STROMBOLIAN ERUPTION DYNAMICS FROM THERMAL
(FLIR) VIDEO IMAGERY

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Abstract

A handheld thermal video camera, or FLIR (Forward Looking Infrared Radiometer), was used to understand the dynamics of strombolian eruptions. Images (at up to 30 frames per second) were collected for 344 eruptive events occurring between 2001 and 2004 at Stromboli volcano, Italy. Two major eruption styles were observed: Type 1, which are dominated by ballistic particles and Type 2, which are dominated by an optically-thick ash plume, with (Type 2a) or without (Type 2b) additional ballistic particles. Eruption styles (Type 1 vs. 2) were generally maintained on a scale of days to weeks for a given vent. Type 1 eruptions indicated a clear magma-air interface, while Type 2 eruptions were likely caused by loose backfill sitting atop the magma column. Type 2a and 2b behaviors are shown to be a function of bubble overpressure and backfill muffling. These results support a broadening of the current paradigm for strombolian behavior.

Strombolian ash plumes (i.e., Type 2) were analyzed in detail, with fine timescale tracking of rise rates, dimensions and temperatures. Type 2a eruptions exhibited gas thrust velocities (>15 m s⁻¹), while Type 2b eruptions rose at buoyant velocities only (<15 m s⁻¹). Velocity trends produced varying plume morphologies which in turn controlled air entrainment rates, as shown by measurements of spreading rates. Plume temperatures began between 300 and 600 °C, and generally dropped to <150 °C in the first 100 m of rise. Determining plume opacity is imperative in using FLIR-derived plume
temperatures. Plume temperatures varied on a daily basis, likely a function of the amount of accidental ash available to entrain. Using a heat conservation approach, explicit mass estimates were made. Due to a lack of knowledge of gas mass fraction, however, the mass results had uncertainties on the order of 60%.

Integrating these ideas and models, an automated approach is presented to process Strombolian FLIR data and determine eruption style, ejecta mass and collimation. In some cases a spot radiometer, instead of the more expensive FLIR, can be used garner ejecta mass. The automated routine could prove useful in identifying anomalous behavior that might presage more hazardous eruptive phases.
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Chapter 1

Introduction

1.1 General overview

In this dissertation, I investigate the utility of handheld thermal infrared cameras for understanding ‘strombolian’ eruptions, a type of small-scale explosive volcanic activity (Macdonald, 1972). Strombolian eruptions have typically been described as consisting of fountains of brilliant incandescent clots of molten lava (Fig. 1.1), with eruptions occurring at intervals of minutes to tens of minutes. At least, this has been a common perception of strombolian eruptions. A fundamental result of this work, however, based upon analyses of the thermal imagery, is that strombolian eruptions can have quite a range of behaviors and eruptions which do not resemble the aforementioned incandescent arcs can still very much be ‘strombolian’.

The thermal camera used in this study is a FLIR, or Forward Looking Infrared Radiometer. Though resembling a typical digital camcorder, the FLIR provides a powerful tool for the volcanologist: images of calibrated temperature. This allowed me to approach the eruption dynamics with some degree of quantitative rigor. My data originated from Stromboli volcano itself, the archetype volcano for ‘strombolian’ activity, situated in the Aeolian Islands of Italy (Fig. 1.2). I used FLIR images of 344 individual Strombolian eruptions collected between 2001 and 2004 to understand the
Figure 1.1. Eruption photos from Stromboli. The top image shows the ballistic arcs of incandescent ejecta, extending ~75 m above the crater rim. The bottom image shows an eruption with an ash plume, also ~75 m above the crater rim.
Figure 1.2. Location of Stromboli island.
variety of strombolian styles, the possible causes for these variations, the dynamics of strombolian ash plumes, and how the FLIR can be used to measure eruption parameters (such as ejecta mass and eruption styles) in an operational setting.

1.2 Strombolian activity

Strombolian eruptions are a relatively mild form of explosive volcanic activity (Fig. 1.3, also Newhall and Self, 1982). In ordering the explosive power of volcanic eruptions there is a progression in common eruption styles (Macdonald, 1972; Walker, 1973; Newhall and Self, 1982), although the particular naming conventions can be variable. Here I choose the most common and simple. At the most powerful end of the spectrum are the ‘plinian’, or ‘ultra-plinian’ as ultimate endmember, eruptions (Cioni et al., 2000). These form the mushroom-shaped ash columns which reach high into the stratosphere (>25 km), darken the sky, and can deposit ash particles well outside the immediate vicinity of the volcano. Exit velocities at the vent are on the order of a few hundred meters per second (Cioni et al., 2000). The historically notable 79 A.D. eruption of Mount Vesuvius, described in detail by Pliny the Younger, was the origin for the term. Skipping the subtype ‘subplinian’ (a weaker plinian eruption; Walker, 1973) we reach the ‘vulcanian’ eruption style. This involves powerful eruptions (exit velocities of several hundred meters per second) producing small to moderate-size ash plumes (3-15 km height) which cast dense ballistic ejecta to great distances (several kilometers) (Morrissey and Mastin, 2000). Unlike plinian eruptions which persist over hours, vulcanian eruptions are discrete events (durations of seconds to minutes). Another Italian volcano
Figure 1.3. The scale of volcanic eruptions. Classification and height ranges from Newhall and Self (1982).
is for the origin of this term - Vulcano, in the Aeolian Islands north of Sicily. The 1888-1890 eruptive sequence at Vulcano formed the basis for this style (Mercalli, 1907). The term strombolian refers to small explosive events (<1 km height) which occur intermittently, and are most notable for the brilliant incandescent parabolic arcs traced by their molten ejecta (Fig. 1.1). They are distinct from vulcanian eruptions in their lower exit velocities (typically <100 m s\(^{-1}\)) and more regular frequency (tens of minutes) (Vergniolle and Mangan, 2000). The eponym here is, naturally, Stromboli, which is several islands north of Vulcano in the Aeolian islands. Stromboli is the archetypal volcano for strombolian activity, having been active in this style for many centuries (Washington, 1912) and probably since at least 700 A.D. (Rosi et al., 2000). Lastly, at the least powerful end of the spectrum are the ‘hawaiian’ eruptions, which involve sustained fountains of lava which feed lava flows (Vergniolle and Mangan, 2000).

As eruptions at the more powerful end of the spectrum (i.e. plinian) clearly have been the most hazardous for society, one might ask: ‘Why study strombolian eruptions if they are so small’? First, strombolian explosions, though small, are still explosions in every sense of the word – a violent release of pressurized gas and its subsequent interplay with solid and liquid lava particles. Thus, strombolian eruptions offer a partial analogue for more powerful, and dangerous, eruptions. Second, strombolian eruptions are the only explosive eruption style that can be studied from a reasonable distance (hundreds of meters) with relative safety. Plinian eruptions can incinerate people and equipment within several kilometers of the vent, and vulcanian eruptions are too unpredictable to make close approach prudent. Strombolian eruptions, due to their mild nature, allow
observers to approach within a few hundred meters (e.g. Chouet et al., 1974; Blackburn et al., 1976; Ripepe et al., 1993) and rarely damage monitoring equipment (Ripepe et al., 2004). Third, strombolian eruptions tend to erupt in a predictable pattern, further ensuring safety for observers. This feasibility makes strombolian explosive behavior more favorable for direct studies, with the hope that it can yield insights into the dynamics of more energetic eruption styles.

Strombolian activity is restricted to basaltic (or basaltic-andesite) magma types because only these have sufficiently low-viscosities to allow a gas slug to coalesce and rise independently from the fluid surrounding it (Vergniolle and Mangan, 2000). The more viscous magmas (dacite, rhyolite) involved in plinian eruptions do not permit gas bubbles to travel so freely. Strombolian activity is a very common basaltic eruption style, and often strombolian activity appears in the waxing or waning stages of a more powerful eruption (Vergniolle and Mangan, 2000). For instance, the sub-plinian eruptions of Shishaldin volcano, Alaska, in 1999 were preceded by several months of strombolian activity (Dehn et al., 2002). Reports from the Smithsonian Institution’s Global Volcanism Program also reveal the commonality of strombolian activity. Strombolian eruptions have been reported for the following volcanoes since my project began in August 2002: Stromboli, Etna, Nyamiragira, Nyiragongo, Anatahan, Kliuchevskoi, Arenal, Yasur, Ambrym, Tungurahua, Soputan, Fuego, and Manam (Smithsonian Institution, 2005). This list includes volcanoes from the Americas, Asia, Europe, Africa and Oceania.
1.3 Stromboli volcano

Given the numerous volcanoes which exhibit strombolian activity, why was Stromboli volcano chosen as the study location? Stromboli is remarkable because it exhibits a quality rarely found with volcanoes: near-perfect dependability. Since the end of Roman times, Stromboli volcano has been experiencing a small eruption every 10 minutes, more or less (Rosi et al., 2000). It remains one of the best-studied, and best-instrumented, volcanoes on Earth because of its easy access, regular activity and consistent eruption sizes. The latter circumstance allows researchers to safely collect high quality data from just a few hundred meters from the active vents – making it an excellent location to study conduit dynamics and explosive behaviors, and a perfect analogue to understanding other ‘strombolian’ volcanoes which are more difficult to access. Tempering its absolute dependability are occasional phases (every 3-10 years) of lava flow activity which interrupt the normal explosive behavior (Barberi et al., 1993). These typically last a few months, after which normal activity resumes. In addition, ‘normal’ strombolian activity can be punctuated by unusually explosive strombolian eruptions, typically several per year. These are the most dangerous events at Stromboli, and have been responsible for several tourist deaths over the last decade. Particularly violent paroxysms have occurred less frequently (e.g. 1930, 2003). These produce island-wide effects, e.g. ejection of ballistic blocks into populated areas, shock waves, and ashfall (Barberi et al., 1993; Calvari et al. 2005).
Stromboli volcano lies at the northern end of the Aeolian Island arc, a chain of volcanoes just north of Sicily, in the Tyrrehnian Sea (Fig. 1.2). It is a small (12 km$^2$) island volcano rising to 924 m above sea level (asl) (Fig. 1.4), or ~3000 m above the floor of the Tyrhennian Sea (Romagnoli et al., 1993). The western side of the island is dominated by a large sector collapse structure called the Sciara del Fuoco ('scar of fire'). At the head of the Sciara, and at an elevation of 800 m asl, lies the active crater terrace. This consists of several active craters, each containing one or more active vents which host intermittent explosions. Normal Strombolian activity is sufficiently mild that most material stays in the vicinity of the craters, and any excess is directed down the Sciara into the sea. Two population centers exist on the island. Stromboli–San Bartolo village rests on a flat plain on the northeast portion of the island and is home to ~400 permanent residents. The summer months see many more people on the island due to the volcano’s status as a prominent tourist attraction. Another town, Ginostra, is home to ~30 people and sits on the southwest corner of the island. These towns can be reached by ballistic particles during paroxysmal events such as those in 1930 and 2003 (Barberi et al., 1993; Calvari et al., 2005).

Volcanism in the Aeolian Islands is the result of microplate subduction related to convergence of the Eurasian plate and the northward-drifting African plate (Ferrari and Manetti, 1993). Aeolian volcanism is generally Pleistocene in age, with the oldest rocks on Stromboli dating to ~100 ka (Hornig-Kjarsgaard et al., 1993). Strombolicchio, a small island 1.7 km northeast of Stromboli, is the eroded neck of a 200 ka pre-Stromboli
Figure 1.4. ASTER satellite image of Stromboli island. In this false-color image, red indicates vegetated areas, brown is areas of open rock, and white is buildings. Image acquired June 26, 2004.
edifice. Present activity represents the seventh stage of Stromboli's subaerial evolution. These stages are as follows (Hornig-Kjarsgaard et al., 1993):

<table>
<thead>
<tr>
<th>Stage (order &amp; name)</th>
<th>Age</th>
<th>Composition</th>
</tr>
</thead>
<tbody>
<tr>
<td>1. Paleostromboli I</td>
<td>61 – 100 ka</td>
<td>High-K, calc-alkaline</td>
</tr>
<tr>
<td>2. Paleostromboli II</td>
<td>55 – 64 ka</td>
<td>Calc-alkaline</td>
</tr>
<tr>
<td>4. Scari</td>
<td>~35 ka</td>
<td>Transitional shoshonite</td>
</tr>
<tr>
<td>5. Vancori</td>
<td>13 – 25 ka</td>
<td>Shoshonite</td>
</tr>
<tr>
<td>6. Neostromboli</td>
<td>6 – 13 ka</td>
<td>Shoshonite</td>
</tr>
<tr>
<td>7. Recent Stromboli</td>
<td>Present – 6 ka</td>
<td>Shoshonite, high-K, calc-alkaline</td>
</tr>
</tbody>
</table>

The Paleostromboli series consists of lava flows and pyroclastics, pointing to alternating phases of effusive and explosive activity, with resulting air fall and pyroclastic flows and surges. The height of the island reached at least 700 m asl in this phase, and eruptive vents were centered south and southeast, and within 2 km, of the current active vents. The Scari formation consists of prominent pyroclastic deposits, representing strombolian as well as phreatomagmatic explosive activity. The Vancori phase, typified by lava effusion, built Stromboli to its greatest height, and the current summit is composed of deposits from this period. Neostromboli, also composed primarily of lava flows, built much of the northwest portion of the island. All of these phases are delineated by compositional changes (see table above) in the erupted materials directly related to structural modifications to the Stromboli edifice (Hornig-Kjarsgaard et
al., 1993). Large scale sector collapses, for instance, mark the transition from Vancori to Neostromboli and the most recent sector collapse, ~6 ka, formed the current Sciara del Fuoco graben (Fig. 1.4). This event initiated the current phase of activity, Recent Stromboli.

