Radiative Temperature Measurements at Kupaianaha
Lava Lake, Kilauea Volcano, Hawaii

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Field spectroradiometer data in the wavelength range of 0.4-2.5 µm and spectral resolution of 1-5 nm have been used to compute the radiative temperature of the surface of Kupaianaha lava lake, Kilauea Volcano, Hawaii. Two sets of observations (a total of 120 spectra) were made on October 12, 1987, and January 23, 1988, when the lava lake was in a period of active overturning. The area of the surface for which temperatures were measured was 0.23-0.55 m². Two numerical models of two and three components have been used to match the measured radiant flux ratios and to describe the surface of the lava pond in terms of radiant area and temperature. Three stages of activity on the lake surface are identified: Stage 1, characterized by magma fountaining and overturning events exhibited the hottest crustal temperatures (180-572°C) and the largest fractional hot areas (> 10⁻³). Stage 1 average flux densities were ~2.2 x 10⁴ W/m², the highest recorded for the three stages of activity on either day. The largest radiant area of fresh magma was 29% at 1100°C, while cooling from magmatic temperatures to newly formed crust at 790°C took place in a matter of seconds. Stage 2, marked by rifting events between plates of crust, exhibited crustal temperatures between 100 and 340°C with fractional hot areas at least an order of magnitude lower than those found for stage 1. Average flux densities calculated for three examples of stage 2 activity were 5.3 x 10³ W/m². Stage 3, which was quiescent periods when the lake was covered by a thick crust, dominated the activity of the lake both temporally and spatially over 90% of the time. The characteristic crustal temperature of stage 3 was 80-345°C with most solutions near 200-300°C and fractional hot areas of ≤ 10⁻³ of the viewing area. Average flux densities for stage 3 were 4.9 x 10³ W/m². For many stage 3 examples, a two-component model was sufficient to describe the spectral data; however, for almost all of the stage 1 and 2 examples and the remainder of the stage 3 examples a three-component model was required. These determinations of lava temperature and radiant area have relevance for satellite and airborne measurements of the thermal characteristics of active volcanoes and indicate that temporal variability of the thermal output of lava lakes occurs on the time scale of seconds to minutes.

INTRODUCTION

Recent advances in the remote thermal investigation of active volcanoes have demonstrated the ability of near-infrared and infrared sensors to estimate the radiative temperature of lava lakes, lava flows, and fumaroles. The instruments used to measure the radiative flux in such studies fall into two categories: satellite-borne sensors, primarily the Landsat thematic mapper (TM) [e.g., Rothery et al., 1988; Glaze et al., 1989; Pieri et al., 1990; Oppenheimer, 1991] and the advanced very high resolution radiometer (AVHRR) [Wiesnet and D'Aguanno, 1982], and field portable single-channel radiometers, such as the Minolta/Land Cyclops infrared thermometers [Jones et al., 1990; Oppenheimer and Rothery, 1991; Oppenheimer, 1991].

Satellite sensors have been particularly useful in advancing the understanding of regional characteristics of active volcanoes. Rothery et al. [1988] pioneered the adaptation of the "dual-band" method developed by Matson and Dozier [1981] and Dozier [1981] of separating two subpixel thermal radiators to LandSat TM data. For example, analysis of a January 1986 TM image of Erta Ale volcano in Ethiopia [Rothery et al., 1988] gave a dual-band solution of 800-1000°C occupying 0.2-0.04% of a pixel for a 60-m-diameter lava lake. Using the same dual-band technique, Glaze et al. [1989] attempted to quantify the energy budget of Lascar volcano, Chile, using Landsat TM data. They recognized that the radiation emitting from the cooler background contributes much more to the overall flux of a radiant pixel than the fractionally small hot, magmatic area. Since the total energy budget increases as T⁴, it is thus very important to choose a background temperature carefully.

To date, ground truth measurements necessary to confirm many of the observations reached using satellite data are surprisingly sparse. The analysis of Oppenheimer and Rothery [1991] is a notable exception, in that they calculate the radiative energy budget of a fumarole at Vulcano, Italy, using the Stephan-Boltzmann relationship and brightness temperatures generated by a Minolta/Land Cyclops thermometer. It was determined that the radiant energy of the fumarole would have been insufficient to produce a noticeable effect in a Landsat TM pixel.

While instantaneous regional views of hot volcanic surfaces have been obtained by satellite observations, and some ground truth brightness temperatures for small area volcanic surfaces have been obtained using single-channel radiometers, critical questions remain to be explored; e.g., what are the radiative components of an active lava lake as a function of temperature, and how do these components fluctuate with time? TM observations provide estimates of lava temperatures and the percentage of the image pixel that is filled by material at this temperature but lack the temporal context or the spatial variability of dynamic lava lakes. Under optimum conditions, a daytime image of any particular volcano can only be obtained from Landsat once every 16 days, so that only approximations to the thermal characteristics of any volcano can be derived [Glaze et al., 1989]. In addition, all Landsat studies of active

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volcanoes are severely limited because only one pair of bands may be used at one time for identifying multiple subpixel areas at different radiative temperatures. For the most part, TM bands 5 (1.55-1.75 μm) and 7 (2.08-2.35 μm), and under extremely limited conditions, bands 4 (0.76-0.90 μm) and 5 [Oppenheimer, 1991], may be used to calculate the radiative response of most volcanic areas [Rothery et al., 1988] so that even a simple solution to single pixel unmixing problems [Dzoyer, 1981] cannot be derived without assuming at least one of the three independent variables (the temperatures of the hot and cool radiators, and the fraction of the pixel occupied by either the hot or cool radiator). Furthermore, band widths for the TM are very wide (0.20 μm for band 5 and 0.27 μm for band 7), which limits the accuracy of the temperature determination due to the uncertainty in the distribution of flux over the particular band width.

In order to address the temporal and spatial distribution of temperatures on an active lava lake, observations were made in 1987 and 1988 of the Kupaianaha lava lake, Hawaii, using a commercial Geophysical Environmental Research (GER) visible/infrared dual-beam spectroradiometer [Flynn, 1992]. This instrument collects 1000 data points in the 0.35-5.00 μm spectral range with a spectral resolution of 1.5 nm. The GER IRIS Mark IV Spectroradiometer is a dual-beam system (with a fixed 3° angular separation between beams) so that two targets may be viewed simultaneously. The spectral window from 0.35 to 3.0 μm is covered in one continuous scan, during which time two spectra are acquired, one for each field of view. The instrument is equipped with a viewing window and superimposed grid which allows one to center the spectroradiometer on a target using the same optics which are then used by the instrument to take a measurement. During spectrum acquisition, gain is automatically adjusted by a microprocessor to accommodate especially strong or weak signals. Scan rate (between 30 s and 2 min for a single pair of spectra) can be varied by the operator [Geophysical Environmental Research Corporation, 1986]. Almost all of the data reported below have 12-bit precision and each measurement took ~30 s. Two pairs of spectra were collected at 15-bit precision over 90 s and one pair of spectra have 14-bit precision measured over 70 s [Flynn, 1992].

Use of this instrument represents a significant improvement over previous remote sensing techniques for several reasons: (1) The very large number of data points (the GER spectroradiometer collects ~200 usable data points within atmospheric windows versus only 2 for the Landsat TM); (2) the increased dynamic range of the instrument (GER ~40,000 versus 256 for Landsat TM); and (3) the greater spectral resolution (~56 times that of Landsat TM).

These three characteristics of the GER spectroradiometer allow the flux distribution of the lava lake to be much more precisely determined than is possible with spaceborne sensors. In addition, the ability to make high spatial resolution measurements (< 0.6 m²) at frequent resampling (~1 min) intervals enabled us to correlate the measured thermal characteristics with visual observations of the dynamics of the lake surface. Observations made over a time period of several hours also provide insights into the temporal variability of the volcano at a time scale unobtainable by spaceborne sensors.

**Field Method**

Most active lava lakes (e.g., Erta Ale, Erebus, Nyiragongo) are almost inaccessible, either due to political reasons or geographic location. An exception was the Kupaianaha lava lake, Hawaii (Figure 1), which was continuously active from July 1986 until August 1990. For more than 4 years, the Kupaianaha lava lake experienced constant semiperiodic overturning, followed by thick, stable, immobile crust formation, making it an excellent field site for radiative temperature measurements. Periods when the lake surface was covered by a slow-moving crust and periods of vigorous magma fountaining provide important end-members for this thermal study.

Field spectra were collected of the active Kupaianaha lava lake on October 12, 1987, and on January 23, 1988. On October 12, 1987, the lava lake was ~50 m in diameter with a small, roughly rectangular, 30 m x 10 m channel. The lake surface was 1–2 m below the crater rim, and the field spectroradiometer was mounted on a 1.5-m-high tripod which was situated on the edge of a ~6-m-high ledge. The instrument was aimed so that the field of view (1.5° x 3.0°) of each of the two beams encompassed relatively small (~0.23 m² each) areas of the lake surface. Twenty spectra were collected over a time period of 1.5 hours. One hundred spectra were collected on January 23, 1988, when the lake was ~50 m in diameter, but the channel had partially crusted over in the interim period. The surface of the lake was now ~10 m below the crater rim resulting in a ~0.55 m² field of view for each of the two beams. The viewing angle of the lake surface was 55° from the vertical on both occasions. Because the lake completely filled the field of view, the values of distance from the spectroradiometer to the lake surface and the area of the lake surface do not enter into the analysis of the blackbody temperatures. These spectral observations covered a wide range of activity of the lake surface: some spectra were obtained when the lake was in a quiescent state, others when small (~2-5 m high) fountaining was taking place, and others when the lake surface was vigorously overturning. An example of the raw data of the quiescent lake surface, corrected for automatic gain changes and plotted with respect to wavelength, is given in Figure 2a.

