GROUND-BASED THERMAL REMOTE SENSING OF ERUPTION DYNAMICS AT SANTIAGUITO LAVA DOME COMPLEX, GUATEMALA

A DISSERTATION SUBMITTED TO THE GRADUATE DIVISION OF THE UNIVERSITY OF HAWAI'I IN PARTIAL FULFILLMENT OF THE REQUIREMENTS FOR THE DEGREE OF

DOCTOR OF PHILOSOPHY

IN

GEOLOGY AND GEOPHYSICS

DECEMBER 2007

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ACKNOWLEDGMENTS

It is hard to believe that the end is here after 5 ½ years. I would like to thank my advisor Luke Flynn who took me in as his graduate student and allowed me to conduct research on one of the most exciting volcanoes I have ever been to. I also would like to thank Andy Harris who really helped me maintain focus and really helped me with all the discussions that we had about everything Santiaguito. I also would like to thank the rest of my committee members, Rob Wright, Bruce Houghton and Matt McGranaghan, for taking the time out of their busy schedules to help me with the dissertation process. I also would like to thank NASA for providing the majority of funding through the Earth Systems Science Fellowship, which along with various other NASA and NSF grants through Luke Flynn and Andy Harris, made the works included in this dissertation possible.

I have met many friends in the 5 ½ years here that have made my journey much more exciting and fun, many of which have gone on to bigger and better things, Eric Haskins, John Bailey, Nancy Adams, Aisha Morris, Constanza Bonnadona, Lucia Gurioli, Julia Sable, Elaine Smid, Rebecca Carey, Matt Patrick, Leon Geschwind, Lizzette Rodriguez, Darren Chertkoff and many more. To my family, my mom Ellen and my step-dad Jim, thank you so much for your support for all these years.

To my wife Heather, thank you for helping me through all the roadblocks and letting me know that I can really finish this dissertation. I really don’t think I
would have gone this far without your love and support. To my son Miles, thank you for letting me know what the most important thing in life is, rolling a tennis ball down the hallway. I would also like to dedicate this dissertation to the memory of my brother and my dad, James and Paul Sahetapy-Engel. I missed you both of you very much. I wish that you were still here to share this huge accomplishment and I hope that have made you proud.
ABSTRACT

Here I describe the use of ground-based thermal infrared data to study eruption dynamics associated with the emplacement of an active lava dome at Santiaguito, Guatemala. Integration of thermal data with seismic and infrasound (Chapter 2) allows for detection and determination of explosion frequency (2.3 per hour). It also provides a systematic method to track relative eruption intensity. Explosion source depths are constrained, using the thermo-infrasound delay, to within a 600 m deep region below the vent. Consistent partitioning of released energy into thermal, seismic and infrasound suggests a single source mechanism consistent with shear-induced fragmentation due to intermittent stick-slip of a rising dacite plug.

Integration of thermal data with data regarding the spectral radiance emanating from the vent reveals cyclic temperature fluctuations (Chapter 3). Application of a two-component thermal mixture model to the spectral radiance data reveals a transition from a pre-explosion thermal structure of predominantly cool crust with fractures at near-magmatic temperatures, to a post-explosion isothermal surface. The change in structure is explained in terms of the removal of the chilled crust of lava during explosions to expose a hotter underlying layer.

The use of thermal video camera allows for summit-wide characterization of the vent thermal structure, revealing a structure with an outer ring of high heat and gas permeability surrounding a cooler central region. The outer ring is the source of ash emissions, and the structure is consistent with the at-vent
expression of the dacite plug defined in Chapter 2. At the surface this comprises a central plug of, extruding, massive lava, surrounded by a marginal shear zone where heat, gas, and mass can preferentially flow.

Integration of ground and satellite-based (ASTER) thermal data in Chapter 5 allows for estimates of the emplaced block lava flow’s dimensions, down-flow heat loss (radiative, convective, conduction, and rainfall evaporation). A total heat flux of 256 – 913 MW corresponds to an extrusion rate of 0.3 – 1.2 m$^3$/s.

The results presented in this dissertation demonstrate the utility of ground-based thermal remote sensing to study a wide variety of eruption processes, from deep within the conduit, to the vent, and beyond.
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CHAPTER 1
INTRODUCTION

1.1. Dissertation Overview

This dissertation demonstrates the utility of ground-based thermal infrared remote sensing to the study of active volcanism. These thermal data are integrated with visual (or video) observations, geophysical datasets (such as seismic and acoustic data) and thermal data derived from satellites to observe volcanic activity at the Santiaguito lava dome complex, Guatemala (Figure 1.1). Santiaguito, like many other active lava domes, poses a number of hazards to the surrounding community. Regular eruptions of ash inundate down-wind communities as well as damage agricultural crops. Emplacement of thick block lava flows on steep slopes result in occasional failure of the leading edge of the flow to form destructive block and ash flows. During heavy rainfall, pyroclastic deposits can be re-mobilized to form lahars, which affect communities tens of kilometers downstream from the volcano.

Despite the hazards posed by this lava domes, like many other volcanoes in developing countries, volcano monitoring remains inadequate to this day due to the lack of funding. No permanent geophysical monitoring instruments have been installed at Santiaguito. The majority of volcano monitoring at Santiaguito is done by visual observations from Santiaguito Volcano Observatory, 5.5 km south of the lava dome. These observations can only be made during daytime and cloud-free periods and are limited to observation of the ash eruption, rockfall
Figure 1.1. Early morning view of Santiaguito lava dome complex and Santa Maria volcano. The cloud above the volcanoes is a remnant of an ash plume erupted from Santiaguito.
and pyroclastic flow frequency and duration as well as the height of the erupted plumes.

The general aim of the study is to explore the role that ground-based thermal remote sensing data can play in understanding conduit and eruption dynamics associated with the persistent extrusive and explosive activity at the Santiaguito lava dome complex, Guatemala. In each of the chapters of this dissertation, I describe how thermal data derived from a variety of ground-based remote sensing instruments can be used to observe different aspects of volcanic activity at Santiaguito in order to gain insights into the eruption dynamics from the sub-surface conduit level to the vent and to sub-aerial eruptions of ash plumes and emplacement of block lava flows. The hazards posed by these volcanic phenomena preclude direct temperature measurement and the spatial and temporal scale at which these processes operate lies beyond the current capability of satellite-based approaches. In addition, these instruments were not designed specifically for volcano monitoring applications and this thesis will highlight the instruments' capabilities as volcano monitoring tools.

I begin by using a combination of thermal infrared thermometer, infrasound and seismic data to define the likely source and conduit ascent conditions for magma feeding the persistent explosive and extrusive activity at Santiaguito's active vent (Chapter 2). I intend to use the integrated system as a way to characterize individual explosions in a consistent manner in order to gain a better estimate of eruption frequency and relative intensity, and to monitor temporal changes in those two parameters. The delay in the arrival between
thermal and acoustic signal will be used to constrain the source depth of the explosion, providing field-based constraints on conduit parameters critical to theoretical modeling of volcanic conduits.

In chapters 3 and 4 I move beyond the sub-surface and look at the vent thermal configuration and its short-term evolution, which is very important when selecting the best type of satellite data to use for this type of volcano. In Chapter 3 I used an infra-red thermometer and a spectroradiometer to observe vent activity at Santiaguito. The high temporal resolution of the instrument allowed me to observe how the surface temperature varied with time and how the surface responded to intermittent ash eruptions that occurred at the vent. In this chapter I also show how multiple thermal components within the field of view can be distinguished, providing a more detailed thermal structure of the vent. I then use thermal infrared image data to map the spatial dynamics of the vent structure, and relate this to the arrival of a plug-like extrusion at the vent. The resulting structure fed emission of an active block lava flow from its eastern sector and explosive emissions from annular fissures that bound the central plug (Chapter 4). Finally, I use thermal camera and high spatial resolution satellite data to examine the heat loss and emplacement properties of the active lava flows that the active vent feeds (Chapter 5), as well as the emission and dynamics of initial ascent of the mildly explosive ash plumes that are emitted every ~30 minutes (Chapter 6).

The use of different ground-based thermal remote sensing instruments in this study (single-band thermal infrared, multispectral short-wave infrared, and
thermal infrared image data) further facilitates comparison between thermal data types. In my conclusions, I am thus able to highlight the strengths and weaknesses of each instrument and provide recommendations regarding how thermal infrared instruments can best be used to study active lava dome processes and dynamics.

1.2. Monitoring volcanoes using thermal remote sensing

Volcanic eruptions involve the release, or transfer, of thermal energy into the surrounding atmosphere. The detection and monitoring of this released thermal energy can provide information which observations made at visual wavelengths lack. There are many examples of detection and monitoring of temperatures of volcanic surfaces to yield insights into the dynamics of the processes involved in generating that heat. Surface temperature measurements can be used, for example, to gain insights into sub-surface processes and structures (e.g. Hardee 1982, Aubert 1999, Calvari and Pinkerton 2004). In addition, heat budgets of volcanoes can be constrained (Oppenheimer 2004, Wright and Flynn 2004) and cooling rates and emplacement properties of active lava flows, lakes and domes can be quantified (e.g. Oppenheimer 1991, Oppenheimer et al. 1993, 2004, Flynn et al. 1994, Hon et al. 1994, Keszthelyi 1995, Wooster et al. 2000, Harris et al. 1998, 2005, Lodato et al. 2007). Further, satellite and ground-based thermal sensors can be used to quantify the dynamics of explosively emitted ash plumes (e.g. Holasek and Self 1995, Ripepe et al. 2002, 2005, Calvari et al. 2006, Rosi et al. 2006, Patrick et al. 2007). One of the
more critical aspects of thermal monitoring is the change in surface temperature that often accompanies heightened volcanic activity and in some cases serves as precursors to volcanic eruptions. Examples of such precursory changes in released heat are the increase in temperatures of fumaroles prior to eruptions (e.g. Stoiber and Williams 1990), increasing temperatures of crater-lakes prior to phreatic explosions, as witnessed at Kelut and Ijen volcanoes, Indonesia (Sudradjat 1991) or increases in the magma level in an open conduit (e.g. Dehn et al. 2001).

Some volcanic processes are sufficiently benign to allow in-situ measurement of temperature, such as measuring surface temperature of basaltic lavas and fumaroles using thermocouples and hand-held infrared thermometers (e.g. Hon et al. 1994, Harris and Maciejewski 2000, Keszthelyi et al. 2003). For most volcanic activity, in particular all explosive types, use of such contact-based approaches to measure temperature will compromise the safety of the observers. Thus remote sensing techniques are often used and relied upon to gather thermal data safely.

Developments in satellite-based techniques for retrieving thermal data over volcanic surfaces since the 1980s (e.g. Rothery et al. 1988, Oppenheimer 1991, Oppenheimer et al. 1993, Flynn et al. 1994, Harris et al. 1997, 1998, Wooster et al. 2000) have allowed for routine measurement and monitoring of volcanic activities (e.g. Harris et al. 2001, Wright et al. 2004). Satellite-based thermal data, however, often lacks the necessary spatial resolution to provide detailed information across surfaces whose temperatures may vary by 100's of
degrees centigrade on a spatial scale of centimeters or of a temporal resolution that allows examination of short-term changes in surface temperature associated with rapidly evolving, transient volcanic events, such as the emission and initial ascent of an ash plume (seconds). In most cases satellite-based methods can be used to extract a single pixel-integrated temperature from pixels with sizes that range from 30 m (Landsat TM data), through 1 km (MODIS, Figure 1.2) to 4 km (GOES data). Across such pixels surfaces are rarely thermally homogeneous, but are instead comprised of many thermal components radiating at different temperatures. Methods have been developed to extract the size and temperature of two to three sub-pixel thermal components over active lavas (Rothery et al. 1988, Oppenheimer 1991, 1993, Harris et al. 1997, 1999, 2003, Wooster et al. 2000, Donegan and Flynn 2004, Lombardo et al. 2004). However, such approaches require assumptions regarding the temperatures of one or two of the thermal components (Wright and Flynn 2003). In addition, temporal resolutions range from 16 days (for Landsat TM) to 15 minutes (for GOES). The 15-minute resolution data have been shown capable of tracking large eruption plumes (e.g. Holasek and Self 1995, Woods et al. 1995) and the progression of effusive eruptions (e.g. Harris and Thornber 1999, Harris et al. 2000, Mouginis-Mark et al. 2001). However, even a temporal resolution of 15 minutes is too slow to fully document short-lived or rapidly evolving volcanic events, such as Strombolian eruptions which evolve over durations of tenths of a second, and can be over in a few tens of seconds. These spatial and temporal limitations have led to development of improved ground-based thermal measurements, using
Figure 1.2. Image taken from MODIS near real-time thermal monitoring tool at http://modis.higp.hawaii.edu/, showing a thermal anomaly at the summit crater of Caliente vent. MODIS has a temporal resolution of 1 day and a spatial resolution of 1 km.

1.3. Measuring temperature from a distance

1.3.1. Radiometer

Early field-based non-contact methods for measuring temperatures of volcanic surfaces involved the use of optical pyrometers. An optical pyrometer operates within the visible portion of the electromagnetic spectrum by matching the color of the observed target with a calibrated glowing filament within the instrument's field of view. The visible-light calibration of this instrument imposes a lower detection limit, this being the longest visible red that the eye can detect. The lowest detectable temperature limit is given by Macdonald (1972) as 475 °C, meaning that any surface at a temperature lower than this cannot be distinguished by an optical pyrometer. This clearly poses a problem given that volcanic surfaces and features can range in temperatures from around ambient to ~1200°C.

The use of radiometers, although not specifically designed for volcanological applications originally, allows for wider ranges of temperatures to be measured from ambient to magmatic (e.g. Birnie 1973, Oppenheimer and Rothery 1991, Hon et al. 1994, Harris and Maciejewski 2000, Harris et al.
2005b). They are also able to sample at rates of 50 – 100 Hz. They have thus been used to track and measure lava flow cooling (e.g. Hon et al. 1994, Keszthelyi et al. 2003), lava lake and tube dynamics (e.g. Harris et al. 2005d, Witter and Harris 2007), fumarolic and open vent degassing (e.g. Harris and Maciejewski 2000, Harris and Ripepe 2007), as well as gas jet and explosive dynamics (e.g. Ripepe et al. 2002, 2005, Johnson et al. 2005, Rosi et al. 2006).

Radiometers that operate in the thermal infrared wavebands are also known as IR thermometers. These radiometers collect radiant energy in the long wavelength infrared portion of the electromagnetic spectrum (8 – 14 µm) using a single optical lens to give a single field-of-view-integrated temperature for a specific target area. This target area is proportional to the angular field of view (FOV) of the optical lens and the distance between the instrument and the target. Single-channel radiometers cannot distinguish multiple thermal components that are smaller than the instrument’s field of view, much as single-band satellite data cannot be used directly to distinguish sub-pixel thermal features. Spectral radiance \( R \) (in \( \text{W/m}^2/\text{sr/µm} \)) emitted at a given wavelength \( \lambda \) (in µm) varies as a function of surface temperature \( T \) (in K) following Planck’s Law,

\[
R(\lambda) = \frac{c_1}{\pi \lambda^3 c_2} \frac{1}{\lambda T} - 1
\]
where \( c_1 \) and \( c_2 \) are constants with values of \( 3.74 \times 10^8 \text{ W m}^{-2} \) and \( 1.44 \times 10^4 \text{ m K} \) respectively. Conversion of measured spectral radiance to brightness temperature \( (T_b) \) is then given by the inverse of Planck's relationship,

\[
T_b = \frac{c_2}{\lambda \ln\left(\frac{c_1 \lambda^5}{\pi R} + 1\right)}
\]

Temperature values calculated are radiant temperatures because they have not been corrected for the emissivity of the object. To revert to kinetic temperature requires a further correction to the imperfect transmission of radiance by the atmosphere.

In this dissertation I used two types of radiometers. In chapter 2, I used data collected with a Raytek Raynger Infrared (IR) Thermometer (Figure 1.3a). This instrument uses a thermopile detector, has a spectral response in the 8 - 14 µm waveband, and a temperature range of 30 - 1200 °C. This instrument features a digitally adjustable emissivity setting between 0.1 - 1.0, a scope optic allowing manual targeting, and a 1° field of view (FOV). The accuracy of measurements is ±1°C or ±1% (whichever is greater) and the response time is 0.7 ms. For chapter 3, I used a second type of instrument. This was a ‘DUCK’ thermal monitoring unit as developed at the Hawaii Institute of Geophysics and Planetology (Harris et al. 2005b). This instrument was designed to be low-cost, robust and portable, allowing for easy deployment in harsh volcanic conditions. This portable instrument is equipped with an Omega™ OS554 IR thermometer sensor housed in a Pelican™ case fitted with a selenium-germanium-arsenide
Figure 1.3. Infrared radiometers used for the work thermal remote sensing studies at Santiaguito. a. Raytek Raynger 3i, image from http://www.rayteknorthamerica.com/, and b. DUCK portable thermal monitoring sensor (image from Harris et al. 2005).
window (Figure 1.3b). The OS554 sensors have a spectral response in the 8 – 14 µm range, a temperature measurement range of -13 to 1371 °C (at a 1 °C resolution) and an accuracy of 1°C or ±1%, whichever is greater. The response time for the sensor is 0.25 s. Three of these units were used in chapter 3, two with a 1° field of view (FOV) and 1 with a 60° FOV. This instrument can be attached to a portable data-logger, battery and solar panel allowing for high (100 Hz) sampling rates. A more detailed description of these thermal sensors can be found in Harris et al. (2005).

1.3.2. Spectroradiometer

A spectroradiometer is a type of radiometer that measures radiant flux over a spectrum of wavelengths, making measurements at a large number of discrete, narrow wavebands (as opposed to a single, broad waveband as is the case for thermal radiometers). I use a spectroradiometer in chapter 2, in conjunction with limited video and radiometer observations, to obtain the thermal structure of the Santiaguito vent area. The spectroradiometer utilized in chapter 2 is an Analytical Spectral Devices (ASD) FieldSpec Pro FR (Figure 1.4a). This spectroradiometer has a spectral range of 0.35 – 2.5 µm. Three detectors are used to measure different portions of the spectrum. The first detector is a 512 element Si photodiode array which measures the Visible/Near Infrared (VNIR) portion of the spectrum, or the 0.35 to 1 µm wavelength range. The sampling interval of the VNIR detector is 1.4 nm with a spectral resolution of 3 nm at the 0.7 µm wavelength. For the shortwave infrared (SWIR) portion of the spectrum
(which extends from 1.0 – 2.5 µm) measurements are made by two separate TE-cooled, graded index InGaAs photodiodes. The first SWIR detector measures radiation between 0.9 and 1.85 µm and the second measures between 1.7 and 2.5 µm. The sampling interval for the SWIR detectors is 2 nm with spectral resolution varying between 10 – 12 nm. Improvements can be made to the signal-to-noise ratio by internal measurement of dark currents prior to the actual measurement period. The instrument can be fitted with a number of different optical lenses depending on the field of view (FOV) required. Radiation can be measured using the bare fiber optic cable which has a FOV of 25°. Alternatively the instrument can be fitted with fore-optic devices, decreasing the FOV to either 8° or 1°.

The data collected by the spectroradiometer produce near-continuous spectra from the ultraviolet (UV) to SWIR (Figure 1.4b). This allows determination of multiple thermal components within the FOV, instead of the single FOV-integrated brightness temperature measured by thermal radiometers. Flynn (1992) developed an algorithm to discriminate sub-FOV thermal components using radiant spectra collected from a lava lake on Kilauea volcano, Hawaii, using a numerical model with two and three-radiating components. I applied the two-component technique developed by Flynn (1992) to data collected from Santiaguito in Chapter 2. Here I summarize the algorithm and refer readers to Flynn (1992) for a complete and rigorous description of the algorithm.
Figure 1.4. a. ASD FieldSpec Pro FR spectroradiometer (image from http://www.uhb.fr/gstb/images/ASD-MATOS.jpg). b. An example of raw spectral radiance curve collected at Santiaguito.
The surface observed within the FOV of the spectroradiometer is assumed to be comprised of two thermal components: a hotter component radiating at temperature $T_h$ and occupying a fractional area $f_h$ of the FOV, and a colder component radiating at temperature $T_c$ occupying the remaining fractional area $(1 - f_h)$ of the FOV. The total radiant flux ($R_{total}$) detected as a function of wavelength is then the sum of the radiant flux from each component multiplied by their fractional areas,

\[
R_{total}(\lambda) = R_h(\lambda) \cdot f_h + R_c(\lambda) \cdot (1 - f_h)
\]

This equation contains three variables ($f_h$, $T_h$ and $T_c$) and one measurement ($R_{total}$). Given the high spectral resolution of the spectroradiometer, these variables can be solved by measuring radiance values at three different wavelengths [$R(\lambda_1)$, $R(\lambda_2)$, $R(\lambda_3)$], allowing me to set up and solve the following system of simultaneous equations,

\[
R(\lambda_1) = \varepsilon \tau [R_h(\lambda_1) \cdot f_h + R_c(\lambda_1) \cdot (1 - f_h)]
\]

\[
R(\lambda_2) = \varepsilon \tau [R_h(\lambda_2) \cdot f_h + R_c(\lambda_2) \cdot (1 - f_h)]
\]

\[
R(\lambda_3) = \varepsilon \tau [R_h(\lambda_3) \cdot f_h + R_c(\lambda_3) \cdot (1 - f_h)]
\]
where $\varepsilon$ is emissivity (assumed to be constant at 0.9) and $\tau$ is atmospheric transmissivity. The algorithm developed by Flynn (1992) generates Planck curves from mixing the two thermal components in varying proportions iteratively until the best fit with the spectral curves measured by the spectroradiometer is obtained. The shape of the spectral curve varies with the temperature of the radiating body and the proportional FOV coverage of the two thermal components with each mixture giving unique solutions at the three wavelengths.

For the analysis in chapter two, I use six wavelengths selected to fully characterize the spectrum, from well within the 0.35 – 2.5 µm atmospheric windows, meaning that transmissivity is maximized. The wavelengths that I selected were 1.50, 1.63, 1.75, 2.08, 2.17, 2.25 µm. The algorithm then varies the three unknowns ($f_h$, $T_h$ and $T_c$) and at each iteration a Planck-derived spectral curve is generated and the ratio of radiant flux from each of the selected wavelengths are compared to the collected spectrum until the weighted error reaches a minimum value.

### 1.3.3. Thermal imaging using a Forward-Looking Infrared Radiometer

Radiometers and spectroradiometers both collect radiant flux from within a single FOV. To observe thermal structures over a large area it requires retargeting of these point-measurement instruments many times. Thermal imagers, such as the Forward-Looking Infrared (FLIR) camera, solve the problem of the limited field of view provided by point temperature measurements. Using the FLIR, numerous discrete temperature values can be measured
simultaneously through acquisition of a 320 X 240 pixel image (76,800 individual data points). Each pixel provides a calibrated temperature corrected for both emissivity and atmospheric effects. These instruments also have an advantage over space-based imaging in their capability of collecting centimeter-resolution thermal images at a rate of 30 images per second. They produce thermal video footage useful in studying dynamics of fast-moving lava flows (e.g. Calvari et al. 2005, Bailey et al. 2006, James et al. 2007, Lodato et al., 2007), measuring heat loss from lava channels and tubes (e.g. Wright and Flynn 2004, Harris et al. 2005, Coppola et al. 2006, 2007, Witter and Harris 2007), understanding degassing dynamics (e.g. Harris and Ripepe 2007) and the thermo-dynamics of small strombolian eruption plumes (e.g. Patrick et al. 2007) as well as more violent explosive emissions (e.g. Calvari et al. 2006).

The thermal imaging camera used in chapters 4, 5 and 6 is a ThermaCAM™ model P40 from FLIR Systems, Inc (Figure 1.5). This thermal video camera is equipped with a 320 X 240 element focal plane array (FPA) of uncooled microbolometers with spectral response in the 7.5 to 13 µm wavelength range. The optical system provides a total FOV of 18° X 24° with a spatial resolution of 1.3 milliradians. The thermal sensitivity of the instrument is 0.08 °C at 30 °C with a total temperature measurement range of -40 to 1500 °C at three different gain settings. The gain settings are -40 to 120 °C, -10 to 500 °C and 250 to 1500 °C. Image frequency is 50/60 Hz non-interlaced, which can be sampled at 30 Hz via firewire connection to a personal computer (Patrick 2006), or can be stored (at a lower frame rate of ~1 image every 3 seconds) on the
Figure 1.5. a. FLIR thermal video camera aimed at the target area, Caliente dome, 4.5 km away. b. A FLIR image showing Caliente dome with a plume emission and Santa María peak to the right of the image.
onboard Random Access Memory or to a CompactFlash™ memory card. Focusing can be achieved automatically or by manual control. The total radiance collected by the detectors is the sum of the emitted target radiance ($R_{obj}$), the reflected target radiance ($R_{ref}$) and emitted atmospheric radiance ($R_{atm}$),

\[ R_{\text{total}} = \varepsilon \tau R_{obj} + (1 - \varepsilon) \tau R_{ref} + (1 - \varepsilon) R_{atm} \]  

(1.7)

Solving for the emitted target radiance is done by rearranging equation 1.7,

\[ R_{obj} = \frac{1}{\varepsilon \tau} R_{\text{total}} - \frac{1 - \varepsilon}{\varepsilon} R_{ref} - \frac{1 - \varepsilon}{\varepsilon \tau} R_{atm} \]  

(1.8)

For emissivity and atmospheric transmissivity corrections the camera requires a user input target emissivity value and runs a built-in LOWTRAN correction routine (Selby et al., 1978). This requires the user to input values for instrument-to-target distance, relative humidity and temperature. $R_{obj}$ is then converted to temperature of the object ($T_{obj}$) using a Planck function (equation 1.2).