The Recent Stromboli phase of activity is thought to have begun with repeated episodes of sustained fire fountaining followed by a period of quiescence (Rosi et al., 2000). Based upon stratigraphic and radiometric studies of deposits on the NE flank of the volcano, the current eruptive style was established between the third and seventh centuries A.D. This activity involves small intermittent explosions which are periodically interrupted by two rarer forms of eruptive behavior. First, major explosions are unusually violent strombolian eruptions which can sometimes affect the settled areas (Barberi et al., 1993 calls major explosions 'paroxysms' if they threaten populated areas). Second, periods of lava effusion involve the eruption of lava flows beyond the crater terrace, sending lava flows from the crater terrace or closely linked flank vents down the Sciara del Fuoco. Observations from the last ~100 years indicate that major explosions occur 1-3 times a year, while paroxysms have a repose period of 4.8 years (Barberi et al., 1993). Lava effusion episodes have a mean duration of 90 days, with a repose period of 3.7 years (Barberi et al., 1993). Recent major effusive periods include 1975, 1985-86, and 2002-03 (Calvari et al., 2005). The last paroxysmal eruption occurred on April 5, 2003 (Calvari et al., 2005).
1.4 Eruptions at Stromboli

What causes normal eruptions at Stromboli? The model almost universally embraced over the last 30 years is based upon two-phase flow – simply the rise of both gas and magma in the conduit at the same time. At low gas concentrations, the gas phase comprises small bubbles which rise incrementally and independently. With an increase in gas concentration, the gas tends to organize itself into large ‘slugs’ (Vergniolle and Mangan, 2000). Eruptions at Stromboli are thought to be caused by such a slug rising in the conduit and bursting at the top of the magma column, ejecting a mixture of gas and clots of molten lava. There is an impressive body of research corroborating this model in the laboratory (e.g. Jaupart and Vergniolle, 1988; Ripepe et al., 2001), in addition to evidence from seismicity (e.g. Ntepe and Dorel, 1990; Chouet et al., 1999), acoustic pressure (e.g. Vergniolle and Brandeis, 1994; Ripepe et al., 2001) and eruption images (e.g. Chouet et al., 1974; Blackburn et al., 1976). Also, scientists have occasionally viewed a bubble push up the surface of the magma column and burst (Foshag and Gonzalez-Reyna, 1956; Vergniolle and Brandeis, 1994).

Chouet et al. (1974) and Blackburn et al. (1976) were among the first to provide quantitative support that large gas bubbles were involved, as they both estimated that the amount of gas being expelled in an explosion was much greater than that which the lava ejecta could conceivably hold in solution. Basaltic magmas in subduction-related settings have been measured to hold up to 6 weight percent gas in solution (Wallace and Anderson, 2000). Chouet et al. (1974) measured that, by mass, there was up to 16 times more gas than lava in a Strombolian explosion. That is, the lava being erupted clearly did
not represent the complete parent magma for the erupted gas. They reasoned that gas segregation must take place at some depth, allowing gas to concentrate into large bubbles that could rapidly outrun their parent magma. This model is in contrast with plinian eruptions, where gas and magma stay closely coupled during their ascent in the conduit due to the magma's high viscosity (Cashman et al., 2000).

A point of debate concerns the origin of the gas slugs, and two main models have been proposed (see Parfitt, 2004, for review). In the 'collapsing foam' model (Jaupart and Vergniolle, 1988; Vergniolle and Brandeis, 1994; Vergniolle and Brandeis, 1996), gas bubbles concentrate at a physical boundary (the roof of a supposed magma chamber, dike constriction, or otherwise) and comprise a 'foam' of highly gaseous magma. As the gas concentration increases, a critical point is reached and the individual gas bubbles forming the foam rapidly coalesce into a single, large gas slug. In the 'rise speed dependent' model (Wilson, 1980; Parfitt and Wilson, 1995), bubbles of many original sizes buoyantly rise within a column of upward rising magma. If the magma rise speed is slow, there is time for the bubbles to travel a sufficient distance to interact with each other, allowing larger bubbles (which travel faster) to overtake and coalesce with smaller ones (traveling slower). This has a compounding effect and a large gas slug is eventually formed.

Seismic evidence points to the gas slug formation occurring at a depth of ~250 m below the active craters (Chouet et al., 1999). The slugs (which may have a diameter of several meters - almost as wide as the conduit itself) can rise faster than the gas inside can equilibrate its internal pressure with that of ambient pressure. The result is that the
gas slug arrives at the top of the magma column with a minor amount of excess pressure (anywhere from 0.5 to 4 bars, or $0.5 \times 10^5 - 4 \times 10^5$ Pa, Ripepe and Marchetti, 2002). Because of the low tensile strength of the fluid bubble wall, the slug easily explodes at the surface as the pressure gradient equilibrates.

The fluid walls of the bubble are torn in the explosion to produce the molten clots of lava ejecta (Fig. 1.1) which travel in parabolic (largely ballistic) arcs (Wilson, 1980), and the gas itself is ejected as a plume. Ejecta velocities are generally $<100\text{ m s}^{-1}$ (Chouet et al., 1974; Blackburn et al., 1976; Weill et al., 1992; Ripepe et al., 1993; Hort and Seyfried, 1998; Ripepe et al., 2001; and Hort et al., 2003). Normal eruptions at Stromboli have ballistic ejection heights $<300\text{ m}$, with lateral ejection distances $<100\text{ m}$. The majority of ballistic ejecta measured have been $<30\text{ cm}$ in dimension (Chouet et al., 1974; Ripepe et al., 1993). Chouet et al. (1974) determined that, in one eruption, 99% of the particles were $<6\text{ cm}$, and the mean particle size was 2.4 cm.

Blackburn et al. (1976) analyzed a variety of Strombolian eruptions which expelled optically-thick, particle-laden plumes (Fig. 1.1). These were observed to have a two-stage evolution (Blackburn et al., 1976; Sparks and Wilson, 1976). Initially the plumes traveled at high velocity with substantial decelerations, indicating their movement was dominated by inertia. This was termed the ‘gas-thrust’ phase. At a height of $\sim100\text{ m}$, the plumes decelerated to a low, constant velocity that was controlled by the plume buoyancy. This was termed the ‘convective-thrust’, or ‘buoyant’ phase. Velocities for the convective phase were 4-8 m s$^{-1}$ (Blackburn et al., 1976). The plumes
were then observed to continue rising to a height of neutral buoyancy (out of the field of view of the camera) or dissipate out of view.

1.5 The FLIR camera

In my study FLIR cameras were used to examine Strombolian activity (Fig. 1.5). Unlike early analog FLIR cameras, the modern models I have used are temperature-calibrated, producing an absolute reading for temperature at every pixel in the field of view (Fig. 1.6). This allows a powerful means to extract quantitative parameters from eruption imagery. Eruptions are fast-moving phenomena, and still images cannot convey the full dynamic nature of ejecta motion. One of the three models was able to acquire thermal imagery at 30 frames per second (or 30 Hz). Coupling a map of calibrated temperatures with high-speed acquisition rates, these data offered a view of explosive volcanic activity that has never before been seen. Nevertheless, there are important limitations of the FLIR in measuring explosions - the most important being the coarse pixel size – that are discussed in this dissertation.

I used three different models of FLIR cameras, all manufactured by FLIR Systems, Inc. (Boston, USA). These models include the ThermaCAM™ 595, 695 and S40, which are similar cameras with many common traits (Table 1.1). The cameras use a focal plane array (FPA) of uncooled microbolometers, with the detector array (and resultant images) comprising 240 x 320 elements (or image pixels). The detectors are sensitive to longwave thermal radiation from 7.5 to 13 microns, where the spectral
Figure 1.5. The FLIR camera. Above: the FLIR Systems ThermaCAM PM595 camera, similar in form and function to the PM695, being used from a helicopter. Bottom: the ThermaCAM S40 camera, mounted on a tripod and aimed at the active craters.
Figure 1.6. FLIR image of an eruption at Stromboli. Image acquired July 11, 2004.
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response is shown in Figure 1.7. The total field of view is $18^\circ \times 24^\circ$ and the instantaneous field of view (spatial resolution) is 1.3 milliradians. The data have 14 bit quantization, producing image sizes of $\sim 150$ kilobytes. Image frequency is 60 Hz, although actual image storage is less frequent and varies by camera model. Thermal sensitivity ($\text{NEAT}$) is 0.08 °C, and the stated accuracy is ±2 °C, or ±2% of the reading. The temperature measurement range extends from –40 to 1500 °C, at three different gain settings which vary by model. The cameras are generally lightweight (<3 kg) and compact (<25 cm in longest dimension), about the size and weight of a conventional camcorder. The models are all battery powered, with battery life extending 2-3 hours.

There are minor differences among the three cameras (Table 1.1). The ThermaCAM™ 595 allows image storage only onto an internal Flash card, which can be done at 0.5 Hz (an image every 2 seconds) at the fastest rate. The gain settings are low: –40 to 120 °C, medium: 32 to $\sim 500$ °C, and high: $\sim 250$ to 1500 °C. The temperature range for each gain setting, however, was observed to extend slightly broader than the values stated in the user manuals. This model weighs 2.2 kg with battery. The ThermaCAM™ 695 is identical in form and function to the 595 with two exceptions. First, the medium gain setting extends from 0 to $\sim 500$ °C. Second, the camera also acquires visible wavelength images, but these are of such poor quality as to be virtually worthless.

The ThermaCAM™ S40 is a newer generation camera than the 595/695 series, having been released around 2002. Its internal storage is on a CompactFlash card, but the camera saves images at a slower (and more unpredictable) rate than the 595 or 695,
Figure 1.7. Normalized spectral response of the FLIR cameras.
usually an image every few seconds. A great strength of this model, however, is its ability to connect to a notebook computer which allows acquisition rates up to 30 Hz (30 images each second). Connection between camera and computer is through a four or six pin Firewire (IEEE 1394) cable. Its medium gain setting extends from −10 to ~500 °C. The S40 is somewhat smaller than the 595/695, weighing just 1.4 kg with battery.

How does a focal plane array of microbolometers measure radiance? Unlike photon detectors which respond directly to incoming photons, bolometer detectors respond to infrared radiation with an increase in temperature, and resultant change in electrical conductivity. The array of microbolometer detectors is sampled by the camera through a current pulse, row by row. In each row, the voltage across each detector is read and stored in an array of 320 capacitors, which are coupled to 320 analog-to-digital converters which digitize the analog voltage. This voltage is factory calibrated to incoming radiance. In changing from the low gain setting to the medium gain setting, the current pulse is increased in order to compensate for the increased radiation reaching the detectors, to avoid saturating the capacitors. When switching to the high gain setting, the camera places an aperture in front of the detector array which reduces incoming radiation uniformly across all wavelengths.

How does the camera convert total measured radiance to temperature of the target? The general measurement formula used by the camera to measure target radiance is as follows:
\[ R_{\text{tot}} = \varepsilon \tau R_{\text{obj}} + (1 - \varepsilon) \tau R_{\text{refl}} + (1 - \tau) R_{\text{atm}} \quad 1.1 \]

where the total radiance \((R_{\text{tot}})\) is a sum of the emitted target radiance \((R_{\text{obj}})\), the reflected target radiance \((R_{\text{refl}})\), and the emitted atmospheric radiance \((R_{\text{atm}})\). The variables \(\varepsilon\) and \(\tau\) denote emissivity of the target and transmissivity of the atmosphere over the measurement path length, respectively. Solving for blackbody emitted target radiance \((R_{\text{obj}})\), the equation is then:

\[ R_{\text{obj}} = \frac{1}{\varepsilon \tau} R_{\text{tot}} - \frac{1 - \varepsilon}{\varepsilon} R_{\text{refl}} - \frac{1 - \tau}{\varepsilon \tau} R_{\text{atm}} \quad 1.2 \]

The emissivity is user input and the transmissivity of the atmosphere is calculated with the camera's built-in LOWTRAN atmospheric correction routine (Selby et al., 1978), where the user input parameters include distance to target, ambient temperature and relative humidity. The blackbody emitted target radiance \((R_{\text{obj}})\) is then used to calculate kinetic target temperature, \(T_{\text{obj}}\), with the Planck function:

\[ T_{\text{obj}} = \frac{c_1}{\lambda \ln \left( \frac{c_2 \lambda^{2.5}}{\pi R_{\text{obj}}} + 1 \right)} \quad 1.3 \]

where \(c_1=0.0144\ \text{m K}\), \(c_2=3.742 \times 10^{-16}\ \text{W m}^2\), and \(\lambda = \text{wavelength}\).
The stated accuracy of the cameras is 2 °C, or ±2% of the reading in Celsius (Table 1.1), and the camera is factory calibrated before shipment. The noise-equivalent ΔT is 0.1 °C. Following purchase, periodic calibration at the factory is necessary to maintain these specifications. Ball and Pinkerton (2005, submitted) performed a detailed laboratory study of ThermaCAM™ S40 performance and found that actual noise and accuracy were well within the published values.