Analysis of the concurrent lake surface video shows that Kupaianaha exhibited three stages of activity. Stage 1, which lasted only a few minutes at a time, was dominated by active and rapid rising (up to 5 m in width and from visual estimation up to 5-10 cm deep) punctuated by the effervescence of gas bubbles and small (~2-5 m high) fountaining events as segments of the lake surface subducted. Stage 2 occurred when the surface began to form long (many meters), but narrow (0.5-2 m) fissures as the crust slowly pulled apart. During stage 2, the surface was cut by several opening rifts which exposed hotter material that quickly solidified. Stage 3, a quiescent stage, which could last up to 30 min at a time, was characterized by a thick, glassy crust moving slowly across the surface.

Using the video, it was possible to study the surface of the lava lake and estimate levels of activity in terms of the three stages identified above. More than 90% of the time, Kupaianaha exhibited a state which could be most accurately described as being ≤ 1% stage 1, ≤ 10% stage 2, and ≥ 90% stage 3. During these periods, stage 1 fountaining or overturning episodes would frequently occur in very limited areas along the perimeter of the lava lake, with narrow rifts (stage 2) developing between large (up to few hundred square meters), rigid plates (stage 3) dominating the rest of the lake surface. During periods of very sporadic (tens of minutes between episodes), increased activity, one plate would subduct beneath another, exposing fresh magma; or two plates would
collide and deform one another until one cracked, again exposing fresh magma. On rare occasions, a magma fountain would erupt onto the surface from a rift between two plates.

Lake surface velocity measurements were approximated from the video data (Figure 3). Distances were estimated from side view pictures of the lake by scaling the size of people operating the spectroradiometer. Velocity values were checked by the timing of movement of objects on the lake surface. Data collected on October 12, 1987 (Figure 3), illustrate the stochastic nature of the motion. The highest "crustal" velocity observed was -18 cm/s for a fresh magma flowing across older crust. Large plates of crust would frequently impede the forward movement of the lake surface. Comparable data were also collected on January 23, 1988 (Figure 3). The video coverage over both days suggests that the ~50-m-wide lake overturned every 20-30 min.

DATA CALIBRATION

In order to produce accurate flux ratios from the GER spectral measurements a number of calibrations were required. The raw data numbers (DNraw) produced by the GER spectroradiometer are influenced by several factors: the flux emitted by the lava lake surface, the solar flux reflected from the lake surface, the atmospheric transmission between the lake surface and the spectroradiometer, the atmospheric transmission from the top of the atmosphere to the lake surface (which affects the reflected solar component), the instrument response function which relates incident flux as a function of wavelength to output DN, and the dark response which is the digital signal produced with no flux entering the instrument. Equation (1) shows this relationship:

$$DN_{raw} = (F_e \times \tau_{ls} + F_r \times \tau_{als}) \times G_i + D$$  (1)

where DNraw is the raw data number in counts; F_e is the emitted flux from the lake surface in W cm^{-2} \mu m^{-1} sr^{-1}; G_i is the instrument response in counts \times (W cm^{-2} \mu m^{-1} sr^{-1})^{-1}; \tau_{ls} is the atmospheric transmission from lake to spectroradiometer; F_r is the reflected solar flux; \tau_{als} is the transmission from the top of the atmosphere to the lake and from the lake to the spectroradiometer; and D is the dark response. We neglect scattered light in this analysis. Our temperature calculations rely upon accurate estimates of the flux ratios from band to band and so each term in the equation must be accounted for and corrected before proceeding.

The dark measurement is a straightforward calibration accomplished by covering the instrument aperture and not allowing any incident flux. This eliminates all terms but D. This value as a function of wavelength was subtracted from all measurements.

The instrument response, G_i in equation (1), was determined by measuring the response of the spectroradiometer to a calibrated spectral source. We used two calibrated blackbody sources: an EG&G Gamma Scientific Incorporated RS10A tungsten-halogen lamp with an RS-70-2 Diffuser and an Infrared Industries, Inc. Model IR-463 Source with a Model 101C Temperature Controller. Two sources were necessary because the wide spectral response of the GER spectroradiometer did not allow sufficient signal to noise over the entire wavelength region.
Fig. 2. Examples of raw data, corrected for automatic gain change, collected with the spectroradiometer at the Kupaianaha lava lake. Horizontal axis plots wavelength between 400 and 2500 nm. Vertical axis is in number of counts, which varied for each spectrum that was collected due to the automatic gain feature of the GER spectrometer. Although data were collected to 3.0 \( \mu \text{m} \), large water and carbon dioxide absorption features occupying 2.5-3.0 \( \mu \text{m} \) severely limited the usefulness of this data for this study. (a) The lake was in a quiescent state (stage 3) when these data were collected on October 12, 1987. (b) The field of view contained a narrow rift between plates of crust (stage 2) collected on October 12, 1987. (c) An example of stage 1 activity, collected on January 23, 1988. The possible causes of "flickering" in the shape of the spectrum at both a small-scale (a few counts) and large-scale (thousands of counts) between 0.82-2.10 \( \mu \text{m} \) is interpreted to be due to temporal variations in the level of activity of the lake during the time interval that the spectrum was collected (~30 s) (d) Spectrum of stage 1 and stage 2 activity occurring almost at the same time. The rapid increase in flux at 1.6 \( \mu \text{m} \) represents the arrival of an active rift (stage 2) in the spectroradiometer's field of view. Saturation of the instrument detectors occurred at 2.00-2.08 \( \mu \text{m} \) due to erupted magma (stage 1) passing through the field of view. This is an excellent example of the transient nature of stage 1 events.

The resultant spectrum is thus corrected for instrument response and the variable effect of reflected sunlight, which generally decreases in importance with increasing wavelength, and is in a form suitable for processing via the numerical model described below. Particularly for the examples of thick-crusted, stage 3 activity, we found that at wavelengths < 1.3 \( \mu \text{m} \) more than 90% of the total flux was due to reflected sunlight. Since the amount of radiation emitted by the lake surface is not comparably significant at wavelengths such as this, and calculations using these data may potentially include large errors, these data were not used in the numerical models for stage 3 enumerated below.

**NUMERICAL MODEL**

**Two- and Three-Component Models**

Our numerical model [Flynn, 1992] takes advantage of the fact that the GER spectroradiometer provides ~200 data in atmospheric windows which can be used for the calculation of Planckian blackbody curves. The amount of spectral data gained in a single measurement is a vast improvement over the two usable channels (bands 5 and 7) available from Landsat TM data. Unlike previous TM studies where the lack of data necessitated an estimation of the temperature of the cool (background) radiating area [Rothery et al., 1988] or the temperature of the hot radiating area [Oppenheimer and...
measurements) of the GER spectroradiometer allow for a determination of the amount of radiation contributed by individual subareas to the overall flux. This can be seen in the following system of equations to determine the three variables multiple fractional portions, \(i\), radiating at different \(T(K)\) is given by Planck's law. A given area displaying different wavelengths in atmospheric transmission zones, where \(Fe(3.1), Fe(3.2), Fe(3.3)\) are measured radiance values at a particular model. This means that in the two-component model, it is assumed that there are only two separate areas (\(Ah, Th, Tc\)) for the two-component model: 

\[ Fe(\lambda) = AhB(\lambda, Th) + AcB(\lambda, Tc) \]  

where \(Ac = 1 - Ah\). The number of variables prevents a unique solution with fewer than three channels measured. Clearly, single-band radiometers can not be used to collect data for a unique solution to the two-component model. 

The large number of channels (discrete wavelength-band measurements) of the GER spectroradiometer allow for a determination of the amount of radiation contributed by individual subareas to the overall flux. This can be seen in the following system of equations to determine the three variables (\(Ah, Th, Tc\)) for the two-component model: 

\[ Fe(\lambda_1) = E_{13}(AhB(\lambda_1, Th) + (1-Ah)B(\lambda_1, Tc)) \]  
\[ Fe(\lambda_2) = E_{13}(AhB(\lambda_2, Th) + (1-Ah)B(\lambda_2, Tc)) \]  
\[ Fe(\lambda_3) = E_{13}(AhB(\lambda_3, Th) + (1-Ah)B(\lambda_3, Tc)) \]  

where \(Fe(\lambda_1), Fe(\lambda_2), Fe(\lambda_3)\) are measured radiance values at different wavelengths in atmospheric transmission zones, leaving \(Ah, Th, Tc\) as the quantities which must be determined. \(E\) and \(\tau_2\) are the emissivity of the lake surface and the atmospheric transmissivity from the lake surface to the spectroradiometer, respectively. Our laboratory spectra of glassy, recently formed (within 3 days of cooling) basalt samples exhibit reflectances of \(\leq 0.5\%\) over the wavelength region from 0.4 to 2.25 \(\mu m\) where the data for this study were acquired. Here we assume that \(E\) is \(\geq 99.5\%\) and does not vary significantly. Hence \(E\) has been treated as a constant which will drop out of subsequent equations ratioing \(Fe(\lambda_1), Fe(\lambda_2), \) and \(Fe(\lambda_3)\). However, if the targets were andesitic or rhyolitic flows, \(E\) would probably vary with wavelength and could not be dropped from flux ratios.