1.4. Study Area

1.4.1. Volcanic Setting

The main focus of this dissertation is the utilization of ground-based thermal remote sensing instruments (and integration with geophysical data and visual observations) to gain an understanding of the dynamics of volcanic eruptions at the Santiaguito lava dome complex in Guatemala. Santiaguito is a
dacite lava dome complex comprised of four lava dome units that have been persistently active since initial extrusion in 1922 (Rose, 1987). The volcanic complex is located approximately 12 km southwest of the city of Quetzaltenango and within the 1902 explosion crater of Santa María volcano. Santa María and Santiaguito are part of the Central American volcanic chain formed by the subduction of Cocos plate beneath the Carribean plate. This volcanic chain, an extension of the Sierra Madre volcanic range, separates the Guatemalan highlands to the north and the low-lying coastal plains to the south. The estimated age of Santa María volcano, based on paleomagnetic anomalies, is ~30,000 years, where the majority of cone construction occurred by 25 ka (Rose et al. 1977). Following extended periods of dormancy and rare extrusion of small-volume lava flows (Conway et al. 1994), Santa María erupted explosively in October 1902 in what is now known to have been one of the largest volcanic eruptions of the 20th century (Stoiber and Rose 1969, Williams and Self 1983). Approximately 7.5 km³ of dense dacite was erupted to feed a plinian eruption column that reached heights of 30 km (Stoiber and Rose 1969, Williams and Self 1983). The eruption removed the southwestern flank of the Santa María cone, creating an explosion crater in this sector. Over the next 20 years (1902 – 1922) the level of activity remained low at Santa María, being limited to weak ash emissions and geyser activity from vents on the floor of the 1902 crater (Rose 1972a). In 1922 lava extrusion began at the center of the 1902 crater, over the location where the main conduit for Santa María is thought to have been located (Rose 1972b).
1.4.2. Volcanic Activity at Santiaguito, Guatemala

The history of Santa María and Santiaguito is divided into three-phases; (1) construction of the Santa María cone, characterized by initial eruptions of basaltic lavas and gradual transition to more silicic lavas; (2) the 1902 cataclysmic plinian eruption of Santa María; and (3) the current phase of dome construction to build the Santiaguito complex (Rose 1987). Persistent lava effusion since 1922 has constructed a lava dome complex approximately 1 km$^3$ in volume, centered on four distinct, westward-younging vents: El Caliente (the principal vent), La Mitad, El Monje and El Brujo.

The effusive activity that has persisted since 1922 includes both endogenous growth and exogenous extrusion in the form of thick block lava flows. Effusive activity at Santiaguito can be defined by cycles of higher extrusion rate followed by longer periods at low extrusion rate (Rose 1972b, Rose 1987, Harris 2003). Six cycles have been defined for Santiaguito during the period 1922 – 1984 (Rose 1972b, Rose 1987). These are detailed in Harris (2003) and updated to 2000, at which point an 8th cycle was identified as beginning. The first cycle began with the first dome extrusion at the Caliente vent in 1922 and lasted until 1929. The initial high extrusion phase for this cycle occurred from 1922 – 1925, during which time a total of 0.2 km$^3$ of lava was emplaced at a peak extrusion rate of 2.06 m$^3$/s. Effusive activity was accompanied by explosive eruptions and nuee ardentes (Rose 1972b). The following four years (up to 1929) was a period of low extrusion (< 0.19 m$^3$/s). The second cycle was again focused at the Caliente vent and began with an initial high extrusion rate phase
between 1929 and 1934 with a peak extrusion rate of 0.57 m³/s. This was followed by a low extrusion rate phase lasting until 1939 (< 0.19 m³/s). The third cycle (1939 – 1949) marked a shift in vent activity to the west of the principal Caliente vent (Figure 1.6). During the third cycle, three years of high extrusion rate activity (0.95 m³/s) emplaced a dome unit and a lava flow from a vent 700 m west of Caliente, building the La Mitad unit (Figure 1.6.). Following the seven-year low extrusion rate phase of the third cycle, the fourth cycle began with another westward shift of vent activity to El Monje (Figure 1.6). This dome and a block flow unit was built and extruded during a six-year-long initial high extrusion rate (1.24 m³/s) phase followed by a low extrusion rate phase lasting until 1958. In 1958, the fifth cycle marked a further westward shift in vent activity to a location 1.5 km west of Caliente at El Brujo (Figure 1.6). The initial high extrusion rate occurred from 1958 – 1963 (1.49 m³/s) and the low phase occurred from 1963 – 1972 (0.03 – 0.13 m³/s). This cycle marked the beginning of increased prominence of block lava flow activity (Rose 1987). The sixth cycle marked simultaneous extrusions at both El Brujo and Caliente from 1972 – 1986. The initial high extrusion rate phase (0.95 m³/s) occurred at both vents with a 2 km long block flow being emplaced from El Brujo. Activity at El Brujo ceased in 1977. Conversely activity continued at Caliente and, since 1975, its activity has been marked by emplacement of block lava flows of increasing length (Harris et al. 2003) along with regular eruptions of vertical ash plumes at a rate of ~ 2 events per hour (Rose 1987).

The seventh cycle began in 1986 with an initial phase extrusion rate of 0.8
* Observation sites used in this study
• Approximate location of El Brujo vent
• Caliente vent
• 1922-34 lava
• 1939-58 lava
• 1958-66 lava
• 1967-75 lava

Figure 1.6. Map of extrusive units at Santiaguito Lava Dome Complex. The three observation sites from which measurements are made for this dissertation are shown as filled stars. Data for chapter 2 was collected from site #2, chapter 3 from site #1, chapter 4 from site #1, chapter 5 and 6 from site #3.
m$^3$/s, during which, two block lava flow units were emplaced (Harris et al. 2003). In 1989 the low extrusion phase began. This phase was characterized by rates fluctuating between 0.07 and 0.37 m$^3$/s, building series of relatively short lava flows around the Caliente vent and, in 1996, breaching the 1902 crater rim. Cycle 8 began in 1996 with an initial phase extrusion rate of ~ 0.48 m$^3$/s. This phase was on-going as of January 2002, by which time a channel-fed block lava flow had reached ~3.9 km from the vent (Harris et al. 2003).

1.4.3. Observation Sites

The persistence of extrusive and explosive activity at Santiaguito means that it represents a location where data for such events can be collected. In addition, the volcanic setting at Santiaguito provides accessible vantage points for observing the activity from a safe distance. The peak of Santa María volcano is located 2500 m to the northeast of, and 500 m higher in elevation than Santiaguito (Figure 1.6). This site provides a clear birds-eye view into the summit crater vent at Caliente where current effusive and explosive activity is focused. Integrated radiometer, spectroradiometer and video datasets presented in chapter 2 and the FLIR datasets of chapter 4 were collected from this observation site. Lateral views of the dome can be achieved from a number of locations to the east, south and west (Figure 1.6), two of which are used in this study. Integrated thermal, seismic and acoustic data used in chapter 3 were collected from a site located atop the El Brujo dome, approximately 1.6 km to the west of Caliente vent. FLIR data of the block lava flows and the plumes are
collected from a site on top of a ridge, 3.8 km to the south of Caliente vent. These data were used in Chapters 5 and Appendix A.

1.5. Thesis Structure and Co-Author Contribution

Each of the following chapters was initially written in formats appropriate for publication in peer-reviewed journals, and is here adapted from that submitted format. However, I have attempted to give additional technical and background information in this opening (Introductory) chapter, and within some of the Chapters themselves (especially Chapter 5 which is expanded from a short Geophysical Research Letters submission). I am the primary source of >90% of the work presented in this dissertation, where all work was planned and completed by me under the guidance of my advisor (Luke Flynn) and committee members (Andrew Harris and Robert Wright). The work was also supported by NASA Earth System Science Fellowship, NASA grants NCC 5-461 and NGT5-40076, NASA Pathfinder Grant 59413 and NSF grants EAR-0106349 and EAR-0207734.

- In Chapter 2 I present the results from an integrated thermal, seismic and acoustic data set for explosions at Santiaguito's Caliente vent collected in January 2003. This work has been submitted to Journal of Volcanology and Geothermal Research and has been accepted pending minor revision. The co-authors for this submission were Andrew Harris and Emanuele Marchetti. While Harris assisted with the experiment design and field deployment, Marchetti helped with discussion and advice regarding
infrasound and seismic data reduction. As primary author, I completed all
the data reduction, analysis, and interpretation. I also developed all
models and ideas guided by my co-authors.

- In chapter 3 I present the results from simultaneous radiometer,
spectroradiometer and video observations of the surface of the Caliente
vent during January 2002. This chapter has now been published in the
Geophysical Research Letters (Sahetapy-Engel et al., 2004). The co­
authors for this paper were Luke Flynn, Andrew Harris, Greg Bluth,
William Rose and Oto Matias. Bluth, Rose and Matias helped with
fieldwork and provided video data. Harris and Flynn assisted with the
collection of all thermal data. Flynn also contributed discussions on
temperature extractions from spectral radiance data. I completed all data
processes, analyses and temperature extraction from spectral radiance
data. I developed all interpretations and conclusions with guidance from
Flynn and Harris.

- In chapter 4 I present the results of thermal imaging of the dome surface
completed during January 2004. This work has been submitted to Bulletin
of Volcanology with Andrew Harris as a co-author. I collected all FLIR
images used in this study with the assistance of Robert Wright. I
completed all data analyses and interpretations. I developed all the ideas
and conclusions presented in this paper through discussion with Harris.

- Chapter 5 deals with the thermal characteristics and heat budget of a
silicic (block) lava flow using thermal image data collected in January 2005
and ASTER satellite data. This chapter has been submitted to Geophysical Research Letters and, at the time of writing, was in the process of being re-submitted with Andrew Harris as co-author. This resubmission will be a distilled version of the chapter focusing on the new heat loss model. For this chapter, I completed all field data collection with the assistance of Robert Wright. I also identified and ordered cloud-free ASTER satellite data. I completed all analyses, interpretations and writing, with Harris advising on extraction of temperature data for lava flow heat loss study and design of a new heat loss model for silicic flows.

- In chapter 6 I present a summary of each of the studies presented in the dissertation and use the results from this set of Santiaguito-focused studies to test the capability of satellite-based data to detect thermal anomalies and explosions at systems characterized by Santiaguito-style activity.

- In Appendix A I present a thermal image data set for the ash plumes erupted from Santiaguito during January 2005. This has been submitted as a short-communication to the Bulletin of Volcanology with Andrew Harris as co-author. I completed all data collection, analyses, interpretations and conclusions, with advice and guidance from Harris.
Chapter 2
Thermal, seismic and infrasound observations of persistent explosive activity and conduit dynamics at Santiaguito lava dome, Guatemala

2.0. Abstract
In January 2003 I deployed a seismo-acoustic array along with thermal sensors to define the dynamics and source characteristics of persistent, intermittent explosive activity at the Santiaguito lava dome, Guatemala. Seismic and acoustic waveforms accompanying thermal transients (i.e. the thermal signature produced by the vertical ash plumes) allowed discrimination of 35 explosions from minor degassing events during the 15.5 hour observation period. Characteristics of thermal transients along with elastic and thermal energy flux from the explosive ash emissions allowed for a standardized measurement of eruption intensity, duration, frequency and repose interval. Using the thermal data I calculated a range of minimum exit velocities of 16 – 76 m/s and convective rise rates of 9 – 26 m/s to heights of 100 and 600 m above the vent. Source depths for the explosions were calculated using the thermo-acoustic delay, indicating a source region 100 – 620 m below the vent. The intensity of the explosions, based on thermal amplitude and proxies for elastic energy released is not dependent on repose interval or depth. The data support the notion for shear-induced fragmentation at the conduit walls due to stick-slip movement of the upper degassed part of the magma column as a source mechanism. Loci of explosion
depths suggest that this operates within a 500 m thick dacite plug in the uppermost portion of the conduit.

2.1. Introduction

Intermittent small-to-moderate-sized explosions (Volcanic Explosivity Index (VEI) of 2) are commonly observed at volcanoes which are characterized by persistent eruptive activity (Francis 1993). This type of activity has persisted at the dacite lava dome complex of Santiaguito, Guatemala, since 1977, where repeated low energy explosions sending ash plumes 1 to 4 km above the vent occur at a rate of approximately 1 event every 30 minutes (Rose 1987). The intensity and frequency of explosions at Santiaguito are similar to the explosions observed at Stromboli, Italy, and have led previous workers to incorrectly classify these explosions as Strombolian or Strombolian-type eruptions (Bluth and Rose 2004, Johnson et al. 2004). The characteristic of the ash plumes produced by these explosions however more resembles that of weak vulcanian eruptions.

Similar styles of repetitive, low-intensity explosions at both basaltic (e.g. Stromboli) and silicic (e.g. Karymsky in Russia and Arenal in Costa Rica) volcanoes have been observed to generate both seismic and acoustic energy in the infrasonic range from an explosion source within the conduit. Integrating seismic and acoustic observations of these explosions have provided insights into the source mechanism of the explosions within the conduit (e.g. Braun and Ripepe 1993, Vergniolle and Brandeis 1996, Johnson et al. 1998, Chouet et al. 1999, Hagerty et al. 2000). Seismo-acoustic studies of such explosions at
basaltic system such as Stromboli, Etna and Erebus suggest that bursting of discrete gas slugs at the magma free surface within the conduit as a source mechanism for generating low-intensity repetitive explosions (e.g. Blackburn et al. 1976, Vergniolle and Brandeis 1994, Braun and Ripepe 1993, Ripepe and Braun 1994, Vergniolle and Brandeis 1996, Chouet et al. 1999, Ripepe et al., 2002). At silicic volcanic systems such as Santiaguito, higher magma viscosity prevents formation of large discrete gas slugs. In addition, the conduit is typically capped by some form of obstruction, such as a dome or a lava plug, making a simple model involving bubble bursting at the free surface improbable. Rose (1987) and Sanchez-Bennet (1992) suggested that, given the catchment-like geometry of Santa Maria and Santiaguito volcanic setting, explosions at Santiaguito maybe phreato-magmatic. Johnson et al. (1998) proposed that similar repetitive explosions at Karymsky (analogous to Santiaguito explosions) are the result of cycles of gas buildup and release beneath an obstruction within the conduit. Recent dynamic conduit flow models (Gonnermann and Manga 2003) have led to a newer interpretation for the explosions at Santiaguito and other similar systems, with the explosions resulting from shear-induced fragmentation due to intermittent rise of the dacite plug (Bluth and Rose 2004).

To define and understand the conduit and eruption dynamics involved in the repeated explosive emissions at Santiaguito, a network of thermal, seismic and acoustic sensors was deployed in January 2003 to record the signals associated with the explosive events. The objectives of this study were twofold. The first goal was to quantify the properties of the explosions, thereby providing a
simple and consistent way for defining the character of these explosions. The second objective was to gain insights into the subsurface source mechanism and conduit dynamics that generate such repeated, mildly explosive events at silicic systems.

Preliminary analysis and review of the dataset was presented by Johnson et al. (2004). Johnson et al. (2004) selected 18 explosions from the dataset and calculated exit and buoyant velocities using the delay in the thermal signal arrivals at two stacked thermal sensors. They also calculated cumulative seismic, acoustic and thermal energy released from these explosions and found positive correlation between thermal energy and buoyant rise rates.

This study builds and expands this preliminary study, and recalculates some of the initially given values. Johnson et al. (2004) considered just a section of the data set (18 explosions). Here I analyzed all 35 explosions within the data set. I also fully detailed and classified the thermal waveforms, while extracting new parameters, such as explosion intensity and duration, as well as velocity. In addition, I extended the Johnson et al. (2004) calculations for plume rise rates and cumulative released energy to examine how total energy released by each explosion is partitioned between thermal, seismic and acoustic energy. This allowed me to determine whether all explosion types are generated by the same source mechanism. Finally, I carried out a new analysis that uses the delay between the arrivals of the thermal signal at sensors aimed at different heights above the vent to calculate the ascent velocity of the plumes. Following Ripepe et al. (2002) I used these velocities, along with the delay between thermal and
acoustic arrivals, to determine the explosion source depth. These results were used to assess the applicability of a variety source models to describe the intermittent explosions at Santiaguito.

2.2. Background: Activity and vent structure at Santiaguito

Santiaguito is a dacite lava dome complex that formed within the 1902 explosion crater on the southwestern flank of Santa María volcano, Guatemala (Rose 1972, Rose 1987). Since 1922, eight cycles of lava extrusion from the central El Caliente vent and three lateral vents (La Mitad, El Monje and El Brujo) have formed a ~1.1 km$^3$ dome complex (Rose 1987, Harris et al. 2003). Since 1977 the locus of volcanic activity has been at the primary El Caliente vent (Rose 1987), the location of the initial extrusive activity during 1922-1939 (Rose 1972). Since this time, intermittent explosions producing vertical ash plumes typically reaching heights of ~1 to 2 km above the vent has been the prominent type of activity (Rose 1987, Figure 2.1) along with extrusions of silicic lava flows (Rose 1987, Harris et al. 2003, Harris et al. 2004).

Regular visual observations of these explosions are made by staff at the Santiaguito Volcano Observatory of the Instituto Nacional de Sismologia, Vulcanologia, Meteorologia e Hidrologia (INSIVUMEH). In addition, a few intermittent observations of the explosions have been made, each of a few hours in duration (Rose, 1987, Sanchez-Bennet 1992, Bluth and Rose 2004, Johnson et al. 2004). However lack of geophysical monitoring capabilities and frequent cloud cover severely limit the amount of observations available for complete
Figure 2.1. Vertical ash plume from an explosion at the active Caliente vent, Santiaguito, Guatemala. Santa María peak and scarp from the 1902 eruption of Santa María is visible to the right of Caliente vent. Small pyroclastic flow travels down a chute on the southeast flank of Caliente dome.
analysis. From visual observations made during 1999 – 2004, the frequency of the explosions is of the order of 1 to 2 events per hour but with occasional clusters of explosions occurring over the time period of a few minutes. The explosions generate vertical ash plumes which rise up to 3 km above the vent, but more typically ~1 km or less. Larger explosions are capable of feeding plumes that rise up to 4 km and generate small pyroclastic flows (Figure 2.1). Bluth and Rose (2004) observed that the plumes generated by the explosions rise from a series of vents which formed a ring around the 180 m-wide summit crater. They proposed that the ring-like configuration of the emission points was the surface expression of the flared top of the conduit. Bluth and Rose (2004) observed changes in the diameter of the ring using frames from video footage collected during the yearly campaigns from 2000 to 2004. They found that the diameter of the ring increased from 80 m in 2000 to 120 m in 2004. Bluth and Rose (2004) attributed this to a possible widening of the conduit. Source mechanisms for these explosions have not been explored in great detail but magma-water interaction (Rose 1987, Sanches-Bennet 1992) and shear-induced fragmentation along the conduit walls due to unsteady flow (Bluth and Rose 2004) have been proposed as potential mechanisms.

2.3. Data Acquisition

The setup for the thermo-seismo-acoustic experiment at Santiaguito follows the design of Ripepe et al. (2002, 2004) for Stromboli Volcano, Italy. I used three thermal sensors operating in the 8 – 14 microns wavelength region,
four band-passed (0.1 – 10 Hz) Panasonic WM-034BY electret condenser microphones and a single vertical short-period L1 Mark Products seismometer. The array was deployed on the summit of El Brujo Dome, approximately 1.5 km to the west of the active vent at the summit of the El Caliente Dome unit.

Data acquisition began at 7:30 pm (all times are local) on January 8, 2003 and ran until 8:02 am January 12, 2003. Data were recorded using an 8-channel DataQ DI-700 acquisition system at a sampling rate of 122 Hz. Due to weather and instrument malfunctions, only the first 15 hours and 21 minutes of the dataset contain coherent signals across all sensors (Figure 2.2a). During cloudy conditions, for example, the thermal data cannot be collected because they require a cloud-free line of sight to the vent. From this subset 35 explosions can be distinguished from their distinctive thermal, seismic and acoustic waveforms (Figures 2.2 and 2.3), giving a time-averaged frequency of 1 explosion every 26 minutes. Comparison with daytime video data available for some of the explosions confirms that seismic, acoustic and thermal signals all correspond to explosive events producing vertical ash eruptions. Thermal, seismic and acoustic waveforms from these 35 explosions were extracted for analysis (Figure 2.2a). All parameters discussed in the following sections are derived from the dataset given in Figure 2.2a unless otherwise noted. The three thermal sensors used were designed specifically for near-vent deployment (Harris et al. 2005). They are equipped with an Omega™ OS554 thermal infrared (IR) sensor operating in the 8 – 14µm wavelength range. At Santiaguito, one of the sensors was a 15° field of view (FOV) instrument (T₁, Figure 2.2b) and the other two (T₂ and T₃,
Figure 2.2. a. First 15 hours and 21 minutes of data during which coherent signals were recorded across all sensors. The 5 channels shown from the top to bottom are seismic, acoustic, and the three channels of thermal data. The field of view targeted by the sensors cover zones marked T_1, T_2 and T_3 in the plume on Figure 2b. Note the effect solar heating and cooling on the thermal dataset. b. Approximate vertical positioning of the thermal sensors field of view and thermal transients produced by rising plumes for each of the sensor. Note the time delay between the onset of the event for the lower (T_1 + T_2) sensors and the upper sensor (T_3).
Figure 2.3. Seismic, infrasonic and thermal record of a single explosion from Santiaguito.
Figure 2.2b) were 1° FOV instruments. The line of sight for instrument T1 was inclined at an angle of 12 ± 3° from the horizontal, so that the field of view was roughly centered over the vent (Figure 2.2b). At a sensor-to-target distance of 1.5 km, this gave a 136,000 m² target area at the vent. This spot included a portion of the dome flank, sky and, during explosive events, plumes (Figure 2.2b). The line of sight for instrument T2 was also inclined at an angle 12 ± 3° from the horizontal, again so that the field of view was roughly centered over the vent. At a distance of 1.5 km, this gave a 670 m² FOV centered at approximately 100 m above the vent (Figure 2.2b). Instrument T3 was inclined at 29 ± 3° from horizontal, giving a 1050 m² FOV centered at 600 m above the vent (Figure 2.2b). This configuration was intended to allow the thermal arrival of the ascending plumes to be obtained at three different vertical positions: at the level of the vent (T1), at 100 m (T2) and at 600 m (T3) above the vent.

The thermal sensors have an instrument response time of 0.25 s, so that, although detection of the thermal signal is instantaneous, a delay of 0.25 s will occur before the full signal is registered by the sensor (Harris et al. 2005). This feature poses no problems for this study because I am only concerned with the relative timing of the arrival and progress of the plumes through the three vertically aligned fields of view. Furthermore, the temperature recorded by the instrument is an integrated brightness temperature ($T_b$) for a mixed field of view comprised of background (sky and/or parts of the dome) occupying a fraction of the field of view and, during explosive emissions, plume occupying the remainder of the FOV. Thus, as the relatively hot plume ascends through the FOV so its
fractional coverage increases and so the integrated brightness temperature climbs. Because the dataset extends from night to day (Figure 2.2a) I minimized the effect of diurnal variation in background temperature by normalizing brightness temperature recorded during an explosion to the background temperature recorded immediately preceding the onset of an explosion.

The explosions from Caliente vent produce seismic infrasonic and thermal transient waveforms, an example of which is shown in Figure 2.3. Seismic records of explosions consist of low frequency (< 5 Hz) weak emergent transients, lasting 20 – 40 seconds. Spectral analysis of the seismic waveforms points to a narrow frequency content with a peak around 1 – 2 Hz (Figure 2.4). However, deployment of a single component vertical seismometer prevented detailed analysis of the spectral characteristics of seismic waveforms. Infrasound associated with the explosive events ranges from a single infrasonic pulse, a few seconds long, to emergent infrasonic signals lasting up to 30 seconds (Figure 2.4). Acoustic pressure at the vent, estimated from recorded transients corrected from geometrical spreading, typically does not exceed 0.25 Pa. Thermal transients for the explosions reveal an initial, relatively gentle (slow) increase in temperature, followed by a very-long-lasting, smooth decrease to give a long tail to the signal (Figure 2.3). Transients last up to few minutes and the first major peak may be followed by secondary pulses. These thermal waveforms are described in greater detail in the following section. There is a large difference in duration between the infrasonic and thermal transients. This can be explained by an initial gas-thrust phase, which explains the infrasonic signal, followed
Figure 2.4. Detail of the waveform and spectral analysis (Power Spectral Density) of the seismic transient shown in Figure 3.4.
by a longer buoyant phase with which most of the longer thermal signal is associated.