FLIR cameras thus provide an image of temperatures: an image of 240 x 320 pixels comprising 76,800 individual data points. This technology is a step forward from radiometers which provide temperature integrated over a single area (i.e. one data point), which I call 'spot radiometers'. Spot radiometers have some advantages, however, in being much cheaper than a FLIR (~$1000 compared to ~$50,000) and generally more robust – enabling long-term, unmanned field deployment (Harris et al., 2005c). A portion of this study compares FLIR results to spot radiometer results.

1.6 Broadband infrared imaging of active volcanoes

A review of past work involving infrared imaging is presented here, and I restrict the scope to imaging systems that were most similar, considering their time period, to the current broadband, handheld models (such as the ThermaCAM™ S40). This is meant to convey the trajectory that similar technology and applications have progressed through. Thus, I limit the discussion to broadband imaging performed on the ground or from
aircraft, which is generally accomplished from oblique viewing angles. Note that I do not include studies using spot radiometers, infrared film, or satellite remote sensing.

I also do not explore multispectral imaging in this summary, which takes a more precise approach to surface temperature measurement than broadband infrared imagers and has provided valuable information on active lava flows. For instance, Horton (1995) used spaceborne and airborne spectroradiometers to determine excess radiative energy flux at Kilauea (9-12 GW in 1991), which was tied to eruptive mass flux. Other studies (Flynn and Mouginis-Mark, 1992, 1994; Flynn et al., 1993) have used field spectroradiometers to measure thermal distributions and radiative heat flux from active flows and lava lakes.

Infrared imaging began its transition into the civilian sector during the 1960s, with technology based upon a single cooled photon detector element augmented by a mechanical scanning apparatus. A review of the early studies (spanning the 1960s and 70s) is provided by Francis (1979). Most likely, the first ground or air-based infrared imaging of an active volcano was at Kilauea, Hawaii, in 1963 (Fischer et al., 1964). The images, acquired during airborne surveys, were crude by today’s standards but nonetheless proved effective in identifying areas of anomalous temperatures produced by fumarolic activity. Subsequent work throughout the 1960s, 70s and 80s used similar technology and airborne survey techniques (e.g. Moxham et al., 1965; Friedman et al., 1969; Moxham 1970; Eichelberger et al., 1976; Kagiyama, 1981; Yuhara et al., 1981; Ballestracci and Nougier, 1984; Tabbagh et al., 1987). All of these used cooled photon detector cameras with a mechanical scanning apparatus. This equipment was not field
portable, as a complete system could comprise the cameras, a liquid nitrogen tank, video recorders and TV monitors (Fig. 1.8). By the 1990s single-detector scanning cameras were using more modern electronics (e.g. digital data storage) and Stirling-cycle coolers (i.e. no liquid nitrogen), and camera systems were reduced to more manageable sizes. Mongillo and Wood (1995) used such a camera, the 9 kg Inframetrics 600, during helicopter surveys of White Island volcano, New Zealand, in 1992. Recently, Matsushima et al. (2003) used the 5.5 kg handheld AGEMA Thermovision 470 to measure heat flux from Iwodake volcano, Japan.

The 1980 eruption of Mount St. Helens, USA, saw one of the first uses of a thermal camera at an erupting volcano. Kieffer et al. (1981) used a variety of different infrared scanning imagers from aircraft to map temperatures on Mt. St. Helens in the two months preceding the May 18 paroxysm. Friedman et al. (1981) used the same equipment to monitor the new dome growth and examine temperatures of pyroclastic flow deposits.

Hawaii has been a natural target for thermal imaging due to the accessibility and frequency of lava effusion. Pieri et al. (1984) used an Inframetrics 525 (a single-detector scanner with liquid nitrogen cooling) from a helicopter to map active a’a lava flows from Mauna Loa, as well as their source fountains. Beginning in 1993, the Hawaii Volcano Observatory (HVO) has used an infrared camera several times a year from plane and helicopter to map areas of active lava at Kilauea (Kauahikaua et al., 2003).

In the late 1980s the technology for focal plane arrays (FPA), originally developed by the US military, appeared on the civilian market. FPAs involve an array of
Figure 1.8. Example of previous infrared imaging system. Airborne system used by Ballestracci and Nougier (1984) at the volcanic Kerguelen Islands. Compare to the modern setup in Figure 1.5.
detectors, much like the charged-coupled devices in digital cameras, eliminating the need for a scanning apparatus. Related to this breakthrough, uncooled microbolometer technology was becoming commercially available. Unlike cooled detectors which react directly to incoming photons, bolometer detectors respond to incoming radiation with a small temperature change and electrical conductivity alteration. The bolometer was originally invented in 1880 by Samuel Langley, but the 1980s saw its first micro-manufacturing and integration into FPAs. Although microbolometers are generally less sensitive (i.e. more noise prone) than photon detectors, they do not require cooling and thus depend on few, if any, moving parts. This reduces the weight of the camera (generally <5 kg) as well as the likelihood of field maintenance. Among current models, shortwave (3-5 micron) cameras tend to use cooled photon detectors while longwave (8-13 micron) cameras use microbolometers.

A recent notable application of thermal imaging occurred in Alaska in 1997 and 1999 during the eruptions of Pavlof and Shishaldin volcanoes. Utilizing a FLIR Systems Safire model mounted on the underbelly of an Alaska State Troopers twin-engine aircraft, McGimsey et al. (1999) identified eruptive activity at these volcanoes during their 1997 and 1999 (respective) strombolian phases. The Alaska Volcano Observatory (AVO) purchased a handheld FLIR Systems ThermaCAM™ 595 shortly thereafter, in 2000, and Italy’s Istituto Nazionale di Geofisica e Vulcanologia acquired a ThermaCAM™ 695 the following year. These cameras, as well as several ThermaCAM™ S40s, were used primarily at Stromboli, Vulcano, and Etna volcanoes, in Italy (Dehn et al., 2001; Patrick et al., 2003; Patrick et al., 2004; Calvari and Pinkerton, 2004; Calvari et al., 2005; Harris
et al., 2005a,b). In addition, the 2001 eruption of Mt. Cleveland, Alaska, produced a lava flow and debris fan which were imaged by Dean et al. (2004) with the AVO ThermaCAM™ 595.

Beyond Alaska and Italy, recent thermal imaging has also occurred in Hawaii, Mexico, and Japan. Lava flow cooling at Kilauea volcano was examined with a ThermaCAM™ 595 (Wright and Flynn, 2002) and ThermaCAM™ S40 (Pinkerton et al., 2004). Kaneko et al. (2002) used an automated ground-based thermal camera (Nikon LAIRD 3A) to monitor small phreatic eruptive plumes at Mt. Usu. Wright et al. (2002) made use of infrared imagery from a monitoring camera at Popocatapetl volcano, Mexico, operated by the Centro Nacional de Prevención de Desastres.

The 2002-2003 eruptive phase at Stromboli volcano and a series of eruptions at Etna between 2001 and 2004 tested thermal cameras in the first prolonged volcanic crisis since Mt. St. Helens in 1980. The cameras were generally judged to be an enormous asset (Calvari and Pinkerton, 2004; Calvari et al., 2005). The cameras were able to perform a number of tasks difficult or impossible to accomplish through any other means, including lava flow effusion rate estimation and inner crater temperature measurements. The cameras were typically used from a helicopter on twice-daily surveys, enabling real-time tracking of changes in eruption parameters which proved invaluable for up-to-date hazard assessment reports.

Handheld thermal cameras have gained further attention with the renewed activity at Mount St. Helens in 2004. With a Safire-I borrowed from the Alaska Volcano Observatory, helicopter surveys were performed to map temperatures of the growing
dome and estimate effusion rates (Gardner et al., 2004; Schneider et al., 2004). The term 'FLIR' entered common usage among reporters covering the story, and FLIR images appeared on news programs, in newspapers and on the world wide web. For instance, FLIR images played prominently in KATU (Portland, Oregon) reports of activity at Mount St. Helens.

Infrared imaging has thus been used for a variety of different objectives in understanding active volcanism. These can be divided into six main activities:

1) **Passive heat loss and fumaroles:** Nearly all of the early imaging studies centered on simply locating thermal anomalies on volcanoes associated with passive heat loss, commonly related to fumarolic activity. This was the objective at Kilauea in Hawaii (Fischer et al., 1964), Mt. Rainier, Mt. Shasta, Lassen Peak, Mt. Baker and Mt. St. Helens in the Cascades (Moxham et al., 1965; Moxham, 1970; Eichelberger et al., 1976; Kieffer et al., 1981), in Iceland (Friedman et al., 1969), at O-shima, Kirishima and Unzen volcanoes in Japan (Shimozuru and Kagiyama, 1978; Kagiyama, 1981; Yuhara et al., 1981), the Kerguelen Islands, South Indian Ocean (Ballestracci and Nougier, 1984), Etna and Vulcano, Italy (Tabbagh et al., 1987) and White Island, New Zealand (Mongillo and Wood, 1995).

Recent studies used a FLIR on Stromboli (Patrick et al., 2003) and Iwodake (Matsushima et al., 2003) volcanoes to examine heat budgets, and Calvari et al. (2005) performed aerial surveys to measure crater temperatures during the 2002-2003 eruption of Stromboli. Patrick et al. (2003) determined pre-eruption heat flux averaged 420 MW, and increased to 2100 MW during the eruptive phase. Matsushima et al. (2003) found
that total heat flux from the Iwodake summit crater dropped from 910 MW in 1996 to 110 MW in 1999. Bonnaccorso et al. (2003) used thermal camera data to reveal hot cracks in the slopes of Stromboli that outlined the failure scarp for a large collapse event in late 2002 which caused a small tsunami. Likewise, Calvari and Pinkerton (2004) used a FLIR to identify hot cracks on a new cinder cone at Etna in order to judge slope instability.

2) Active lava flows and lava lakes: The Hawaii Volcano Observatory has been using a FLIR camera in their monitoring efforts at Kilauea since 1993, mapping active flow areas (Keszthelyi, 1993; Kauahikaua et al., 2003). Spatter-fed lava flows on Shishaldin were imaged by the Alaska Volcano Observatory (McGimsey et al., 1999). The active lava lake at Erta Ale, Ethiopia, was imaged by Oppenheimer and Yirgu (2002) to estimate total heat loss (100-200 MW), which they determined was much lower than that previously determined through coarser resolution satellite imagery (e.g. Harris et al., 1999). Bailey et al. (2002) and Harris et al. (2005a) positioned a stationary FLIR camera beside an active flow at Mount Etna to examine lava flow dynamics over the course of a few hours in 2001, and recognized periodic surging and overflows on the scale of minutes. Wright and Flynn (2003) analyzed lava flow surface temperatures at Kilauea to explore the accuracy of the popular dual-band method (Rothery et al., 1988) for deconstructing satellite image radiance values into multiple thermal components. They found that a two-component model of lava flow surface temperatures was inadequate for characterizing the complex spectrum of temperatures on cooling pahoehoe lobes. Pinkerton et al. (2004) and Ball and Pinkerton (2005) have used the FLIR to make
improved estimates of basalt emissivity, which varies by roughness and viewing angle. For normal viewing angles, emissivity was 0.97-0.98 but dropped to 0.86 for a viewing angle of 80° from vertical.

The emergence of lava flows at Stromboli in 2002-2003 allowed Harris (2004) to devise a method for estimating lava flow effusion rates from the imagery by adapting a thermal approach initially developed for satellite data by Harris et al. (1997) (Dehn et al., 2003; Harris et al., 2005b), the results of which were integrated into daily hazard assessments (Calvari et al., 2005). Lava effusion rates in May-June 2003 were generally in the range of 0.3-0.5 m³ s⁻¹, or 810-1350 kg s⁻¹ using a density of 2700 kg m⁻³.

3) Dome growth: Mount St. Helens has been the primary focus for FLIR work on active lava domes due to its recent activity and accessibility. Friedman et al. (1981) examined the growing dome at Mt. St. Helens following the 1980 sector collapse. In August 1980 radiant heat loss from the dome exceeded 18 MW. With the resurgence of activity in 2004, attention has been centered on the new phase of dome growth (Gardner et al., 2004; Schneider et al., 2004). FLIR proved effective in identifying areas of active effusion that were difficult to identify with the naked eye, from which effusion rate estimates were made using changes in the dimensions of thermally radiant portions of the dome. Extrusion rates ranged from 2 to 8 m³ s⁻¹ during October 2004. A very similar volcano, Bezymianny in Kamchatka, has also been a target for FLIR surveys. Its active dome periodically effuses lobes of fresh dacitic lava with the relative ages of the dome surface material recognizable in FLIR imagery acquired from a helicopter (Ramsey and Dehn, 2004; Steffke et al., 2004).
4) Strombolian explosions: Glaze et al. (1991) acquired a limited data set for strombolian activity of Mt. Etna in 1989. McGimsey et al. (1999) acquired thermal imagery of Pavlof and Shishaldin volcanoes between 1997 and 1999, imaging both the ejecta and warm deposits. Dehn et al. (2001) used the first FLIR camera on Stromboli volcano, which was followed with work by Patrick et al. (2003; 2004). These studies recognized two primary eruption styles at Stromboli, one of which contains only gas and coarse ballistics, and another which erupts an optically-thick ash plume. Patrick et al. (2004) present initial results for 344 strombolian eruptions which formed the basis for this dissertation.