To investigate the effects of changes in spectral emissivity of two radiative surface components at different temperatures, the emissivity was allowed to vary by \(5\%\) and \(10\%\) between the three atmospheric windows (1.2-1.3 \(\mu m\), 1.52-1.79 \(\mu m\), and 2.03-2.34 \(\mu m\)) from which data were chosen. The best fit crust component temperature varied by \(< 20^\circ\)C for stage 1 activity and \(< 6^\circ\)C for stage 3 activity. By allowing the emissivity of the hot component to vary by up to \(20\%\) from the emissivity of the crust component, the best fit results varied by \(< 3^\circ\)C for stage 1 activity and \(< 2^\circ\)C for stage 3 activity. Changes in emissivity between channels in different atmospheric windows resulted in larger potential errors than those associated with changes in emissivity due to different radiative temperature components. The atmospheric transmissivity was not a concern for this study since data points were chosen in only well-defined atmospheric windows between 0.9 and 2.25 \(\mu m\) in which \(\tau_2\) is very nearly equal to 1 [Wolfe and Zissis, 1985]. \(\tau_2\) was assumed to be unity.

The equation which Planck used to describe blackbody curves has an interesting feature in that the shape of each temperature curve across the electromagnetic spectrum is different from the shapes of other curves exhibiting different temperatures. Wien's law, which can be used to calculate the wavelength at which the maximum flux is emitted by a blackbody of any given temperature, shows that the cooler a blackbody gets, the farther out into the infrared the wavelength of maximum emission occurs. The properties of blackbodies are such that the shape of a blackbody curve or the ratio of fluxes at any two given wavelengths is unique for each emitting object exhibiting a different temperature, or

\[ K_{12} = B(\lambda_1, T)/B(\lambda_2, T) \]

where \(B(\lambda_1, T)\) and \(B(\lambda_2, T)\) are the radiance values of a blackbody exhibiting temperature \(T\) at wavelengths \(\lambda_1\) and \(\lambda_2\), respectively, and \(K_{12}\) is the ratio of these values. \(K_{12}\) is unique for each object at wavelengths \(\lambda_1\) and \(\lambda_2\) at any temperature \(T\). In this study, the accurate measurement of absolute fluxes would have been extremely difficult due the number of calibrations involved and the automatic gain change feature of the spectroradiometer. However, the relative fluxes are all that are necessary for flux ratios such as equation (7). Due to the repeatability of the calibration measurements, it is inferred that the relative fluxes obtained are very accurate.

In each of the numerical models, the number of temperature-exhibiting emitters is limited to the number of components in a particular model. This means that in the two-component model, it is assumed that there are only two separate areas (\(Ah, 1 - Ah\)) radiating at two separate temperatures (\(Th, Tc\)). The result is that the radiative flux as measured by the spectroradiometer at a particular wavelength is the sum of two
components exhibiting different radiative temperatures. Using equations (4) – (6), one can solve the system of equations simultaneously in order to determine the three unknowns in the two-component problem: $A_h$, $T_h$, and $T_c$.

The numerical model of this study [Flynn, 1992] uses the characteristics of blackbody curves exhibited in equation (7) to solve for the three unknowns in the two-component case. Four wavelengths, each located in atmospheric windows, were chosen to represent the spectrum. These data were used to calculate flux ratios which represent the shape of the spectral curve. Next, using Planck’s law, the iterative model generates theoretical fluxes and flux ratios to most closely match those calculated with the spectral data by supplying temperatures for the molten component and the lake surface crust. Then the area of the hot emitting object which results in the best fit to the calculated flux ratios is determined. The best fit is ascertained by minimizing the sum of individual flux ratio error values, where one of these values represents the absolute value of the spectral-derived flux ratio minus the theoretical flux ratio divided by the spectral-derived flux ratio. Best fit parameters to spectral data are determined solely by the lowest sum of individual error values for a given hot component temperature. None of the parameters in our model have to be assumed. The hot component temperatures (at 20-50°C intervals) used in our calculations are listed below and represent possible magmatic temperatures.

An important question is whether or not a two-component thermal model is an accurate representation of the lava lake surface. By increasing the number of discrete temperature radiating areas from two to three or four components and observing the behavior of the resulting fit to the data, one can find the number of radiating areas which most accurately represents the spectral data. In the three-component case, equation (4) becomes

$$F_\lambda(A_1) = E_{\lambda 1}(A_hB(\lambda_1,T_h)) + A_mB(\lambda_1,T_m)$$

$$+ [1-(A_h + A_m)]B(\lambda_1,T_c)$$

where $F_\lambda(A_1)$ is the measured flux from all three radiating areas at wavelength, $\lambda_1$; $A_m$ is the mid-temperature radiating area; and $B(\lambda_1,T_m)$ is the Planck law calculated radiative flux at wavelength, $\lambda_1$, of an object at temperature $T_m$. The number of wavelengths used for the study was increased from four for the two-component model to six for the three-component model [Flynn, 1992], but the analysis of the three-component model is similar to that described above for the two-component case.

### Results of the Models

The spectral determination of the two and three thermal components of selected areas of the lava lake crust are most significant when considered according to the three Stages of activity observed in the concurrent video data. Spectra were characterized as belonging to stage 1 (active overturning or fountaining of magma), stage 2 (limited rifting), or stage 3 (solid crust), and the results of the numerical model were used to determine representative temperatures.

**Stage 3.** Stage 3 was the easiest to measure as the temperature of the solid crust did not vary greatly during each observation. Also, as a result of plates of crust buckling together, the movement of the lake surface frequently slowed to a standstill, which allowed a measurement to be completed on the same piece of crust. Figure 2a shows a typical spectrum (out of the 13 collected) of stage 3 quiescent crustal activity. In order to appreciate how the two-component model determines a result for a given hot temperature and set of spectral parameters, we present the abbreviated results of a typical run (Figure 5) using data from Figure 2a. In this case, the temperature for the hotter portion of the lake surface being measured was arbitrarily set at 1100°C.

As seen in Figure 5, the surface of the quiescent lake is remarkably cool. The best fit matches with a surface crust having a temperature of 264°C with glowing cracks at 1100°C comprising only 0.0015% of the total viewing area. As can be seen from the errors, the choice of the correct solid crust temperature for the model is very important as the error for numerical fits increases 18-fold if the crust temperature is 50°C too hot or too cold.

![Graph comparing crust temperature and total error](image)
Figure 6a shows the results of the 2-component best fits to the spectrum presented in Figure 2a for sample magmatic basalt temperatures of 900°C, 950°C, 1000°C, 1050°C, 1100°C, 1130°C, 1150°C, 1200°C, and 1250°C. The best fit result of \( T_h = 900°C \) and \( T_c = 260°C \) is plotted in terms of radiance in Figure 7a. The results of Figure 6a are typical for most of the other stage 3 examples in that the crack temperature, \( T_c \), could vary from 900 to 1250°C with little effect on the surface crust temperature, \( T_c \), or the hot area, \( A_h \) [Flynn et al., 1990]. The solid crust temperature of the lava lake may be summarized as \( 265 ± 5°C \) and include all of the best fit results. With the hot cracks making up only \( 5 \times 10^{-5} \) total pixel, the surface crust (\( T_c = 260°C \)) contributes \( >50\% \) of the total flux for wavelengths greater than 1.49 \( \mu m \) for \( T_h = 900°C \) (Figure 7a and see discussion below). The model best fits give almost the same all-important solid crust temperature (\( T_c \)), irrespective of what magmatic temperature is chosen between 900 and 1250°C. This is due to two factors which are the very small size of the hot component and the small change in the radiance of the hot component with respect to wavelength. Wavelengths shorter than 1.5 \( \mu m \) contain significant reflected sunlight and cannot be used as data for the numerical models. The remaining spectral range (1.5-2.5 \( \mu m \)) is not as sensitive to changes in \( T_h \) as would simulations using data in the 0.9-1.5 \( \mu m \) range, where the increase in radiance with wavelength is large. The error values obtained from the best fits in Figure 6a were all very low (< 0.01), which means that all of the fits were good, and again suggests that for stage 3 quiescent activity, the accurate determination of \( T_c \) is most important.

Table 1 includes a summary of all two-component stage 3 measurements. Some data were recorded within minutes of one another while viewing the same plate of crust on the surface of the lake (e.g., spectra 1-4). Slight movements of the nearly stagnant crust prevent us from being absolutely certain that the same piece of crust occupied the field of view for both sets of measurements, but it is certain that the measurements were completed within 3 min of one another and that the target areas were very close to one another (≤ 1 m). Of course, the two beams of the spectroradiometer provided two spectra which were taken of portions of crust that were separated by only 0.68 m and 1.05 m on October 12, 1987 and January 23, 1988, respectively. Spectral pairs 3 and 4 and 12 and 13 suggest that the radiative character of solid plates of crust was uniform on a 1-m scale.