2.4. Thermal Dataset

2.4.1. Thermal waveforms

The emergence and ascent of each ash plume emitted during the acquisition period was recorded in the thermal dataset as a transient thermal waveform. The shape and smoothness of the waveforms vary with the size and the vertical position of the field of view (FOV) (Figure 2.2b). $T_1$ has a FOV with a diameter of $\sim 400$ m. Given a plume as wide as the crater itself ($\sim 180$ m), the fractional area of the FOV that can be occupied by the plume ($a_p$) can reach a maximum of only $50\%$ of the total area of the FOV if the plume ascends to the top of the FOV as a non-spreading cylinder. In contrast, the FOV widths for $T_2$ and $T_3$ are narrow enough ($\sim 26$ m) for $a_p$ to be $100\%$ once the plume has ascended through the entire height of the IFOV. Thus the wider field of view ($T_1$) shows a longer rising limb as the plume ascends the full $300$ m height this FOV. During this ascent period the relatively hot plume will fill an increasing portion of the FOV causing the integrated signal to rise. Because the FOVs for the narrower field of view are smaller, the signal climbs more rapidly because it takes less time for the FOV to move from empty to full. I note, also, that the waveforms for the narrower FOVs are less smooth and show more thermal detail than the wider FOV, and that the onset of the thermal signal is later for the wider FOV (Figure 2.2b).
Thermal waveforms recorded during the experiment could be split into three groups based on waveform shape: (i) single-peaked transient asymmetric (short rising limb and long waning limb, Figure 2.5a), (ii) clustered (multiple peaks, Figure 2.5b) and (iii) symmetric transients (Figure 2.5c). Single-peaked transients represent emission of a single eruption plume, whereas clustered transients result from groups of explosions occurring within minutes of one another to send a series of emissions through the FOVs. In the latter case, the onset of each new thermal waveform from subsequent explosions overprints the waning limbs of the previous explosions. Symmetric transients (Figure 2.5c) have bell-shaped waveforms with shorter duration than the impulsive events. They have a gently sloping onset phase with much lower, but quite broad, peaks (Figure 2.5c). There were no seismic or acoustic signals associated with this type of thermal signal. This type of thermal signature is caused by small gas emissions that rise very slowly from the vent. These were weak events that were observed to rise occasionally from the vent during repose intervals between explosive events or at the tail end of explosions. Given the absence of seismic and acoustic signals, I do not consider these events to be explosion events and they are omitted from my complete analysis. In contrast the impulsive events, both single events and clustered events, have associated seismic and infrasound signal.

I stacked the thermal transients from all 35 explosion events recorded by each of the three instruments to obtain an average waveform showing the typical shape of the thermal transient for the Santiaguito plumes (Figure 2.6a). This
Figure 2.5. Three observed thermal features along with corresponding seismic and acoustic signals, a - simple impulsive event, b - a cluster of impulsive events, c - non-impulsive event with no associated seismic nor acoustic signal. a and b have correlated footage to an explosion event with a single emission and an event with multiple explosions and emissions.
showed an asymmetric shape, with a rapid waxing phase followed by a longer waning phase and a short, flat peak (Figure 2.6a). The waveform can thus be separated into 4 phases, pre-explosion, onset, peak and decay (Figure 2.6b). The pre-explosion phase shows a flat response, with the signal stable at background (ambient) temperature levels ($T_a$, Figure 2.6b). For the background temperature-normalized dataset this defines the zero level, which is subtracted from the elevated levels to achieve normalization (Figure 2.6a). Some explosions showed a gradual increase in temperature prior to the main explosion onset (Figure 2.7a). I interpret these as small gas puffs expelled prior to the eruption of the main plume.

The onset phase of the waveform signifies the start of the main emission (Figure 2.6b), and thus marks the arrival of the plume within the FOV and records its ascent through the FOV (Figure 2.7b). For the lowest FOV ($T_1$), the time at which the signal begins to rise marks the point at which the emission begins to exit the vent. This phase is marked by a rapid rise to a peak temperature ($T_{peak}$). The amplitude of the thermal waveform ($\delta T$) is defined by the background-normalized $T_{peak}$ (Ripepe et al. 2005). The slope of the onset phase was relatively constant in some explosions but not in others, having a number of inflections within the rising phase (Figure 2.7b). The average duration of this phase was 18, 6.5 and 10 seconds for $T_1$, $T_2$ and $T_3$ respectively.

The peak phase of the waveform begins when the onset phase reaches its peak value (Figure 2.6b). Some explosions have a sustained peak phase lasting as long as 60 s while others fail to maintain peak temperature for an extended
Figure 2.6. a. Stacked thermal waveforms of 34 individual single transients. Solid black waveform is the average waveform for each of the three sensors. b. Generalized thermal waveform for Santiaguito, subdivided into 4 stages: pre-explosion, onset, peak and decay. $T_a$ is ambient temperature, $t_o$ is the time at which the plume first enters the sensors FOV. Broadness of peak phase is a measure of plume sustainability.
Figure 2.7.a Thermal signature of an explosion event where a small puff precedes the main emission. b. Comparison between an onset phase with an inflection and one without an inflection. The time taken to reach the first inflection is assumed to be the time it takes for the plume front to travel from the bottom to the top of the IFOV and the remaining part of the phase reflects changes in the plume temperature and optical thickness.
period of time. These two peak types represent the difference between two end-member emission cases: sustained emissions versus short bursts where sustained emissions resulting in broader peaks. The average duration of the peak phase was 9.5 s, with a minimum of < 1 s and a maximum of 62 s.

The final phase is represented by a slow decay to background temperature (Figure 2.6b). The slow return to background temperature is likely a result of the time necessary for the plume material to cool and drift clear of the FOV, as the plume expands and rises. It therefore represents the cooling period of the plume as it ingests ambient air and begins to disperse. It may also be enhanced by the extended degassing or fuming that was observed to follow each vertical ash eruption tracked by Bluth and Rose (2004).

2.4.2. Eruption parameters from thermal data

Variations in integrated brightness temperature ($T_b$) recorded by the thermal sensor during an event are a function of the rate at which the fractional area covered by the plume ($a_p$), plume temperature ($T_p$) and optical thickness / density, all of which relate to how much ash or pyroclasts are present in the FOV. Thus I assume that the maximum temperature recorded ($T_{peak}$) during an event is at the time when there is a maximum amount of ash present within the FOV (Ripepe et al. 2005). Thereafter, the ash begins to disperse and/or cool as it expands and drifts unless further pulses cause ash loading and temperature to recover again. The amplitude of the thermal waveform $\Delta T$, calculated by subtracting background temperature from $T_{peak}$, can then be used as a proxy for
relative plume intensity (in terms of mass loading, temperature, and optical
density). In the cases recorded here $\delta T$ varies between 6 and 48 °C with a mean
value of 19 °C for $T_1$. For $T_2$ it varies between 12 and 62 °C, with a mean of 25
°C, and for $T_3$ it has a range of 2 to 28 °C with a mean of 17 °C. The decrease in
$\delta T$ between the lower FOV ($T_2$) and the higher FOV ($T_3$) is likely the result of
cooling of the plume during ascent as well as plume expansion/incorporation of
air as it rises, expands and disperses.

Although the duration of each emission can be extracted from the length
of the thermal waveform, the actual explosion duration is best determined using
the length of the seismic and/or infrasound signal. In the context of this study,
explosion duration refers to the sub-surface process and emission duration
represent the length of time from initial exit of the plume from the vent to the
cessation of plume emission. Transients recorded by $T_2$ best represent the
duration of the plume emission from the explosion as the drifting plumes will
quickly move through the relatively narrow and low FOV. Plume dispersion
duration (i.e. thermal transient length) for points between 100 and 500 m above
the vent ranges from 0.5 to 15 minutes with a mean of 2.8 minutes. The majority
of the plumes analyzed (31 out of 35) have duration of less than 3.4 minutes.
Longer duration events exhibit extended degassing periods following the main
explosion phase, reflected by an extended waning tail in the thermal waveform.
Duration of the explosions that produce the emissions, as extracted from the
duration of the seismic and acoustic signals ranged from 12 to 49 s (mean of 26
s) for seismic and 1.4 – 93 s (mean of 18 s) for acoustic.
The number of seismic, acoustic and thermal waveforms recorded over a given period can be used to determine the frequency of the explosions at Santiaguito. In order to achieve this, I must first distinguish actual explosions from other low-energy degassing events. I distinguished explosions from low-energy or lazy degassing events based on the presence of seismic and infrasonic signals along with thermal transients. Thus using thermal data alone for an explosion count will yield higher explosion frequency as it will include non-explosive gas exhalations. Conversely using seismic or acoustic signals alone will yield high apparent number of explosions due to the inclusion of other sources such as rockfalls for seismic and wind noise for acoustic, as well as instrument malfunctions for both. There were 35 actual explosions during the 15 hour and 26 minute acquisition period, giving an explosion frequency of 2.3 explosions per hour with repose periods of 1 to 56 minutes. If I distinguish single explosion events and events with multiple explosions, there are 26 events with 6 being clustered events. This gives a frequency of 1.7 events per hour and repose periods of 8 to 56 minutes. Within the same timeframe there were a total of 73 thermal waveforms. This represents the total number of thermal emissions (explosive and non-explosive) and equates to an emission frequency of 4.7 per hour, meaning that 50% of emissions are non-explosive gas exhalations.

2.5. Plume Rise Rate

Previous methods of measuring plume rise rates include the use of photographs (Chouet et al. 1974, Blackburn et al. 1976, Wilson and Self 1982),
acoustic sound detection and Doppler radar (Weill et al. 1992, Seyfried and Hort 1999, Hort and Seyfried 1998) plus the use of video sequences (Formenti et al. 2003, Bluth and Rose 2004). Previous estimates of exit velocity for the plumes at Santiaguito were made by Bluth and Rose (2004) using video data collected from the peak of Santa María volcano. They reported exit velocities of 5 to 30 m/s, with most of the plumes rising at 10 m/s. These values are underestimates of the exit velocity as they represent an averaged rise rate over 1500 m of ascent from the vent level to the height from which their observations were made.

In this study I obtained plume rise rates from the delay in onset of the thermal waveforms between two thermal sensors positioned at different heights in the plume, as well as from a single sensor using the time it takes the signal to move from background to peak. This allowed me to estimate ascent velocities at three levels above the vent and create a rough vertical profile of the rate of ascent for individual plumes. Given the different vertical positions of the FOVs for the three thermal sensors, the delay in the arrival of the thermal transient between the three instruments (Figure 2.3) can be used to derive plume front ascent velocities. Onset of the thermal transient at $T_1$ and $T_2$ records the plume exiting the vent and the arrival of the plume at 100 m level above the vent, respectively. The ascent velocity between these two points can be obtained from the separation distance (100 m) divided by the delay in the arrival of the waveform between the two sensors, $T_1$ and $T_2$. This rise rate is the best approximation to the exit velocity ($V_e$) of individual plumes, being the average velocity over the first 100 m of ascent. $V_e$ values calculated from the $T_1 - T_2$ delay
range from 16 to 76 m/s, with an average of ~40 m/s. Johnson et al. (2004), in an initial analysis of the first 18 explosions on this data set calculated plume rise velocities of 5 to 20 m/s using the onset delay between $T_2$ and $T_3$ and a separation distance between these two sensors of 420 m. I find that the separation distance between $T_2$ and $T_3$ instead to be 520 m and thus recalculated the rise rates for those 18 explosions, as well as for the 17 subsequent explosions, accordingly to give an average ascent velocity between 100 and 620 m above the vent ($V_{123}$) of 10 to 20 m/s, with a mean of 15 m/s.

The second method for calculating plume rise uses the onset phase of the thermal waveform from a single radiometer (Figure 2.7b). This method allows me to measure plume rise rate at a particular height (the height at which the radiometer is aimed) above the vent (Harris et al. 2005b, Rosi et al. 2006). Peak temperature of the thermal waveform corresponds to a time period of maximum plume temperature and plume opacity which correspond to the maximum amount of ash in the field of view or 100% coverage of the field of view by the plume. (Harris et al. 2005b). Using this assumption, the time it takes for the signal to rise from background to peak, i.e. the duration of the onset phase (Figure 2.7b), should be equal to the time it takes for the plume to move from the bottom to the top of the FOV. The average rise rate across the IFOV can then be obtained by dividing the diameter of the FOV by the duration of the onset phase. The average duration of the onset phase for $T_2$, located ~100 m above the vent, was 6 s. Given the 30 m diameter of this FOV, the average plume rise rate is 5 m/s, which appears rather low when compared with the velocities of 17 – 76 m/s obtained
from the $T_1 - T_2$ delay. As mentioned previously, I observed that most of the onset phases of the waveforms had one or more inflections prior to reaching the peak thermal amplitude (Figure 2.5b), suggesting multiple pulses of emission increasing in relative intensity culminating with maximum intensity emission at peak amplitude. Time to the first of the inflections is then the time it takes for the initial plume front to ascend the length of the IFOV (Figure 2.5b). I therefore assume that the time it takes for the plume to ascend the IFOV diameter is just the time it takes to reach the first inflection during the onset phase (Figure 2.5b). This approach was limited by the fact that not all of the thermal waveforms displayed straightforward onset phase; hence, there was some difficulty in determining where the first inflection occurred. The results, however, gave a range in ascent velocities of 15 to 39 m/s, with a mean of 27 m/s. This velocity seems more consistent when compared to the exit velocities ($V_e$) obtained from the $T_1 - T_2$ delay. I would expect the velocities at the 100-m level to be a little lower than those obtained from the $T_1 - T_2$ delay, because the latter velocity is integrated over the first 100 m of rise, which should include faster velocities right at the vent. Similar calculations can be made using the onset phase from $T_3$ waveforms although with much less success as onset phase and/or inflections are even less straightforward compared to $T_2$ waveforms. Nevertheless this gives an idea of plume rise rates at 620 m level above the vent, which ranges from 9 to 26 m/s with a mean of 17 m/s. These ascent velocities ideally should be lower than the $T_2 - T_3$ delay velocities as they represent the upper end of the ascent region, thus should be the slowest values. However the higher velocities
calculated can be reconciled with higher velocities maintained in the lower part of this upper ascent region. Johnson et al. (2004) calculated plume rise rates using data from individual thermal sensors using a slightly different method. Using the average of waveforms from T₂ and T₃, they calculate average rise rates at the 100 m and 500 m levels to be 11 and 12 m/s, respectively. These values are low compared to the calculated values, even when compared to velocities obtained from the T₁ – T₂ delay. The reason for this is two-fold, the first of which is the use of the average of all the waveforms for calculation and the second is from selection of the peak to define the duration of the onset, rather than the first inflection. This lack of deceleration between 100 and 600 m above the vent led Johnson et al. (2004) to conclude that the plume ascent was buoyantly driven between these two levels. In my calculations I observe that for the majority of the observed plumes there is a significant decrease in rise rate between the 100 m and 500 m levels above the vent but given the large height interval between the two measurements, the deceleration rate is extremely low.

The result from the experiment allowed me to define a rough vertical plume-front ascent velocity profile for some of the observed plumes (Figure 2.8.). The range of exit velocity (Vₑ) for the plumes is significantly larger than the range of velocity at 100 m above the vent, which suggest a significant deceleration in the first 100 m of rise. Conversely, there is no significant change in the range of velocities between 100 and 600 m position above the vent and thus the plume maintain its velocity within the two height intervals. This generalized plume velocity profile is consistent with plume ascent involving an
Figure 2.8. Crude vertical plume rise rate profile constructed by plotting the range of exit velocity ($V_e$), velocity at 100 m and velocity at 600 m height above the vent, derived from the delay in the arrival of the thermal signals at the three different thermal sensors.
initial high velocity gas-thrust phase, followed by rapid deceleration and transition to steady (but slower) ascent driven by buoyancy.

2.6. Insights from integrating thermo-seismo-acoustic data

2.6.1. Elastic and Thermal Energy Flux

The use of integrated seismic, acoustic and thermal data allows me to quantify proxies for the total amount of elastic and thermal energy emitted during each explosive event. I can use these to gain a comparative measure of eruptive intensity and to gain some insights into the source dynamics. Following Johnson et al. (2004) I quantified the proxies for total seismic energy flux \( E_s \) and acoustic energy \( E_i \) via time-integrating the squared velocity \( U \) and excess pressure \( P \) trace for the duration of the explosion,

\[
(2.1) \quad E_s = \int U(t)^2 dt
\]

\[
(2.2) \quad E_i = \int \Delta P(t)^2 dt
\]

Thermal energy released by individual explosions are calculated by converting background-normalized temperature to radiant flux using the Stefan-Boltzmann equation and then taking the time-integrated radiant flux for the duration of the explosion or duration of the thermal waveform,

\[
(2.3) \quad E_t = \int \sigma \varepsilon (T_r^4 - T_b^4) dt
\]
where $\sigma$ is the Stefan-Boltzmann constant ($5.6703 \times 10^{-8} \text{ W/m}^2/\text{K}^4$) and $\varepsilon$ is emissivity (0.9). Johnson et al. (2004) found good correlation between thermal flux and convective velocity because hotter plumes will become more buoyant and thus will rise faster than cooler plumes. They also found a moderate correlation between convective velocity and acoustic energy where a longer-duration source generates a larger amount of hot gas, effectively increasing the overall plume temperature and thus increasing its buoyancy. I recalculated the energy proxies to account for the expanded dataset of eruptions and compared the energy partitioning for individual explosions (Figure 2.9).

A first-order observation is that the majority of the explosions are clustered in the low end of the energy spectrum, where there is reasonable positive correlation of released thermo-seismo-acoustic energy at the lowest energies. The positive correlation extends to the higher-end of the energy spectrum, but with much scatter. I attribute the overall moderate positive trend in energy partitioning to a similar source mechanism for all of the explosions. However, the correlations from the larger data set are not as strong as those given by Johnson et al. (2004) using a smaller sub-set of events. There is a de-coupling of relative energy partitioning in the case of the more energetic explosions. Excess heat energy relative to both seismic and acoustic energy can be attributed to extended degassing events contributing more heat, for which no seismic and acoustic energy are generated. An explanation for the excess infrasound energy relative to seismic energy is beyond the scope of this study.
Figure 2.9. Normalized energy proxy for seismic ($E_s$), infrasound ($E_i$) and thermal ($E_t$) obtained during each explosion.
but perhaps can be attributed conduit or atmospheric conditions that promote the enhancement of infrasonic energy, or conversely, impede seismic energy.

2.6.2. Thermo-acoustic delay

The delay between the seismic, acoustic and thermal signals emitted by a single explosive event can be used to estimate the depth of the explosion source if the propagation path conditions and propagation rates from the source to the sensor are known (Figure 2.10) (Chouet et al., 1999, Ripepe et al., 2001, Ripepe et al., 2002, Gresta et al., 2004). In the dataset, the seismic signal always arrives first, followed by the infrasound and then by the onset of thermal waveform. The onset of the thermal waveform heralds the exit of the plume out of the vent into the atmosphere, whereas the former are sourced by the explosion itself within the conduit (Figure 2.10). The complete range of onset delays between seismic and acoustic ($\Delta t_s$) or the time the infrasound lags behind seismic signal, is 0.9 to 6s with a mean value of 2.5 s. However for the majority of the explosions, seismo-acoustic delay is in a tight range around the mean value, falling between 2 and 3s. The onset delay between infrasound and thermal signal at sensor $T_1$ ($\Delta t_t$) or the amount of time that the thermal signal lags behind infrasound, has a wider range of values of -1.4 to 7.4 s with a mean value of 1.2. Negative $\Delta t_t$ values indicate the few cases (5 out of 35 explosions) where the thermal onset precedes the arrival of acoustic waves at the sensor location.

Variations in thermo-acoustic delay ($\Delta t_a$) have been attributed to either changes in the source depth of the explosions or variations in the rise rate of the
ash-gas cloud within the conduit (Ripepe et al. 2002). In a silicic setting, the
details of the source conditions and mechanics of infrasound generation from low
magnitude explosions have not been as fully explored as in the basaltic
Strombolian case. However, one interpretation for the source of infrasound in
silicic systems is that they are generated by some rapid gas release or degassing
mechanism (Johnson 2003). I assume that this rapid gas release generates a
pocket of pyroclastics and gas, thus implying a co-located thermo-infrasonic
source, as previously suggested by the energy-partitioning trend. Assuming a
hypothetical explosion right at the vent level, infrasound generated will travel the
distance \( x \) from vent to the sensor (1600 m) at the speed of sound \( s \)
corresponding to the local ambient pressure and temperature. At Caliente’s
summit altitude (~ 2500 m) with ambient temperatures in the order of 10 – 20 °C,
\( s \) has a value of ~ 330 m/s. In contrast, thermal energy generated by the plume
as it exits the vent travels the distance \( x \) at the speed of light and will thus be
detected instantaneously. As a result infrasound waves will reach the sensor
approximately 4.8 s following the arrival of the thermal signal if the explosion
source is located at the vent level. All but one of the \( \Delta t_i \) values measured for the
35 explosions recorded are less than 4.2 s, suggesting that the explosion source
must be located at a certain depth \( h \) below the vent level.

The total travel time of the infrasound \( (T_i) \) pulse generated by the
explosion at depth is the time it takes for the infrasound waves to propagate up
the length of conduit between the source depth and the vent surface \( h \) at the
speed of sound in the conduit \( s' \), and the time it takes to propagate from vent
surface to the sensor at sound speed at ambient pressure and temperature (s).
Total travel time of the thermal signal ($T_e$) is the time it takes for the package of
ash and gas generated by the explosion to rise up the same length of conduit ($h$)
at velocity ($V_g$). $V_g$ in this case is assumed to be equivalent to the exit velocity of
the plume ($V_e$). Thus the thermo-acoustic onset delay ($\Delta t_i$) can be used to
calculate the depth of the source if the exit velocity is known (Ripepe et al. 2002)
using the following equation,

\begin{equation}
\Delta t_i = t_e - t_i = \frac{h}{V_g} - \left[ \frac{h}{c'} + \frac{x}{c} \right]
\end{equation}

Because I assume that the gas rise rate within the conduit $V_g$ is equal to the exit
velocity ($V_e$), which varies between explosions but is measured here, variable
source depth can be calculated as a function $\Delta t_i$ and $V_e$ by using,

\begin{equation}
h = \left[ \frac{\Delta t_i + x}{c} \right] \frac{V_e c'}{(c'-V_g)}
\end{equation}

The speed of sound in the conduit is an elusive parameter to define, varying
greatly with the type and temperature of the medium that the sound is
propagating through. Similar calculations done at Stromboli assumed that sound
waves propagate through air heated to magmatic temperature which gives a
sound speed of 705 m/s (Weill et al. 1992, Ripepe et al. 2002). Bluth and Rose
Figure 2.10. Infrasound and thermal travel path and velocities from a source at h depth below the surface. Infrasonic waves generated by gas acceleration travels up the conduit at speed of sound within the conduit (s') and, once it exits the vent, propagates at speed of sound at ambient pressure and temperature (s). The explosion simultaneously generates a packet of gas and pyroclasts which travels up the conduit at the gas/jet velocity (Vg) and as it exits the vent detected instantaneously by the thermal sensors (Ripepe et al., 2002).
observed that most of the plumes emanate from circular rings of vents which are thought to be the surface expression of the conduit walls. This boundary will most likely provide the same path that the sound waves will propagate along. I assume that infrasound is generated and propagates ahead of the mixture of ash and gas (which rises at sub-sonic speeds) and thus travel through 'clean' conduit filled with water vapor at dacitic magma temperature (950° C). The speed of sound in gases \( V_s \) is given by:

\[
V_s = \sqrt{\frac{\gamma RT}{M}}
\]

where \( \gamma \) is the adiabatic constant of the gas (1.28 for \( H_2O \)), \( R \) is the universal gas constant (8.314 J/mol K), \( T \) is the absolute temperature of the gas (1173 K) and \( M \) is the molecular weight of \( H_2O \) (0.018 kg/mol). The resulting speed of sound in the conduit \( (s') \) is 832 m/s. However changes in gas content, \( SO_2 \) in particular, in the conduit may significantly change the speed of sound in the conduit (Hagerty et al., 2000). Although Santiaguito overall is a relatively low emitter of \( SO_2 \) (20-year average of 80 tons/day; Andres et al. 1993), it has been noted that \( SO_2 \) emission can increase prior to and during explosions (Rodriguez et al. 2004). I therefore consider an end member case where infrasound waves generated by the explosions are propagating through a \( SO_2 \) saturated conduit (where \( \gamma \) is equal to 1.26, and \( M \) is equal to 0.064 kg/mol). In such a case, the speed of sound in the conduit decreases to 441 m/s. In reality, the conduit will more likely be occupied by varying proportions of water vapor and \( SO_2 \) (and
perhaps other minor volcanic gases), thus the actual speed of sound in the conduit will fluctuate between the two end member cases. Using the $c'$ value from the water vapor-filled conduit model in equation 2.5, I calculated that the source depth of the explosions ($h$) ranges from 100 to 620 m (with a mean of 257 m) below the vent level. For comparison, using the SO$_2$-filled conduit $s'$ value, $h$ ranges from 105 to 673 m (with a mean of 272 m). The error increases with increasing depth but with a maximum change of less than 10% (+57 m). Temporal and statistical distribution of calculated source depth is shown in Figures 2.11a and 2.11b. Most of the explosions (23 out of 35) have their sources at depths of between 100 and 250 m, 10 explosions are sourced between 250 and 460 m and only two are at depths of around 600 m. Although the dataset is too short to establish any credible daily patterns, I observe that the deeper explosions seem to be concentrated in the earlier (nighttime) segment of the dataset (Figure 2.11a).

2.7. Discussion

Here I discuss the results of the source depth calculations in the context of previous models that have been presented as source mechanisms for repetitive, low to moderate intensity explosions. From the source depth calculation I can define a ~ 500 m thick region within the shallow conduit from which explosions at Santiaguito originate (Figure 2.11). The results position the upper boundary of this explosion source region at 100 m below the vent and the lower boundary at ~ 620 m below the vent. Almost all of the explosions are thus sourced within the
Figure 2.11. a. Plot of mean source depth of the explosions with time of the explosions with errorbars representing minimum source depth calculated using speed of sound through vapor-filled conduit and maximum source depth using speed of sound through SO₂-filled conduit. b. Histogram showing the distribution of the source depths, most of which is located within the top 250 m of the conduit.
Caliente dome edifice itself above the pre-extrusion surface. The depth of the explosion source also varies non-systematically by up to 340 m between subsequent explosions. Comparison between explosion depth and eruptive intensity parameters such as thermal amplitude ($\Delta T_2$) and calculated thermal and elastic energy released ($E_t$, $E_s$ and $E_i$) shows no significant trends or correlations (Figure 2.12), thus indicating that explosion source depth does not control how energetic the explosion will be.