5) Plumes: Kaneko et al. (2002) used a remote thermal camera, broadcasting data via the internet, to monitor plumes at Mt. Usu, Japan. A simple algorithm counting temperature-elevated pixels was used to quantify activity levels and track trends in gas output. Wright et al. (2002) also used data broadcast over the internet from a remote camera in observing Popocatepetl volcano to assess activity styles, primarily using satellite data, during dome growth and explosive activity. Plumes exhibited low temperatures (9-12 °C) due to the entrainment of air.

6) Cooling deposits: Thermal cameras have been effective at differentiating deposits of different ages based upon their temperature. Friedman et al. (1981) analyzed pyroclastic flow deposits at Mt. St. Helens after the 1980 paroxysm, which were observed to be slightly warmer than ambient three months after formation. Gully deposits on the upper flanks were 6 °C warmer than ridges covered by thin ash. McGimsey et al. (1999)
was able to differentiate between days-old and weeks-old debris flows at Shishaldin volcano in 1999.

1.7 Dissertation outline

In Chapter 2, I use FLIR imagery to classify and describe different styles of eruptions at Stromboli. I then explore the conduit scenarios which could be responsible for these variations. In Chapter 3, I take a detailed look at Strombolian ash-plume thermodynamics. Using the trend of plume temperatures over time, I try to understand the changing dynamic behavior of explosively ejected mixtures of volcanic gas, ash and entrained air. I also compare the observed plume dynamics with existing conceptual models of plume formation. In Chapter 4, I take the mass and heat budget ideas developed in Chapter 3 and implement them in a simulated operational scenario. That is, if there was an operational feed of FLIR data from a permanent camera installation, how might eruption parameters be estimated? Note that Chapters 2-4 are meant to be self-standing works equivalent to a published article. Therefore, each offers some common, introductory discussion on the FLIR, Stromboli, and strombolian activity. Chapter 5 offers a summary of the dissertation, some discussion of the implications of this work, and suggestions for future work for strombolian and other eruption styles. In the appendices, I have included a paper which I wrote on Mount Belinda, in the South Sandwich Islands, in collaboration with a number of people. This has been published as Patrick et al. (2005).
Chapter 2

Strombolian explosive styles and source conditions: Insights from thermal video (FLIR) imagery

2.1 Introduction

Stromboli volcano, in the Aeolian Islands of Italy, offers an ideal location to study mildly explosive basaltic volcanism - i.e. 'strombolian' activity. The volcano provides a venue for constraining the characteristics of shallow conduit dynamics relevant for the numerous other volcanoes that periodically exhibit similar behavior (e.g. Shishaldin, USA; Villarrica, Chile; Etna, Italy; Karymsky, Russia). Through seismicity (e.g. Ntepe and Dorel, 1990), infrasound (e.g. Ripepe and Marchetti, 2002), geochemistry (e.g. Francalanci et al., 2004), laboratory models (Jaupart and Vergniolle, 1988; Ripepe et al., 2001), physical volcanology (Lautze and Houghton, 2005) and assorted other tools, our understanding of Stromboli's eruption dynamics has increased substantially over the last 30 years. Ground-based imaging (Chouet et al., 1974; Blackburn et al., 1976; Ripepe et al., 1993) has proven to be an important member of this suite, providing information on the surface manifestation, and end product, of the explosive process.

Previous imaging studies at Stromboli (Chouet et al., 1974; Blackburn et al., 1976; Ripepe et al., 1993), however, suffer from two shortcomings. First, the inherent
limitations of visible and near-infrared photography preclude simultaneous observation of major ejecta constituents. For instance, if viewing at night, the incandescent ballistic particles will be visible but the generally cold ash plume will be largely invisible. During daylight, the plume will be conspicuous but ballistic particles, due to their small size, can only be viewed at a high magnification – thus losing a synoptic sense of the eruption. Second, each study considered a small dataset collected over a similarly short time period, and may therefore not characterize the entire suite of normal strombolian activity. In spite of the small data sets, the results from these studies have served as supportive data, and at times a fundamental basis, for a number of models of Stromboli’s eruption and conduit dynamics (e.g. McGetchin and Chouet, 1979; Wilson, 1980; Giberti et al., 1992; Allard et al., 1994; Vergniolle and Brandies, 1994; Harris and Stevenson, 1997; Parfitt, 2004). This underlines the importance of exploring the degree to which these imaging studies convey the full range of typical strombolian activity.

In this chapter I use a Forward Looking Infrared Radiometer (FLIR) handheld camera to characterize strombolian eruptions. As described in Chapter 1, FLIR data consist of temperature-calibrated thermal images measuring long-wave infrared radiation (7.5 – 13 μm), and are capable of observing the thermal expression of the ballistic particles as well as any ash plume within a field of view (~130 m) that conveys the scale of the eruption. During May 2001, May 2002, September 2003, and June-July 2004, 344 Strombolian eruptions were imaged and these data were analyzed to characterize the variability in eruptive behavior. This chapter presents observations on the range of
eruptive styles and provides some possible explanations for the observed variation in eruption style.

2.2 Conceptual model of Strombolian eruptions

Eruptions at Stromboli, as well as strombolian eruptions at other volcanoes, are commonly thought to originate from the bursting of a large bubble, or slug, of gas at the magma free surface (Chouet et al., 1974; Blackburn et al., 1976; Vergniolle and Brandeis, 1989). As early as the eruption of Paricutin in 1943, Foshag and Gonzalez-Reyna (1956) remarked on ‘Huge bubbles...presumably filled with gases. When they burst, huge glowing masses of viscous lava were flung over the cone’. The mechanism of gas slug formation has been a point of debate centered around two primary models (see Parfitt, 2004 for review). In the first - a collapsing foam model (Jaupart and Vergniolle, 1988; Vergniolle and Brandeis, 1994; Vergniolle and Brandeis, 1996) - the slug originates from an accumulation of small bubbles, or foam, at the roof of a magma reservoir. At a critical point, the foam collapses and a large gas slug coalesces, and subsequently travels up the conduit. In the second – a rise speed dependent model (Wilson, 1980; Parfitt and Wilson, 1995) - bubbles of many original sizes buoyantly rise within a column of upward rising magma. If the magma rise speed is slow, there is time for the bubbles to travel a sufficient distance to interact with each other, allowing larger bubbles (which travel faster) to overtake and coalesce with smaller ones (traveling slower). This has a compounding effect and a large gas slug is eventually formed.
Regardless of the model, seismic evidence suggests that slug formation at Stromboli occurs at an approximate depth of 250 m (Chouet et al., 1999). Viscous forces inhibit perfect equilibration of the pressure inside the slug during its ascent and it arrives at the magma free-surface with a minor amount of overpressure, typically $0.5 - 4 \times 10^5$ Pa or ~0.5-4 bars (Ripepe and Marchetti, 2002). The overpressure bursts the gas slug, with the force of the gas explosion ejecting coarse fragments of molten lava which had surrounded the slug (Walker, 1973; Blackburn et al., 1976). Many of the particles are incandescent, resulting in the brilliant parabolic trajectories that are characteristic of strombolian eruptions. These coarse particles are lapilli to bomb size (~2-30 cm, Chouet et al., 1974; Ripepe et al., 1993), are typically ejected to heights of several hundred meters and generally fall within a hundred meters of the vent.

In some strombolian explosions finer particles are erupted and couple with the gas phase, creating an ash plume. The eruption carries the ash plume upward at a high initial velocity (~10-100 m s$^{-1}$; Blackburn et al., 1976), entraining increasing amounts of air during its rise. Momentum conservation, related to the entrained air mass, forces the plume to decelerate (Woods, 1988). By the time that this deceleration approaches zero, the plume has normally entrained and heated sufficient air to greatly reduce its bulk density, so that it becomes buoyant. At this point, the plume transitions from a gas-thrust phase into a buoyant convective phase (Sparks and Wilson, 1976). During this convective phase, velocities are roughly constant (~4-8 m s$^{-1}$; Blackburn et al., 1976) and the ash plume continues rising until it reaches a height of neutral buoyancy.
Strombolian eruptions thus eject a mixture of both volcanic gas and solid particles. The solid component comprises a range of different size fragments from ash to bombs. Very fine particles remain mechanically coupled to the gas phase and can form an ash plume. Coarser particles follow a trajectory that is largely independent of the dynamics of the gas phase, although the path can be slightly modified by the influence of the ascending gas (Self et al., 1980). I regard these as ballistic particles, as they remain for the most part mechanically decoupled from the gas phase, following roughly parabolic trajectories. Note that my distinction between the two types of fragments is based purely on their different post-ejection motion, paths, and dynamics, which are in turn tied to size. Also note that ‘ballistic’ particles here can be much smaller than those found in plinian eruptions, due to much lower gas velocities encountered during strombolian eruptions.

2.3 Data collection

2.3.1 FLIR ThermaCAM™ cameras

I used three different FLIR camera models to image the eruptions. The first instrument was a ThermaCAM™ PM 595 model, produced by FLIR Systems, Inc (Fig. 2.1). The camera uses an uncooled microbolometer detector array with a field of view of 24° x 18°, an instantaneous field of view of 1.3 milliradians, and a thermal response from 7.5 to 13 μm. Acquisition occurs at 60 Hz, but actual image storage on this model can only be achieved at a two second interval (0.5 Hz). The images are 320 pixels wide by
Figure 2.1. Recording setup on Stromboli. Top: The FLIR ThermaCAM S40, about the size and weight of a conventional video camera, at the Rocette (ROC) site on Stromboli volcano. The camera is connected via Firewire to a notebook computer, in the orange Pelican case, to collect at 30 Hz. Bottom: The two recording sites relative to the crater terrace. The inset shows a larger scale view of the western side of Stromboli and the Sciara del Fuoco, and extent of the main photograph. Photo taken in May 2003.
240 pixels high (76,800 data points) at 14 bit quantization, producing an image file size of 158 kilobytes which includes accessory information (date and image settings). The second camera was a ThermaCAM™ PM 695, which is essentially identical to the PM 595. The third camera, a ThermaCAM™ S40, is also similar to the PM 595 in most respects but with one important exception, this being that it can save imagery at 30 Hz (or 30 frames per second) via Firewire to a notebook computer. On several occasions I used two S40 models simultaneously, with one imaging with the standard 24° lens, and the other using a 12° zoom lens pointed at the vent area.

The FLIR is well-suited for the field (Dehn et al., 2001). Including the battery it weighs just ~2 kg, and is about the size of a standard hand-held video camera (longest dimension is 22 cm). Because there is no scanning mechanism, there are no moving parts other than the shutter mechanism that periodically corrects the image scale. And being uncooled, there is no need for cooler maintenance.

2.3.2 FLIR measurement setup

FLIR data used in this study were collected over four field seasons: 17 May – 02 June 2001, 09 May – 25 May 2002, 16 Sept 2003, and 05 June – 25 July 2004 (Table 2.1). A total of 344 eruptions were recorded, with the majority (240) being obtained during the period June-July 2004. As this latter dataset was collected with the S40 model acquiring at 30 Hz it offered parameterization at an extremely fine timescale. The other large dataset includes 64 eruptions acquired during a very low-activity period in May.
Table 2.1. Explosions imaged by the FLIR camera

<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
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<tr>
<td>NE eruptions</td>
<td>8</td>
<td>35</td>
<td>2</td>
<td>170</td>
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<td>C eruptions</td>
<td>0</td>
<td>2</td>
<td>5</td>
<td>9</td>
<td>16</td>
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<td>18</td>
<td>27</td>
<td>7</td>
<td>61</td>
<td>113</td>
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<td>Total eruptions</td>
<td>26</td>
<td>64</td>
<td>14</td>
<td>240</td>
<td>344</td>
</tr>
<tr>
<td>Imaging frequency (Hz)</td>
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<td>0.5</td>
<td>0.5</td>
<td>30</td>
<td></td>
</tr>
<tr>
<td>Camera</td>
<td>595</td>
<td>595</td>
<td>695</td>
<td>S40</td>
<td></td>
</tr>
<tr>
<td>Location*</td>
<td>PZZ</td>
<td>PZZ</td>
<td>PZZ</td>
<td>ROC+PZZ</td>
<td></td>
</tr>
</tbody>
</table>

* see Fig. 2.1 for locations
2002. These eruptions, along with those from May 2001 and September 2003, were imaged every 2 seconds (0.5 Hz). This timescale is much coarser than the 30 Hz data, but nonetheless provides a serviceable characterization of the eruptions. The camera was always run continuously, for typically ~4 hours on each observation day, to ensure capture of the eruption onsets.