![Figure 6](image_url)

Fig. 6. Two- and three-component model results for hot temperature starting conditions between 900°C - 1250°C. Lines connecting data points illustrate trends in solutions; however, data points have only been calculated for the hot temperatures mentioned in the text. Only these data points are actual results. Hot temperature, \( T_h \); medium temperature, \( T_m \); and crust temperature, \( T_c \) represent the temperatures of magmatic cracks in the surface of the solid plates, of recently formed solid crust, and of older thicker crust, respectively. Hot area, \( A_h \), and medium area, \( A_m \), identify the fractions of the measurement area exhibiting temperatures of \( T_h \) and \( T_m \), respectively. The remainder of the measurement area radiates at temperature \( T_c \). \( T_m \) was allowed to vary by a minimum increment of 5°C in all three-component numerical simulations, while \( T_c \) was allowed to vary by a minimum of 2°C. Note that for the two-component results, \( A_m = 0 \) by definition. "Total error" is the sum of the weighted differences between the data derived flux ratios and the model generated flux ratios, and measures the quality of the model fit to the spectroradiometer data. (a) Two-component results (stage 3) for hot temperature starting conditions between 900°C and 1250°C for spectral data shown in Figure 2a, (b) and (c) three-component model results (stage 3) for hot temperature starting conditions between 900°C and 1250°C for spectral data shown in Figure 2a, (d) two-component model results for stage 2 activity (data presented in Figure 2b), note that in this case, the best fit occurs for a hot temperature of 1000°C; (e) and (f) three-component temperature model for stage 2 activity (data presented in Figure 2b), error values were higher when the hot temperature was set below 1100°C; (g) two-component temperature model for stage 1 activity (data presented in Figure 2c). In this case, the error increases from a minimum at a hot temperature of 900°C.
At first glance at the two-component best fits in Table 1, it seems that the magma filling the cracks in the surface plates exhibited a temperature, $T_h$, in many cases of 900°C. This number is a bit misleading. For 12 out of the 13 model results shown for stage 3 activity, $T_h$ could be varied from 900 to 1250°C with a corresponding range in the solid crust temperature $T_c$ of less than 10°C. Best fit numerical model results 2-5, 7-9, and 11 in Table 1 showed variances in $T_c$ of 4°C or less with corresponding changes in the best fit errors of less than 6% for $T_h$ varying between 900 and 1250°C. This means that for these eight observations of stage 3 activity where Kupaianaha was covered by a solid crust, a particular value for $T_h$ was not important in reducing the error of the fit to the spectral curve. On the other hand, small variations in $T_c$ would result in large increases in the error values (Figure 5).

The fractional area of the hot radiator, $A_h$, varied by less than two orders of magnitude for all two-component stage 3 examples in Table 1.

Errors listed for three-component best fits are in no way correlative with the errors calculated for the two-component models. The three-component model uses six ratios producing six individual error values which are summed to get a total error, while the two-component model total error is the sum of three error values. It is not easy to compare the two best fit error quantities of the two- and three-component models for the same spectrum. A further discussion of best fit errors and the factors which influence them is detailed below. Three-component numerical model best fit results for sample magmatic temperatures of 900-1250°C (at 50°C intervals) using input spectral data from Figure 2a are given in Figures 6b and 6c. These model results contain a number of trends that were present in the majority of the stage 3 numerical model best fits.

Some general characteristics of stage 3 activity may be derived from the numerical model results (Tables 1 and 2). The choice of a particular temperature representing microcracks in the surface of the solid crust plates, $T_h$, did not seem to be an important factor in determining the best fit to the data points chosen for the numerical model. The fractional radiating area of the microcracks, $A_h$, was $10^{-5}$ or less of the total measured area for both the two- and three-component models. For many of the best three-component solutions, the difference between $T_m$ and $T_c$ was very large (> 90°C) suggesting that a two-component solution may provide an inadequate fit to the data. The largest change in the error value occurred in all stage 3 examples when $T_c$ was allowed to vary from the best possible numerical fit value. The proper choices of the crust temperatures $T_m$ and $T_c$ were the most important factors resulting in the lowest possible error value. Typical values of $T_c$ for stage 3 for both models were generally found to be between 80 and 345°C with most solutions near 200-300°C.

Stage 2. Stage 2 was more difficult to measure since the activity was continuous with fresh crust forming from recently erupted magma. The three examples of stage 2 activity (Tables 1 and 2) represent developing rifts which were stationary on the lake surface. Distinct differences can be seen in the shapes of the spectral curve collected for stage 3 (Figure 2a) and stage 2 (Figure 2b). Stage 2 rifting spectra show much higher fluxes in the 1.5-1.8 μm region relative to stage 3 spectra. Also, the ratios of measured fluxes between 1.5 and 1.8 μm to the fluxes between 2.0 and 2.25 μm are less in the stage 2 spectra relative to stage 3 spectra. Reflected sunlight contributes a smaller percentage of the measured flux in stage 2 activity (3% at 1.53 μm) than in stage 3 activity (42% at 1.53 μm).

A summary of the best two-component numerical model fits to the data presented in Figure 2b for magmatic temperatures ranging from 900 to 1250°C is given in Figure 6d. Unlike the stage 3 example (Figure 6a), the temperature of the solid crust, $T_c$, varies considerably depending on the magmatic temperature, $T_h$, chosen for the numerical model (Figure 6d). Over the range of values for $T_h$, the best fit for $T_c$ varies by more than 130°C. The best fit errors for stage 2 two-component models (Table 2) also show wide differences of up to 1000% depending on the value for $T_h$ chosen for the model. As would be expected, the hot emitting area, $A_h$, for rifting stage 2 activity increased relative to quiescent stage 3 activity.

It is evident from the three-component best fit results (Figures 6e and 6f) that the temperatures of the two solid crust components, $T_m$ and $T_c$, vary considerably depending on the value chosen for the hot component temperature, $T_h$. The three-component results also show that the difference between $T_m$ and $T_c$ is 40°C or less for most best fits between 900°C and 1250°C. However, this trend was not consistent for the other two examples (Table 2) where the difference between $T_m$ and $T_c$ was ≥ 80°C. Again, this suggests that a two-component solution would be sufficient to describe the spectral data.
presented in Figures 6e and 6f, but would be inadequate for the other examples of stage 2 activity.

Some very general assumptions about rifting events on the lava lake surface associated with stage 2 may be gained from our examples (Tables 1 and 2). For the three-component best fit examples, the difference between $T_m$ and $T_c$ was small enough in one case that a two-component model would be sufficient to describe the radiative characteristics of the lake surface. Solid crust temperatures ($T_m$ and $T_c$) ranged from 100 to 340°C. The area radiating at magmatic temperatures, $A_h$, was of the order of $10^{-4}$-$10^{-6}$ for all stage 2 examples.

Stage 1. Stage 1 occurs when the surface of the lava lake experiences rapid, wide rifting exposing fresh magma and is marked by the effervescence of gas bubbles and small (3 m high) fountaining events caused by the subduction of segments of the lake surface. Sixteen spectra of stage 1 activity were collected at Kupaianaha. Of these spectra, eight were collected while the spectroradiometer was aimed at very active rifting and fountaining events, and eight were collected while the instrument was pointed at a glowing lava tube at an edge of the lake. Stage 1 events proved to be the most difficult to measure because these events were extremely transient. The duration of these events was frequently much less than the time that the instrument required to complete a measurement (>30 s). Figure 2c displays the data collected while observing a fountaining region on the surface of the lava lake. The small-scale features superimposed on the blackbody curves are due to both the instrument response with a neutral density filter in the field of view [Flynn, 1992] and the dynamic nature of the lake surface as blobs of magma erupted and cooled in the field of view. Note the low reflected sunlight peak (~0.554 μm) which is a result of the lake surface lying in cloud shadow, and instrument autocalibration with the very large fluxes measured at longer wavelengths.

Due to the small features appearing throughout stage 1 spectra, errors in the numerical models can be particularly large.
due to two factors: (1) the high normalizations (5.47 x 10^{-4} at 0.554 μm and 3.41 at 2.115 μm) required to correct the data would accentuate small anomalous features, and (2) the amount of flux emitted at a given wavelength may change rapidly due to the dynamic activity in the field of view. With these possible problems in mind, Figure 6g summarizes the best fits for Figure 2c. The two three-component examples of stage 1 (Table 2) contain some of the highest temperatures for the solid crust, T_c, that have been found in this study. This is not too surprising considering that a fountaining event and very fresh crust occupied the field of view. The fractional hot area, A_h, was also very large when compared to the other data. We calculate that ~1.5% of the area measured by the spectroradiometer was covered by magma. The errors were moderately high, suggesting that the numerical fit was not entirely satisfactory for the data chosen from Figure 2c.

The stage 1 two-component best fit results (Table 1) show examples of the temporal and spatial brevity of these events. Examples 22 to 24 were collected within 6 min of one another with no visible apparent change in activity during the time that the measurements were collected. However, the result for example 22 suggests that the crust of the lake was much warmer than it was when examples 23 and 24 were collected only minutes later. This could be the result of the surface cooling dramatically in 4 min (not supported by video data), or the hot surface could have shifted out of the field of view. Examples 25-27 were spectra of the same lava tube taken 30-65 min after examples 23 and 24. The video record shows that the tube became very active during this time. It is interesting that example 26, which was taken with the reference beam of the spectroradiometer, monitored a crust over 200°C cooler than examples 25 and 27 (target beam spectra of the same surface). Examples 25 and 26 were collected at the same time within 1.05 m of one another. These examples show the spatially limited extent of stage 1 activity. These results are similar to those obtained for examples 30 and 31 for the same reasons.

The results of this study (Tables 1 and 2) suggest that stage 1 areas exhibit solid crust temperatures (T_m and T_c) between 180 and 575°C and hot radiating fractional areas frequently greater than 10^{-3} of the viewing area. Examples 19-21, 23, 24, 26, 30, and 32 are not representative of stage 1, but they do help to illustrate the temporal and spatial brevity of stage 1 events. The two three-component examples of stage 1 (Table 2) activity which were not collected with the filter suggest that at least 3 radiative components may be necessary to describe this type of activity. Some of these results have high errors which may be due to the inability of the spectroradiometer to record temporally brief events or due to the effects of the filter used on January 23, 1988.