The first model that I consider is that of Rose (1987) and subsequently Sanchez-Bennet (1992), where they suggested that the explosions at Santiaguito are phreato-magmatic in origin. In the context of this model, the explosion source region defined by the results of this work could correspond to water-saturated zone within the shallow conduit where magma-water interaction can occur. The source for the flux of water into the dome can either be groundwater flow or meteoric water percolating down, pooling above the contact between the 1902 crater floor and the base of the Caliente dome, i.e. at and above the contact with the pre-extrusion surface, creating the 'wet' zone which should limit the eruption source to positions above the pre-extrusion surface. Shallow water-saturated zones have been found in other volcanoes (e.g. Mt. Unzen in Japan (Kagiyama et al. 1999) and Masaya Volcano in Nicaragua (Connor et al. 2005) using results from geophysical studies. No such study has been conducted at Santiaguito, thus I lack the knowledge of the subsurface hydrology to confirm the presence or, if present, the depth of such a layer. Large fluctuations in calculated source depth
Figure 2.12. Comparison between eruptive intensity parameters (normalized thermal amplitude and thermo-seismo-acoustic energy proxies) and calculated source depth, showing no distinct trends.
thus it would imply extreme fluctuation in the water table level which is unlikely. A more reasonable explanation is that the level in which the water-magma interaction can occur varies randomly within the explosion region due to the presence of localized groundwater pockets at different depths. The validity of this model can be confirmed by conducting geophysical studies at Caliente dome to determine the subsurface hydrology. In the absence of such study, a simple way to validate this model is to compare the frequency of eruptions during wet season versus dry season. If these explosions are indeed phreato-magmatic, then the explosions should be more frequent and/or violent during the wet season.

The second model that I consider is the gas-pressure buildup beneath an obstruction in the conduit, which has been proposed to be the source mechanism for similar low-intensity explosions at Karymsky (Johnson et al. 1998, Ozerov et al. 2003). Both Santiaguito and Karymsky are characterized by repeated, mildly explosive activity. Both systems involve a more viscous magma and both systems display both effusive eruption and eruptions of ash-rich plumes and produce detectable seismic and infrasound waves. Johnson et al. (1998) proposed gas buildup under a confining plug for an explosion mechanism at Karymsky. In their model, gas pressure steadily increases beneath a degassed, highly viscous cap and when that pressure reaches a certain threshold, ‘uncorking’ of the plug occurs as well as subsequent release of a certain quantity of gas up through the conduit. Applying this model to Santiaguito limits the source location of each explosion to the boundary between viscous plug and
underlying magma and may thus be expected to be fairly stable. Alternatively, if I assume that the source of the explosions varies in depth as the analysis indicated, it would require the rheological boundary to migrate up and down at fairly rapid pace. Assuming a stationary source depth \((h)\), the thermo-acoustic onset delay then is a function of sound speed within the conduit and/or the rise rate of the gas-particle cloud. My previous estimates suggest that reasonable sound speed variations do not significantly affect the calculations of the depth of the explosion, thus I focus on the rise rate of gas and particles cloud within the conduit. At \(h = 620\) m (the lowest depth calculated), for the given range in thermo-acoustic time delays, the rise rate of the gas and particles cloud within the conduit will have to vary between 60 and 130 m/s to explain the thermal-infrasound delays. This is significantly higher than the calculated exit velocities \((V_e)\). This discrepancy may be resolved given that the estimate for exit velocities is in fact an average velocity along a 100 m vertical profile immediately above the vent and that the true exit velocities may be much higher, and/or that gas jet velocity may decrease significantly as the gas/pyroclast mixture travels up the conduit. One argument against such a mechanism operating at Santiaguito is that the model implies that the relative size and intensity of the explosion should be proportional to the time for pressure to build up or the repose period between explosions. Data from this study shows that there is no correlation between repose intervals and any of the parameters used as a measure of intensity (Figure 2.13).
The third model that I consider was proposed by Bluth and Rose (2004). In this model, the repetitive emissions at Santiaguito are fed by fragmentation at conduit walls induced by cyclic movement of the upper part of the magma column. An example of the shallow system manifestation of such cyclic conduit flow has been observed as cycles in deformation, earthquakes and explosive events at Soufriere Hills during the period of 1996 – 1998, with a cyclic period range of 4 and up to 30 hrs (Voight et al. 1999) Cyclic conduit flow models can be divided into two general classifications, rheology-based (Voight et al. 1999, Melnik and Sparks 1999, Barmin et al. 2002), or stick-slip model (e.g. Denlinger and Hoblitt 1999). In the rheology-based model, degassing and crystallization of silicic magma as it ascends up the conduit leads to very large vertical gradients in viscosity across the uppermost part of the column (Sparks et al. 1997, Melnik and Sparks 1999, 2005, Barmin et al. 2002), creating a viscous upper part with high yield-strength typically termed as the plug. The presence of this degassed plug inhibits flow, reducing the rate of extrusion thereby increasing the pressure at the rheological boundary. Once the pressure buildup exceeds the pressure threshold or the yield strength of the plug, the degassed plug is extruded or expelled explosively and conduit flow resumes at high extrusion rate. The stick-slip model of Denlinger and Hoblitt (1999) assumes that rheology is not the dominant factor controlling the conduit flow cycles. In their model, constant flux of compressible magma into the conduit produces laminar flow with stick-slip boundary conditions along the upper part of the magma column, equivalent to the
Figure 2.13. Repose interval versus normalized thermal amplitude and thermoelastic energy proxies showing the lack of correlation between relative intensity and the time interval between explosions.
plug in the rheology-based model. Conduit flow occurs so long as the shear stress along conduit walls exceeds the yield strength of the magma. If the flux of magma into the conduit increases, there is an increase in the shear stress at conduit boundary exceeding the yield strength and slip occurs along the detachment point. This leads to enhanced flow and increased extrusion rate which eventually exceeds the magma supply rate into the conduit. Once extrusion rate exceeds supply rate, shear stress decreases and once it falls below the yield-strength threshold, slip ceases and extrusion rates decrease. As extrusion rate falls below supply rate, shear-stress increases and the cycle is repeated. Regardless of which is the dominant controlling factor (rheology or boundary conditions), both models agree that there is a period of stagnation and slippage of the upper part of the magma column along a detachment point and that during the stagnant phase there is a large buildup of pressure or shear stress along the conduit boundaries. Recent numerical modeling of conduit flow by Gonnermann and Manga (2003) took into account both rheology and boundary conditions. Magma rising through the conduit as an isothermal flow with constant supply rate loses pressure due to magmastatic and dynamic pressure loss, resulting in volatile exsolution and bubble growth. This in turn leads to an increase in the melt viscosity thereby increasing shear-strain rates along the conduit walls. Large shear-strain rates at conduit boundaries can eventually exceed the critical strain rate of silicate melts, leading to a ‘shear-induced’ fragmentation, opening up fractures and pathways for the magma to outgas. Following fragmentation the decrease in shear-strain allows for re-
annealing of fractures in the magma, shutting down magma degassing and renewing the cycle.

I agree with Bluth and Rose (2004) and prefer the shear-induced fragmentation model as the most appropriate model for explosions at Santiaguito. If I consider the variability in the depth of the explosion source, it is much easier to vary the boundary conditions and magma influx into the conduit than the migration of rheological boundary or water table level over short time scales. The results suggest that a stick-slip flow regime in a shallow plug region extends between 100 and 620 m below the vent (Figure 2.14). Plug slips can occur either as a single event or in a series of successive slips resulting in either a single explosion event or a cluster event. Within this model, I have constrained the thickness of the viscous, degassed upper part of the magma column or the plug, within which the shear-induced fragmentation events are taking place to be approximately 500 m thick.

The results of this study highlight the importance of monitoring these explosions as changes in their character, be it frequency or relative intensity, may provide clues to imminent changes in the behavior of the volcano such as decreasing magma supply, or the rheological conditions in the upper section of the conduit and/or possibility of a larger explosive eruption.

2.8. Summary

During a short thermal and seismo-acoustic experiment at Santiaguito I tracked the dynamics of the intermittent, low to moderate intensity explosions
Figure 2.14. Schematic of conduit dynamics at Santiaguito based on stick-slip motion of the upper part of the magma column (Denlinger and Hobblit, 1999) and shear-induced fragmentation at conduit boundary (Gonnermann and Manga, 2003).
that occur at this silicic system over a period of ~15.5 hours. A total of 73 thermal emissions were recorded as thermal transient waveforms (4.7 emissions per hour), produced by plumes of gas and pyroclast mixtures ascending through the field of view of the thermal sensors. This yields an emission frequency of 4.3 emissions / hour with emission duration of 1 – 8 minutes and repose period of 1 – 56 minutes. Thermal emissions produced by actual explosions are distinguished from gas exhalations by the presence of seismic and acoustic signal. Thirty-five explosions were detected on the basis of near-synchronous thermal, seismic and infrasonic signals giving an explosion frequency of 2.3/hour. These explosions are either simple single explosion (20 events) or part of multiple-explosion events in which sequential explosions are 1 – 5 minutes apart (6 events). The duration for the individual explosions feeding the plumes as evident from the seismic and infrasonic data were shorter than the emission duration (as obtained from thermal waveform duration) at an average of 18 – 26 seconds.

The rise rates of the plume fronts generated by 35 explosions observed were measured from onset delays between different thermal sensors to obtain exit velocities of 16 – 77 m/s over the first 100 m above the vent decelerating to relatively stable convective velocities of 10 – 20 m/s for the next 500 m of ascent. Using thermo-acoustic onset delays I defined an explosion source zone within a 500 m thick layer located 100 to 600 m below the vent level. Interpretations of my data favor shear-induced fragmentation as the most likely source mechanism for the explosions, with slip occurring within a 500-m-long shallow plug that is undergoing stick-slip cycles.
Chapter 3

Surface temperature and spectral measurements at Santiaguito lava dome, Guatemala

3.0. Abstract

An infrared (IR) thermometer, a spectroradiometer and a digital video camera were used to observe and document the short-term evolution of surface brightness temperature and infer changes in the morphology of the vent crater at Santiaguito lava dome, Guatemala. The dataset from the IR thermometer shows 40 – 70 minute-long cooling cycles, each defined by a cooling curve that is both initiated and terminated by rapid increases in temperature during cyclic ash venting. The average cooling rates calculated for each cycle range from 0.9 to 1.6 °C/min. I have applied a two-component thermal mixing model to the spectroradiometer (0.4 – 2.5 µm) dataset. The results suggest that the vent crater morphology changed from a cool (120 – 250 °C) crust-dominated surface dissected by high temperature fractures (> 900 °C) in the first time segment of the measurement period to an isothermal surface at moderately high temperature (350 – 500 °C) during the second segment. I attribute the change in the thermal state of the surface to the removal or partial removal of the overlying lava crust, exposing hotter layer below during the most energetic of the ash eruptions.
3.1. Introduction

The activity at Santiaguito dome complex since 1975 has included regular eruptions of short-duration ash-rich plumes that typically occur at every 0.5 – 1 hr (Rose, 1987). This accompanied constant low-level degassing with persistent extrusion of lava that has built a 1.1 km$^3$ dome complex (Rose 1987, Harris et al. 2002). Current activity is focused on the Caliente vent where lava extrusion has created a rubbly surface of dacitic lava blocks within a 300 m wide crater. Continuing lava extrusion is evident from the slow-moving active block-lava flows that have extended up to 3.8 km from the vent (Harris et al. 2002). Periodic vertical ash eruptions continue at intervals of ~5 min to < 1 hr from a ~150 m wide ring of vents within the crater (Bluth et al. 2002).

In January 2002, for a 5 hr period, I collected remote thermal infrared temperature and spectral radiance measurements as well as digital video footage of the dome surface, recording 17 ash eruptions. The objectives of this study were to observe and record the short-term changes in the temperature and morphology of the lava dome surface on a time-scale of seconds and to study how dome temperature and morphology are affected by the periodic ash eruptions. This data provide constraints on initial time-dependent parameters at the vent such as temperature, heat flux and crust coverage, all of which are essential inputs for models that extract thermal / mass flux information from satellite images, as well as lava emplacement models applicable to silicic lava domes and flows (e.g. Harris et al. 2002). For example, thermal remote sensing studies of lava domes typically assume that the dome surface is composed of
one, two or three thermal components to retrieve sub-pixel temperatures (e.g. Rothery et al. 1988, Oppenheimer 1993, Wooster et al. 2000). Thermal structures derived from this data will thus help determine and guide the number of components to best represent the lava dome surface at Santiaguito.

3.2. Data Collection

Continuous brightness temperature measurements were collected from 4:22 AM until 8:39 AM on January 11, 2002, using a single-channel (8 – 14 μm) Raytek IR thermometer at 2 s sampling rate. These yield integrated brightness temperatures of surfaces that may or may not be thermally homogenous. They do not represent the absolute temperatures of the surfaces because I did not take into account emissivity or atmospheric effects. Five hundred and ninety radiance spectra were collected at 3 s intervals during two nighttime periods within the brightness temperature time series between 4:31 – 4:46 AM and 5:29 – 5:44 AM, using an Analytical Spectral Devices (ASD) FieldSpec FR narrowband spectroradiometer. The ASD collects the radiant flux in the 0.4 – 2.5 μm range at a 3 – 10 nm spectral resolution, producing a continuous spectrum averaged from 10 collected spectra. Each measurement period was preceded by a dark current measurement intended to decrease the amount of noise. All measurements and observations were made from the summit of Volcán Santa María (3700 m asl), 2.5 km away from the target area. The target area is located in the NW sector of the summit crater and was chosen because it was having the most activity at the start of the observation period. Both instruments have 1°
field of view (FOV), equivalent to a 44 m-wide spot (1520 m²) on the dome surface, or approximately 2.1% of the total area of the summit crater. Both instruments were mounted on heavy-duty tripods secured with rocks at the base to stabilize the FOV for the observation period. Both also had aiming optical scopes fine-tuned by maximizing radiative response through the FOV. The two FOVs probably were not identical which may result in a discrepancy in the area of the dome observed and the derived temperatures between the two instruments. Digital video data were collected from sunrise onwards to give a 2hr-long data set beginning at 6:29AM ending at 8:22AM, overlapping with the latter part of the IR thermometer dataset.

3.3. Cooling Cycles

The temperature time series from the IR thermometer reveals cyclic fluctuations in the apparent temperature of the target area (Figure 3.1). Four complete cooling cycles, each between 42 and 72 minutes long, can be seen in this dataset. A temperature spike, $T_s$, ranging from 168 – 242°C, started each cycle and is quickly followed by a 1 – 3 min period of temperature drop to as low 35°C before recovery to the more typical decaying trend of a cooling curve. Two to three more intervals of drastic temperature drops occurred in each cycle, at roughly 5-minute intervals. In each cycle the temperature decayed exponentially down to an apparent equilibrium final temperature, $(T_e)$ of 117 – 122 °C prior to the onset of the next cycle. In cycles 2, 3 and 4 there appear to have been periods of small-scale temperature fluctuations (< 20°C) from the general
Figure 3.1. IR thermometer dataset showing the cyclic variations of temperature with time. Four complete cooling cycles are distinguished, designated as cycles 1 – 4. Best-fit exponential functions for each individual cooling cycle and the $r^2$ values are shown in the inset. Temporal coverage of the spectroradiometer and video footage are shown in gray-shaded and white rectangle respectively.
cooling trend in the beginning to near the mid-point of each cycle.

Digital video footage is only available for the entire duration of cooling cycle #4 given that the three previous cycles occurred prior to daybreak. This footage allowed the linkage of thermal signatures to specific events. Figure 3.2 provides a summary of the observed events during cooling cycle #4. The temperature spike at the onset of the cycle coincides with a relatively energetic ash-rich plume emission event (Figure 3.2i). The dark ash-rich plume dominating the IFOV during the event appears to be the source for the sudden temperature drop, when ash particles within the plume cool very rapidly as air at ambient temperature is entrained in the vertical plume. Two less energetic ash plumes were erupted 7 minutes after the first eruption. These plumes formed approximately ~5 minutes apart, each with durations of approximately three minute and both decay in relative intensity in a systematic fashion. Both plumes lack initiating high-T spikes, but similar to the first eruption, correspond to periods of low temperature in the Raytek dataset. A 30 minute-long period of intermittent emission of pale, gas-rich, weak plumes from a vent within the observation area occurred mid-cycle (Figure 3.2ii). This phase can be correlated with the period of small-scale temperature fluctuations. Two isolated exhalations of pale plumes occurred towards the end of the cycle but fail to register any thermal signature in the dataset (Figure 3.2iii). The absence of a thermal signature suggests that the two plumes may be roughly at the same temperatures as the dome surface or that these plumes failed to intersect the FOV. The cycle then is terminated by an
eruption of a darker plume with an associated temperature spike, temperature drop and cooling curve pattern.

The general shape of the cooling curve for the observed surface remained relatively constant throughout the cycle. Temperature recovery followed by cooling following each major event suggests that thermal renewal of the surface occurred as a result of the larger of the ash eruptions, where a large fraction of the observed surface is suddenly disturbed. As a result, new hot material may be exposed and blocks become rotated so that hotter surfaces now face outwards and high temperature cracks form: thus increasing the integrated temperature. The smaller events were insufficiently energetic to create such temperature changes and/or only affected limited areas beyond the measurement area considered. A comprehensive characterization of the cooling curve was hindered by the presence of the cool ash plume masking the first few minutes of the cycle; thus, I am unable to determine conclusively whether $T_s$ is equal to the starting temperature of the surface for a particular cycle. If I assume that $T_s$ is equal to the initial temperature, each cooling curve can then be fitted fairly well with logarithmic functions (Figure 3.1). Alternatively, if I ignore the first three minutes of each cycle to accommodate the lack of surface temperature data, a slightly better fit to the data can be obtained by quadratic expressions. I calculated the average cooling, excluding the initial ~3 minutes of each cycle, to be 0.9 – 1.6 °C/min. This compares to 1.2 – 3 °C/min calculated for the proximal section of the Santiaguito block lava flow (Harris et al. 2002), and to 2.7 °C/min calculated for basaltic pahoehoe surface by Hon et al. (1994) (taking into account only data
Figure 3.2. a. IR temperature measurements for thermal cycle #4 with associated physical events observed from digital video footage. b. Three images captured at three different times during this cycle, (i) ash eruption at the initiation of the cycle, (ii) mid-cycle semi continuous gas puffing and (iii) small plumes with no thermal signature. Note the approximate IFOV shown by black circle in the photo sequence.
beyond the first three minutes which represents the initial rapid cooling phase).

3.4. Radiance Spectra

Three hundred radiance spectra were collected for a 15-minute period before the start of cooling cycle 1 and 290 were collected after the start of cycle 2 (Figure 3.1). The large number of wavebands available to the spectroradiometer allowed me to collect radiant flux as a near-continuous spectrum from 0.4 to 2.5 µm. Where a number of thermal sources are present in the FOV, an integrated radiance spectrum will be produced which can then be de-convoluted to provide an estimate of the temperature and fractional area of each source (Flynn et al. 1993). I applied a simple two-component mixing model developed by Flynn et al. (1993) to the radiance spectrum. This model calculates a Planck curve that fits best the measured radiance spectrum by varying the temperatures of the hot component \( T_h \), and the cool component \( T_c \), as well as the areal proportion of the hot component \( f_h \). Temperatures for each component were allowed to vary from 0 – 950 °C and proportion of hot material from 0 to 1 for model calculations. I assumed that 950°C is a reasonable maximum eruptive temperature based on previous values of 800 – 850 °C from Fe-Ti oxides and optical pyrometer measurements and taking into account recent increase in SiO₂ content by adding 100°C (W.I. Rose pers.comm. 2003). I refer the readers to chapter 1 for complete treatment of the model used in this work.

In Figure 3.3 I display three spectra from the data. Spectra A and B were collected during the first segment and spectrum C was collected during the
second segment. Two atmospheric absorption bands at 1.3 – 1.5 µm and 1.8 – 2.0 µm are evident by the decrease in radiant flux at those wavebands. Very small amounts of radiant flux detected for both spectra A and C at 0.9 – 1.3 µm indicate the presence of high temperature radiative component in the IFOV. The notable increase in radiance flux at the same wavelengths in spectrum B suggests an increase in the fractional area of the high temperature radiator. More significant changes in flux for the three spectra occur at the wavelength ranges of 1.5 – 2.5 µm, varying from low (spectrum A), to intermediate (spectrum C) and to high (spectrum B). Changes in the slope of the spectral curve suggest changes in temperature and fractional areas of each component, which can be elucidated by applying a mixing model to these spectra. For spectrum A, I obtained $T_c = 187^\circ\text{C}$, $T_h = 947^\circ\text{C}$ and $f_h = 2.59 \times 10^{-5}$, which is consistent with a surface composed of a fairly cool crust with rare fractures radiating at or near magmatic temperature. For spectrum B, $T_c = 467^\circ\text{C}$, $T_h = 468^\circ\text{C}$ and $f_h = 0.39$, which suggests an isothermal surface. Solution for spectrum C also yields an isothermal surface with $T_c = 355^\circ\text{C}$, $T_h = 359^\circ\text{C}$ and $f_h = 0.54$.

Applying the model to the rest of the dataset illustrates how the two solutions for spectra A and C represent a fundamental difference between the two segments of the spectral dataset (Figure 3.4). The two-component structure is consistent throughout most of the first segment, with $T_c$ at temperatures of 120 – 250 °C cooling with time while $T_h$ approximates 850 – 950 °C. The $f_h$ values remain extremely small (< 0.02 %). This suggests that, at this stage, the surface is comprised of an extensive blocky crust that is cooling slowly with time, with
Figure 3.3. Three representative radiance flux spectrum from segment 1 (A and B) and segment 2 (C). Best-fit curves calculated from my model for each spectrum are shown in solid black curves.
localized fractures radiating at high temperature $T_h$ due to heat escaping from hotter underlying material. Conversely, the 2nd segment is characterized by an isothermal surface for nearly the entire time. Almost all of the solutions for $T_c$ and $T_h$ fall within 6 °C of one another in this segment. There is a significant increase in temperature of the dominant radiating component from the previous segment, with temperatures ranging from 350 to 500 °C during the second segment, compared to 120 – 250 °C during the first segment.

Departures from the two-component structure for the 1st segment, represented by radiance spectrum B, occurred at the beginning and towards the middle of the segment. Neither of these events is registered in the IR thermometer data and no video footage is available for visual confirmation of the actual activity responsible for the events. These events are characterized by rapid convergence of $T_c$ and $T_h$ to an intermediate value (400 – 600 °C) and after ~3 minutes all parameters returned to their pre-event trends. This suggests the rapid development of an isothermal surface and subsequent breakdown to a two-component surface. Such an event may be similar to the thermally indistinct ash/gas venting events observed from video footage. The rapid recovery to the overall cooling trend is consistent with cooling of a surface composed of fine fragments that reached thermal equilibrium with the underlying surface in a short period of time.

The spectra-derived crustal temperature appears to be 50 – 200 °C higher than the IR thermometer measurements. I explain this discrepancy partly as a result of having slightly different FOVs for the two instruments, where the IR
Figure 3.4. Temperature values derived from radiance flux dataset. \( T_h \) is shown in solid circles and \( T_c \) is shown in gray solid triangles. 15s moving average for \( T_h \) and \( T_c \) are shown in solid black and gray solid lines respectively. Note that \( T_h \) and \( T_c \) values for segment 2 generally are almost the same and nearly overlap on the graph.
thermometer may be viewing part of the dome that is relatively cooler in temperature, as well as due to the differential emissivity and atmospheric effects at the different operational wavebands of the two instruments.

3.5. Discussions and Conclusions

The combined use of IR thermometer, spectroradiometer and video footage at high data collection frequencies (0.3 Hz) have allowed the documentation and characterization of cooling cycles and changes in the thermal state and morphology of the surface at Santiaguito. Using the IR thermometer dataset and video observations, I recognized that each cooling cycle is initiated and terminated by a ‘thermal resurfacing’ event which coincided with the periodic ash eruptions. However, not all of the ash eruptions result in resurfacing, at least not of the dome sector observed. Temperature retrievals from spectral radiance datasets confirm that during thermal resurfacing, the surface was transformed from a two-component crust-crack surface to an isothermal surface at intermediate temperature. Thereafter cooling dominates as the isothermal surface cools by radiation and convection via a logarithmically decaying cooling curve. The low temperatures of the ash plumes coupled with the absence of nighttime visual observations of incandescent ejecta eliminate plume fallout as a source of the thermal resurfacing.

From video data I recognized a progressive decrease in relative intensity of subsequent ash eruptions in each cycle. The weaker events tend not to be associated with thermal renewal. Therefore I assume that only ash eruptions
beyond a certain threshold of intensity that can cause change in thermal state of
the surface, and/or smaller events only affect localized areas beyond the
measurement area. I propose that ash eruptions of sufficient energy and intensity
can cause the change in the thermal state of the surface by creating more
fractures or widening the fractures on the surface crust, and/or rotating the lava
blocks on the surface, effectively removing the chilled crust and exposing the
slightly hotter layer below. The relatively low temperatures of the newly exposed
layer show that, although elevated, they are cooler than dacite lava, and thus do
not represent exposure/emplacement of fresh lava.