The camera was placed at either the Pizzo sopra la Fossa (PZZ) or Rocette (ROC) site (Fig. 2.1), which are approximately 250 and 450 m away from the active craters, respectively. PZZ is about 150 m above the craters, while ROC is approximately level with them. At 250 m and 450 m distance, each pixel in the FLIR image is 33 cm and 59 cm in dimension, respectively. Each observation site had advantages and disadvantages. PZZ provides views of all the active craters and was used exclusively in May 2001, May 2002, and September 2003. Being closer to the craters it allowed features to be imaged at a finer resolution, however, its field of view was limited. The vertical height of the image was just ~75 m, so that many eruptions quickly extended out of the field of view. The ROC site was used most often in June 2004. Although it was farther from the crater terrace, and could only image the NE crater without obstruction, its vertical field of view was ~140-180 m (depending on camera orientation). This allowed a longer observation of the rising ash plumes. Also, due to a crater floor collapse and loss of the NE wall of the crater terrace in December 2002, the ROC site in 2004 allowed a rare opportunity to see deep into NE crater, and thereby allowed me to capture the ejecta much closer to the vent than during other periods. Prior to the collapse, higher crater rims meant that ejecta
was out of sight for a longer period. Eruptions in 2001-2002 were captured after dark,
while those in 2003 and 2004 were imaged during the daytime.

Each explosion originated from one of three main craters (SW, Central and NE) set in a 300 m long crater terrace (Figs. 2.1+2.2). At the onset of the 2002-2003 effusive phase in December 2002, much of the terrace collapsed destroying the septa and inner flanks which had distinguished the Central and NE craters. Although the crater edifices were no longer in place, the approximate vent locations and eruptive behavior remained intact, and by summer 2004 the craters were reforming in their pre-2003 locations.

Of the 344 eruptions imaged, 118 were from SW crater. The Central crater only rarely produced what would be considered typical Strombolian explosions. Just 11 eruptions were observed at Central crater, and these were extremely small in size. The NE crater was separated into several sub-craters, which I refer to as NE-1, NE-2, and NE-3 (Fig. 2.2). During May 2002 and June-July 2004 the NE crater was the most active, and the camera was pointed at this crater more often than the others, with a total of 215 eruptions imaged. Because the FLIR field of view could only accommodate 1-2 craters at a time, the number of imaged eruptions for each crater is merely a measure of where the camera was pointed, and not the relative eruption frequency.

2.3.3 FLIR sensitivity to ballistic particle size

Chouet et al. (1974) and Ripepe et al. (1993) determined that most Strombolian ballistic particles are less than 30 cm in diameter, suggesting that the majority of the ballistic particles I image with the FLIR are subpixel in size. The ballistic particle size
Figure 2.2. Crater terrace viewed from the Pizzo sopra la Fossa (~250 m distance, see Fig. 2.1). Crater/vent locations are shown for May 2002 (A) and June 2004 (B). The elevated temperatures on the outer flanks of the crater terrace in (B) result from solar heating. Each image is ~300 m wide.
sensitivity was calculated using the Planck function, incorporating the response function provided by FLIR Systems, Inc (Chapter 1). Considering the variance in background temperatures, I judged that a pixel-integrated temperature difference of 5 °C was the threshold for reliable visual identification of ballistic pixels. At a distance of 450 m, this equates to particles sizes of 2.1 and 3.5 cm for particle temperatures of 1000 and 500 °C, respectively. At 250 m, this equates to 1.2 and 1.9 cm for temperatures of 1000 and 500 °C, respectively. This calculation only applies to solitary ballistic particles. In ash plumes the particle size is much smaller than these values, however, the particle concentration typically produces an opaque plume that spatially fills the pixel.

2.4 Methods of analysis

2.4.1 Visual observations

Naked-eye (and audible) observations comprised the first step to characterizing eruption styles, and were done most rigorously in 2004 when acquisitions were performed in daytime. During other periods (2001-2002) the eruptions were captured after dark and visual descriptions were ineffective. Important parameters were 1) the presence or lack of an optically thick ash plume, 2) the general abundance, height and collimation of ballistic ejecta, 3) relative velocities, and 4) audible characteristics. These visual and audible observations were then compared to the FLIR video sequences viewed with FLIR Systems ThermaCAM™ Researcher software. The FLIR video consistently

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agreed with general naked-eye observations. Most importantly, it was clear which eruptions were ballistic-dominated and which were ash-dominated. The FLIR was much more effective than visual observations, however, at characterizing the full dispersal of ballistic ejecta, as a large portion of these are too small to see by eye at a distance of 450 m.

2.4.2 Plume front tracking

Ash plume heights were measured at the leading front of the plume for ash-rich explosions where the main plume front could be tracked continuously. In ballistic-rich (ash-poor) eruptions, plumes became diffuse and invisible too early to track for any appreciable amount of time. The method was applied to 80 ash plumes in June 2004 (NE-1 crater), in which the vertical field of view was either ~130 m or ~180 m depending on camera orientation. Height values were measured every 3-5 frames (or every 0.08-0.16 s at 30 Hz acquisition) in the gas thrust region and every 10-30 frames (0.33-1 s) in the slower convective region. These apparent height values were then corrected for minor vertical distortions (<3%) due to camera viewing angle. For calculating velocity and acceleration, the height values were then smoothed with a fifth-order polynomial in order to remove minor erratic variations (all fits had $R^2 > 0.99$).

2.4.3 Initial velocities

The velocity of the gas/ash plume onset for 134 eruptions from NE-1 crater during June-July 2004 was measured from the FLIR imagery, generally over a small
distance (mean: 9.5 m) and short time interval (mean: 0.7 s) just above the crater rim. The plume was chosen because it is a single, easily-trackable object with a spatial scale larger than the FLIR pixels. Ballistic velocities, on the other hand, depend on particle size (Chouet et al., 1974; Steinberg and Babenko, 1978) which could not be measured accurately from the FLIR images due to the large size of the pixels. The actual initial gas velocity at the source would be greater than my measured plume velocity at the crater rim, however, as long as the plume is measured at a consistent location, as it is here, and the magma free surface is relatively stationary, the plume velocity would offer a good proxy for initial gas velocity. Although ballistic-rich eruptions lacked an ash-rich plume, their gas plume was discernable very close to the crater rim where gas concentrations were high.

2.4.4 Ballistic height

The maximum observed height of ballistics was measured from the FLIR imagery by eye for 138 eruptions from NE-1 crater during June-July 2004. A limitation of the FLIR image is that the field of view extended to only 115-180 m above the crater rim. Forty-three (31%) of the eruptions had maximum ballistic heights which exceeded the top of the field of view, which changed (115-180 m) depending on camera orientation. To be consistent, I assigned a value of 115 m to maximum height values that exceeded the field of view, be it 115 or 180 m.
At 450 m distance, the FLIR particle size sensitivity extends down to 2.1 and 3.5 cm for ballistic temperatures of 1000 and 500 °C, respectively (section 2.3.3). Using a drag coefficient of 1 (Sparks et al., 1997) and mean ballistic density of 1100 kg m⁻³ (N. Lautze, personal communication, 2005), initial velocities <100 m s⁻¹ will require that ballistics which reach >115 m in height must be >7 cm in diameter. Thus, I can be confident that no unaccounted for particles were exceeding the field of view.

2.4.5 Eruption duration

Eruption duration was calculated for 135 eruptions from NE-1 crater in June-July 2004 using the FLIR imagery. The onset of all eruptions in the FLIR was impulsive and easily identified, however, the end of each eruption was sometimes difficult to discern with great precision. The cessation of a ballistic-rich eruption was indicated relatively clearly by the end of visible ballistic ejecta being thrown upward, and this marked a generally abrupt end. Ash-rich eruptions were more troublesome as the trailing portions of the ash plumes often stagnated inside the crater and made it difficult to tell if vigorous ejection had ended. Several approaches were used to identify the end of ash-rich eruptions, and the most effective and objective was to plot the maximum temperature in a small (~20 m diameter) window, placed at the crater rim, throughout the course of the eruption. This seemed to produce a good proxy for the mass flux of hot material passing the crater rim, with an abrupt drop in maximum temperature marking the end of eruption.
2.5 Results

As the 2004 field season resulted in the highest quality dataset of my study, I relied on this period for most of the quantitative results. The 2001-2003 data were used mainly for qualitative assessments of eruption style. This section thus integrates results from all three major craters during multiple field seasons for qualitative examinations (sections 2.5.1-2.5.2) as well as results for a particular crater (NE-1) during one field season, 2004 for a more quantitative analysis (sections 2.5.3-2.5.9).

2.5.1 Styles of eruptions at Stromboli

Eruptions varied widely in style and appearance, but could be broadly categorized into two main groups: Type 1 and Type 2 eruptions. This partly follows the classification of Chouet et al. (1999), who attributed Type 1 eruptions to NE crater and Type 2 eruptions to SW crater. My results show that both eruption styles were common at either crater (Table 2.2).

Type 1 eruptions were ballistic-dominated with little to no visible plume component (Fig. 2.3). Although some plume must be present due to the gas slug, it is largely invisible to the naked eye and is semi-transparent in the FLIR imagery due to the paucity of ash-sized particles. Because of the low optical thickness, reliable temperature measurements of the plume cannot be made and it becomes difficult to discern with increasing height and dilution from entrained air.
Table 2.2. Eruption styles at the active craters

<table>
<thead>
<tr>
<th></th>
<th>SW</th>
<th>C</th>
<th>NE-1</th>
<th>NE-2</th>
<th>NE-3</th>
</tr>
</thead>
<tbody>
<tr>
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<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
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<td>15</td>
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<td>2</td>
</tr>
<tr>
<td>Type 2a</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>Type 2b</td>
<td>12</td>
<td>0</td>
<td>0</td>
<td>0</td>
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</tr>
<tr>
<td>Type 1</td>
<td>4</td>
<td>1</td>
<td>14</td>
<td>5</td>
<td>1</td>
</tr>
<tr>
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<td>18</td>
<td>0</td>
<td>0</td>
<td>1</td>
<td>0</td>
</tr>
<tr>
<td>Type 2b</td>
<td>5</td>
<td>0</td>
<td>10</td>
<td>3</td>
<td>2</td>
</tr>
<tr>
<td><strong>Sept 2003</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>Type 1</td>
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<td>0</td>
<td>3</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>Type 2a</td>
<td>0</td>
<td>4</td>
<td>0</td>
<td>0</td>
<td>0</td>
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<tr>
<td>Type 2b</td>
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<td>0</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td><strong>Jun-Jul 2004</strong></td>
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<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
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<td>53</td>
<td>12</td>
<td>0</td>
</tr>
<tr>
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<td>2</td>
<td>49</td>
<td>1</td>
<td>0</td>
</tr>
<tr>
<td>Type 2b</td>
<td>55</td>
<td>3</td>
<td>44</td>
<td>11</td>
<td>0</td>
</tr>
</tbody>
</table>

*Dual vent eruptions were counted separately for the purposes of eruption style
Figure 2.3. The main types of eruptions at Stromboli. Type 1 eruptions are spatter-dominated, with little to no visible plume due to the absence of ash-sized particles. Type 2 eruptions involve emission of an ash plume, with Type 2a containing spatter visible above the crater rim and Type 2b lacking any visible spatter. Notice the skirt of fallen spatter at the base of the Type 2a plume, and several large airborne particles marked by arrows. White boxes show corresponding photo area, and the spatial scale is identical in all three FLIR images.
Type 2 eruptions were also common in my study period. This style involves a conspicuous ash plume, which often visually overwhelmed the ballistic particles due to its high optical thickness (Fig. 2.3). I found that these eruptions can be subdivided into two groups: Type 2a, which exhibited a visible gas-thrust phase and usually significant ballistics, and Type 2b, which only exhibited convective velocities and usually lacked appreciable ballistic particles. Type 2a velocities overlapped those of Type 1 activity. Type 2b eruptions, however, were consistently smaller in size with lower velocities. While Type 1 and 2a eruptions were often very loud, Type 2b eruptions were commonly inaudible.

Visually, both Type 1 and 2 eruptions at Stromboli varied in many different ways (Fig. 2.4). Some were poorly-collimated explosions that seemed to originate close to the crater rim (Fig. 2.4a), while others were well-collimated and hence are assumed to come from much deeper levels (Fig. 2.4c). Some were simply small in size (Fig. 2.4f), while others were moderately (Fig. 2.4d) to impressively large (Fig. 2.4b). Some Type 2 eruptions contained a small (Fig. 2.4f) to moderate (Fig. 2.4b) amount of ash in their plumes, while others were visibly ash-laden and displayed significant sedimentation (Fig. 2.4c).

Accurate discrimination of these styles is difficult in visible or near-infrared imagery. As mentioned above, these imagery miss either the ash plume at night or ballistic particles during the day (at a reasonable magnification). Longwave infrared imagery responds well to both the large and warm ash plume and the small and hot ballistic
Figure 2.4. The varying styles of eruptions at Stromboli. A) Intense Type 1 eruption, with poor collimation possibly due to high level of magma in the conduit, 5 June 2004, B) Type 2a eruption with high-velocity ash plume, 7 June 2004, C) Type 2b eruption with low-velocity, particle-laden plume and high degree of visible sedimentation, 13 June 2004. Arrow points to sedimentation cloud. D) Mild Type 1 eruption with moderate collimation, 7 July 2004. E) Intense Type 1 eruption, with excellent collimation possibly due to low level of magma in conduit, 14 July 2004. F) Trivial Type 2b eruption, 18 July 2004. Spatial scale is identical in all images. All eruptions originate from NE-1 crater.
particles. When the explosions are seen against the typically cold sky temperatures, the full range of ejecta styles is easily apparent.