Special case: Stages 1 and 2. One spectrum (Figure 2d) of a moving spreading center collected on October 12, 1987 did not neatly fit into any of the three stages mentioned above. A large spike occurring at 2.00-2.08 μm shows where the flux from freshly erupted magma saturated the instrument detectors. Since this spectrum was collected in a 70-s (speed factor 3) measurement time, this eruptive event, which saturated 18 channels, must have been present in the field of view for only ~1.25 s (assuming an equal time interval for each channel measurement). Less obvious is the arrival of the recently formed crust from the spreading center (stage 2) in the field of

**TABLE 2. Summary of Three-Component Model Best Fits**

<table>
<thead>
<tr>
<th>Example</th>
<th>Activity</th>
<th>T_h,°C</th>
<th>T_m,°C</th>
<th>T_c,°C</th>
<th>A_h</th>
<th>A_m</th>
<th>Error</th>
<th>Q/A, W m^{-2}</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>crust after rifting</td>
<td>950</td>
<td>270</td>
<td>138</td>
<td>1.3 x 10^{-5}</td>
<td>0.29</td>
<td>0.0371</td>
<td>2.6 x 10^{3}</td>
</tr>
<tr>
<td>2</td>
<td>crust after rifting</td>
<td>900</td>
<td>305</td>
<td>125</td>
<td>2.2 x 10^{-6}</td>
<td>0.99</td>
<td>0.0647</td>
<td>6.2 x 10^{3}</td>
</tr>
<tr>
<td>3</td>
<td>plate crust</td>
<td>1100</td>
<td>275</td>
<td>270</td>
<td>9.7 x 10^{-7}</td>
<td>0.95</td>
<td>0.6350</td>
<td>5.1 x 10^{4}</td>
</tr>
<tr>
<td>4</td>
<td>plate crust</td>
<td>1100</td>
<td>285</td>
<td>275</td>
<td>2.2 x 10^{-8}</td>
<td>0.62</td>
<td>0.4733</td>
<td>5.3 x 10^{3}</td>
</tr>
<tr>
<td>5</td>
<td>fresh crust</td>
<td>1100</td>
<td>345</td>
<td>245</td>
<td>6.4 x 10^{-8}</td>
<td>0.30</td>
<td>0.2076</td>
<td>5.3 x 10^{3}</td>
</tr>
<tr>
<td>6</td>
<td>fresh crust</td>
<td>1100</td>
<td>280</td>
<td>170</td>
<td>8.3 x 10^{-8}</td>
<td>0.35</td>
<td>0.3958</td>
<td>3.4 x 10^{4}</td>
</tr>
<tr>
<td>7</td>
<td>plate crust</td>
<td>1100</td>
<td>300</td>
<td>170</td>
<td>4.6 x 10^{-7}</td>
<td>0.40</td>
<td>0.0828</td>
<td>3.7 x 10^{3}</td>
</tr>
<tr>
<td>8</td>
<td>plate crust</td>
<td>1100</td>
<td>220</td>
<td>80</td>
<td>8.5 x 10^{-7}</td>
<td>0.95</td>
<td>0.6015</td>
<td>3.2 x 10^{3}</td>
</tr>
<tr>
<td>9</td>
<td>plate crust</td>
<td>900</td>
<td>270</td>
<td>180</td>
<td>4.2 x 10^{-8}</td>
<td>0.30</td>
<td>0.3180</td>
<td>3.1 x 10^{3}</td>
</tr>
<tr>
<td>10</td>
<td>plate crust</td>
<td>1100</td>
<td>280</td>
<td>160</td>
<td>9.4 x 10^{-6}</td>
<td>0.30</td>
<td>0.1074</td>
<td>3.0 x 10^{3}</td>
</tr>
<tr>
<td>11</td>
<td>plate crust</td>
<td>900</td>
<td>180</td>
<td>80</td>
<td>2.2 x 10^{-6}</td>
<td>0.70</td>
<td>0.7731</td>
<td>1.9 x 10^{3}</td>
</tr>
<tr>
<td>12</td>
<td>crust near rifting</td>
<td>1100</td>
<td>215</td>
<td>195</td>
<td>5.0 x 10^{-6}</td>
<td>0.10</td>
<td>0.3133</td>
<td>2.8 x 10^{3}</td>
</tr>
<tr>
<td>13</td>
<td>crust near rifting</td>
<td>1100</td>
<td>295</td>
<td>155</td>
<td>1.1 x 10^{-5}</td>
<td>0.40</td>
<td>0.0379</td>
<td>3.5 x 10^{3}</td>
</tr>
<tr>
<td>14</td>
<td>beginning rift</td>
<td>1100</td>
<td>290</td>
<td>285</td>
<td>3.0 x 10^{-4}</td>
<td>0.30</td>
<td>0.1130</td>
<td>5.6 x 10^{3}</td>
</tr>
<tr>
<td>15</td>
<td>beginning rift</td>
<td>1100</td>
<td>200</td>
<td>120</td>
<td>1.3 x 10^{-4}</td>
<td>0.15</td>
<td>0.1503</td>
<td>1.6 x 10^{3}</td>
</tr>
<tr>
<td>16</td>
<td>active rift</td>
<td>1100</td>
<td>340</td>
<td>100</td>
<td>3.7 x 10^{-6}</td>
<td>0.90</td>
<td>0.4682</td>
<td>7.3 x 10^{3}</td>
</tr>
<tr>
<td>17</td>
<td>very active, magma</td>
<td>900</td>
<td>390</td>
<td>180</td>
<td>5.5 x 10^{-4}</td>
<td>0.05</td>
<td>0.1332</td>
<td>2.9 x 10^{3}</td>
</tr>
<tr>
<td>18</td>
<td>very active, magma</td>
<td>1100</td>
<td>550</td>
<td>270</td>
<td>8.4 x 10^{-4}</td>
<td>0.30</td>
<td>0.2640</td>
<td>1.1 x 10^{4}</td>
</tr>
<tr>
<td>19-32</td>
<td>no result</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>33</td>
<td>active rift, magma</td>
<td>1100</td>
<td>790</td>
<td>200</td>
<td>2.9 x 10^{-1}</td>
<td>0.45</td>
<td>0.3666</td>
<td>9.1 x 10^{3}</td>
</tr>
</tbody>
</table>

Summary of the three-component model best fits to the spectra obtained at Kupaianaha lava lake on October 12, 1987 and January 23, 1988. Examples are presented in the same order as in Table 1. Column headings are consistent with those described for Table 1, with the exceptions of T_m, and A_m, which refer to the temperature and fractional area, respectively, of the component radiating at a temperature between that of the other two components. Error refers to the lowest sum of the six individual flux ratio errors of the three-component model and is the best fit criterion. Q/A is the average flux density radiated by a surface exhibiting the parameters (T_h, T_m, T_c, A_m, and A_h) of the best fit.
view, which manifests itself as a rapid increase in flux after 1.6 μm.

The spectral data (Figure 2d) used for the two- and three-component fits (Tables 1 and 2, example 33) were located in two atmospheric windows between 1.61 and 1.77 μm and 2.08 and 2.23 μm; hence it was the arrival of spreading center which was modeled. Both two- and three-component results show that the surface of the lake was very hot when this part of the spectrum was recorded. The two-component fit shows that 13% of the field of view radiated at 1100°C with the remainder at 710°C, while the three-component result shows that 29% of the surface radiated at 1100°C, 45% of the surface radiated at 790°C, while 26% radiated at 200°C. These results represent the highest crust temperatures and hot component areas identified in this study. Both numerical fits are rather poor considering that the data chosen to represent the spectrum were relatively close together in two atmospheric windows (1.5-1.8 μm and 2.0-2.4 μm).

**DISCUSSION**

**Cool Surface Temperatures**

Remote sensing studies of active volcanoes using Landsat TM data [Rothery et al., 1988; Glaze et al., 1989; Pieri et al., 1990] conflict with one another over the lower temperature limit of detection using bands 5 and 7. Rothery et al. [1988] claim that the lowest temperature that may be derived using the dual-band method on Landsat TM bands 5 and 7 is 160°C. Glaze et al. [1989] assume a background temperature of 100°C in their study because this is the lowest detection limit of TM band 7. Conversely, Pieri et al. [1990] use 125°C as a background temperature because they claim that it is the “detection threshold for thermally emitting pixels”. Looking closely at these studies, we conclude that the lowest derivable temperature in the other two studies depends on the lowest DN (digital number) value that is considered to be significant above the background DN. Tables 1 and 2 show that the Kupaianaha lava lake exhibited solid crust stage 3 temperatures of 80-345°C, which means that parts of solid crust would not appear to be thermally significant in TM band 5 (> 225 - 270°C to be detectable [Glaze et al., 1989]), and a dual-band method temperature for the background crust could not be derived under some conditions, especially if the radiant pixels including the lava lake included significant portions of the surrounding nonthermally radiant terrain.