The results of this study have a significant implication for satellite
monitoring of active silicic volcanoes which typically involve relatively low surface
temperatures. The periodic thermal resurfacing of the summit crater surface by
the ash eruptions allows for at-vent activities at Santiaguito to be detectable in
the near-IR as well as in the thermal IR wavelengths. This observation is
consistent with the analysis of Harris et al. (2002) where the vent appears
thermally anomalous in TM data but the crust-dominated block flow does not.
The results suggest that the lava dome surface can be modeled with a two-
component surface (chilled crust broken by hot cracks). However, given the very
small fractional area of the hot component, and the presence of isothermal crust,
a single cool crust component may well be sufficient to model the dome surface
at Santiaguito. This in turn facilitates the use of satellite data sets to extend my
observations beyond the temporal limits of field-based campaigns to reveal hour-
to-day long as well as long-term temperature variations and activity cycles. The
analysis of the automated thermal alert information using GOES or MODIS data (Harris et al. 2001, Wright et al. 2004) with their 15 minute to 6 hr temporal resolutions would provide a complementary long-term thermal data set to examine cyclic behavior of Santiaguito for week-to-month long time scales. Both data sets could be used to provide constraints for models that simulate cyclic conduit dynamics, rheological changes and pressurization cycles (e.g. Denlinger and Hoblitt 1999, Barmin et al., 2002).
Chapter 4

Thermal structure and heat-loss at the summit crater of an active lava dome

4.0. Abstract

Forward-Looking Infrared (FLIR) nighttime thermal images were used to extract the thermal and morphological properties for the surface of a blocky-to-rubbley lava mass active within the summit crater of the Caliente vent at Santiaguito lava dome (Guatemala). Thermally, the crater was characterized by three concentric regions: a hot outer annulus of loose fine material at 150 – 400 °C, an inner cold annulus of blocky lava at 40 – 80 °C, and a warm central core at 100 – 200 °C comprising younger, hotter lava. Intermittent explosions resulted in increase in temperature of some surfaces, mostly across the outer annulus where loose, fine, fill material was ejected to expose hotter, underlying, material. Surface heat flux densities (radiative + free-convection) were dominated by losses from the outer annulus region (0.3 – 1.5 x 10⁴ J/s/m), followed by the hot central core (0.1 – 0.4 x 10⁴ J/s/m) and cold annulus (0.04 – 0.1 x 10⁴ J/s/m). Overall surface power output was also dominated by the outer annulus region (31 – 177 MJ/s), but the cold annulus contributed more power (2.6 – 7 MJ/s) than that of the hot central core (1.2 – 3.5 MJ/s) due to its greater area. Cooled surfaces (i.e. the upper thermal boundary layer separating the surface from underlying material at magmatic temperatures) across the central core and cold annulus had thicknesses of ~0.6 – 2 m and 2.7 – 7 m, respectively, as estimated using simple
conduction model. The stability of the thermal structure through time and between explosions indicates that it is linked to a deeper structural control likely comprising a central massive plug, feeding lava flow from the SW rim of the crater and surrounded by an arcuate, marginal fracture zone up-which heat and mass can preferentially flow.

4.1. Introduction

The 1902 eruption of Santa Marfa volcano (Guatemala) was one of the largest explosive eruptions of the twentieth century. In a 36-hour period 7.5 km$^3$ of dacite was erupted to feed an eruption column that reached a height of at least 28 km (Williams and Self 1983, Rose 1987). The eruption also removed a significant portion of the southern flank of Santa Marfa to form a crater on its southeastern flank (Figure 4.1a). Twenty years of repose ensued until June 1922 when a new lava dome emerged within the 1902 crater (Rose 1972). Eight decades of persistent extrusion of dacite at a time-averaged rate (over the period 1922 – 2000) of 0.44 m$^3$/s built a 1.1 km$^3$ dome complex with four main centers: El Caliente, El Brujo, La Mitad, El Monje (Rose 1987, Harris et al. 2003). The locus of activity since 1977 has been at the Caliente vent atop the dome unit of the same name. This is considered to be the principal vent of Santiaguito given its location coincident with the approximate location of the main conduit (Rose 1972, Rose 1987). The latest and ongoing activity at Caliente is characterized by relatively high extrusion rates of $\sim$0.5 m$^3$/s (Harris et al. 2003) and extremely well insulated silicic lava flows which have attained lengths of $\sim$3.9 km (Harris et al.
2002, 2004). This extrusive activity has been accompanied by intermittent low to moderate intensity explosions producing vertical ash plumes up to 2 km high (Rose 1987, Bluth and Rose 2004, Johnson et al., 2004). These occur from the summit vent at a typical frequency of 1.7 explosions per hour (Chapter 2).

The location of Santiaguito within the 1902 crater of Santa María is advantageous for observing and measuring at-vent activity, where an unobstructed view of the Caliente vent (2520 m asl) is possible from the peak of Santa María volcano (3752 m asl). This vantage point is 1220 m vertically above, and 2500 m horizontally away from the vent, giving a line of sight view of 2780 m (Figure 4.1). Using digital video footage obtained from this vantage point Bluth and Rose (2004) proposed a hypothetical conduit model at Santiaguito where a dacite plug is rising through a conduit 50 m in diameter. Based on the work of Gonnermann and Manga (2003), Bluth and Rose (2004) further proposed that shear-induced magma fragmentation at the conduit boundaries gave rise to the ash eruptions. Based on the integrated thermal and infrasonic observations, in Chapter 2 I proposed that these explosions originate from shear failure (plug slippage) at depths between 100 – 600 m below the crater surface.

Field-based thermal infrared remote sensing attempts to describe the surface thermal structure at Santiaguito were first conducted by Birnie (1973) using a radiation thermometer. Birnie (1973) recorded maximum surface temperatures, corrected for emissivity and atmospheric effects, of 20 – 30 °C located at the Caliente summit crater, the summit of El Brujo dome (which was actively extruding during this period) and areas corresponding to fumarolic
Figure 4.1. a. View of Santa María volcano and the Santiaguito Lava Dome Complex growing within the collapse crater of the 1902 Santa María plinian eruption. View is from the east. Height difference between the Caliente and Santa María summits is 1223 m.
activity. Birnie (1973) calculated excess radiant heat flow and estimated that the depth to molten rock at El Brujo at 11 m. In Chapter 3 I used a combination of single-channel IR thermometer and a portable spectroradiometer data to establish the thermal structure of the Caliente summit crater and its temporal evolution during 2002. Using a two-component thermal mixture model I reported that the summit can be characterized by a 120 – 250 °C surface broken by fractures radiating at high temperatures (> 950 °C ). Thermal renewal of the surface occurs following an explosion where the thermal structure changes from a two-component surface to an isothermal surface radiating at 350 – 500 °C as a result of the removal of the overlying chilled crust and subsequent exposure of slightly hotter underlying layer. In this study I explore the use of a high-temporal resolution thermal video camera to document the spatial and temporal features of this surface, and establish its relation to plug-like ascent of the magma to feed ash plume emissions and a southward extending lava flow.

4.2. Thermal Video Camera and Data Acquisition

The thermal video camera used in this study was a Forward Looking InfraRed (FLIR) ThermaCAM™ P40 model, manufactured by FLIR™ Systems. This camera operates in the 7.5 – 13 µm spectral range using a focal plane array of uncooled microbolometer detectors, producing a 320 X 240 pixel calibrated thermal image. The accuracy of the instrument is ± 2 °C, or 2 % of the reading, whichever is greater. The camera has an optical system with a field of view (FOV) of 1.3 mrad. The actual pixel width in the acquired image varies with
distance from the instrument to the target (d) and can be calculated from \(2d \tan(1.3 \text{ mrad} / 2)\). Over the \(\sim 2500 \text{ m}\) line of sight, this gives a pixel width of 3.3 m, yielding thermal images approximately 1056 m x 792 m in dimensions. The instrument has the capability to perform automatic corrections to take into account surface emissivity and atmospheric effects using a LOWTRAN atmospheric model and user-specified variables for target emissivity, path length (distance from sensor to target), relative humidity and ambient temperature. Collected images can either be stored to a removable CompactFlash™ media card, the instrument’s limited onboard memory or, for streaming image acquisition at 15 to 30 images per second, by connection to a laptop via Firewire™.

I collected thermal images of the summit vent of Santiaguito’s Caliente dome unit during the early morning of January 16, 2004, from the summit of Santa María. Thermal images were acquired at a rate of 30 images per second between 2:52 A.M. to 6:38 A.M. (all times are local), where nighttime imaging minimized the effects of solar heating on recorded surface temperatures. This also took advantage of the clear viewing conditions of the early morning, which typically only last until around 10 A.M. during the dry season. Relative humidity and ambient temperature were measured in-situ at 15-minute intervals for atmospheric correction. For emissivity correction I used value of 0.9 following Birnie (1973). In addition, 52 individual images were collected around sunrise (during 7 – 8 A.M.) at random intervals to observe eruptive phenomena and
surface structures. Digital photos at 3 to 5 megapixel resolution were also collected to facilitate comparison with the thermal data.

A total of six explosions producing vertical ash plumes occurred during the nighttime acquisition period. Thermal data for these explosions appear to be of high quality only for the first two explosions (explosions #1 and #2). Subsequent explosions (#3 to #6) occurred during periods of increasing relative humidity (to greater than 85 %) culminating in some atmospheric condensation during the last two explosions.

4.3. Surface Thermal Distribution

4.3.1. Spatial arrangement of thermal structures

Figure 4.2 gives, from the summit of Santa Marfa, a panoramic view of the four lava dome units that comprise the Santiaguito dome complex. The mosaic is constructed using three FLIR daytime images. The thermal image mosaic confirms that current thermal activity at Santiaguito is strictly confined to the vent atop the Caliente unit, with no significant thermal anomalies being detected elsewhere. On Caliente itself the main thermal feature is located within the summit crater. However, two additional thermal anomalies are also present on the flank of the dome. The first is evident on the upper flank, immediately outside of the summit crater, and comprises an apron of relatively hot surfaces that appear to be radiating at elevated temperatures of 60 – 100 °C. Previous workers have observed, from video footage, the presence of fumarolic activity on the upper flank of the Caliente dome near the summit (e.g. Stoiber and Rose 1969,
Figure 4.2. a. Panoramic view of the four lava domes comprising the Santiaguito lava dome complex taken from Santa Maria summit showing the four main extrusive centers and a vertical ash plume emitted from Caliente vent. The distance from Caliente to Brujo is approximately 1.5 km. b. A composite of three FLIR thermal images of the same view as given in a showing the main flank and summit crater thermal anomalies of the Caliente in January 2004.
Chapter 3). Stoiber and Rose (1969) measured temperatures of fumaroles near the Caliente vent to be as high as 700 °C but decreasing down to 300 °C only ~0.9 m away from the fumarole vent. These fumarole vents will be sub-pixel features in the pixel sizes relevant to this study. Thus the pixel-integrated temperature will be somewhat lower than that of the vent residing within the pixel.

The second flank anomaly is located on the left hand side of the image, where an elongate feature is radiating at temperatures of 30 to 40 °C. The location of the feature corresponds to one of the main paths that minor pyroclastic flows, accompanying the vertical ash eruptions, extend and is thus most likely comprised of warm, cooling, pyroclastic flow deposits. During daytime periods this feature may not be thermally distinguishable as the heat from the sun warms the surrounding dome temperatures to a similar temperature range (Harris et al. 2002).

Figure 4.3 displays one of the thermal images collected immediately after sunrise, along with the corresponding digital photo, and focuses on the summit crater of the Caliente dome. From this image I can distinguish three different thermal zones within the summit vent area. These three zones are concentrically arranged and comprise: (1) an outer annulus that is radiating at relatively high temperatures, (2) an inner annulus that is relatively cool, and (3) a central core that is radiating at intermediate-to-high temperatures. To describe fully the temperature distribution of the summit vent I extracted temperature data along a horizontal cross-section marked L – L’ on Figure 4.4. This profile shows the three thermal regions in the horizontal plane. The hot outer annulus appears as the two
Figure 4.3. Thermal image taken using the FLIR camera of Caliente dome and its summit crater and a co-located digital photo for comparison. The thermal image shows the three-component thermal structure within the summit crater comprising an outer hot annulus, a cool inner annulus and a hot central core.
outer peaks in the profile, with peak temperatures ranging from 175 to 225 °C. The cool inner annulus appears as two temperature troughs with temperatures as low as 80 °C. The central temperature peak in the profile represents the central core region with temperatures of up to 150°C. I measured the dimensions of these thermal regions using this horizontal profile because, being orientated at right angles to the line of sight, it provides the least amount of error due to viewing angle effects. The maximum width of the outer annulus is approximately 10 pixels or 33 m. The maximum width for the inner annulus is 11 pixels or 36 m, and the central core has a diameter of approximately 10 pixels or 33 m. The thermal image (Figure 4.3) also shows the presence of what appear to be linear fractures, radiating outward from the central core and dissecting the inner annulus. These linear features are narrow (approximately 1 pixel or 3.3 m in width) and, in contrast to the rest of the inner annulus, have elevated surface temperatures ranging from 120 – 200 °C. The inner annulus appears to be horse-shoe-like in shape, due to a discontinuity on the back side (southwest portion) of the crater where the pixels are radiating at temperatures similar to those of the hot central core, effectively extending the hot central core region out towards the crater rim across that portion of the crater. The location of this discontinuity corresponds to a block lava flow (SW block lava flow) active at the time of image acquisition exited the vent to extend onto and down the dome’s outer flank.
Figure 4.4. Temperature profile across the summit crater illustrating the three thermal regions.
4.3.2. Temporal stability of thermal structures

The explosions produce intermittent changes in temperature or 'thermal renewals' of the surface thereby disrupting simple cooling of the surface and setting up a cyclic (cooling and renewal) thermal phenomena that was first observed and described in chapter 3. Six explosions were captured during the nighttime portion of my dataset and their effects on thermal structure of the vent region were examined. Given the lower humidity levels and lack of condensation, data for explosions 1 and 2 were the most reliable. Data for explosions 3 and 4 were collected during elevated humidity levels (>90 %) and may thus not be reliable in an absolute sense. During explosions 5 and 6 thermal data were collected in a condensing atmosphere and therefore the absolute temperature data are completely unreliable. Regardless of the deteriorating conditions leading to unreliable absolute temperature data, the relative (temporal and spatial) patterns observed around the times of all of the explosions are repeatable and remain valid for the purpose of observing the dynamics of the thermal structure of the vent surface.

Figures 4.5 and 4.6 give the thermal image sequences acquired during explosions 1 to 4 showing the evolution of the vertical ash plumes produced by each of the explosions. Note the temperature scale adjustment between the explosion 1 and 2 sequences and explosions 3 and 4 to accommodate the lower apparent temperatures of the latter explosions, which was probably caused by the higher humidity levels. Plumes emitted by the explosions were all initiated by emission across the inner annulus. A few seconds later the main plume emission
followed from the outer annulus. In the case of explosion 4 (Figure 4.6b) initial emissions from the inner annulus preceded the main plume by at least a minute. These observations are consistent with previous video observations of Bluth and Rose (2004) of an outward migration of the emission source from the inner to the outer annulus. Explosions 1 and 4 emitted plumes relatively evenly from across the entire outer annulus, producing cylindrical plumes. Explosions 2 and 3, however, showed most intense and prolonged emission from the right-side of the outer annulus, producing non-symmetrical plumes that deposited material preferentially on the right-hand flank.

Temperature time series constructed for the duration of an explosion can be used as a simple way to illustrate the evolution of the plume emission during the event, as well as the effect of the explosion on the surface temperature. I extracted two temperature time series for the duration of explosion 1, this being the explosion imaged under clearest conditions. The first time series involved tracking the maximum temperature recorded from the whole image, and the second tracked the maximum temperature recorded from the whole summit area (Figure 4.7). I tracked the whole-image maxima to identify any temperature anomalies beyond the summit crater, which should be indicative of elevated temperatures within the plume itself (i.e. the maximum is not at the vent region, but higher in the plume). Masking of the vent by cold ash during the emission may cause maximum temperatures to occur within the plume at some height above the vent itself, which will be missed if just vent temperature is tracked.
Figure 4.5. Sequence of thermal images taken during the duration of explosions #1 and #2.
Figure 4.6. Sequence of FLIR thermal images taken during explosions #3 and #4. Recorded temperatures are much lower overall compared to images from explosions #1 and #2 due to higher relative humidity.
The time series show multiple temperature peaks corresponding to pulses of varying intensity during emission. The time series also shows that, although the maximum plume temperature is not always recorded at the vent, for most of the time the maximum temperature does occur in the vicinity of the vent (Figure 4.7). For explosion #1, the initial plume emission has a maximum temperature of \(\sim 400 \, ^\circ C\), but the maximum recorded temperature (520 °C) occurs \(\sim 15\) seconds after initial emission from a plume area four to six pixels (25 – 40 m) above the vent area. This particular higher-temperature pulse was emitted from the back of the crater on the right-hand side (southwest portion) of the summit crater, and may actually had higher exit temperatures at the vent as the initial emission, and early ascent, was obscured by ash from initial cooler pulses from the front of the vent. The source of the maximum temperature pulses during ash emissions for all the explosions is consistently located within this region which corresponds to the location of the inner annulus discontinuity at the location where the active block lava flow exits the summit crater. The temperature time series for explosion 1 also shows that the highest temperature recorded at the vent is approximately 175 °C higher after the explosion when compared with the pre-explosion level (Figure 4.7).

A more spatial manner to document the thermal changes on the surface due to the explosions is to use the temperature images. The temperature distribution across the summit and its temporal evolution are presented as a series of isotherm maps in Figure 4.8. Here I use the thermal images acquire
Figure 4.7. Maximum temperature time series for the duration of explosion 1. Solid black line is the maximum temperature recorded from the vent area and the solid red line is the maximum temperature recorded from the whole image. Six emission pulses during the explosion can be identified by temperature spikes on the time series.
immediately before (a) and after (b) each of explosions 1 through 4 to show the changes caused by the explosion. I then produce a third temperature map (c) that shows the change in temperature caused by each explosion. Explosion 1 occurred 14 minutes into the dataset, and no previous explosion occurred during the instrument setup period of 20 minutes; meaning that there had been no explosion for at-least 34 minutes. I therefore assume that this temperature profile represents that characteristic of the end of a fairly long (>30 minute) cooling cycle, and approximates the maximum amount by which the surface is able to cool prior to disruption by a new explosion. In spite of this, the outer annulus can still be clearly identified outlining the margins of the summit crater. Most of the surface of the outer annulus at this point is, thermally, relatively homogenous and within the 100 to 150 °C temperature range, with the exception of the edges of the region, where 150 to 250°C temperatures are encountered. The inner annulus can also be distinguished by the relatively cool temperature range of 50 – 100°C just inside the hot zone of the outer annulus. Within the inner annulus, the central core is radiating at 100 – 200 °C. Significant temperature changes occurred across the outer annulus as the surface was transformed by the explosion from a relatively homogeneous surface into one marked by several clusters of higher temperatures arranged around the annulus. Typical temperatures increase to 200 – 300 °C, with some zones increasing to more than 300 °C. The inner annulus, on the other hand, experiences no significant changes in temperature, showing only slight increases in temperature at only localized areas. The central core shows a decrease in temperature by ~50 °C.
Figure 4.8. Temperature maps of Caliente summit for explosions 1 to 4 (E1 – E4). The three maps are given covering the zone marked by the black box on the digital photo, where each row gives isotherm maps for (a) before and (b) after each explosion, with (c) giving the temperature change map.
There was then approximately 10 minutes between explosions 1 and 2. Just prior to explosion 2 I find that temperatures, over this short time, had not cooled to the pre-explosion 1 level, and that the thermal structure across the outer annulus was still quite heterogeneous. The inner annulus was again relatively cool and homogenous, with localized areas of elevated temperature. The central core was still apparent as a zone of elevated temperature, with a similar temperature level and structure to that observed prior to explosion 1. The thermal structure following explosion 2 shows similar, if slightly less dramatic, changes in temperature to those encountered following explosion 1. Sources of elevated temperature are encountered in the same locations as mapped following explosion 1. I suspect that the locations mark more permeable fracture zones allowing more efficient gas-transfer and heat loss than less permeable zones, and likely are rooted to pathways extending downwards from the summit vent surface and into the upper conduit.

I included data from explosions 3 and 4 to illustrate that, although deteriorating atmospheric conditions affected absolute levels of the thermal observations, the same structures and temporal trends remained apparent around each explosion. Overall the poor conditions caused the explosion 3 and 4 temperature maps to be cooler and more homogeneous than during explosions 1 and 2. Additionally, only portions of the outer annulus can be distinguished on the explosion 3 and 4 maps. However, the post-explosion 3 map reveals the expected thermal structure, albeit at much lower temperatures. The explosion 4 maps again show identical structures, but contain a slightly higher temperature
contrasts when compared with the explosion 3 maps, consistent with improved atmospheric conditions at that time (i.e. relative humidity was lower than compared to that measured during explosion 3).

Although the absolute temperatures extracted across the vent area vary as a result of temperature changes caused by explosive events and changes in viewing conditions, the thermal structure remains stable through time, and between explosions. The resulting stable thermal structure is comprised of a thermally static inner lava region of low and a thermally dynamic outer annulus fill region. This thermal arrangement is consistent with a plug-like central region comprised of relatively static lava surface surrounded by marginal shear-zone or the outer annulus which is thermally dynamic. The assumed background state of the vent as represented by temperature map E1a (Figure 4.8) show that the outer annulus fill maintained higher temperatures than the inner region, which suggests preferential heating of the outer annulus from below and the stability of the sub-surface structure that is controlling the surface thermal pattern.

4.4. Shallow Level Structure from Thermal and Visible Images

Based on the surface thermal characteristics and their response to explosions I can make some correlation between the different thermal regions and the actual surface morphology. To do this I use the corresponding digital photography and in turn make inferences regarding the shallow level structure of the summit crater and vent/conduit system. Figure 4.9 shows a digital photo of the summit crater with thermal data overlain. The outer, hot, annulus
corresponds to a summit-wide circular depression that contains series of smaller pits. Compared with other regions of the summit crater, the entire area of the depression is filled with relatively fine clastic material. The pits have been observed by previous workers (e.g. Bluth and Rose 2004) and were proposed as vents within the ring-like depression of the outer annulus. The location of these individual vents is consistent with the location of the thermally elevated zones observed following each of the explosions, confirming them as locations of concentrated venting of hot gas and ash.

The inner, cold, annulus is located over a portion of the crater occupied by a blocky surface that appears to slope away from the center of the vent. Hot fractures dissecting the inner annulus observed in the thermal data are also apparent in the digital photo. Given its relatively cool temperatures and structure, the inner annulus is most likely to represent the surface of a thick, relatively mature, lava crust that has undergone some cooling. The observed fractures radiating at higher temperatures suggest that higher temperature lava resides just beneath this carapace of older, blocky, lava. The discontinuity in the portion of the inner annulus on the back of the crater is radiating at the same temperature range as these fractures. The central, hot, region enclosed within the inner annulus appears to be composed of a relatively smooth surface but radiates at temperatures in the same range as those encountered across the inner annulus fractures and at the back-wall discontinuity. I suggest that the discontinuity and the central hot thermal region represents surface exposure of active and most recent lava extrusion which feeds the active SW block lava flow.
Figure 4.9. Isotherm map overlain on top of a digital photo of the Caliente summit crater showing the relationship between the different thermal regions and morphological features of the summit.
This would suggest upwelling over the central core, before flow out of the crater to the southwest. Maximum plume emission temperatures observed across the region of discontinuity further support the presence of active lava extrusion in that portion of the vent: where fragmentation of the active lava in shallow conduit contribute to higher temperatures. A generalized cartoon summarizing the inferred surface structure along with two cross-sections across the summit crater is given in Figure 4.10.

The temperature of the outer annulus fill is significantly higher than that of the exposed lava surfaces. Further, the zones of greatest increase are at the proposed vent zones within the outer annulus (Figure 4.8). These high temperatures can be explained either by deposition of hot pyroclastics from the explosions and/or excavation of the annulus fill by the explosion thereby exposing hotter underlying surface. The non-uniform temperature increase across the summit area does not support pyroclastic deposition as it is difficult to justify deposition of hot materials strictly along the outer annulus. Furthermore the location of the highest temperature zones is identical following all explosions. I suggest that the explosions thus excavate the fill material of the outer annulus, thus penetrating higher temperature, previously buried, layers which then begin to cool. In addition, the occurrence of highest temperatures at the localized "vent" regions may be due to clearance of material to a greater depth, as well as more efficient heat supply, to these high permeability regions. I suggest that the high temperatures encountered around the outer annulus in general are a result of this region being located atop of a marginal fracture zone surrounding a
Figure 4.10. Summit crater and shallow conduit structure in plan view and cross-section showing the outer annulus fill heated by the underlying lava, carapace of older extruded lava and the recent lava extrusion exposed at the center and flowing to the southwest to feed the SW-flank block lava flow. Total width of the crater is approximately 180 m and cross-sections are not to scale vertically.
central lava plug of much lower permeability.