2.5.2 Style timescales and crater behavior

Over a long-term period (months), each crater could exhibit the full range of styles. For any particular crater, however, I observed temporal consistency in style on a short term; a given vent typically maintained a particular style on a timescale of days to weeks. For instance, in May 2002 SW crater erupted in Type 2 style for the entire three weeks while the NE subcraters alternated between Types 1 and 2. NE-1 erupted in Type 1 style on May 09 while NE-2 was erupting in Type 2 style. By May 14, this was reversed and remained so through May 18. On May 25, NE-1 had returned to Type 1 activity.

In June-July 2004, SW crater exhibited large but low-intensity Type 2 eruptions for the entire two months. NE-1 began June by erupting in Type 1 style but switched to Type 2 during June 07-27. By July 07, NE-1 had returned to Type 1 activity which lasted until July 18, after which it switched back to Type 2 eruptions.

With the exception of July 25, 2004, on only one occasion was a change in major eruption style (Type 1 vs. 2) observed on the same day for a particular crater. On July 18, 2004 at NE-1 crater, a series of 10 low-intensity Type 1 eruptions were followed, after an interval of 22 minutes, by three robust Type 2 eruptions within the next 30 minutes. Forty minutes after the last Type 2 eruption, activity reverted to the original
Type 1 style. Type 2a and 2b eruptions on the same day, however, were common, and could alternate in succession on a timescale of minutes.

2.5.3 Plume height, velocity and acceleration trends

Type 2a eruptions are defined as those having a gas thrust phase, and Type 2b eruptions as those with only buoyant velocities. Buoyant phases, evident from their lack of deceleration, had a peak velocity of 11 m s\textsuperscript{-1} among the eruptions I analyzed, and so I choose the rough lower limit for gas thrust velocities as 15 m s\textsuperscript{-1}.

Figure 2.5 shows plume height trends for 80 NE-1 crater ash plumes in 2004. The height versus time plots show the original data points, while the velocity and acceleration plots were produced from polynomial-smoothed data. Some of the anomalous sharp curves (notably, Type 2a accelerations at heights of 100-150 m) are end-point artifacts of the smoothing.

The fundamental difference between Type 2a (gas-thrust) and Type 2b (buoyant) is shown in Figure 2.5. Velocities for Type 2a eruptions could reach up to 50 m s\textsuperscript{-1} at the crater rim and decelerated (at a rate of up to 30 m s\textsuperscript{-2}) to buoyant velocities (<15 m s\textsuperscript{-1}) within 100 m above the rim. Notice that the velocity versus height slope in the gas thrust phase is roughly linear, whereas velocities remained roughly constant with height after buoyant velocities were attained. Type 2b eruptions lacked any significant deceleration phase and maintained nearly constant buoyant velocities throughout their observed lifespan (rise). Time-averaged velocities (last measured height in field of view divided
Figure 2.5. Height, velocity and acceleration for Strombolian ash plumes. Results are from NE-1 crater in June-July 2004, and separated by eruption type. Original height values are shown, while velocity and acceleration were produced from smoothed height data (fifth-order polynomial).
by time) of the 42 Type 2b profiles indicate mean buoyant velocities between 1.4 and 10.9 m s\(^{-1}\), with an average of 4.5 m s\(^{-1}\) (st. dev.: 2.1 m s\(^{-1}\)).

2.5.4 Initial plume velocity

Initial plume velocities ranged from 3 to 101 m s\(^{-1}\) (Table 2.3, Fig. 2.6a), agreeing well with ejecta velocities measured at Stromboli by Chouet et al. (1974), Blackburn et al. (1976), Weill et al. (1992), Ripepe et al. (1993), Hort and Seyfried (1998), Ripepe et al. (2001), and Hort et al. (2003). The overall mean velocity was 24 m s\(^{-1}\), but this is better examined by eruption style. Type 1 eruptions had a mean velocity of 34 m s\(^{-1}\) (range 9-101 m s\(^{-1}\)), while Type 2 eruptions had a mean of 19 m s\(^{-1}\) (range 3-58 m s\(^{-1}\)). The Type 1 velocity distribution is unimodal (20-30 m s\(^{-1}\)) and more normal than the Type 2 distribution, which is bimodal (0-10 and 20-30 m s\(^{-1}\)) and highly skewed towards lower velocities, due to the differences between Type 2a and Type 2b velocities (Fig. 2.6b). The mean Type 2a velocity is 31 m s\(^{-1}\) (range 16-58 m s\(^{-1}\)), close to the mean Type 1 velocity, however the mean Type 2b velocity is 7 m s\(^{-1}\) (range 3-11 m s\(^{-1}\)). The two Type 2 velocity modes, therefore, reflect the Type 2a and Type 2b means (Fig. 2.6b).

The trend of initial plume velocity with time, for NE-1 crater in 2004, is shown in Figure 2.7a. Although my data are not continuous (~4 hours of data on each observation day), the consistency of eruption style (Type 1 or 2) indicates a persistence on a timescale of days to weeks, as mentioned above (section 2.5.2). Note the general overlap of Type 1 and 2 velocities, with the exception that Type 2b velocities were lower than the majority of the lowest Type 1 velocities. Also note that the transition to Type 2 style on June 07
### Table 2.3. Explosion parameters for June-July 2004, NE-1 crater

<table>
<thead>
<tr>
<th>Overall</th>
<th>Count</th>
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<th>Max</th>
<th>Mean</th>
<th>Stdev</th>
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<td>3</td>
<td>101</td>
<td>24</td>
<td>19</td>
</tr>
<tr>
<td>Maximum ballistic height, m</td>
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<td>0</td>
<td>115* (67)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Eruption duration, s</td>
<td>135</td>
<td>6</td>
<td>41</td>
<td>15</td>
<td>6</td>
</tr>
<tr>
<td><strong>Type 1</strong></td>
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<td></td>
<td></td>
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<td></td>
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<tr>
<td>Initial velocity, m/s</td>
<td>47</td>
<td>9</td>
<td>101</td>
<td>34</td>
<td>21</td>
</tr>
<tr>
<td>Maximum ballistic height, m</td>
<td>53</td>
<td>37</td>
<td>115* (33)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Eruption duration, s</td>
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<td>6</td>
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<td>5</td>
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<td></td>
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<td>58</td>
<td>19</td>
<td>15</td>
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<tr>
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<td>0</td>
<td>115* (34)</td>
<td></td>
<td></td>
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<tr>
<td>Eruption duration, s</td>
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<td>6</td>
<td>41</td>
<td>15</td>
<td>6</td>
</tr>
<tr>
<td><strong>Type 2a</strong></td>
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<td></td>
<td></td>
<td></td>
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<tr>
<td>Initial velocity, m/s</td>
<td>44</td>
<td>16</td>
<td>58</td>
<td>31</td>
<td>11</td>
</tr>
<tr>
<td>Maximum ballistic height, m</td>
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<td>42</td>
<td>115* (34)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Eruption duration, s</td>
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<td>6</td>
<td>41</td>
<td>15</td>
<td>6</td>
</tr>
<tr>
<td><strong>Type 2b</strong></td>
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<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Initial velocity, m/s</td>
<td>43</td>
<td>3</td>
<td>11</td>
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<tr>
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<td>96</td>
<td>16</td>
<td>24</td>
</tr>
<tr>
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<td>40</td>
<td>8</td>
<td>31</td>
<td>15</td>
<td>6</td>
</tr>
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</table>

* Ejecta extended above the minimum field of view, 115 m, in (X) eruptions
Figure 2.6a. Initial velocity, duration and ballistic height results (Type 1 vs. 2). Histograms showing parameters for NE-1 crater eruptions captured by the FLIR during the June-July 2004 field season.
Figure 2.6b. Initial velocity, duration and ballistic height results (Type 2a vs. 2b). Histograms showing parameters for NE-1 crater eruptions captured by the FLIR during the June-July 2004 field season.
Figure 2.7a. Initial velocity through time. Observed initial velocity (at the crater rim) for 134 NE-1 crater eruptions captured by the FLIR during the June-July 2004 field season.
Figure 2.7b. Duration and ballistic height through time. Eruption duration (134 eruptions, top) and maximum ballistic height above the crater rim (138 eruptions, below) for eruptions at NE-1 crater in June-July 2004.
remained in a relatively high velocity regime, whereas the pronounced drop in velocities between July 14 and July 18 was not marked by an immediate transition in style.

The three Type 2b eruptions of July 18 are explained in section 2.5.2 as a real and conspicuous shift in style. Conversely, the one Type 1 eruption on July 25 represents only a limitation of my visual approach to discriminating eruption styles, as the July 25 eruptions appeared to be transitioning between Type 2 and 1 eruption styles. Thus, the July 25 eruptions were essentially the same eruption style, but so close to my visual threshold for Type 1 vs. Type 2 that one eruption was considered Type 1.

2.5.5 Maximum ballistic height

Table 2.3 and Figure 2.6a show that a large percentage (62%) of the Type 1 ballistics exceeded the minimum field of view (115 m height), while a smaller proportion (40%) of Type 2 ballistics did the same. Type 2a eruptions account for all of the Type 2 eruptions which exceeded 115 m (Fig. 2.6b). As with velocity, the Type 2 height histogram is bimodal with Type 2b eruptions representing the bulk of the lowest mode (0-10 m) and high energy Type 2a eruptions comprising the high mode (>115 m).

The daily trend of maximum ballistic height shows that ballistic height simply reflects initial velocity recast on a scale that is limited by the field of view (Fig. 2.7b). Initial velocities greater than ~20 m s\(^{-1}\), which account for a major portion of my eruptions, generally cast particles above the 115 m height.
2.5.6 Eruption duration

Eruptions had an overall mean duration of 15 s (range 6-41 s), and eruption duration was similar among the different styles. Type 1 eruptions had a mean duration of 14 s (range 6-27 s), while Type 2 eruptions had a mean of 15 s (range 6-41 s) (Fig. 2.6a). Type 2a eruptions had a mean duration of 15 s (range 6-41 s) and Type 2b had a mean value of 15 s (range 8-31 s) (Fig. 2.6b). The daily trend of eruption durations for NE-1 crater in 2004 shows no significant variation in values for Type 1 or 2 eruptions (Fig. 2.7b), or for Type 2a versus 2b eruptions.

2.6 Discussion

2.6.1 Model for strombolian eruption styles

2.6.1.1 Origin of Type 1 and 2 styles

The Type 1 eruption style follows the common perception of strombolian eruptions: an ejection of coarse, incandescent scoriae following roughly ballistic trajectories. These molten clots have been explained as the fluid lava forming the bubble wall (Walker, 1973; Blackburn et al., 1976). Type 2 behavior, in which a fine particle plume dominates, cannot be explained as such. In ‘dry’ eruptions (i.e. lacking external water), the production of fine particles (ash, <2 mm) is an internal process resulting from vesiculation reaching a critical volume fraction at which point thinning bubble walls break and magma is ripped apart. Ash would represent the countless bubble wall
fragments of the magmatic foam (Cashman et al., 2000). In the slug flow geometry ascribed to Stromboli, however, a single large slug or series of large slugs are bursting, necessarily limiting the amount of bubble-wall material and, therefore, the capacity to produce primary ash. Strombolian deposits from Heimay in 1973, for instance, showed that a very small proportion of magma was fragmented to <1 mm in size (Self et al., 1974). Walker (1973) commented on the general dearth of fine particles (<1 mm) in over 100 strombolian deposits analyzed. Typical strombolian activity, therefore, should not produce large amounts of ash-sized particles. In 'wet' eruptions, fine fragmentation can be effected by the high explosive energy of the water-melt interaction (Wohletz, 1986).

As I have shown, Type 2 eruptions exhibit typical strombolian ejecta velocities (Fig. 2.6a, 2.7a; see also Chouet et al., 1974; Blackburn et al., 1976; Weill et al., 1992; Ripepe et al., 1993; Hort and Seyfried, 1998; Ripepe et al., 2001; and Hort et al., 2003), indicating that the ash is not produced by anomalously explosive activity. Instead, the roughly similar velocities suggest that the same normal strombolian eruption mechanism is in place during Type 2 phases, but modified to the extent that fine particles can be involved.

What produces the ash in Type 2 eruptions? Several studies have observed anomalous ash-rich behavior during strombolian phases to be linked to loose material sitting atop the magma column (Murata et al., 1966; Booth and Walker, 1973; Self et al., 1974; Guest, 1982; Houghton and Schmincke, 1989). In this case, an otherwise normal strombolian eruption passes through and/or displaces a layer of loose material and mobilizes fine particles which already exist in the layer. Backfill can be produced
through two means. First, discrete collapses may occur which maintain the crater’s inner angle of repose and shift material onto the vent. Second, ejecta may fall within the rim of the cinder cone, roll down, and accumulate on top of the vent. The former mechanism would likely be accelerated during high activity periods (and high cinder cone growth rates), while the latter mechanism would be encouraged during low activity periods when clast dispersal is limited.