During field and video observations, we observed that the margins of the lake were most likely to experience stage 1 and stage 2 activity, due to the vertical fluctuations in the level of the lake. Rothery et al. [1988] claim that a 30-m TM pixel, when blurred by radiance “bleeding” into the adjacent pixel and geometric-correction effects, will actually be ~40 m in size. Thus it is conceivable (but unlikely) that a TM pixel occupying the center of the lava lake would show little or no activity if it were covered by common, thick, stage 3 crust, while the pixels representing the fringes of the lava lake would show hot anomalies due to fountaining or rift events. Using TM data, one may easily seriously underestimate the activity in the scene. This study suggests that future satellites should have increased sensitivity to be able to unequivocally detect 100°C anomalies. In order to achieve this, future satellite sensors should have a greater dynamic range than 256 (8-bit) of Landsat TM, for instance. In addition to this, more spectral data beyond 2.0 μm, where the radiative flux from the cooler crust dominates that of the small fraction radiating at near magmatic temperatures, should be collected. A combination of these two improvements over that which the current generation satellites has to offer will allow for the solution of multiple subpixel radiating areas with greater precision and accuracy than is presently possible. Instruments such as MODIS-N [Salomonson et al., 1989], and especially ASTER [Kahle et al., 1991] and HIRIS [Goetz and Herring, 1989], on the EOS platforms will provide these capabilities.

**Choice of Wavelengths for Study and Effect on Errors**

All of our radiative results from Kupaianaha suggest that the fractional area of the magmatic radiator was very small. Stage 3 activity exhibited magmatic areas of ≤ 5 x 10⁻² (which corresponds to ≤ 5.5 mm²) of the total viewing area. This is almost nonexistent, and one has to consider whether or not a two- or three-component solution using lower than magmatic temperatures would result in better fits to the spectral data. Many of the best fit results occur for hot component temperatures of 900°C or 950°C, which again suggests that temperatures below 900°C may provide better two- and three-component solutions. Most three-component examples of stage 3 activity exhibited a fractional hot area an order of magnitude lower than the lowest derivable fractional hot area using the dual-band method with data from Landsat TM bands 5 and 7 (which is 10⁻² of the total pixel for a hot area radiating at ~1100°C [Oppenheimer, 1991]). Video observations collected at Kupaianaha include frequent periods of tens of minutes duration in which the lava lake is covered by stage 3 crust, which means that it is probable that a Landsat TM scene acquired over Kupaianaha lava lake would not be suitable for dual-band calculations because of the very small hot fractional areas of the crusted surface. Stage 1 activity showed the highest fractional hot areas to be ~10⁻² of the total area.

While the areas exhibiting magmatic temperatures are very small, the radiation emitted by these sources dominates the total flux at shorter wavelengths, after the reflected sunlight component is removed from the data. At some point of the collected spectrum, the magmatic cracks and the solid crust will both contribute approximately equal amounts of flux to the total radiant flux. At shorter wavelengths than this transitional wavelength, the magmatic crack will be contributing more of the radiant flux, while at longer wavelengths, the solid crust will be contributing more of the total radiant flux. Figures 7a and 7b are interesting in that they show the best fit two-component radiance curves which are summed to match the collected spectra. For the two-component best fit model to a stage 3 example (Figure 2a), the contribution from magmatic microcracks dominate the radiative flux up to ~1.49 μm (Figure 7a). In this stage 1 example of magma fountaining (Figure 2c), the hot component dominates the total flux up to 1.27 μm (Figure 7b). One might expect that the radiance from the hot component in the stage 1 example to dominate further into the infrared than the radiance from the stage 3 hot area, which is 3 orders of magnitude smaller. However, the stage 1 example has a much higher crust temperature (and hence radiative flux from the crust component) than that of the stage 3 example. The transition wavelength occurred between 1.25 μm and 2.00 μm in all examples, except example 33 in which the magmatic component filled almost 13% of the field of view, and the transition wavelength is located at 2.20 μm.
The wavelengths used in the models for this study were chosen to most accurately represent the magmatic component, the solid crust component(s), and the summation of the two curves. The 3-component model requires six flux values at certain wavelengths while the two-component model requires four. Input data for the numerical models were chosen from atmospheric windows in the 1.2-1.3 µm (magmatic component), 1.5-1.8 µm (transition wavelength zone), and 2.0-2.5 µm (solid crust component) ranges. Ideally, for the three-component model, two data were chosen to represent each of the three windows mentioned above. In the two-component model, a number of combinations of input data were tried with the best result that, if possible, one input datum each should be chosen from 1.2-1.3 µm and 1.5-1.8 µm, with the remaining two coming from the 2.0-2.5 µm range.

Errors should be viewed only as a rough relative indicator of comparable goodness of fit. It was not possible to use the same wavelengths for all examples in this study because of the large reflected sunlight component at < 1.5 µm and the filter effects in most data collected on January 23, 1988. The three-component model also had more data points (and hence more ratios) to match successfully to lower error values. For these reasons, the two- and three-component model errors are not correlateable between examples or between two- and three-component models for the same example. Errors are only an indicator of fit for a given set of data input values. In a very general sense, error values between measurements for the same component model may be compared. For instance, a two-component best fit error of 0.0047 (example 1, Table 1) certainly fits the data values chosen for one spectrum better than a 2-component best fit error of 0.3205 (example 8, Table 1), but one factor which must be considered is whether the wavelength distributions of chosen data were the same in both cases. The low error may be due to the fact that only data values which were close together in wavelength space could be used, thus making it easier to fit a curve to these data. A case in point is example 33, which exhibited moderate error values for both two- and three-component best fits. However, data values used for this example were taken from 1.62, 1.66, 1.70, 2.10, 2.18, and 2.23 µm because, as pointed out above, a spreading center (rift) entered the field of view as the instrument was collecting data. This event, which manifested itself in the spectrum as a rapid increase in flux longward of 1.60 µm, limited which wavelengths could be used in the numerical models for that particular example.

**Reflected Sunlight in Measurements**

Earlier, we stated that the amount of radiation from reflected sunlight dominated the measured flux at 0.554 µm, but we must now prove that this is valid for stage 1 and 2 activity. The amount of solar irradiance at 0.554 µm striking an area at sea level is 135 mW cm^-2 µm^-1 [Wolfe and Zissis, 1985]. If the lake surface reflects only 0.5% of the radiation, the reflected solar irradiance becomes 0.107 mW cm^-2 µm^-2 sr^-1. The stage 2 example with the largest fractional magmatic radiating area (example 14, Table 1) exhibited a radiant flux at 0.554 µm of 1.2 x 10^-4 mW cm^-2 µm^-1 sr^-1, which is still only about 0.1% of the flux from the reflected solar component. The stage 1 example with the highest radiance at 0.554 µm (example 18, Table 1) was 2.6 x 10^-3 mW cm^-2 µm^-1 sr^-1, which is about 2.4% of the flux from the reflected solar component. The amount of reflected sunlight at the wavelengths used for stages 1 and 2 in this study were at most only 5% and 10% of the total flux, respectively. This results in an error in the estimate of the subtracted reflected solar flux of the order of 0.1% at most. Thus the assumptions that were made earlier concerning the flux at 0.554 µm being entirely reflected sunlight are still valid.

One factor in particular that made it difficult to choose data values representative of the hot component was the amount of reflected sunlight present in the data. Especially for stage 3, where the radiative contribution of the magmatic component was very small, the flux of reflected sunlight frequently dominated the total radiative flux at wavelengths < 1.5 µm. Thus, in some stage 3 cases, it was impossible to choose data where the hot component dominated as input for the numerical model. Figure 8 shows the effect of reflected sunlight on examples of the various stages identified in this study. As a comparison between the different stages of activity, reflected sunlight comprises 13% of the thermal flux at 0.988 µm for stage 1, 12% at 1.23 µm for stage 2, and 19% at 1.70 µm for stage 3. Even more surprising is the fact that in the 1.2-1.3 µm window, from which input data representing the magmatic radiators were chosen, reflected sunlight makes up only about 3.5% of the total radiative flux for stage 1, of the order of 10% for stage 2, and above 90% for stage 3. Stage 3 solid crust activity exhibits hot areas of 10^-5 or less, which is 2 to 3 orders of magnitude below that found for stage 1 activity. Since the amount of sunlight reflecting from the surface was fairly constant on October 12, 1987, and examples of all three stages were collected on this date, the radiative area of the hot component made a critical difference in our ability to reliably choose input data from the 1.2-1.3 µm window.

**Average Radiative Flux and Total Energy Budget of Lake**

Recent studies [e.g., Glaze et al., 1989; Pieri et al., 1990; Oppenheimer and Rothery, 1991; Oppenheimer, 1991] of active volcanoes using Landsat TM data have attempted to quantify the total energy budget, Q, of volcanic areas. The two-component equation for the total energy budget, as discussed by Pieri et al. [1990], is

\[
Q = (E_{cTh4}) \times A_{h} + (E_{oc4}) \times A_{c}
\]  

(9)