4.5. Surface Heat Flux

The calibrated thermal data obtained under good conditions allows me to quantify the proportion of heat loss associated with each thermal zone identified across the vent area. I assume that cooling and crystallization of magma ascending the conduit to feed the explosive emissions, lava flow extrusion and permeable gas flux provide the heat lost at the summit vent. The extruded magma is exposed in the hot central core area, the fractures within the inner annulus, and the intra-crater portion of the SW block lava flow marked by the discontinuity. Dissipation of heat from this extruded lava is assumed to occur via radiative ($q_{\text{rad}}$) and convective heat transfer ($q_{\text{conv}}$) into the atmosphere. Lateral heat transfer via conduction through conduit walls may also contribute to the overall heat budget but was not quantified in this study. Given the FLIR-derived surface temperature ($T_{\text{surf}}$) and measured ambient temperature ($T_{\text{air}}$), the heat flux density $q_{\text{rad}}$, in J/s/m$^2$, can be calculated using the Stefan-Boltzmann equation:

\begin{equation}
q_{\text{rad}} = \varepsilon\sigma\left(T_{\text{surf}}^4 - T_{\text{air}}^4\right)
\end{equation}

where $\varepsilon$ is the surface emissivity (0.9) and $\sigma$ is the Stefan Boltzmann constant ($5.67 \times 10^{-8}$ J/K$^4$/m$^2$/s$^1$). Convective heat loss can be calculated as free convection ($q_{\text{free}}$) in cases where there is no wind or forced convection ($q_{\text{forced}}$)
during windy conditions. I did not collect wind measurements at this time thus I can only estimate free convective heat flux density \( (q_{\text{free}}) \) while noting that, during the measurements, winds were observed to be light to minimal. Free convection can be calculated from:

\[
q_{\text{conv}} = h_c (T_{\text{surf}} - T_{\text{air}})
\]

where \( h_c \) is the free convection heat transfer coefficient calculated for each of the thermal zones, where I obtained \( h_c \) of 8.4 – 11.4, 6.2 – 7.6 and 7.7 – 9.5 J/s/m\(^2\)/K, for the outer annulus, inner annulus and central core, respectively (see Appendix 1).

I estimated heat flux density for each of the thermal component identified within the summit crater (Figure 4.11). The first component of the thermal model is the young lava, radiating at 100 – 200 °C. This component occupies the 830 m\(^2\) area extending across the hot core zone, the five fractures within the inner annulus (total area of 437 m\(^2\)), and the intra-crater portion of the SW-flank lava flow which dissects both the inner and the outer annulus (13 m wide with a total surface area of 455 m\(^2\)). This gives a total area for of 1722 m\(^2\). The second thermal component occupies the inner annulus zone which is mostly comprised of the cooled crusts of older lava radiating at 40 – 80 °C. The total surface area of this region (minus the fractures and the lava flow) is 6530 m\(^2\). The third component occupies the outer annulus which is filled with loose material radiating at 150 – 400 °C, with a total surface area (minus the discontinuity of the
Figure 4.11. Generalized model of the summit thermal surface used to calculate surface heat flux density and power loss.
lava flow sector) of 12,904 m$^2$. Multiplying area by heat flux density, allows the total heat flux or power loss from each zone (in J/s) to be calculated. The results of the calculations are summarized in Table 1. The minimum and maximum limits of calculated values are dictated by the minimum and maximum of the range of temperatures used for each region. Maximum heat flux density occurs at the outer annulus and at the inner annulus, which is comprised of cool crust of older lava, having the lowest heat flux densities. Young lava surfaces (central core + fractures + discontinuity thermal zones) have values intermediate between those of the outer and inner annulus (Table 1). Total power loss ($Q_{surf} = Q_{rad} + Q_{free}$) from the summit vent (in MJ/s) is dominated by the outer annulus fill, where $Q_{surf}$ ranges between 31 – 177 MJ/s. This accounts for 90 – 94 % of the total power loss from the entire vent, highlighting the permeability of this region for heat and gas flow. Despite its low temperature, the inner annulus (old lava) contributes more to the overall power loss (2.6 – 6.9 MJ/s or 8 – 4 % of the total) than the much higher temperature zones of the fractures, inner core and discontinuity. These together contribute 1.2 – 3.5 MJ/s or ~2 % of the total. This is a result of the January 2005 surface configuration, wherein the young lava surface zones exposed at the core, fractures and discontinuity comprised the smallest fractional area of the summit crater.

4.6. Surface Crust and Depth to Magmatic Temperatures

The young lava surface component at the summit radiates at temperatures of 100 – 200 °C, significantly lower than the expected magmatic
Table 1. Calculated heat flux densities \( (q_{rad} \text{ and } q_{free}) \) in J/s/m\(^2\) and power loss \( (Q_{rad}, Q_{free} \text{ and } Q_{surf}) \) in MJ/s) for each thermal zone within the summit crater for the observed minimum and maximum FLIR-derived temperatures \( (T_{surf}) \).

<table>
<thead>
<tr>
<th></th>
<th>Core, fractures discontinuity (Young Lava)</th>
<th>Cold Inner Annulus (Old Lava)</th>
<th>Outer Annulus Fill</th>
<th>TOTAL</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>min</td>
<td>max</td>
<td>min</td>
<td>max</td>
</tr>
<tr>
<td>( T_{surf} ) (^{\circ}\text{C})</td>
<td>100</td>
<td>200</td>
<td>40</td>
<td>80</td>
</tr>
<tr>
<td>( q_{rad} \times 10^3 ) J/s/m(^2)</td>
<td>0.68</td>
<td>2.25</td>
<td>0.19</td>
<td>0.49</td>
</tr>
<tr>
<td>( q_{free} \times 10^3 ) J/s/m(^2)</td>
<td>0.72</td>
<td>1.86</td>
<td>0.22</td>
<td>0.57</td>
</tr>
<tr>
<td>( q_{surf} \times 10^3 ) J/s/m(^2)</td>
<td>1.41</td>
<td>4.11</td>
<td>0.41</td>
<td>1.06</td>
</tr>
<tr>
<td>( Q_{rad} ) MJ/s</td>
<td>0.6</td>
<td>1.9</td>
<td>1.2</td>
<td>3.2</td>
</tr>
<tr>
<td>( Q_{free} ) MJ/s</td>
<td>0.6</td>
<td>1.6</td>
<td>1.4</td>
<td>3.7</td>
</tr>
<tr>
<td>( Q_{surf} ) MJ/s</td>
<td>1.2</td>
<td>3.5</td>
<td>2.6</td>
<td>6.9</td>
</tr>
</tbody>
</table>
temperature of dacite (800 – 1100 °C, Cas and Wright 1987). In Chapter 3 I applied a two-component thermal model to hyper-spectral data and found that the summit surface comprised a 120 – 250 °C 'cold component' broken by (areally) very small fractures radiating at near magmatic temperature of ~950 °C. The temperature range for the cold component I obtained in Chapter 3 is consistent with the measurements given here, suggesting that the measurements are dominated by same type of “cold” crusted surface. The FLIR failed to resolve the higher temperature surface mainly due to the low spatial resolution and the single-channel spectral resolution of the FLIR coupled with the small areal extent of the high temperature “cracks”. The relatively low temperature compared to the magmatic temperature can be attributed to the initial rapid cooling of extruded lava, forming an outer crust that insulates the underlying material that is at magmatic temperature. A similar phenomenon is observed down the silicic flows that this vent feeds, where exceedingly cool (near-ambient) crust insulate the flow core (Harris et al. 2002, 2004). In fact, these at-vent crust temperatures match the near-vent surface maxima obtained for the lava flow surfaces by Harris et al. (2004) of ~150 °C.

From the heat flux density estimates I can estimate the depth to the 900 °C isotherm for the lava regions of the summit using Fourier’s Law. Using this approach, depth to the 950 °C isotherm may be viewed as roughly equivalent to the thickness (δ) of the outer crust on any lava body (e.g. Oppenheimer 1991, Harris et al. 2002). The main assumption is that the rate of heat loss at the
surface is equal to the rate of heat conducted from the magmatic surface upwards through an overlying boundary layer, so that,

\[ q_{\text{cond}} = k \frac{(T_{\text{core}} - T_{\text{surf}})}{\delta} \]

where \( k \) is the thermal conductivity of the surface, assumed for Santiaguito rocks to be 2 ± 1 W/m\( ^{\circ} \)K (Robertson and Peck, 1974), \( T_{\text{core}} \) is the core temperature and \( T_{\text{surf}} \) is the surface temperature. \( T_{\text{core}} \) can be assumed to be equivalent to the eruption temperature (\( T_{\text{erupt}} = 950 ^{\circ} \text{C} \), chapter 2). For the young lava surface of the core, fracture and discontinuity zones with \( T_{\text{surf}} \) range of 100 – 200 °C, the depth to magmatic layer (or the crust thickness) ranges from 0.6 to 2 m. The depth to magmatic surface below the old extruded lava exposed across the inner annulus radiating at 40 – 80 °C, ranges from 2.7 – 7 m.

Figure 4.12 shows the depth to magmatic temperature layer below the inner lava regions, or relative thickness of the lava units along the summit cross-section calculated using equation 4.3. The temperature profile used is extracted prior to explosion 1, which is assumed to be the stable state of the summit crater. For the calculation of free-convective heat flux density I calculated heat transfer coefficients for each pixel along the cross-section (Appendix I), with values ranging from 3.8 to 10.2 J/s/m/K. The thickness of the carapace on the older lava that comprised the inner annulus range from 1 to 3 m, The crust thickness of the young lava of the inner annulus is much lower between 0.3 to
Figure 4.12. Estimated crustal thickness of the old lava (inner annulus) and young lava (central core) regions in the summit crater. Error bars represent the uncertainty in thermal conductivity of dacite lava (2 ± 1 W/m/K).
1.5 m. The thickness of the fill layer of the outer annulus was not calculated given the uncertainty in the particle size and the porosity of the granular deposit, complicating the calculation of total heat conducted through deposit.

4.7. Conclusions

Following Gonnerman and Manga (2003), intermittent explosions and associated vertical ash plumes at Santiaguito have been attributed as resulting from shear-induced fragmentation along the conduit walls, where the outer annulus is a surface expression of an annular fracture zone at the margins of the broadly cylindrical, conduit (Bluth and Rose 2004, Chapter 2 of this dissertation). In Chapter 2 I further suggested that fragmentation occurs at depths between 100 – 500 m below the vent and that generates pyroclastic material that propagates up the fracture zone at conduit boundary, exiting along the outer annulus. As such the conduit boundary provides an easy pathway, intermittently ‘cleaned out’ by the explosions, up which heat and mass from the shallow-surface magma column can escape. This is consistent with the observation that the outer annulus is stable in space and time and is the most productive heat generator at the summit crater.

This sub-surface structure results in a plug like feature within the crater, comprising a central core of lava, which spills out of the crater to feed flank lava flows, surrounded by the hot, outer annulus that marks the top of the marginal shear zones surrounding this plug. Interestingly, this thermal and structural make-up means that intermittent explosions result in elevated temperatures
across some of the vent surface (mainly at the outer annulus) and not across the central core or plug. Simply thermal renewal is only effective at the outer annulus, where heat and mass flow is more efficient by virtue of the high permeability of this marginal fracture zone when compared with the central plug. This in turn drives the wide range of total summit power output observed at Caliente vent (154 - 828 MJ/s), where heat loss is most efficient (highest) at the marginal zones and reduced across the plug.
4.8. Appendix I

Convective heat transfer coefficient

Free-convection heat transfer coefficient \( (h_c) \) is calculated following Holman (1992) using the Nusselt Number \( (Nu) \),

\[
(4.4) \quad Nu = \frac{h_c H}{k_{air}}
\]

where \( H \) is the thickness of the hot fluid overlying the surface (m) and \( k \) is the thermal conductivity of air (J/s/m/K). In free convection cooling can be calculated using the Rayleigh number \( (Ra) \),

\[
(4.5) \quad Nu = 0.16Ra^{1/3}
\]

where \( Ra \) is the product of the ratio of buoyancy force to viscous force (or the Grashof, \( Gr \)) number and the of ratio between momentum diffusivity and thermal diffusivity (or the Prandtl number, \( Pr \)):

\[
(4.6) \quad Ra = Gr \cdot Pr
\]

Grashof number can be calculated by,

\[
(4.7) \quad Gr = \left[ \frac{g \beta (T_{surface} - T_{ambient}) H^3}{v^2} \right]
\]
Where \( g \) is gravitational acceleration (9.8 m/s), \( \beta \) is the inverse of the mean temperature of the overlying thermal boundary layer \([(T_{\text{surface}} + T_{\text{ambient}})/2]\), and \( \nu \) is the kinematic viscosity of the overlying fluid (m\(^2\)/s). \( Pr \) is given by,

\[
Pr = \frac{\nu}{\alpha}
\]

where \( \alpha \) is thermal diffusivity (m\(^2\)/s). This sequence of equations is used to calculate \( hc \) on a pixel-by-pixel basis using the pixel temperature for \( T_{\text{surface}} \) and the measured atmospheric air temperature of \( T_{\text{ambient}} \).
Chapter 5

FLIR and ASTER observations of the thermal structure, heat-loss and discharge rate of an active block-lava flow at Santiaguito, Guatemala

5.0. Abstract

An active silicic block lava flow at Santiaguito (Guatemala) was observed using both a ground-based Forward-Looking InfraRed (FLIR) video camera and the satellite-based Advanced Spaceborne Thermal Emission and Reflection Radiometer (ASTER) during January 19 and 20th, 2005. These images provided thermal data for the 700 m long flow at spatial resolutions of 5 and 90 m, respectively. Down-flow temperature profiles revealed a 175-m long proximal zone across which surface temperatures declined from 200 to 70 °C (to give a surface cooling rate of 0.74 °C/m). A 250-m long medial zone was characterized by stable and low surface temperatures (35 – 70 °C), whereas the distal zone was marked by a 275-m long zone of higher surface temperatures. Calculated cooling rates at this well-insulated flow are 135 to 600 °C/km across the proximal section, decreasing to 140 – 200 °C/km across the medial-distal section, leading to total (down-flow) core cooling of ~100 °C. Down-flow, crust thickness is calculated as increasing from 1 – 2 m across the proximal zone, to up to ~4.5 m across the medial-distal zone; across which there are also localized decreases due to exposure of hotter underlying surfaces during localized rock falls. Estimated discharge rates for January 2005 (0.15 – 1.2 m³/s) are consistent with
previous estimates for the 1999 – 2002 flow, and show that extreme thermal insulation allows flow extension to several kilometers in spite of low discharge rates.

5.1. Introduction

The extrusion of dacite lava at the Santiaguito dome complex, Guatemala, has been ongoing since 1922 (Rose 1987). Extrusive activity has been characterized by cycles comprising three to six year-long periods of high extrusion rate (0.5 – 2.1 m$^3$/s) followed by three to eleven-year periods of low extrusion rate (< 0.2 m$^3$/s). By 2000, this persistent extrusive activity had built a dome complex, comprising 4 extrusive centers, at a time-averaged extrusion rate of ~0.45 m$^3$/s (Rose 1987, Harris et al. 2003).

Since 1970, extrusive activity has been marked by emplacement of a series of increasingly longer silicic (62 – 64 wt % SiO$_2$) block lava flows, which have extended 3.9 km down the flanks of the dome complex (Harris et al. 2003, 2004). These flows are typically slow moving, with flow front advance rates of 2 – 13 m/day, and of high viscosity and yield strength (Harris et al. 2004). Ground-based thermal infrared thermometer and satellite-based (Enhanced Thematic Mapper Plus) observations of the flow active between 1999 and 2002 showed that the flow surface is characterized by low surface temperatures - typically less than 100 °C (Harris et al. 2002, 2004). The presence of a cool, thick, stable crust provides effective insulation for the flow core, reducing cooling rates to ~0.5 °C/day). As a result, the 1999 – 2001 flow was capable of attaining cooling-
limited distances of 3 to 4 km despite relatively low extrusion rates (0.5 – 1.6 m³/s) and high viscosities (4 – 70 x 10⁹ Pa s) (Harris et al. 2002).

Measurements of surface temperature of the lava flow constrain estimates of surface heat losses, thereby allowing parameters such as core cooling, crystallization and rheological changes to be modeled (e.g. Dragoni 1989, Crisp and Baloga 1990, Cashman et al. 1999, Harris et al. 2005a). This, in turn, allows estimates of how far a flow can extend before cooling forces the flow to stop (e.g. Ishihara et al. 1990, Keszthelyi and Self 1998, Harris and Rowland 2001, Hikada et al. 2005). In the case of Santiaguito, effective thermal insulation reduces core cooling so as to allow the flows to extend several kilometers at low extrusion and advance rates (Harris et al. 2002). In this study I present the results of thermal observations of the surface temperature of the silicic block flow active on the SW flank of Caliente Dome (Santiaguito) during January 2005. Thermal data for this active block flow were acquired using both ground-based and satellite-based thermal imagers, the former being a tripod-mounted Forward-Looking InfraRed (FLIR) thermal video camera and the latter being the Advanced Spaceborne Thermal Emission and Reflection Radiometer (ASTER) instrument flown aboard NASA’s Terra satellite. Here I describe the spatial variations in surface temperature and use these to provide well-constrained surface heat loss estimates. I use these to assess and cross-check crust thickness, core cooling rates and extrusion rates calculated from FLIR and ASTER, respectively. This allows me to confirm, by comparison with the 1999 – 2002 flow, the predictable emplacement and cooling behavior of Santiaguito’s lava flows.
5.2. Thermal Dataset

The data collection aims of this study were two-fold. First, to use a ground-based thermal video camera to obtain surface temperature maps of an active block lava flow. These image data will provide high spatial resolution thermal maps covering the entire flow surface, as opposed the point-based measurements taken for selected down flow locations as have previously been made for Santiaguito's flows using thermal infrared thermometers (Harris et al. 2002, 2004). Second, to collect the field-based thermal data simultaneous with an overpass of a high spatial resolution satellite-based thermal sensor, allowing comparison of the two data sets.

The ground-based camera used here was a Forward-Looking InfraRed (FLIR) thermal video camera (model ThermaCam™ P40 from FLIR™ Systems). The FLIR collects emitted radiance in the 7 – 13 micron spectral range using a focal plane array (FPA) uncooled microbolometer detector. The instrument produces a 320 x 240 pixel image, where each pixel gives a calibrated pixel-integrated temperature with an accuracy of ±2 °C or 2% of the reading (whichever is greater). Spatial resolution of the image varies with distance between the sensor and target (d) and is given by 2d tan(1.3 mrad / 2), 1.3 mrad being the instantaneous field of view (equivalent to the pixel size). Specifying the target's emissivity, ambient temperature and relative humidity during acquisition, as well as the distance to the target, allows corrections for emissivity and atmospheric effects.
The acquisition site was located on a ridge (UTM Zone 15N, 0653448
1626301, elevation 1343 m asl) located 3.8 km south of Santiaguito's Caliente
vent (Figure 5.1a). The active flow was moving SW away from the vent, with a
center-line orientated approximately at right-angles to the camera's field of view.
For the 3.8 km camera-to-flow distance, spatial resolution for pixels along the
flow will be approximately 5 m. To avoid problems associated with solar heating,
the FLIR thermal images were collected on 19th of January 2005 at 5:07 A.M.
(Figure 5.1b) local time (i.e. just before sunrise). At the acquisition time, ambient
temperature and humidity were 20 °C and 62%, respectively, and wind speed
was zero. Based on previous emissivity measurements made of banded andesite
from a lava dome that is considered to be similar to rocks of Santiaguito (Birnie,
1973), the flow surface emissivity was assumed to be 0.9. All corrections, and
analysis of thermal data, were carried out post-acquisition using the FLIR
analysis software (ThermaCAM Researcher™).

ASTER (Advanced Spaceborne Thermal Emission and Reflection) is a
multi-spectral imager on-board NASA's Terra platform. Terra is in a sun-
synchronous orbit with repeat cycle of 16 days. The multispectral imager collects
data with three separate instruments that operate in the Visible and Near Infrared
(VNIR, 3 bands, 0.52 – 0.86 µm), Shortwave Infrared (SWIR, 6 bands, 1.6 – 2.43
µm) and Thermal Infrared (TIR, 5 bands, 8.13 – 11.65 µm). These respectively
have spatial resolutions of 15, 30 and 90 m. The ASTER image used in this study
was a Level 1B product (in which radiometric and geometric calibration has been
applied to the original Level 1A data) collected on January 20, 2005, at 10:40
Figure 5.1a. ASTER image (VNIR bands 3,2,1) false color composite (vegetation = red) of the Santiaguito lava dome complex showing the four centers comprising the complex (Brujo, Monje, Mitad and Caliente). The Santa María peak to the northeast and the observation site where FLIR images of the 2005 flow were collected are marked. FLIR image coverage is shown as a yellow swath. b. ASTER TIR (thermal infrared) band 14 image. c. FLIR thermal image of the Caliente and the 2005 SW flank lava flow with co-located digital photo.
Figure 5.1a is the ASTER image band combination 3, 2 and 1, showing the whole extent of the Santiaguito lava dome complex within the 1902 eruption crater of Santa María. The summit crater of the Caliente dome and the active lava flow appears as thermally anomalous pixels only in the TIR bands (Figure 5.1a); their surfaces being too cool to emit detectable radiation in the VNIR and SWIR bands. The radiance data provided by ASTER’s TIR bands were converted to brightness temperature ($T_b$) by applying an approximation of the inverse Planck function that uses the central wavelength ($\lambda$) to convert spectral radiance ($R$) to $T_b$ (Alley and Jentoft-Nilsen 1999):

$$T_b = \frac{c_2}{\lambda \ln \left[ c_1 \lambda^{-5} / \pi R \right] + 1}$$

where $R$ and $\lambda$ are in units of W/m/sr/µm and µm. Constants $c_1$ and $c_2$ have values of $3.742 \times 10^{-22}$ W m$^2$ and 0.0143879 m K. Spectral radiance values are corrected for an emissivity of 0.9 and for atmospheric effects using the MODTRAN 4 atmospheric model run using a winter tropical atmosphere for a surface elevation of 2500 m.

5.3. Flow Surface Thermal Structure, Dimensions and Extrusion Rate

The FLIR and ASTER thermal images for the January 2005 lava flow at Santiaguito are shown in Figures 5.1a and b. What is immediately apparent is that the FLIR image is of a higher spatial resolution than the ASTER image, but
the view is oblique with respect to the flow, so that only the SE facing section of
the flow is captured. In contrast, the ASTER image is of lower spatial resolution
but the view is vertical so that the entire flow surface is imaged. Together,
however, these two images allow the dimensions and thermal character of the
flow to be determined.

The summit vent of Caliente is clearly the hottest feature in the ASTER
image, the surface thermal structure of which has been described in Chapter 4.
As is apparent from the images (Figure 5.1), the flow leaves the vent across the
SW rim to extend down the southwestern flank of the Caliente. The thermal
anomaly dimensions of the flow in the FLIR and ASTER images defined an
active flow unit 700 m long, with the ASTER image giving a typical width of one
TIR pixel (i.e. 90 m). This flow had been active for 2.5 months, giving a time-
averaged flow front advance rate of 9.1 m/day. This flow velocity compares to
the flow front advance rate of 12.5 m/day obtained for the 1999 – 2002 flow over
the first ~6 months of its advance (Harris et al. 2002). Extending down slope from
the flow toe is an apron of loose, warm clastic material produced by collapses
and rock avalanches from the active flow front. This has an average FLIR-
derived surface temperature of 20 °C. A second warm rock-fall trail is also visible
in the FLIR-image running down the left-bank of the flow. This trail is caused by
frequent collapses and rock falls from the upper part of the flow. These then
extend down the flow margin. Some larger blocks can be identified in this warm
talus region with FLIR-derived temperatures of 37 to 50 °C. Such rock falls are
common from the levees and flow fronts of these Santiaguito flows which, being
emplaced on the ~30° flanks of the Caliente, are quite unstable. Over a 6.7 hour observation period on January 25, 2000, a total of 231 rock fall events from the active block flow were observed by Harris et al. (2002).

Two other thermal anomalies are apparent in both the FLIR and ASTER images (Figure 5.1a and b). The first of these are defined by warm surface zones in the proximal section of the inactive, but still cooling, 1999 – 2002 flow unit. In addition, a region of slightly elevated surface temperatures on the south-eastern flank of the Caliente is a chute down which small pyroclastic flows (associated with vertical ash plumes emitted by the vent every ~30 minutes) extend. This anomaly is thus associated with accumulations of still-warm material deposited by these pyroclastic flows.

At basaltic flows, a two-component thermal model is typically applied to describe the surface thermal structure in which a hot crust is broken by cracks which expose higher temperature interior material (e.g. Crisp and Baloga 1990, Oppenheimer 1991). However, at the Santiaguito vent, the fractional area occupied by hot cracks is very small (< 0.02%) and contributes insignificant amount of heat relative to the total surface heat loss (Chapter 3). Thus the lava surface at Santiaguito can be modeled as a single-component thermal radiator. I therefore assume that the pixel-integrated temperatures approximate that of a crack-free surface crust and use these to characterize the surface temperature down the length of the flow.

Surface temperature profiles, taken down the center of the active flow using both the FLIR and ASTER data, are given in Figure 5.2. From these
profiles I can divide the flow into a three distinctive thermal regions on the basis of temperature: proximal, medial and distal. The proximal region extends 175 m from the vent and contained the highest surface temperatures encountered along the flow. This region is characterized by rapid decrease in surface temperature, from 200 °C at the vent to 70 °C after 175 m, giving a rate of decay of 0.74 °C/m. The medial region extends between 175 and 425 m and is characterized by low and scattered but generally stable (with distance) temperatures, some of which are as low as 35 °C, with an average rate of decay of 0.16 °C/m. The distal region covers the final 350 m of the flow and is marked by a general increase in surface temperature, again with scatter. Temperatures are still, however, relatively low, spanning a typical range of 30 – 80 °C (Figure 5.2).