This backfill mechanism was observed by Murata et al. (1966) to produce ash-rich (i.e. Type 2) eruptions interrupting normal (i.e. Type 1) strombolian activity at Irazu volcano in 1963-64. With clasts falling back into the crater, and occasional vent wall collapses, they viewed the active conduit as “a veritable rock crusher and dust-making machine”. Booth and Walker (1973) noted a similar mechanism during the formation of Etna’s Southeast crater in 1971, causing a transition from typical strombolian explosions to ash-rich plumes. As they noted, “The style of eruption changed from one dominated by the escape of gases through relatively fluid lava to one dominated by the passage of gases through loose solid debris”. It was also recognized as the mechanism for producing occasional “dense black clouds of much finer grained ash” during the 1973 strombolian eruptions at Heimay (Self et al., 1974). Guest (1982) noted how collapse events preceded ash-rich eruptions during the growth of Northeast crater on Etna in 1976. From a depositional perspective, recycled backfill was thought to cause the fine-grained and poorly-sorted beds within the excavated Rothenberg scoria cone in the Pleistocene East Eifel Volcanic Field (Houghton and Schmincke, 1989). Supporting this association, inner wall strombolian facies at Rothenberg preserve evidence of clasts tumbling and
avalanching into the crater. Similar recycling mechanisms have been proposed in modifying hawaiian (Wilson et al., 1995) and surtseyan (Kokelaar, 1983) eruption styles as well.

In the backfill scenario the production of fine particles (ash, submillimeter-scale) is the result of secondary fragmentation of originally coarse brittle clasts (lapilli-bombs, cm-scale), through either a passive or active fragmentation mechanism. In passive fragmentation, the ash could be produced by breakage and grinding during collapse events or the ball-mill action (Francis, 1993) of loose material sitting atop a perturbed magma surface. At Stromboli, the constant gas flux through the conduit (Ripepe and Gordeev, 1999) would continually disrupt the magma free-surface, and the magma column would be expected to accumulate a surface covering of auto-brecciated fragments and fine particles. In active fragmentation, ash could be produced from the fracturing of loose, brittle material sitting atop the magma column during the explosion itself.

Several particular events indicated a backfill control on eruptive style in this dataset. In May 2001, eruptions at SW crater were simultaneously exiting from two adjacent vents, separated by ~20 m (Fig. 2.8). Eruptions imaged on May 18+19 show the south vent (left) erupting in Type 1 style while the north vent (right) erupted in Type 2b style. The next day, May 20, the north vent erupted in ash-free Type 1 style. Their coordinated eruptions would suggest these two vents were joined at a very shallow depth, and drew from the same major conduit — eliminating magma rheology as an explanation for the differing styles. Another instance that likely reflected a backfill control on eruption style occurred on 18 July 2004, at NE-1 crater. On that day, ten eruptions imaged over a three
Figure 2.8. FLIR evidence for backfilling. May 2001 saw several eruptions at SW crater which simultaneously exited adjacent vents, presumably joined at a shallow depth. A) On May 18 and 19, the north vent (right) erupted in Type 2 style. B) On May 20, this vent changed to Type 1 style. This likely reflects ash or scoria falling in the vent for a short time before being cleared. On both occasions the south vent (left) erupted in Type 1 style.
hour period were very small Type 1 style eruptions. Twenty-two minutes after the last Type 1 eruption, an ash-rich, low-velocity Type 2b eruption appeared from the same vent, followed by two other Type 2b eruptions over the next 30 minutes. Forty minutes after the last Type 2b eruption, NE-I reverted to typical Type 1 style. Again, this short-term Type 2 behavior argues against a rheological change and likely reflected a small-scale collapse of material into the conduit that was rapidly cleared. More prolonged Type 2 phases (e.g. NE crater in June 2004, Fig. 2.7) would be produced from 1) large volume collapses or progressive slumping of the inner walls or 2) accumulation of ejecta over the vent during extended low intensity periods.

I can discount the influence of weather on the transitions observed during this period. The weather was generally stable during the two-month observation period. Rain was recorded only on June 17, and did not occur with any clear changes in activity, although there was a gap in weather data during 01-16 July. The fact that significant changes in eruptive styles were observed in my dataset in the absence of notable weather fluctuations may argue against the weather-related influences suggested by Urbanski et al. (2002) and Hort et al. (2003) for activity at Stromboli.

2.6.1.2 Origin of Type 2a and 2b styles

Type 2a eruptions involve a gas-thrust phase as well as many ballistic particles, whereas Type 2b eruptions involve only buoyant velocities and few, if any, ballistics. Given only the height vs. time profiles in Figure 2.5, it is tempting to think that Type 2b eruptions are simply Type 2a eruptions observed higher above the magma-air interface
due to a drop in magma level. In this reasoning, Type 2b have simply hidden their gas thrust phase and ballistic particles below the crater rim. This is clearly not the case, however, as Type 2a and 2b eruptions were interlaced in time on a scale of seconds to minutes, much shorter than reasonable magma level change timescales. Supporting this, seismic energy for 75 of the 80 Type 2 eruptions shown in Fig. 2.5 indicates that Type 2a eruptions involve more initial energy (Fig. 2.9, courtesy of M. Ripepe at University of Florence), which can be linked to bubble overpressure (Ripepe and Marchetti, 2002). The high abundance of ballistic particles coincident with gas-thrust velocities in Type 2a eruptions (and lack of both in Type 2b eruptions), therefore, is consistent as both ballistic particles and the ash plume rely on the bubble overpressure for their initial velocities.

If Type 2 eruptions represent normal strombolian eruptions with the addition of loose backfill on top of the conduit, then Type 1 and 2 eruptions should have roughly similar distributions of bubble overpressures, or by proxy, initial velocities. Type 1 and Type 2a distributions largely overlap (with the exception of velocities >50 m s\(^{-1}\)) and have similar mean velocities (Type 1 mean: 34 m s\(^{-1}\), Type 2a mean: 31 m s\(^{-1}\)). Type 2b eruptions, however, appear to form a population distinct from Type 2a eruptions. If the Type 2a/2b difference is a function of bubble overpressure only, why do Type 2b eruptions form a distinct mode in the velocity and ballistic height distributions (Fig. 2.6)? I speculate that the distinction may be caused by the loose material imposing a threshold force on the overpressure-velocity relation. The threshold must be related to the downward force of the overfill relative to the overpressure and size of the gas slug. If bubble overpressure is greater than some threshold, the loose material is disrupted on a
Figure 2.9. Proxy values for seismic energy for Type 2a and 2b eruptions. Type 2a eruptions involved greater seismic energy, and thus bubble overpressure, than Type 2b eruptions.
significant scale and a Type 2a eruption is formed. Incorporation of fine material into the plume, by way of conservation of momentum, would decrease the plume velocities somewhat – possibly explaining the lack of significant Type 2a velocities >50 m s⁻¹. If bubble overpressure is lower than the threshold, the loose material absorbs much of the force of the explosion without significant displacement (again, through conservation of momentum), and gas is relegated to streaming through the backfill layer. In this case, a low-velocity Type 2b plume is formed.

2.6.1.3 The 2001-2004 dataset

Anecdotal evidence, gathered from those who have worked on the volcano for years to decades (e.g. M. Ripepe & A. Harris, personal communications, 2002-2004), indicates that Type 1 activity is the most common eruptive style at Stromboli. Type 2 activity is generally rare, and most commonly associated with SW crater. These observations are in contrast to my dataset of 344 eruptions, in which 64% of the total number were Type 2 style, and of which a significant number (122) originated from NE crater.

Why is there a disproportionate number of Type 2 eruptions in my dataset? The two main datasets, May 2002 and June-July 2004, can be explained separately. In May 2002 activity was at a remarkably low intensity, and 61% of the 64 eruptions were Type 2. Reduced explosivity would limit the dispersal of ballistic clasts and result in more material falling within the crater walls and accumulating on top of the vent. This would encourage backfill establishment and Type 2 eruptions.
The June-July 2004 activity was generally moderate in intensity, with 71% of the 240 eruptions being Type 2. Following the December 2002 crater terrace collapse, and beginning with the return of normal Strombolian activity at the end of the 2002-2003 effusive phase, the craters have been in a prolonged constructional phase. NE-1 crater, in particular, developed a nascent cinder cone between June 2003 and June 2004 (Fig. 2.10). It is likely that new material building the developing cinder cones was unstable and prone to collapse events, shifting more loose backfill onto the vent. Aggravating their instability, the cinder cones are nested within the indurated outer structure of the crater terrace. A normal cinder cone, emplaced on a nominally flat surface, places most of its ejecta in the outer walls of the cone (Fig. 2.11). At Stromboli, and clearly visible at NE-1 crater, the outer walls of the cinder cone are truncated by the outer structure. This would result in accelerated upward growth compared to a normal cinder cone (Fig. 2.11) and increased collapse frequency. Furthermore, that the NE-1 crater walls are not uniformly buttressed by the outer structure (the cinder cone extends through a breach of the crater terrace, Fig. 2.10) would further destabilize the inner walls and encourage more frequent collapse events.

2.6.1.4 Model synthesis

In summary, my model involves backfill on the magma column as a major control on eruption styles at Stromboli, and by extension, other volcanoes which exhibit strombolian activity (Fig. 12). During normal strombolian activity, the magma-air free surface is unobstructed. When a bubble reaches the top of the magma column and bursts, the fluid
Figure 2.10. Cone construction in NE crater. Following the end of the 2002-2003 effusive phase (July 2003), the collapsed crater terrace (top) began rebuilding itself. By June 2004, a nascent cinder cone had developed in NE-1 crater. Top photo courtesy of Elske van Dalffen.
Figure 2.11. Uniqueness of NE-1 crater. The cinder cone at NE-1 crater, truncated by the indurated outer structure, would exhibit greater upward growth rates compared to a typical cinder cone emplaced on a nominally flat surface. Higher growth rates would lead to more frequent collapses, allowing backfill accumulation and encouraging ash-rich (Type 2) eruptions.
Figure 2.12. Scenarios to explain observed differences in eruption styles. Type 1: gas slug bursts at free surface with no overlying brittle material. Type 2a: slug burst produces large scale disruption of brittle overfill, producing ash and ballistics. Type 2b: slug burst results in minor disruption of overlying brittle material, creating ash but few ballistics.
bubble wall is torn apart creating coarse, molten ballistic clots. The gas ejected from the bubble remains relatively free of fine particles and is thus largely transmissive in the longwave infrared and to the naked eye. The result is a Type 1 eruption, which is probably the most common activity style at Stromboli (Fig. 12).

In some instances, loose material may exist on top of the magma column, sourced from either collapses of the cinder cone or direct accumulation of ejecta. Some of this material is fragmented to a fine scale (e.g. <1 mm) due to secondary fragmentation mechanisms (impacting, rolling, ball-milling, etc.). In these cases, the explosion pushes through the loose material and the gas plume entrains fine particles to create an optically-thick ash plume, and Type 2 eruption.

During backfill phases, higher energy eruptions significantly disrupt the backfill and cast up coarse ballistic particles, which could be primary fluid portions of the bubble wall or coarse clasts from the backfill. The high energy is also responsible for gas-thrust velocities of the ash plume, producing a Type 2a eruption (Fig. 12).

Lower energy eruptions offer insufficient force to disrupt the backfill on a large scale. Gas is ejected at a low velocity, and may even stream through a largely in-place backfill layer. The low explosive energy is also insufficient to throw coarse ballistic particles to any great height. The backfill itself may offer resistance to the meager force of the explosion, in effect muffling whatever excess pressure existed in the bursting bubble, further ensuring low initial velocities of both the gas plume and coarse particles. In this case, a Type 2b eruption is formed (Fig. 12).
2.6.2 Implications for classification and modeling

2.6.2.1 Strombolian classification

This work explores the variability in eruption styles at Stromboli, aided both by my comparatively large dataset as well as the improved ability of longwave infrared imagery to visualize a wider range of ejecta types. The Type 1/Type 2 classification was introduced by Chouet et al. (1999), and used to explain different seismic waveforms. Their classification scheme, however, is obfuscated by the fact that two largely independent factors (eruption style and crater) are used to characterize the two styles; they define Type 1 eruptions as those being ballistic-dominated and originating from the NE crater, and Type 2 eruptions as those being ash-dominated and originating from SW crater. It is therefore impossible to discern which factor (style or crater) is responsible for the unique waveform features.

The prior lack of a properly illustrated classification has made it difficult to appreciate that two seminal works on strombolian activity, Chouet et al. (1974) and Blackburn et al. (1976), were analyzing two very different eruption styles. Chouet et al. (1974) describe ‘molten lava fragments’ and ‘small quantities of ash’, suggestive of Type 1 activity, while Blackburn et al. (1976) mention the ‘convective cloud of gas and small pyroclasts’ consistent with Type 2 style. These are often cited interchangeably in the literature, when it should be clear that each was studying a distinct aspect of strombolian eruptive behavior.
If Type 2 eruptions involve significant amounts of non-juvenile material (i.e. recycled backfill material), they may present a semantic dilemma as they could overlap with overly loose definitions of vulcanian eruptions. Vulcanian eruptions are often casually characterized as discrete, intermittent explosions which involve a large non-juvenile component (e.g. Wilson and Self, 1980). Type 2 strombolian eruptions might also qualify under this broad definition, however, even though the source mechanism for Type 2 eruptions is no different than that of more 'normal' Type 1 strombolian eruptions. In the field, it is tempting to explain intermittent, ash-rich plumes which appear to lack incandescent ballistics as being related to a vulcanian eruption mechanism. Lacking more descriptive information, however, they might easily be representative of a Type 2 strombolian regime.

