for capturing the form and magnitude of this signature. If the size of volcanic edifices (and midplate swells) are an indication of both the anomalous basal heat flux and the prevalence of debris flows, then capturing the maximum basal heat flux may be confined by fluid flow within archipelagic aprons, just the place measurements would be the most definitive. The sedimentation history and resulting permeability within the archipelagic apron are complex, and a more complete and detailed knowledge of the geologic environment will likely be required to sort these signals out.

11. Conclusions

On the basis of 180 new heat flow measurements collocated with seismic reflection lines we make the following conclusions.

1. The mean value of heat flow along the Oahu profile is $63.9 \pm 11.9$ mW m$^{-2}$ ($74.2 \pm 14.6$ mW m$^{-2}$ with sediment corrections) with an measurement uncertainty of 3 mW m$^{-2}$. The mean value of heat flow along the Maro Reef profile is $59.8 \pm 5.0$ mW m$^{-2}$ ($64.0 \pm 4.9$ mW m$^{-2}$ with sediment corrections) with a measurement uncertainty of 2 mW m$^{-2}$.

2. Heat flow anomalies along both the Oahu and Maro Reef profiles are negligible with respect to offswell values of heat flow [Von Herzen et al., 1989] and mean values of oceanic heat flow [Stein and Stein, 1993]. The absence of a significant heat flow anomaly along the Maro Reef profile is consistent with the results of Von Herzen et al. [1989]. Overall, these results are inconsistent with simple conductive models of lithospheric reheating.

3. Significant short-wavelength variability exists in both profiles and is greater along the Oahu profile. Heat flow variability is likely related to shallow processes such as fluid flow within the archipelagic apron.

Acknowledgments. We thank the officers and crew of the RV Roger Revelle for shipboard operations and station keeping. We gratefully acknowledge the assistance of J. Bottles, J. Gage, R.S. Hallinan, R. Kelly, J. Kirklin, R. Lammer, S. Naidu, G. Pelletier, J. Sclater, and I. Young for assistance with heat flow instrumentation and measurements. We thank Chris Gurnis for providing us with unpublished U.S. Geological Survey GLORIA survey data to help in selection of the profile locations. The work benefited from constructive reviews by D. Denning, A. Fisher, and H. Villinger and was supported by the Division of Ocean Sciences, NSF grants OCE-9996275 (MBARI) and OCE-9626037 (WHOI). WHOI contribution number 10207.

References


G. Garven, Department of Earth and Planetary Sciences, Johns Hopkins University, Baltimore, MD 21218. (gavrov@jhu.edu)

R. N. Harris, Department of Geology and Geophysics, University of Utah, Salt Lake City, UT 84102. (nharris@mines.utah.edu)

K. Jordahl and M.K. McNutt, Monterey Bay Aquarium Research Institute, 7700 Sausalito Road, Moss Landing, CA 95039. (keks@mbari.org; mcnutt@mbari.org)

R. P. Von Herzen, Woods Hole Oceanographic Institution, Woods Hole, MA 02541 (rvon@whoi.edu)

(Received August 30, 1999; revised March 30, 2000; accepted May 4, 2000.)