ASTER-derived temperatures are lower than FLIR-derived values especially across the proximal section (Figure 5.2). This is a result of the coarser spatial resolution. Given that ASTER integrates temperature from a much larger pixel (90 m) compared to the FLIR (5 m), there is some influence from the ambient surfaces surrounding the flow in the ASTER pixel-integrated temperature. This reduces ASTER pixel-integrated temperatures when compared with those obtained from the pure, flow-containing, FLIR pixels, and indicates that the ASTER pixels are not entirely filled with lava. The flow width must therefore be narrower than the ASTER pixel width. The down-flow thermal trend, however, is consistent between the two datasets and surface temperatures are actually in excellent agreement across the medial and distal sections (Figure
Figure 5.2. Surface temperature data along the length of the 2005 lava flow derived from FLIR and ASTER images along and temperature profile of non-lava surface adjacent to the flow. Intra-crater portion of the flow is excluded from this profile.
5.2). Together, the two images allow the area to be obtained from the ASTER pixel-integrated temperature \(T_{\text{ASTER}}\) and the FLIR-derived flow surface temperature from the following two component mixture model:

\[
L(\lambda, T_{\text{ASTER}}) = pL(\lambda, T_{\text{surf}}) + (1 - p)L(\lambda, T_{\text{back}})
\]

in which \(L\) is the Planck function for a blackbody radiating at temperature \(T\) and wavelength \(\lambda\), \(p\) is the pixel portion occupied by the lava flow, and \(T_{\text{back}}\) is the temperature of surrounding ambient surface. I obtained \(T_{\text{surf}}\) from the mean FLIR-derived surface temperature for each 90-m segment of the flow, \(T_{\text{back}}\) from flow free ASTER pixels adjacent to the flow. This gives flow area of 0.05 km\(^2\), which (when divided by the flow length of 700 m) gives a typical width of 55 m. Assuming a flow thickness of around 20 m based on the measurements of Harris et al. (2003), the cross-sectional area of the flow is 1109 m\(^2\) which, with the time-averaged advance rate (9.1 m/day), gives a time-averaged discharge rate \((E_T)\) of 0.15 m\(^3\)/s.

5.4. Heat Loss

For active basaltic flows, the high surface temperatures mean that radiation, followed by convection, are the dominant heat losses (e.g. Flynn and Mouginois-Mark 1994, Harris et al. 2005). The lava flow heat budget is therefore typically calculated using surface heat losses due to radiation and convection, with conduction occasionally included (e.g. Oppenheimer 1991, Harris et al.
Such a model has been applied to Santiaguito (Harris et al. 2002, 2003). However, given the low surface temperatures at Santiaguito, radiation will have to take into account the contribution due to solar heating, and heat loss due to rainfall and conduction through the base of the flow will have to be considered.

Radiant heat flux \( q_{\text{rad}} \) in W/m\(^2\) down the length of the flow can be calculated using the Stefan-Boltzmann equation:

\[
q_{\text{rad}} = \varepsilon \sigma (T_{\text{surf}}^4 - T_{\text{air}}^4)
\]

where \( \varepsilon \) is the surface emissivity (0.9), \( \sigma \) is the Stefan-Boltzmann constant (5.67 \( \times \) 10\(^{-8}\) W/m\( \cdot \)K\(^4\)), \( T_{\text{surf}} \) is the flow surface temperature, and \( T_{\text{air}} \) is the ambient air temperature. In order to better display down-flow variations in heat flux, \( T_{\text{surf}} \) profile used to calculate down-flow heat flux values is taken as the average FLIR-derived temperature for corresponding ASTER pixels. The FLIR-derived temperature data of the flow used in this study was acquired at nighttime to avoid contamination from solar radiation. To confirm this I extracted a temperature profile for non-lava surfaces immediately adjacent to the flow and parallel to its center line. Temperatures of the ambient surfaces around the flow were lower than the recorded ambient temperature during acquisition time (20 °C), which resulted in negative radiation values or cooling of the non-lava surface by the ambient air. This confirms that emitted radiance due to solar heating did not contribute to the total radiant heat flux detected from the flow surface at FLIR acquisition time. To consider the maximum possible contribution of solar
radiation to total radiative heat flux during the daytime, I assume that the greatest contribution of solar heating to the total radiation will be in the early evening when ground temperature is still relatively warm, and the air temperature is relatively cool. Assuming an average ambient temperature of 30 °C and ambient temperature similar to that recorded during the early morning acquisition time (20 °C), re-emitted radiant energy from solar heating can possibly account for up to 10% of the total radiation detected from the flows.

At the recorded surface temperatures of the block flow (30 – 200 °C), variation in the ambient air temperature becomes significant in the calculation of heat loss due to radiation. The average daily temperature high and low recorded for Quetzaltenango which is the closest city (10 km to the north) to Santiaguito, are 22 and 35 °C, respectively (data from http://weather.msn.com). In Chapter 4 I observed ambient temperature as low as 5 °C. Figure 5.3a shows a plot of the radiative heat flux density envelope calculated given an ambient temperature range of 5 to 35 °C, and the intermediate radiative heat flux density obtained using the ambient temperature at acquisition time (20 °C). Higher surface temperatures in the proximal region resulted in maximum radiative heat flux values of between 450 – 2300 W/m². In the medial region radiative heat flux density decreased from 450 down to as low as 62 W/m². The distal region is marked by an increase in heat flux density up to 450 W/m².

Given the lack of wind during acquisition, convective heat flux density ($q_{conv}$) can be calculated for the free-convection case using the following equation,
Figure 5.3. a. The calculated radiative heat flux \( (q_{\text{rad}}) \) envelope given by the range of air temperature at Santiaguito (5 to 35 °C) and compared to radiative heat flux calculated for the temperature observed at acquisition. b. Convective heat flux \( (q_{\text{conv}}) \) calculated for the free convection case (no wind) where heat transfer coefficient \( (h_c) \) was calculated from individual FLIR pixels, and the forced convection case for gusty condition, where \( h_c \) is assumed to be 50 W/m\(^2\). c. Conductive heat flux from the base of the flow \( (q_{\text{base}}) \) for the range of thermal conductivity \( (k) \) for dacite.
where $h_c$ is the free convection heat transfer coefficient calculated for each pixel following the methods outlined in Chapter 4. Executing this series of equations for each ASTER and FLIR pixel gives $h_c$ of 7 – 10 W/m K. In windy conditions, the free convection heat transfer coefficient is replaced by the forced convection coefficient. Keszthelyi et al. (2003) measured $h_c$ values of 40 – 50 W/m K during wind gusts over active lava at Kilauea (Keszthelyi et al. 2003). To consider both free and forced convection cases, I use $h_c$ values calculated from the ASTER and FLIR data for the wind-less case and the maximum value of 50 W/m/K for a gusty wind case. For the proximal region, $q_{conv}$ ranged from 100 – 2000 W/m$^2$ for the windless case and 560 – 9000 W/m$^2$ during gusty wind scenario. In the proximal region, convective heat flux density decreased from 2000 – 9000 W/m$^2$ down to 580 – 3500 W/m$^2$. In the medial region, convective heat flux density decreased to 200 – 1500 W/m$^2$, followed by an increase in the proximal region to 200 – 3000 W/m$^2$.

Heat loss from the lava flow due to conduction from the base of the flow into the underlying basal layer ($q_{base}$) was estimated using the method outlined in Wooster et al. (1997) and based on equations of Turcotte and Schubert (1982). In this method the temperature of the contact between the flow's base and underlying surface ($T_b$) is first calculated using the following two equations from Turcotte and Schubert (1982),

$$q_{conv} = h_c (T_{surf} - T_{air})$$
where $T_o$ is the ambient surface temperature ($10 - 35 \, ^\circ C$), $T_o$ is the eruptive temperature ($950 \, ^\circ C$), $erf$ is the error function, $\lambda$ is a dimensionless scaling factor, $\varphi$ is the mass fraction crystallization (0.45), $F$ is the latent heat of fusion ($4 \times 10^5$ J/kg), and $C_p$ is the specific heat capacity of dacite ($1150 \pm 250$ J/kg K). $T_b$ is calculated by first evaluating the left-hand side of equation 5.6 and solving for $\lambda$ using numerical iterations. Using the calculated $\lambda$ value in equation for yields $T_b$ value of 857 $^\circ C$. Finally, $q_{base}$ (W/m$^2$) can be calculated by,

$$q_{base} = k(T_b - T_o) / \pi \alpha^2 \sqrt{t}$$

where $k$ is the thermal conductivity ($2 \pm 1$ W/m$^2$), $\alpha$ is the thermal diffusivity of dacite ($4 \times 10^{-7}$ m$^2$/s) and $t$ is the time since emplacement or the flow age for a particular flow pixel. Time $t$ is calculated by measuring the distance of the flow pixel from the actual center of the vent, not the edge of the crater, and dividing that distance with the flow velocity. $q_{base}$ calculated for the full range of thermal conductivity values are plotted in Figure 5.3c. $q_{base}$ decreased from 1600 – 2400 W/m$^2$ to 1100 – 1600 W/m$^2$ in the proximal region, down to 800 – 1200 W/m$^2$ at the end of the medial region and continued to decrease in the distal region down to 700 – 1000 W/m$^2$. 
Heat loss due to rain falling on the surface of the flow can be calculated using the method outlined by Keszthelyi (1995). Rain falling on and around the flow is assumed to percolate down to the 100 °C isotherm and boils away. The heat flux due to rain ($q_{\text{rain}}$ in W/m$^2$) is then calculated by (Keszthely 1998),

\begin{equation}
q_{\text{rain}} = \partial R / \partial t \cdot \rho_{H2O} \cdot L_{H2O}
\end{equation}

where $\partial R / \partial t$ is the rainfall rate, $\rho_{H2O}$ is the water density and $L_{H2O}$ is the latent heat of vaporization ($2.5 \times 10^6$ J/kg) plus the energy it takes to heat water to 100 °C ($3 \times 10^5$ J/kg). The annual average daily rainfall rate from Quetzaltenango was obtained from http://weather.msn.com and was used to calculate variations in $q_{\text{rain}}$ (Figure 5.4). The average $q_{\text{rain}}$ is 60 W/m$^2$ with a maximum value of 180 W/m$^2$ during periods of heaviest rain.

5.5. Thermally-derived Core Cooling, Crustal Thickness and Discharge Rate

Following Oppenheimer (1991) I assume that $q_{\text{surf}}$ represents the heat flux conducted from the hot core, across a surface crust of thickness $\delta$, to be lost from the surface by radiation and convection (i.e. $q_{\text{cond}} = q_{\text{total}} = q_{\text{rad}} + q_{\text{conv}}$). The thickness of the crust can now be calculated at any point along the flow using Fourier's law:

\begin{equation}
q_{\text{cond}} = k \frac{(T_{\text{core}} - T_{\text{surf}})}{\delta}
\end{equation}
Figure 5.4. Annual variation of heat flux from the flow surface due to rainfall.
in which $k$ is the thermal conductivity ($2 \pm 1$ W/m$^2$) and $T_{core}$ is the temperature of the lava core. As heat is lost from the flow surface, core temperature will decline down-flow, moving away from the at-vent eruption temperature ($T_{erupt}$) to lower values at the flow front. Harris et al. (2002) estimated the core cooling rates using the flow heat budget method of Keszthelyi and Self (1998), in which core cooling per unit distance ($\delta T/\delta x$) can be calculated using the following equation,

$$\frac{\delta T}{\delta x} = \frac{q_{tot}}{dV \rho C_p} \quad (5.10)$$

Here $q_{tot}$ is the total heat flux from the core (in Watts), $d$ is flow depth (20 m), $V$ is flow velocity ($1.05 \times 10^{-4}$ m/s), $\rho$ is lava density (2600 kg/m) and $C_p$ is lava specific heat capacity ($1150 \pm 250$ J/kg/K). For this calculation I assume that $q_{tot}$ is the sum of the surface heat flux (radiative + convective) and the heat flux to the base of the flow ($q_{base}$). Using equation 5.10 I calculated core cooling of 135 – 600 °C/km across the proximal section, decreasing to 200 – 140 °C/km across the medial-distal region (Figure 5.5a). Given an eruption temperature of 950 °C and these calculated core cooling rates, core temperature at the distal end of this flow is calculated to be 822 °C (Figure 5.5a), meaning that it has cooled by ~122 °C over 700 m. Using these core temperature profiles I can now use equation 5.9 to calculate down-flow variation in crust thickness. This is modeled as increasing from 0.5 to 1.7 m in the proximal region, to 1.7 – 4 m across the medial region, decreasing to 2 m in the distal region (Figure 5.5b).
Figure 5.5. a. Down-flow core cooling rate and core temperature. b. Estimated flow crust thickness using simple conduction model.
Following Harris et al. (1998, 2000, 2003, 2005), total heat loss ($Q_{tot}$, in watts) estimated from the flow can be used as an alternative method of determining time-averaged discharge or extrusion rate ($E_T$), using the following equation,

\[
E_T = \frac{Q_{tot}}{\rho [Cp\Delta T + fL]}
\]

in which $\Delta T$, $f$ and $L$ are total flow cooling (which for this flow is 122 °C), mass fraction of crystallization (0.45) and latent heat of crystallization ($3.5 \times 10^5$ J/kg). $Q_{tot}$ is calculated for each lava flow pixel in both the FLIR and ASTER images by multiplying the heat flux densities by pixel area (Figure 5.6). These values are then summed for all pixels give $Q_{tot}$ for the whole flow.

The oblique viewing angle of the FLIR results in occlusion of portions of the flow, resulting in an under-estimate of the total flow area and hence heat flux. The near-nadir view of ASTER, however, yields a complete view of the lava surface. Harris et al. (1999), however, observed that pixel-counting in satellite thermal data with coarse spatial resolutions (where by the lava area is assumed to be equal to the number of thermally anomalous pixels multiplied by pixel area) may lead to overestimate of lava area due to mixed pixels containing only a small portion of active lava. The number of pixels occupied by the lava flow in the FLIR image is 931 pixels, which is equivalent to a total surface area of 0.10 km$^2$. The total area calculated using the ASTER data on the other hand is ~0.26 km$^2$. Thus I assume that the true discharge rate lies between the minimum estimate from
the FLIR-data and the maximum provided by ASTER. From the FLIR thermal data, I calculate $Q_{\text{tot}}$ of $\sim256$ MW, yielding to an extrusion rate of $\sim0.3 \text{ m}^3/\text{s}$. For the ASTER data, $Q_{\text{tot}}$ is estimated to be $\sim913$ MW, corresponding to a calculated extrusion rate of $1.2 \text{ m}^3/\text{s}$. In addition to the previous estimate of extrusion rate given in section 5.4, this gives a full estimate range of extrusion rates between 0.15 to 1.2 $\text{m}^3/\text{s}$.

5.6. Conclusions

Down-flow FLIR and ASTER-derived temperature profiles for the 2005 dacitic block lava flow agree with those obtained for the 1999 – 2002 flow by Harris et al. (2002, 2004) showing extremely low surface temperatures. Together, these results show that this low surface temperature emplacement style is characteristic of Santiaguito. This seems consistent with prolonged surface cooling and crust thickening during slow lava transit from the vent to the flow toe, with a thick, stable crust forming with time and distance from the vent. In general, the form of the down-flow surface temperature profile is effectively that of a cooling curve, with surfaces having sufficient time to cool to near-ambient temperatures by the medial section. Localized increases in surface temperature are likely due to small rock falls which remove a portion of the surface layer exposing warmer material beneath, thereby decreasing the thickness of the thermal boundary layer overlying the core and slightly increasing surface temperature and heat fluxes over localized regions. This effect thus introduces
Figure 5.6. Total heat loss (radiative + convective + base conduction, in Watts) calculated for the SW-flank block lava flow.
noise/scatter into the trend.

The eighth and current eruptive cycle at Santiaguito began in 1996 with a Landsat-derived extrusion rate increasing from 0.1 – 0.4 m³/s during 1990 – 1996 to ~0.5 m³/s (Harris et al. 2003). High extrusion rates of 0.5 – 1.6 m³/s persisted throughout the emplacement of the 1999 – 2002 flow (Harris et al. 2004). Beginning in October 2004, a new flow began to extend down the southwest flank of the Caliente, reaching ~1.4 km in length by January 2005. The range of extrusion rate estimates for this time (0.15 – 1.2 m³/s) suggest that no significant decline in extrusion rate has occurred in almost a decade since the eighth cycle started, which is well beyond the typical range of the initial high-extrusion rate phase of Santiaguito’s cycles. These results may support the scenario presented by Harris et al. (2003) in which persistent extrusion at intermediate rates may be replacing the alternating high-low extrusion rate cycles implying of increasingly steady lava supply through the conduit.
CHAPTER 6
CONCLUSIONS

6.1. Dissertation Summary

The studies presented in this dissertation show how a variety of ground-based thermal remote-sensing instruments can be used to monitor dynamic processes at an active silicic lava dome, such as that active Santiaguito's Caliente dome unit. Ground-based approaches provide the necessary spatial and temporal resolution to study the spatially and temporally dynamics of at-vent structures and explosive emissions at Santiaguito from a safe distance. In Chapter 2 I integrate thermal datasets with infrasound and seismic data at high sampling rates (122 Hz) to detect event onset, parameterize Caliente's persistent intermittent explosive activity and infer likely subsurface conduit scenario's. In chapters 3 and 4, I describe the thermal structure of Caliente's summit vent and its short-term temporal evolution using thermal data collected using a radiometer, hyperspectral spectroradiometer and an thermal video (imaging) camera (FLIR). In Chapter 5 I again use the FLIR, along with analysis of an ASTER image to describe the thermal characteristics of the Caliente's active silicic (block) lava flows. Finally in Appendix A, I use the FLIR to image vertical ash plumes and derive some plume ascent parameters, such as vertical velocity profiles.
6.1.1. Chapter 2 Summary

Radiometers utilized in Chapter 2 record plume emission as transient thermal waveforms. Duration of plume emission is estimated from the length of the transients (0.5 – 15 minutes). The amplitude of these transients can be used as a proxy for the relative intensity of the emission. Exit velocities for the plumes measured (using the delay in the arrival of the plume at two vertically stacked thermal sensors) ranges from 16 to 76 m/s. The buoyant rise rates of the plumes range from 9 – 26 m/s. Rise rates at 100 and 600 m above the vent estimated from the onset phase of the thermal waveforms yield velocities of 15 – 39 m/s and 9 – 26 m/s, respectively. Crude vertical velocity profiles from these data show rapid deceleration within the first 100 m, suggesting the transition between momentum-dominated (or gas-thrust phase) to buoyant phase.

For explosion event detection, using peaks in the thermal dataset alone will yield event numbers that are much higher than the actual explosion events (73 thermal events were counted vs 35 actual explosions within the analyzed timeframe). This is due to other non-explosion events that can trigger a signal in the thermal sensor such as a degassing events or rock-fall generated plumes. I find that it is better to use corresponding peaks in thermal, seismic and acoustic signals to properly detect an explosion event. The explosion frequency calculated for this period is approximately 2.3 explosions per hour.

Source depths of the explosions are estimated using the delay in the arrivals of the thermal and infrasound signals. I have constrained the explosion source to a region 100 to 600 m below the vent surface. I find no short-term
systematic temporal variations in source depth of the explosions and no correlation between depth and relative eruptive intensity based on measured parameters such as thermal amplitude and proxies for released thermal and elastic energy.

Using the results from this study I compared three models that have been previously evoked as possible mechanisms for Santiaguito’s explosions, or explosions in similar volcanic settings. The three models discussed are phreatomagmatism (Rose 1987, Sanchez-Bennet 1992), pressure buildup under a degassed viscous plug (Johnson et al. 1998) and shear-induced magma fragmentation (Gonnermann and Manga 2003). The phreato-magmatic model suggests that explosions are the result of interaction of groundwater with magma in the conduit and therefore the source depth should represent the depth of the water table within the edifice. From my study I find that there are extreme fluctuations in source depth where explosion depth can vary up to 300 m over a very short time scales. It is nearly impossible for water table level to fluctuate at the rate observed in this study. The pressure-buildup model suggests that explosions are the result of build-up of gas beneath an obstructing plug due to constant degassing of magma below. Given sufficient overpressure, the plug ‘uncorks’ and the sudden depressurization causes explosive release of the pressurized gas volume. This model requires that the source of the explosion be fairly stationery. Also, relative intensity of the explosions should be proportional to the amount of time it takes for pressure to buildup or the repose intervals; which is not the case observed. In the third model, explosions are the result of
fragmentation along the conduit walls due to shearing of a rising dacite plug within the conduit. This model is consistent with the observed ring-like emission of the plumes. It also better explains the variability in source depth, where the source depth variation can result from varying boundary conditions and magma fluxes into the conduit. Thus I conclude that the shear-induced fragmentation model best fits my observations. In addition I am able to define the thickness of the rising dacite plug as 600 m. This is an important parameter for input into modeling of conduit dynamics at extruding silicic systems.

6.1.2. Chapter 3 summary

Thermal structure and morphology of the summit crater surface, plus the effects of explosions on this structure, are documented using radiometer and spectroradiometer in Chapter 3 and with a thermal video camera in Chapter 4. Temperature time series from a radiometer aimed at a section of the summit crater show thermal cycles in which temperature peaks coincide with explosive events. These peaks are followed by cooling curves. The large number of wavebands available using the spectroradiometer allows for discrimination of multiple thermal components, as opposed to a single integrated temperature measurement available from the radiometer. I find that the thermal structure changes from a two component surface comprised of cool material (120 – 250 °C) broken by fractures radiating at magmatic temperature (950 °C), to a homogeneous surface radiating at intermediate temperature (350 – 500 °C). I attributed this evolution to the removal of the overlying chilled crust during the
more energetic explosions which expose hotter underlying surfaces. This can be achieved by rotating blocks as well as fracturing the chilled crust and/or removal/partial removal over layers of cool, surface material. Periodic thermal renewals of the vent surface produce sufficient increase in temperature to allow the use of shortwave infrared bands of satellite-based sensors over the vent to detect emission from the highest temperature sources.

6.1.3. Chapter 4 summary

The measurements of Chapter 3 were all made over a relatively small portion of the summit, which may (or may not) be representative of the whole summit surface. In order to obtain summit-wide temperature distribution I used a thermal video (imaging) camera (FLIR) which takes 76,800 simultaneous temperature measurements distributed across the vent surface in the form of a 320 x 240 pixel image. The derived thermal structure shows three concentric regions. The first is a high temperature outer annulus (150 – 400 °C) filled with fine materials. This runs around the perimeter of the summit crater and is the source for most ash plumes. Thus surrounds an inner hot core of extruded lava (100 – 200 °C) which has a rim of cool older lava (40 – 80 °C) blocks. A portion of the hot outer annulus was targeted in Chapter 3. Thermal renewal observed in chapter 3 within this region is also observed in the FLIR dataset following an explosion. Based on the FLIR observations I reinterpret the surface thermal renewal as a result of the removal of fill material by the explosion, exposing hotter lava below. The observed surface thermal structure is consistent with a
conduit setting comprising a central plug of rising lava (that feeds block lava flows down the southwest flank) and a highly-fractured conduit boundary zone repeatedly reamed by ash emissions.

6.1.4. Chapter 5 Summary

Ground-based thermal imaging integrated with a satellite-based (ASTER) dataset provides a more complete picture of the thermal characteristics of Santiaguito’s active block lava flows. While the thermal (FLIR) image gives accurate temperature measurements at high spatial resolutions, the overhead view obtained from the satellite allows for better estimates of the total area of the lava flow. Surface temperatures were used to estimate the full heat budget (radiation + convection + conduction + rain) for the flow, with rainfall causing significant heat losses in the cool, thick, crusted carapace. Extracted down-flow temperature and heat-loss profiles suggest extended cooling of the surface and thickening of the overlying crust, insulating the core and minimizing cooling rates. I calculated an extrusion rate of the block flow using the total heat loss in both the FLIR and ASTER and obtained 0.3 m$^3$/s and 1.2 m$^3$/s from the two data sets respectively. This extrusion range confirms that the extrusion rate has been more or less steady since 1996, a period of over 9 years which is much longer than the typical length of time of the high extrusion phase of extrusion cycle at Santiaguito. This further support the previous notion that alternating high-low extrusion cycle that has persisted since 1922 maybe changing to sustained
extrusion at an intermediate cycle, which suggest a steady magma influx into the conduit.

In summary I have shown, through these studies, that ground-based thermal data integrated with complementary seismic, acoustic and satellite-based thermal data, can be used to track and parameterize extrusive and explosive activity at Santiaguito: from the explosion source at 0.5 km below the surface, up through the upper-most section of conduit to the vent and through the surface emplacement as slow-moving block-lava flows.

6.2. Recommendations for future thermal remote sensing of Santiaguito and other silicic systems

Based on the studies presented in this dissertation, I can make some recommendations on how ground-based thermal data can be best deployed and used for future tracking of eruption dynamics at Santiaguito and other, similar, silicic systems. I also discuss the implications of the results obtained using ground-based instruments to test the capabilities of lower (spatial and temporal) resolution satellite-based data for detecting explosions and thermal anomalies at Santiaguito.

6.2.1. Ground-based sensors

Radiometers that operate in the infrared region (8 – 14 μm) are an extremely robust and economic option for detection and monitoring of thermal signals emitted during explosive and effusive activity at silicic systems. From the
studies included in this dissertation I have shown that they have the capability to observe shallow conduit processes, vent processes, and emissions produced by explosions.

Hand-held IR thermometers such as the Raytek Raynger used in chapter 3 provide an option for rapid observations at high temporal resolutions (~ 1 s to 100 Hz). These instruments would be useful in campaign-style monitoring of the summit vent temperature to establish surface cooling rates and to detect any changes in surface temperature due to an explosion. They also have the capability of detecting the emission onset and recording ascent details. The main drawbacks of this type of instrument is that you can only observe a certain part of the vent at any one time and that the temperature reported is converted from total radiance integrated over the field of view (FOV), which can be quite an underestimation of maximum temperatures given that a large FOV may include large number of thermal components radiating at different temperatures.