Fig. 8. Effect of reflected sunlight on the three stages of activity present at Kupianaha lava lake. Data points are actual amounts of reflected sunlight extracted from the spectra presented in Figures 2a, 2b, and 2c.
result that we found for fresh basalt samples. Two-component component and three-component average Q/A values for rifting 4.9 x 10^3 W/m^2. Three-component (Table 2) average flux budget of Kupaianaha by assigning fractional areas of the lake densities for active overturning and magma fountaining events x 10^3 W/m^2, respectively. Two-component average flux densities for quiescent periods were within the range of 1.9-6.2 (stage 3) events are almost identical at 5.3 x 10^3 W/m^2 and 3.0 x 10^4 W/m^2 for the active basaltic lava lake of Erta Ale and the phonolitic lava lake of Erebos, respectively. In our Kupaianaha study, these values were found to be slightly lower than or in agreement with those of stage 3, quiescent activity. Pieri et al. [1990] calculate an average flux density for a 1984 active lava flow on Etna of 1.5 x 10^3 W/m^2 which also falls below the range of values that this study identified for stage 3 crusted surfaces. In agreement with our stage 3 results, Oppenheimer [1991] reports a range of average flux densities of 1.7-4.2 x 10^3 W/m^2 for an andesite flow on Volcan Longquimay, Chile. The only other study that we know of to calculate a Q/A value for a volcanic thermal source is that of Oppenheimer and Rothery [1991], who used Cyclops thermometers to measure brightness temperatures of a 40-m-long fumarole at Vulcano, Italy. They reported a Q/A of 23 W/m^2 for the fumarole, which expectedly is 2 orders of magnitude lower than the average flux densities reported here for stage 3. The total energy budget, Q, is another important parameter in discussing the level of activity of a particular erupting volcano. We calculated Q values for stage of activity distributions identified in the video record (Table 3). More than 90% of the time, the lava lake surface could be described by the first three examples in Table 3. In other words, for the majority of the time, the lava lake was quiescent with very

\[ Q = \text{total energy flux (in W/m}^2\text{)}; \]
\[ E = \text{the emissivity;} \]
\[ \sigma = \text{the Stefan-Boltzmann constant;} \]
\[ T_h = \text{the temperature (in K) and area (in m}^2\text{)} \]
\[ T_c = \text{the temperature and area of the solid crust, respectively.} \]

This study provides important additional data to the growing information on Q values for active volcanoes. Since the spectroradiometer's field of view was very limited (< 0.6 m^2) compared to the 900 m^2 covered by Landsat TM pixels (bands 1,5 and 7), the question of calculating the total energy budget of Kupaianaha on the days when observations were made will be considered in two ways. First, the average flux density, (Q/A), as discussed in Glaze et al. [1989] and Pieri et al. [1990], has been calculated for each of the best fit examples listed in Tables 1 and 2. This allows the results of this study to be directly compared with the results of the larger-scale Landsat studies mentioned above. Next, using the video data collected on both days, we have calculated values for the total energy budget of Kupaianaha by assigning fractional areas of the lake to the various stages observed.

Tables 1 and 2 list the two- and three-component average flux densities for the three stages of activity observed at Kupaianaha. We assumed an emissivity of 0.995 given the result that we found for fresh basalt samples. Two-component (Table 1) average flux densities for quiescent periods (stage 3) were within the range of 3.0-6.3 x 10^3 W/m^2 with an average of 4.9 x 10^3 W/m^2. Three-component (Table 2) average flux densities for quiescent periods were within the range of 1.9-6.2 x 10^3 W/m^2 with an average of 3.8 x 10^3 W/m^2. Two-component and three-component average Q/A values for rifting (stage 2) events are almost identical at 5.3 x 10^3 W/m^2 and 4.8 x 10^3 W/m^2, respectively. Two-component average flux densities for active overturning and magma fountainning events (stage 1) were between 9.8 x 10^3 W/m^2 and 3.0 x 10^4 W/m^2 (excepting examples 19-21, 23, 24, 26, 30, and 32 which, as discussed above, were not representative of stage 1 activity) with an average of 2.2 x 10^4 W/m^2.

The contribution of the hot area to the average flux density was found to be very small for all three stages of activity. Radiating microcracks in stage 3 activity accounted for only 5 W/m^2 of the total average flux density. Larger rifting cracks characteristic of stage 2 contributed up to 50 W/m^2 which was less than 1.5% of the total average flux density. Magma fountains and larger overturning events of stage 1 contributed of the order of 100-1000 W/m^2 to the total average flux density. Example 33 in Tables 1 and 2 exhibited the highest proportional hot area (~13% of the total field of view) in this study. In this case, the area radiating at magmatic temperature accounted for 36% of the average flux density. For all examples of all stages of activity, the solid crust radiated most of the flux detected by the spectroradiometer because of its relatively larger surface area.

The average flux density values calculated from our measurements are generally higher than values derived for other volcanoes using Landsat TM data. Glaze et al. [1989] report a Q/A range of 0.7-2.2 x 10^3 W/m^2 for 10 Landsat images collected over Lascar volcano, Chile, between December 1984 and November 1987. This set of average flux densities is lower than the range found from the three-component results of this study for stage 3 activity. Of particular relevance to our study, Glaze et al. [1989] also report Q/A values of 1.6 x 10^3 W/m^2 and 2.3 x 10^3 W/m^2 for the active basaltic lava lake of Erta Ale and the phonolitic lava lake of Erebos, respectively. In our Kupaianaha study, these values were found to be slightly lower than or in agreement with those of stage 3, quiescent activity. Pieri et al. [1990] calculate an average flux density for a 1984 active lava flow on Etna of 1.5 x 10^3 W/m^2 which also falls below the range of values that this study identified for stage 3 crusted surfaces. In agreement with our stage 3 results, Oppenheimer [1991] reports a range of average flux densities of 1.7-4.2 x 10^3 W/m^2 for an andesite flow on Volcan Longquimay, Chile. The only other study that we know of to calculate a Q/A value for a volcanic thermal source is that of Oppenheimer and Rothery [1991], who used Cyclops thermometers to measure brightness temperatures of a 40-m-long fumarole at Vulcano, Italy. They reported a Q/A of 23 W/m^2 for the fumarole, which expectedly is 2 orders of magnitude lower than the average flux densities reported here for stage 3.

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<table>
<thead>
<tr>
<th>Example</th>
<th>Stage 1, %</th>
<th>Stage 2, %</th>
<th>Stage 3, %</th>
<th>Total Q/A, W m^2</th>
<th>Q x 10^7 W</th>
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<tr>
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<td>0</td>
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<tr>
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<tr>
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<td>1</td>
<td>5</td>
<td>94</td>
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<td>1.2</td>
</tr>
<tr>
<td>4</td>
<td>1</td>
<td>10</td>
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<tr>
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<td>5</td>
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<tr>
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<td>20</td>
<td>75</td>
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<tr>
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<td>20</td>
<td>65</td>
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Calculations of the total energy budget, Q, of Kupaianaha lava lake using video and spectral data collected on October 12, 1987, and January 23, 1988. Stage 1(2), stage 2(2), and stage 3(3) refer to the percentage of the total surface area occupied by one of the three types of activity. Total Q/A is the overall average flux density radiated by the lava lake exhibiting the distribution of activity. Q is the total energy budget of the lava lake and is the product of the overall average flux density and the total radiating surface of the lava lake. Since the lake was roughly 50 m in diameter with a 30 m x 10 m channel, we have assumed a total area of 2300 m^2. Our video observations suggest that the activity of Kupaianaha could be best described by one of the first three examples more than 90% of the time.
limited rifting occurring between massive plates in the center of the lake and very limited (<1 m high) fountains erupting along the perimeter of the lake. Thus, according to our numerical model fits to our spectral data, more than 90% of the time Kupaianaha would have exhibited a Q value of 1.1-1.2 x 10^7 W. However, the lava lake did experience periods of increased radiative output, both through rifting events and large scale subducting (one plate sliding beneath the other) events. Examples 4-10 summarize some of the typical stage distributions observed for these temporally limited events. Increased stage 1 activity was always accompanied by increased stage 2 activity as is suggested by the examples in Table 3. The most active period (example 10) at Kupaianaha occurred on October 12, 1987, when a large (3-4 m high and occupying ~50 m^2) fountain erupted from between two plates near the center of the lake while a large subducting event was exposing even more fresh magma near the edge of the lake. Even with these relatively large events occurring at the same time, stage 1 could not have an area of greater than 15% of the total lake surface and a momentary Q value of 1.8 x 10^7 W. However, these instances of large increases in the surface activity of the lava lake show that Q for Kupaianaha could almost double in the span of seconds, which would be difficult to monitor with the Landsat satellite.

A number of previous studies have attempted to quantify Q for volcanic sources. Le Guern et al. [1979] which reported a Q value of 1.13 x 10^8 W for the two lava lakes (total area = 3800 m^2) of Erta Ale during 1973. Le Guern [1987] reports Q values for Nyiragongo of 5.4 x 10^8 W (area of anomaly = 13,125 m^2) and 1.22 x 10^10 W in 1959 and 1972, respectively. These lava lakes were reported to be much larger than Kupaianaha at the time of our visits. Oppenheimer and Rothery [1991] quote a Q for a fountain at Vulcano, Italy, of 37 kW, placing this type of low thermal activity in a category of its own as far as remote measurements of Q are concerned.

Several investigators have explored the potential for satellite measurements of Q. Glaze et al. [1989] used Landsat TM data to derive Q values of 0.3–7.6 x 10^7 W for Lascar, 5.8 x 10^7 W for Erebus and 6.9 x 10^7 W for Erta Ale volcanoes, respectively. The value for Erta Ale lava lake is greater than our estimates for Kupaianaha, but is actually consistent with our results since the surface area of Kupaianaha was smaller than that of Erta Ale. Further Landsat TM results presented by Pieri et al. [1990] calculate a Q value of 12.5 x 10^7 W for a June 1984 lava flow on Mt. Etna, which is again higher than our estimates for Kupaianaha because of the larger radiating surface presented during this eruption. Another important factor in the calculation of Q (and Q/A) is the emissivity. We have measured the reflectance of basalt to be < 0.5% in the 0.4-2.5 μm range, and thus assume that the emissivity is 0.995. Oppenheimer [1991], which uses an emissivity for andesite of 0.95, reports Q/A values which overlap the three-component results found for stage 3 examples. The emissivities used in our study and that of Oppenheimer [1991] are higher than those used in previous studies such as Rothery et al. [1988] which assumed emissivities in bands 5 (1.55-1.72 μm) and 7 (2.08-2.35 μm) of 0.8 and an emissivity in band 4 (0.76-0.90 μm) of 0.6 based on their unpublished data.