The deployment of semi-permanent or even permanent stations comprised of the DUCKS thermal sensors (Harris et al. 2005c), microphones and seismometers (chapter 3) is the most useful tool for tracking explosion and ash plume dynamics. The spatial resolution of the DUCKS thermal sensors varies depending on the optical lenses used, where it can be as narrow as 1° or as wide as 60° FOV (Harris et al. 2005c). Connected to a dedicated data-logger, this system can achieve sampling rates as high as 122 Hz. Instead of the El Brujo site used in Chapter 2, I would probably deploy the system at the jungle observation site used in Chapter 5 for ease of maintenance. This system can
then be used to detect the onset of an explosion event with high precision, distinguish between actual explosions, degassing events, rock falls and block flow collapses to obtain correct event counts for explosion frequency monitoring. It can also be used to measure intensity of the explosions (from the amplitude of the waveforms generated). One thermal sensor should be aimed as close to summit vent as possible to obtain exit time and velocity for each the emission (with velocity derived using the time-to-first-inflection method of Chapter 2). A second sensor should be aimed immediately above the first one so that the bottom of the FOV of sensor 2 touches the top of the sensor 1 FOV. This will allow a check on exit velocity using the delay between the thermal onset on the two, stacked, sensors. I would add a third sensor aimed ~2 km above the vent to monitor relative heights attained by the plumes. A fourth sensor, with a wider (15 – 60°) FOV aimed at the flanks of the volcano (just below the summit area) can then be used to detect whether an event generates simultaneous pyroclastic flows or not.

What I hope to be able to accomplish with long-term observations using such an integrated geophysical system would be to answer questions such as whether there are seasonal variations of intensity and frequency of explosions, especially during the rainy season. This would help establish any phreatic influence on the explosions (as discussed in Chapter 2). Another question that can be addressed is how the range of explosion depths changes with time, which offers insights into how the rheology of the upper portion of the magma column changes with time and influences extrusive and explosive activity.
The FLIR thermal video camera provides true imaging capability for thermal monitoring, which is useful for obtaining spatial variations in temperature. Continued improvements on the design of the camera enhance the portability and durability of the FLIR, allowing it to endure the rigors of volcano-monitoring. Long term use of the FLIR for monitoring activity at Santiaguito should include weekly or monthly observations of the Caliente summit vent structure from the Santa María summit. Such a time scale provides the best compromise between temporal resolution and data storage issues, given the high data output rates of the FLIR at such high spatial resolution. These campaigns would allow generation of thermal maps of the summit at 6 m resolution. Analysis of these maps will reveal any changes to the overall thermal structure of the summit, which may correspond to increase or decrease in lava supply to the conduit. It would also allow monitoring of the diameter of outer annulus/conduit margin, as well as the diameter of the central plug, allowing for tracking of changes in the dimensions of the conduit and ascending plug. FLIR imaging also allows tracking of the heat-loss budget of the summit, which changes with time. This would also lend insight into increasing or decreasing supply of magma into the system.

Thermal imaging of Santiaguito from the jungle site (3.8 km south of the Caliente vent) with the FLIR provides simultaneous observations of any block flows present on the south-facing flank of the dome, ash emissions, and any pyroclastic flows generated. The issue of power and storage limitations that hamper the FLIR thermal imaging can be remedied by moving the observation site to the Santa María Observatory, approximately 5.5 km away. This site yields
a respectable pixel resolution of ~ 7 m, and also allows provision of power and installation of a PC-based image-logging system. A daily thermal map of the southern flank of the dome can be generated to track the advance of block lava flows or any new flow units emerging from the summit. This can also be used to monitor changes to the thermal structure of the flow and to detect any recent collapses of the front. Such maps should be made from data collected in the cloud-free early hours of the morning, when solar heating effects are minimized. Using such data I can calculate flow-wide heat loss and extrusion rate. Routine imaging of ash emissions would also extend the plume dynamics dataset, where the intermediate gain setting should be used to include a wide range of temperature and allowing the capture of initial plume temperatures.

Moving the imaging site to the Santiaguito Volcano Observatory (5.5 km south of Caliente vent) will increase the total dimensions of the instantaneous field of view. The larger field of view will allow the tracking of the final heights attained by the plumes to explore the inferred relationship between heat flux into the plume and the potential rise height of the plume (Wilson and Head 2007). Heat loss from the plumes can be estimated by extracting a temperature profile immediately above the summit, converting temperature to heat flux and then integrating heat fluxes for the duration of the emission. Initial plume temperatures exceed 160 °C and thus the FLIR gain setting needs to be set to the appropriate temperature range (as described above), which varies depending on the model of FLIR or other thermal video camera being used. This temperature range should also be taken into consideration when choosing radiometers to use as a
monitoring tool for the plumes (e.g. Chapter 2), where temperature for block lava flow surfaces are as low as ambient, but a potential maximum of at least 500 °C needs to be considered.

6.2.2. Implications to satellite-based approach

Thermal remote sensing on the ground provides superior temporal and spatial resolution over satellite-based sensors, allowing the tracking of dynamic, short-term, temperature changes associated with explosions, ascending plumes and thermal renewal of the summit surface. Ground-based approaches, however, are limited in that measurements are often made at an oblique angle, resulting in slightly distorted measurements in temperature and area of the feature. It also means that only partial coverage of a flow or plume surface may be achieved, introducing errors (underestimates) in the derivation of lava flow area, heat loss and extrusion rate, for example. Given the overhead or near overhead field of view obtained from satellite-based sensors, complementary use of satellite-based sensors can help address this issue. Satellite-based thermal monitoring, however, is most effective during periods of low cloud cover and at nighttime when acquisitions lack the contamination from solar-heating of the surrounding surface. It is also limited to satellite overpass times. Satellite-based thermal remote sensing can be divided into high temporal resolution or high spatial resolution systems.
High Temporal Resolution

The Geostationary Operational Environmental Satellite (GOES) Imager and the Moderate Resolution Imaging Spectroradiometer (MODIS) offer the best temporal resolutions for explosion and effusive event tracking at the cost of lower spatial resolution. These instruments also have the added advantage of being available relatively cost-free. GOES and MODIS thermal time series can be obtained freely on-line at http://hotspot.higp.hawaii.edu, and raw MODIS data can be ordered directly from http://edcns17.cr.usgs.gov/EarthExplorer/. The Advanced Very High Resolution Radiometer (AVHRR) imager falls within the same category but requires access to a receiving station.

The GOES Imager acquires images every 15 minutes from a geostationary orbit with a 4 km pixel size. GOES has a total of 5 channels, 3 of which can be used for temperature observations: band 2 (3.78 – 4.03 µm), band 4 (10.2 – 11.2 µm) and band 5 (11.5 – 12.5 µm). Given the calculated explosion frequency of 2.3 events per hour (Chapter 2) and the imager’s sampling rate of 4 images per hour, GOES provides the best opportunity for explosion detection at Santiaguito. But at such a low spatial resolution, can the thermal changes associated with the explosions be detected by GOES? In Chapter 4 I established that thermal renewal occurred at the outer annulus, which in 2005 had a total area of 5.4 x 10⁴ m², with an average pre-explosion temperature of ~150 °C and a maximum temperature change of 250 °C following an explosion. If I assume an average background temperature of 40 °C, the change in GOES band 2 pixel-integrated temperature will be 5 °C. Thus explosion-induced changes involving
outer annulus temperatures greater than 215 °C should be detectable in theory. Applying one of the surface cooling curves fitted to the data in Chapter 3, modified to have an initial peak temperature of 400 °C \( (T = -30\ln(t) + 370 \, ^\circ C) \), we then have a time window of approximately 13 minutes after explosion onset when there will still be sufficient, residual thermal signal to allow for change detection.

The summit surface, however, is obscured by the emitted plume within the first few minutes (as observed in Chapter 3). The plume front presents an initial positive thermal change (increase in observed temperature) but cools rapidly to produce a negative thermal change, until the plume disperses or is blown away from a position above the summit vent. Given the observed emission duration of 3 minutes (Chapter 2), the thermal signal from the vent will be obscured for this period.

MODIS is an imaging spectroradiometer onboard the Terra and Aqua Earth Observing System platforms. MODIS provides up to 4 images per day. This means that there is a significant decrease in the probability of detecting an explosion event compared to GOES. However, it has higher spatial resolution than GOES at 1 km, thus providing a better sensitivity to surface temperature changes. For example, using band 21 of MODIS (3.929 – 3.9891 \( \mu m \)) and the same scenario presented above for GOES produces a temperature increase of greater than 12 °C. Based on the observed maximum cooling rate, MODIS band 2 can be used to detect explosions up to 45 minutes after the onset of the explosion.
Future work should involve comparison of GOES and MODIS datasets acquired during the periods of ground-based data acquisition to test the actual capability of GOES and MODIS to detect explosions at Santiaguito and other systems.

How effective can high temporal resolution systems such as GOES and MODIS in detecting thermal anomalies at Santiaguito? Based on the thermal structures observed at Santiaguito, I am now able to test the ability of GOES and MODIS to detect thermal anomalies on this volcano. Given the spatial resolution of the two instruments (1 km for MODIS, 4 km for GOES) most of the active thermal features on the lava dome will be contained within a single pixel. The pixel integrated temperature \( T_{int} \) of this pixel will thus be a function of wavelength \( \lambda \) and the relative contribution of five thermal components. The pixel integrated radiance \( L(\lambda, T_{int}) \) is the sum of radiant flux from each component multiplied by the fractional area of the component within the pixel:

\[
L(\lambda, T_{int}) = L(T_{vf}, \lambda) \cdot p_{vf} + L(T_{vh}, \lambda) \cdot p_{vh} + L(T_{vc}) \cdot p_{vc} + L(T_{flow}) \cdot p_{flow} + L(T_{back}) \cdot (p_{back})
\]

\( L \) is the planck function for temperature \( T_{int} \) and wavelength \( \lambda \), \( T_{vf} \) and \( p_{vf} \) are the temperature and pixel portion of the outer annulus vent fill (Chapter 4), \( T_{vh} \) and \( p_{vh} \) are the temperature and pixel portion for the high temperature cracks at the vent (Chapter 3), \( T_{vc} \) and \( p_{vc} \) are the temperature and pixel portion of the cold blocky crust at the vent (Chapter 4), \( T_{flow} \) and \( p_{flow} \) describe the block lava flow
contribution, and $T_{\text{back}}$ and $P_{\text{back}}$ are the temperature and pixel portion of the background (ambient, non-active) dome surface where,

\begin{equation}
P_{\text{back}} = (1 - P_{\text{uf}} - P_{\text{sh}} - P_{\text{vc}} - P_{\text{flow}})
\end{equation}

Using data from this dissertation, I applied these equations and converted the integrated radiance values to pixel integrated temperature using the inverse Planck's function for the mid-infrared and thermal infrared bands of GOES (bands 2 and 4 respectively) and MODIS (bands 22 and 32 respectively). The integrated temperature is then compared to a background pixel at 30 °C to see whether or not the pixel will be thermally anomalous. Theoretically Santiaguito should appear strongly anomalous in both the mid-infrared and thermal-infrared bands of MODIS, in which a pixel-integrated temperature 40 to 90 °C is expected, causing an anomaly of 10 – 60 °C. For the coarse spatial resolution GOES images, the Santiaguito pixel should be anomalous only in the mid-IR band, where the pixel-integrated temperature will be 18 °C greater than background levels.

MODIS and GOES mid-IR data can thus be used to obtain daily-to-hourly heat flux time series to establish a background trends. Short-term deviations (spikes) can be interpreted as due to occurrence of an explosion event. Longer-term trends may reflect changes in magma supply and hence extrusion rate. (e.g. Harris et al. 2003). This will allow further tracking and parameterization of Santiaguito’s cyclic extrusion.
**High Spatial Resolution**

Satellite-based thermal datasets with high spatial resolution (90m pixel size or less) are useful for periodic detailed mapping of changes in the surface thermal structure and mapping of lava flow extent. They are, however, less accessible and more costly when compared to their low-spatial resolution counterparts. Landsat TM and ETM+ has been used at Santiaguito for calculation of extrusion rates from 1990 – 2000 (Harris et al. 2003). The Advanced Spaceborne Thermal Emission and Reflection Radiometer (ASTER) provide the most cost-effective of the high-resolution instruments. ASTER has an acquisition interval of 16 days with spatial resolution of 15, 30 and 90 m for the three VNIR bands (0.52 – 0.86 µm), 6 SWIR bands (1.6 – 2.4 µm) and 5 TIR bands (8.125 – 11.65 µm) respectively. The relatively high resolution of ASTER data can be used to map the extent of block lava flows over time to derive their advance rates. As shown in Chapter 5, heat flux derived from ASTER data can be used to derive extrusion rate. This will provide long-term (bi-monthly) extrusion rates if cloud-free data are available. Often, in the summer (wet season) all overpasses will be cloud-covered. ASTER data also provides the capability of constructing Digital Elevation Models (DEM) at 15 m. This provides an alternative means for calculations of extrusion rate by using DEM subtraction to track changes in the volume of the block lava flows with time, which then can be used to test the validity of the thermal-based approach.

Acquisition of radiant spectra in the 0.4 to 2.5 micron range allows for discrimination of multiple radiating components within the field of view for better
characterization of surface thermal structure (chapter 3). A field-based approach however is limited to imaging a portion of the summit at a time, requiring re-targeting of the FOV to collect data from other portions of the summit. This issue can be addressed using a satellite-based imaging spectroradiometer such as the Hyperion. Hyperion, is a high-resolution (30 m) spectroradiometer, onboard the EO-1 platform, with 220 spectral bands from 0.4 to 2.5 microns. This data will allow for detection of surfaces radiating at or near magmatic temperature as well as measurement of the fractional areas covered by these high temperature components.

6.3. Concluding Remarks

Ground-based thermal remote sensing instruments provide a way to observe and monitor dynamic processes that occur at an active silicic lava dome system from a safe distance. There is a wide variety of thermal remote sensing instruments that can potentially be used, each having their own advantages and disadvantages. I find that the full combination of different instruments is the most effective way to study systems such as Santiaguito where both explosive and effusive type of activity occurs at the same time. Continued work on developing permanent ground-based thermal monitoring hubs and improving their cost-effectiveness is critical given that many dangerous volcanoes are located in developing countries with limited manpower and financial resources.
A.0. Abstract

Eruptions of vertical ash plumes at Santiaguito, Guatemala, were observed using a FLIR thermal video camera at 30 frames per second to capture plume ascent dynamics, morphology, temperature and cooling rate. These plumes assume the form of a starting plume, in which the leading plume front is fed by a steady feeder plume. Twenty-four out of the 25 plumes analyzed showed a gas-thrust to buoyant-ascent transition at 20 – 50 m above the vent. Rise rates in the gas-thrust and buoyant-ascent regions were 10 – 100 m/s and 4 – 18 m/s, respectively. Maximum cooling rates (~ 50 °C/s) and lateral spreading rates (> 800 m/s) occurred within the gas-thrust region. Plumes that reached greater heights (> 1 km) had higher relative heat contents and wider feeder plumes, leading to greater convective-rise velocities compared to the plumes that attained lesser heights.

A.1. Introduction

Field observations of volcanic plumes play an important part in understanding the dynamics of explosive eruptions by providing the opportunity to quantify their characteristics and to test theoretical plume dynamic models. Plume imaging allows for extraction of parameters such as plume height, ascent
rate and expansion rate over narrow time intervals. These data can be used in plume dispersal models for hazard mitigation purposes (e.g. Searcy et al., 1998). Previous volcanic plume measurements and modeling involved the use of photographic images (e.g. Wilson and Self, 1980; Ripepe et al., 1993), video camera (e.g. Clarke et al., 2002; Formenti et al., 2003), and satellite images (e.g. Holasek and Self, 1995; Woods et al., 1995). More recently the use of thermal cameras has been demonstrated for understanding the dynamics of Strombolian emissions (Patrick et al. 2007). The use of calibrated thermal data has the added advantage of allowing measurement of plume temperatures during ascent, as well as the ability to capture plumes at nighttime.

Repeated, low to moderate intensity vertical ash eruptions have been a characteristic of activity at the Caliente vent (Santiaguito, Guatemala) since 1975 (Rose, 1987). Extremely regular ash emissions occur at a frequency of ~ 0.5 to 2 events per hour, with observed ascent heights of 1 to 4 km above the vent (Bluth and Rose, 2004). On January 25 2005, 25 vertical ash plumes at Santiaguito were observed using a Forward-Looking Infrared (FLIR) thermal video camera (Figure 6.1), and derived plume front ascent rates, lateral spreading rates and temperature during ascent.

A.2. FLIR camera and dataset

I used a FLIR Systems ThermaCam™ model P40. This camera operates in the 7 – 13 micron range, producing a 320 x 240 pixel image at 15 – 30 frames per second. The thermal video camera was mounted on a tripod approximately
3.8 km south of the Caliente vent. Both emissivity (0.9) and atmospheric transmissivity corrections (for the appropriate path length, atmospheric temperature and humidity) were applied to the dataset. The spatial resolution of the FLIR images over this line of sight distance is 5 m per pixel. For each plume, plume front width and temperature were measured at 10 m intervals over the first 200 m of rise and every 20 m beyond 100 m. The full image field of view was 1600 x 1200 m, allowing me to track plume motion and temperature up to ~ 900 m above the vent.

There are several different plume forms observed at Santiaguito (Figure A1). They include jets, starting plume, rooted and discrete thermals. Jets are defined as momentum-driven plumes with high velocities (Morton, 1959). Thermals, on the other hand are buoyancy-driven, detached fluid vortex (Turner, 1972). Starting plume morphology is where a buoyant plume is capped by a thermal, continuously fed by a feeder plume below. Rooted thermals are starting plumes where the thermal portion of the plume display vortex rings (Patrick 2006).

The plumes are divided into two types on the basis of relative heights they reach: type A plumes (n = 10) are relatively high plumes that ascend beyond the field of view (therefore having final ascent heights > 1km); type B plumes (n = 15) are relatively low plumes which remained within the field of view during ascent and dispersal (Figure A.2).
Figure A.1. Plume morphologies observed at Santiaguito (after Turner 1962 and Patrick 2006).
Figure A.2. FLIR image sequence of two types of plumes defined for Santiaguito in this study. Sequence A shows the transition of a type A (height > 1km) plume from initial individual jets coalescing into a single rising mass before forming a starting plume and then a rooted thermal. For this particular plume a ring vortex can be observed its base. Sequence B shows an example of a type B (height < 1 km) plume showing the transition from starting plume to a thermal.
A.3. Results

A.3.1. Plume Rise rate

Plume front parameters that can be measured directly from the FLIR images include vertical position with respect to time following emission onset (Figure A.3a), plume front radius (Figure 3.2b) and plume front temperature (Figure A.3c). From these parameters I can derive the ascent rate and the spreading rate in m/s (Figures A.3d and A.3e), as well as plume cooling rate (Figure A.3f). Figure 6.3 plots each parameter against plume types (A and B).

The ascent rate profile (Figure A.3d) shows two distinct velocity regimes common to both plume types in which rapid initial ascent (and significant deceleration) occur within the first 25 to 50 m. This is followed by 800 m of steady (non-decelerating) ascent rates. The transition can be attributed to the change from momentum-dominated ascent (i.e. gas-thrust phase) to buoyant ascent. All but one of these 25 plumes displayed a gas-thrust phase, with maximum (at-vent) velocities ranging from 15 – 55 m/s with an average of 25 m/s. There was no correlation between the at-vent velocity and the height of the transition between the two velocity regimes, this being fairly stable and constant at 35 ±15 m above the vent. The buoyancy-driven phase had ascent rates in the range 5 – 20 m/s with an average of 15 m/s (Figure A.3). Type A plumes had the highest buoyant ascent velocities (6 – 15 m/s) and showed very little deceleration during the observed 800 m of buoyant ascent. Type B plumes had a lower range of buoyant ascent velocities (4 – 10 m/s) and showed more deceleration over the ascent path (Figure A.2d).
Figure A.3. Plume front parameters with ascent height measured from FLIR images for type A (black lines) and type B (gray lines) plumes. (a) Plume front position with time, (b) plume front radius, and (c) plume front temperature. First order derivative of these parameters give plume front (d) ascent rate, (e) lateral spreading rate and (f) cooling rate with height.
A.3.2. Morphology and Spreading Rate

The emission source at Caliente is a series of vents arranged in a ring around the circumference of the summit crater. Plume emission either begins at the vents within the ring, before spreading to comprise summit-wide emission, or can simply begin with a summit-wide emission (Bluth and Rose 2004, Chapter 4). Nine of the observed plumes were initiated by single or multiple jets, 5 – 10 m in radius, from the annular source. The subsequent summit-wide emission phase then overtook these initial jet(s), and formed a starting plume (Figure A.2). This is a morphology whereby the plume front behaves as a thermal current linked to the source vent by a feeder, steady-state plume (Turner, 1962). The largest of the observed plumes had a rooted thermal morphology, defined by Patrick (2006, 2007) as having ring vortex at the base of plume front thermal.

The lateral growth of the plume front is shown in Figures A.3b and A.3e. Plume front radius plotted with height above the vent shows two distinct regions. The first is marked by rapid lateral spreading and characterizes the first 25 – 50 m of ascent (i.e. the gas thrust region), so that the plume reaches a maximum radius of 100 m by 50 m height above the vent. Above this the second region is coincident with the region of buoyancy ascent and is marked by reduced lateral spreading rates, where spreading continues at a relatively steady rate of 4 m/s.

A.3.3. Plume front temperature and cooling rate

Figure A.3c and A.3f shows plume front temperature and cooling rate with height. During image acquisition the camera was set at a low-gain setting giving
a sensor saturation temperature of 160.6 °C. The majority of the plumes have plume front temperatures exceeding saturation temperature in their initial stages. Thus the temperature data set precludes measurements of initial temperatures and estimates on the total heat content of individual plumes. A proxy for the relative heat content of the plumes was estimated by measuring the height at which the plumes could maintain temperatures above saturation (160.6 °C). Twenty-three (both type A and B) plumes maintain temperature above saturation levels by the time they have extended to 100 m above the vent. The two remaining plumes, both of which were type A plumes, sustained plume front temperatures above saturation levels to heights of 150 and 250 m, respectively (Figure A.3).

All plumes experienced exponential decay in temperature as the plume ascends (Figure A.3.a). The majority of the temperature decrease thus occurs within the first 100 m of rise with maximum cooling rates of 50 °C/s. Above the initial 100 m of rise the cooling rate profile shows oscillations with some negative cooling rate values implying an increase in plume front temperature with height. These periodic increases in plume temperature corresponds to multiple pulses during emission where fresh hot materials are supplied into the then much cooler plume front via the established feeder steady plume. Average cooling rates for all 25 plumes ranged from -0.7 to -4.0 °C/s (Figure A.4), with type A plumes typically having a higher range (1.5 – 4.0 °C/s) than type B (0.7 – 2.0 °C/s).
A.3.4. Pf-generating plumes

Two out of the 25 plumes observed in this study generated small pyroclastic flows that traveled down a well-defined chute to the right of the image (Figure A.2). The parameters for these two plumes are highlighted in Figure A.3 plots as thicker black or gray lines. The two plumes were very different in their characteristics. The first plume was a type A and easily the largest of all the plumes studied. This plume had a relatively low maximum gas-thrust velocity of 25 m s$^{-1}$ given its relative size. This plume, however, was able to maintain one of the highest buoyant velocities of all the plumes imaged (10 – 15 m/s) with no deceleration within the IFOV. It seems likely that the low velocity and heavy loading promoted the edges of the plume to collapse and feed the flow.

Although the second plume was a low (type B) plume, it possessed a higher initial exit velocity of 34 m/s. As with most type B plumes, this plume experienced significant deceleration during its buoyant phase, decreasing from 15 m/s to 4 m/s by 800 m above the vent. Again, lower volumes and heat contents promoted cooling, such that buoyancy was not as maintained as effectively as in the high volume case. In this case, velocity and loading cannot be called upon to feed flow, so that some other mechanism must have cause flow such as failure of blocky material at the edge of the vent which then tumbled and fragmented as it fell down the steep dome flank.
Figure A.4. Comparisons of selected parameters for type A plumes which reach heights > 1 km (circles) and type B plumes with heights < 1 km (triangles). Filled circle and triangle denote pf-generating plumes. Tsat height is the height at which the plume front maintained temperature values above upper limit of the camera’s gain setting.
A.4. Discussion and Conclusions

High temporal resolution thermal-imaging of plumes at Santiaguito has allowed for extraction of plume ascent dynamics and temperature parameters. Lower (type B) plumes underwent transitions from starting plumes to thermals. Higher (type B) plumes underwent a more complex transition from jets, which coalesced into starting plumes with sustained feeding then resulting in an emission phase marked by formation of a rooted thermal, whereby a wide feeder column feeds a leading plume front (Figure A.4).

I find that whether a plume ascends to heights above or below 1 km is independent of the initial exit velocity. Instead, higher (type A) plumes are characterized by higher buoyant rise rates, lateral spreading rates, plume front cooling rates and by radii (Figure A.4), as well as by the development of rooted thermals. Higher buoyant rise rates can thus be maintained by steady influx of fresh pyroclasts from the vent through the center of a wider, voluminous, feeder plume into the leading plume front. Larger feeder plumes may facilitate better transport and insulation of hotter material fluxing into the plume front, thereby enhancing the ability of the plume front to maintain higher internal temperatures. This prolongs the phase of buoyant ascent. This is consistent with Wilson and Head (2007) who state that, for convective plumes, the potential rise height is proportional to the heat flux to the fourth root. In this case, heat flux scales with supply of mass from the feeder plume. Unfortunately I was not able to thoroughly test this statement given the lack of temperature values at the emission onset (due to saturation). Unsaturated temperature measurement and heat flux
calculations, along with imaging of more plumes, should fully address the apparent relation between plume types, mass, feeder column width, heat content, buoyancy and height at Santiaguito.
REFERENCES


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