Our investigations of the total energy budgets and the average flux densities of our observations have led us to an important conclusion. Both quantities are important in describing the activity of a volcanic eruption. While Q provides a comparable estimate of the overall activity of the volcano [Glaze et al., 1989], we have found that the average flux density Q/A measures the intensity of the eruptive phenomenon. Our calculations of Q are lower than those of other eruptions. However, when Q/A values are compared, our results for stage 3 activity are higher than those results obtained for the Landsat TM studies mentioned above. Volcanic areas composed primarily of fountaining events or other stage 1 activity would exhibit Q values much higher than those reported. Thus we would point out that while Q is an important indicator of the volume and extent of an eruption, Q/A is critical to the understanding of the intensity of the ongoing activity. Furthermore, remotely sensed estimates of Q depend on the accurate determination of the overall size of the radiative anomaly, which, as shown above, can be arbitrary depending on the assumed background temperature [Oppenheimer, 1991]. Q/A can be calculated on an individual pixel basis and only depends on the fractional area of the emitting blackbodies. Thus, even when the total extent of an eruption cannot be determined (due to partial cloud cover, for instance), Q/A, calculated from thermally emitting pixels, can provide comparable estimates of the degree of severity of an eruption.

CONCLUSIONS

Spectral data collected on October 12, 1987, and January 23, 1988, have been analyzed using numerical models in which the ratios of fluxes at wavelengths chosen from atmospheric windows were matched to theoretical values at the same wavelengths for two and three blackbody curves. This method is innovative in that it attempts to match the shape of the spectral curve as closely as possible. For many of the spectra analyzed, it was determined that a two-component model would be insufficient in describing the radiative output of the lava lake, since the two surface crust components of the three component model were different by ≥ 90°C. For these spectra, a numerical model using three or more components is necessary.

With the help of the video film obtained while the spectral data were collected, we have identified three characteristic stages of activity for the Kupaianaha lava lake. Stage 1, characterized by magma fountaining and overturning events exhibited the hottest crust temperatures (> 370°C) and largest fractional hot areas (> 10^-3) and may require a three or more component thermal model to describe the 0.4-2.5 μm spectrum. Stage 1 average flux densities were about 2.2 x 10^8 W/m^2, the highest recorded for the three stages of activity that we observed. Stage 2, marked by rifting events between plates of crust, exhibited crust temperatures between 100 and 340°C with fractional hot areas about an order of magnitude lower than those found for stage 1. Average flux densities calculated for the three examples of stage 2 activity were typically 5.3 x 10^7 W/m^2. Stage 3, quiescent periods when the lake was covered by a thick crust, dominated the activity on the surface of the lake both temporally and spatially over 90% of the time. Numerical model results for stage 3 suggest characteristic crust temperatures of 80-345°C with fractional hot areas of 10^-5 or less of the total viewing area. The hot fractional areas representative of stage 3 were so small that the reflected solar flux dominated the total thermal flux up to about 1.5 μm and the variation of the hot component temperature between 900°C and 1250°C did not have a significant effect on the solution of the crust temperature [Flynn et al., 1990]. Average flux densities for stage 3 were generally within the range of 3.6 x 10^7 W/m^2, which were consistent with a recent study [Oppenheimer, 1991] of active volcanoes using Landsat TM data. Results obtained
for stages 2 and 3 suggest that newly erupted magma from cracks cools extremely rapidly and that the main radiative differences between the two types of activity is the area of the cracks in the field of view. The radiative energy emitted by both types of activity are comparable because $Q$ is dominated by the radiant energy of the surface crust.

Our spectroradiometer data provide the most complete analysis of the thermal radiative properties of lava lakes that has been done to date. However, as has been noted by Oppenheimer and Rothery [1991], further instrument developments will have to be made before satellite observations of thermal anomalies can become adequate for the complete description of surface temperatures of volcanic phenomena. For instance, the data that we have presented lack the spatial information of the thermal anomalies. Although the two spectroradiometer beams provide some information on the variability of surface temperatures at a separation of $\sim 1$ m or less, this information lacks the synoptic view afforded by satellite data to completely describe the thermal structure of either a dynamic lava lake or a moving lava flow. Another problem with the spectroradiometer involves the time that is required to collect a spectrum. Transient stage 1 events are manifested in the spectral data as flickering (Figure 2c) lasting fractions of seconds or as large spikes (Figure 2d) lasting on the order of 1 s. We used the most rapid method of data collection with the spectroradiometer, which was 1000 data points in approximately 30 s, but this was still far too slow to record some events when data were collected in the 0.4-2.5 $\mu$m range. Instruments with more rapid data collection (< 1 s), perhaps having fewer very narrow channels located in atmospheric windows, are required to provide the speed and volume of data necessary to describe transient fountaining events or quickly cooling lava flows. Finally, we also recognize that the orientation (55$^\circ$ from the vertical) of the spectroradiometer with respect to the observed surface may have prevented us from “seeing into” cracks in the crust on the surface of the lava lake. Thus, we conclude that our estimates of the fractional area of the hot component in our examples are lower bound values.

Our spectroradiometer data also show some of the potential problems in using infrared thermometers operating in the 0.8-1.1 $\mu$m range [e.g., Jones et al., 1990; Oppenheimer and Rothery, 1991; Oppenheimer, 1991] or using Landsat bands 4 (0.76-0.90 $\mu$m) and 5 (1.55-1.75 $\mu$m), to study active lava flows and lakes. Solar contamination of the spectrum may be greater than 12% of the total flux at wavelengths shorter than 1 $\mu$m for even highly active stage 1 events (Figure 8). Even worse, the flux from reflected sunlight increases rapidly at shorter wavelengths making estimations of the distribution of the reflected sunlight component across the band difficult for Landsat band 4 and almost impossible for the automatic infrared thermometers. In both cases, the distribution of reflected sunlight shortward of 1 $\mu$m suggests that brightness temperatures measured with infrared thermometers (operating at 0.8-1.1 $\mu$m) or calculated using Landsat TM band 4 will result in higher temperatures than are actually present. A similar problem could be present when using TM band 5 to estimate the temperatures of the surface crust (stage 3 and to a lesser extent stage 2) on lava lakes and lava flows. Our raw spectral data of numerous examples of quiescent activity suggest that reflected sunlight contributes a significant percentage of the total radiative flux at 1.55-1.75 $\mu$m. Again, the reflected solar flux decreases with increasing wavelength and has been shown (Figure 8) to be about 50% of the total radiative flux at 1.55 $\mu$m and 19% at 1.70 $\mu$m for stage 3 activity. This uneven distribution of reflected solar flux would again adversely affect dual-band solutions using TM bands 4 and 5 or 5 and 7.

Nighttime measurements [Flynn and Mouginis-Mark, 1992] are the best method to avoid these solar contamination issues. As noted by Rothery et al. [1988] in the case of Landsat data, it is important to use nongeometrically corrected data for the temperature determinations, because pixel resampling will destroy the radiometry of each pixel. In this instance, the identification of reference points in the image, where most objects in the nighttime scene are dark, makes the comparison between the thermal properties and the morphology of the flow a difficult task. We also note that the emissivity of fresh basalt which we measured in the laboratory was $> 0.995$ across the entire wavelength range of this study (0.4-2.5 $\mu$m). Thus we could treat the emissivity as a constant in our models without compromising the results. We caution that this would not hold true for volcanoes erupting different magma types. To use this ratio technique for nonbasaltic volcanoes, the emissivity of fresh samples must be measured over the same wavelength range as the spectral data.

Imaging spectrometers, such as the ASTER [Kahle et al., 1991] and HIRIS [Goetz and Herring, 1989] instruments planned for the Earth Observing System (EOS), will prove to be much more useful in the study of active volcanoes than the currently available Landsat TM. ASTER will have nine bands in the 0.5-2.5 $\mu$m range with 15-30 m spatial resolution [Kahle et al., 1991], while HIRIS will acquire data in 192 bands from 0.4 to 2.5 $\mu$m with 30-m spatial resolution [Goetz and Herring, 1989]. What ASTER and HIRIS both lack is the temporal resolution that a field instrument can provide. The fact that our spectroradiometer data (Figures 2c and 2d) show radiative thermal changes in the area of the lake that is characterized as stage 1 over time scales of at most a few tens of seconds indicates that an instantaneous view of a lava lake is not sufficient to determine the level of activity that an observer would find on the ground. Fortuitous satellite observations of overturning events on the lava lake surface could lead to an overestimation of the normal representative total energy budget from the lake, whereas an episode of stage 3 activity may be interpreted as an inactive lake. For example, the Landsat TM measurements of the total energy budget, $Q$, that were determined by Glaze et al. [1989] indicated changes in $Q$ for Lascar volcano, but these changes are partially due to the varying size of the anomaly and could equally well be related to short-term (minutes to hours) variations in the thermal characteristics of the volcano. We note that even when the EOS is deployed and operational, this lack of temporal coverage will still exist. Only measurements made from an as yet unapproved geostationary platform, or from a sensor in a Molniya orbit (i.e., an orbit that is highly elliptical, and that enables the spacecraft to “hang” over a particular spot on Earth’s surface for several hours each orbit) will be able to address this dynamic aspect of the total energy budget of a volcano.

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