For instance, Wilson and Self (1980) describe eruption clouds at Fuego volcano, Guatemala, in February 1978 (Smithsonian Institution, 1978) as being vulcanian. The extremely low velocities (most had maximum observed velocities <12 m s⁻¹), low heights (100-500 m), and strombolian-like frequency (3-15 per hour), suggest these may actually have represented a Type 2 strombolian regime which followed the more violent, and almost certainly vulcanian, eruptions of January 1978 (Smithsonian Institution, 1978). This may appear to be a minor semantic argument but it is significant to the extent that a particular eruption mechanism is inferred from the ascribed name. Therefore, it is important to make the distinction that vulcanian eruptions typically involve 1) a high proportion of dense lithics as the non-juvenile component, 2) larger ejecta mass values, around 10²-10⁶ metric tons (Self et al., 1974), compared to Type 2 Strombolian eruptions.
which are probably $\sim 10^6$-$10^7$ metric tons, 3) higher pressures, $\sim 10^0$-$10^2$ bars (Wilson, 1980), compared with Strombolian events which are 0.2-4 bars (Ripepe and Marchetti, 2002), and 4) a propensity for ash column collapse due to the entrainment of cold lithic material. Column collapse was never observed for Type 2 eruptions at Stromboli, likely because the ash is well heated from sitting on or near the magma column.

My results also challenge the common physical volcanology maxim that higher fragmentation necessarily results from more violent explosive activity. For example, the rigorous analysis of cone-building strombolian deposits at numerous volcanoes by Walker (1973) identified minor components of fine fragmentation which were all assumed to be related to 'violent' strombolian behavior. The average buoyant velocities I observed (1.4-10.9 m s$^{-1}$) in my dataset equate to terminal velocities of $\sim 0.1$-$6.0$ mm (using a drag coefficient of 1 and density of 1500 kg m$^{-3}$), suggesting there is a significant amount of material fragmented to less than 1 mm in many of these plumes. As I have shown, this fine fragmentation is produced during periods of normal explosive intensity (Fig. 2.6a+2.7a). My results suggest that, in general, some fine proximal strombolian deposits could be related to Type 2 phases of 'normal' strombolian activity.

2.6.2.2 Insights on modeling

The initial ejecta velocities measured here, along with those from other studies, do not agree with the predicted velocities of Wilson (1980) for strombolian eruptions. My velocity results ranged from 3 to 101 m s$^{-1}$, with a mean of 24 m s$^{-1}$. Velocities measured by Chouet et al. (1974), McGetchin et al. (1974), Blackburn et al. (1976), Weill et al.
(1992), Ripepe et al. (1993), Hort and Seyfried (1998), Ripepe et al. (2001), and Hort et al. (2003) are all <200 m s\(^{-1}\), with most in the general range of 20-50 m s\(^{-1}\). The adiabatic expansion model of Wilson (1980) predicts velocities of 230-990 m s\(^{-1}\) for gas mass fractions of 0.2-0.95 (Blackburn et al., 1976; Chouet et al., 1974) using typical gas overpressures of 0.5 \times 10^5 to 4 \times 10^5 Pa (Ripepe and Marchetti, 2002), indicating that this model does not reproduce realistic velocities for Stromboli. The modified Bernoulli equation dismissed by Wilson (1980) produces velocities (10-25 m s\(^{-1}\)) more consistent with those observed, though it cannot predict the higher velocities (>50 m s\(^{-1}\)) that have been measured. The method of Blackburn et al. (1976) uses an approach similar to that in Wilson (1980) in relating gas overpressure and velocity, and is therefore flawed for the same reasons. This explains why Blackburn et al. (1976) predicted unrealistically low gas overpressures (200-3000 Pa, much lower than those of Ripepe and Marchetti, 2002, mentioned above) for very typical Strombolian ejecta velocities (28-65 m s\(^{-1}\)). Clearly, more research into the relationship between initial pressure, gas content and ejecta velocity is needed for strombolian scenarios.

This study shows how many of the eruptions originate from vents which may be covered by significant amounts of backfill. The model of Vergniolle and Brandeis (1994; 1996) and Vergniolle et al. (1996) essentially requires that the free-surface be clear, in that gas overpressure and volume are calculated using an estimated bubble film thickness (several centimeters). With backfill covering the emergent bubble, their results might be questionable in Type 2 phases. Therefore, caution must be taken when applying their
model universally (e.g. Vergniolle et al., 2004) if one cannot be certain that a clear interface is present.

2.7 Conclusions

In this chapter I imaged 344 eruptions at Stromboli volcano, Italy, with a FLIR thermal video camera. These data provide a new and powerful means to unravel the dynamics of explosive events. The specific insights provided by this work are summarized as follows:

1) Eruptions at Stromboli exhibit a wide range of styles, and explosions can be dominated by ballistic particles (Type 1) or ash (Type 2). Type 2 eruptions can be subdivided into Type 2a style, which exhibits a visible gas-thrust phase (>15 m s\(^{-1}\)) and significant ballistic particles, and Type 2b style which shows only convective velocities (<15 m s\(^{-1}\)) and rarely many ballistics.

2) Type 1 eruptions reflect an unobstructed magma-air interface, while Type 2 eruptions likely indicate backfill material residing in the vent. Type 2a and 2b eruptions are caused by varying levels of bubble overpressure. Changes in eruption style were not caused by hydrothermal interaction or weather fluctuations in my observation period.

3) Individual vents at Stromboli maintain Type 1 and Type 2 phases on a timescale of days to weeks, due to cycles of backfill accumulation, residence and clearance.
on that timescale. Type 2a and 2b eruptions could appear within minutes of each other, because bubble overpressure can vary significantly on that timescale.

4) This dataset covers an atypical period at Stromboli when a relatively high proportion of Type 2 eruptions were evident, a result of low activity levels in May 2002 and crater terrace construction in June-July 2004.

5) Some fine-grained proximal strombolian deposits may result from normal intensity Type 2 strombolian activity, not necessarily more explosive, or ‘violent’, strombolian activity.

6) Intermittent ash-rich eruptions observed elsewhere are commonly classified as vulcanian, but may in fact reflect Type 2 strombolian activity.

Previous FLIR studies have largely focused on effusive volcanic activity (e.g. Wright and Flynn, 2003; Calvari et al., 2005). This paper marks one of the first rigorous studies of explosive activity with the FLIR camera and, overall, the FLIR provides a unique view of explosive dynamics. We have shown how FLIR offers an improved view of different ejecta components, compared to visible or near-infrared imagery. The major shortcomings of the FLIR include the limited field of view from a reasonable distance and the very coarse pixel size. The limited field of view was unable to capture the entirety of ballistic particles throughout their rise for many eruptions. The pixel size limitation results in the FLIR being capable of observing, but not measuring, ballistic particles. A 45° FOV could improve the area of coverage, though it would produce greater distortions in apparent positions and increase the pixel size further. Future work
coupling a high frame-rate conventional camera (as in Chouet et al., 1974) to image ballistic particles, with a FLIR camera to measure the ash plume, could improve upon this study.
Chapter 3

Smallscale ash plume thermodynamics from thermal (FLIR) video

3.1 Introduction

Imaging eruption plumes is an essential component of understanding plume dynamics and their hazards, which includes air fall and derivative pyroclastic flows. Previous imaging studies include those at Mt. St. Helens (Sparks, 1986; Sparks et al., 1986; Calder et al., 1997), Redoubt (Woods and Kienle, 1994), Soufrière (Sparks and Wilson, 1982), Soufrière Hills (Clarke et al., 2002; Formenti et al., 2003), Fuego (Wilson and Self, 1980) and Stromboli (Blackburn et al., 1976) volcanoes. Much of this photography, however, was serendipitous – overall, relatively few eruption plumes have been imaged with the original intention of making scientific measurements (Sparks et al., 1997). Furthermore, very few of these have been acquired at a frame rate sufficient to capture dynamic processes in the plume (Clarke et al., 2002; Formenti et al., 2003). In particular, emergent plume behavior and small-scale motions have generally eluded a thorough description in the literature. Another major limitation in our understanding of plume behavior is temperature, due to the use of visible wavelength imagery in all previous studies. This has required using observed plume motions to infer density, and through this, temperature (e.g. Sparks and Wilson, 1982).
New technology, specifically the handheld FLIR (Forward Looking Infrared Radiometer) camera, now allows direct measurement of temperature and motions at a high frame rate. In this study I utilize a FLIR to measure, map and track ash plume dimensions and temperatures at Stromboli volcano. The FLIR data were acquired at a rate of 30 Hz (30 frames per second), allowing me to characterize the plume at a fine timescale and observe small-scale dynamic behavior. Using a dataset of 344 Strombolian eruptions acquired over four years (2001-2004), I determined in Chapter 2 that eruptions at Stromboli can be either ballistic-dominated (Type 1) or ash-dominated (Type 2). This work takes 25 ‘ideal’ Type 2 eruptions from that dataset, all acquired in June 2004, and performs a detailed examination of the ash plume dynamics and temperatures. A further 68 plumes from the same dataset were used for qualitative assessments.

3.2 Background

Stromboli volcano has been experiencing small, intermittent explosions for many hundreds of years (Washington, 1917; Rosi et al., 2000). These originate from several major (Southwest, SW; Central, C; and Northeast, NE) and minor (e.g. NE-1, NE-2) craters within a 300 m long crater terrace near the summit (Fig. 3.1). The popular perception of Strombolian eruptions is often limited to fountains of incandescent ballistic ejecta, with an absence of significant ash-sized particles (e.g. Chouet et al., 1974). In Chapter 2, however, I showed that ash-rich eruptions were quite common among my dataset of 344 eruptions and, in fact, many eruptions involved ash-sized particles exclusively. Eruptions were either ballistic-rich, with no significant ash plume (Type 1)
Figure 3.1. FLIR setup at Stromboli. The FLIR was operated at the Rocette station (ROC), ~450 m from the NE crater. SW crater was mostly obscured. ROC is at an elevation of ~750 m. Inset: The FLIR camera acquired data at 30 Hz, by connecting to a notebook computer in the orange case.
or ash-rich, with a significant plume (Type 2). Type 2 eruptions were further subdivided into ash-rich eruptions with a distinct gas-thrust phase (>15 m s\(^{-1}\)) and typically many ballistics (Type 2a) and those characterized by only convective velocities (<15 m s\(^{-1}\)) and usually very few, if any, ballistics (Type 2b). Ash plumes would generally rise to a few hundred meters height before drifting away or dissipating.

In June-July, 2004, 240 Strombolian eruptions were imaged with a FLIR camera (Fig. 3.1), and during that time NE crater exhibited the highest activity levels and most variability in eruption style. The main vent, NE-1, exhibited a sustained period of Type 2 eruptions from approximately June 07 to at least June 27, followed by another Type 2 period during July 20-25.

A number of plume forms were observed in this study and should be explained (Fig. 3.2). A ‘jet’ is a high velocity plume driven by initial momentum (Morton, 1959; Crapper, 1977), while a ‘discrete thermal’ is a detached fluid vortex powered solely by buoyancy (Turner, 1962). The total above-background heat content of a discrete thermal remains constant because the thermal is detached from any source and no further heat is introduced. A ‘starting plume’ is a buoyant plume which is topped by a kind of thermal (i.e. the thermal forms the plume front of the starting plume). This thermal is fed continuously by the steady plume, or what I call here a feeder plume. The starting plume does not, however, develop a vigorous entrainment vortex at the base of its thermal, this conclusion being based upon photos of laboratory starting plumes in Turner (1962). I observed another common plume form in my dataset, that of a starting plume whose thermal exhibits a strong ring vortex at its base. I term these ‘rooted thermals’, which are
Jet Starting plume Rooted thermal

\[ \text{Discrete thermal} \]

Increasing time and height
Decreasing velocity

Figure 3.2. Some different plume forms. Partially adapted from Turner (1979). The virtual point source underneath the crater is not shown. We propose the ‘rooted thermal’ form as an intermediate.
distinct from discrete thermals in that their heat content increases with time, and yet they are different from a starting plume in that they display a well-developed ring vortex.

These morphological distinctions are important for understanding plume dynamics. Of prime importance is the relation between plume form and the air entrainment coefficient, which controls the rate of air influx into the plume. Air influx generally plays a major role in determining temperature and bulk density (Woods, 1988) and, therefore, collapse potential (Sparks and Wilson, 1976). According to laboratory measurements, a jet involves an air entrainment coefficient of 0.065 (Turner, 1979; Sparks et al., 1997), a starting plume has an entrainment coefficient of 0.09 (Turner, 1962), and a discrete thermal has a value of 0.25 (Morton et al., 1956). The jet value is low because it entrains air in many small disorganized eddies along its exterior, while the discrete thermal value is high due to the large, organized ring vortex at its base (Sparks et al., 1997). Spreading angles, which are directly linked to the air entrainment coefficient (Turner, 1979), are also different for the plume types. Starting plumes exhibit a spreading half-angle of 10° whereas discrete thermals, due to their greater air entrainment rates, spread at a half-angle of 17° (Sparks et al., 1997).

My basic model for the plumes I observed involves two parts: the plume front and the feeder (or steady) plume (Fig. 3.3). The plume front forms the leading portion of the plume. In jet morphologies, the plume front is simply the uppermost head of the feeder plume. In starting plumes and rooted thermals, the plume front is a discrete circulating
Figure 3.3. Examples of Type 2 Strombolian plumes. A) A high velocity 'jet', B) a 'starting plume' and C) what we call a 'rooted thermal'. Quantitative parameters were calculated by measuring the dimensions and temperature of the plume front, which was continually fed by the feeder plume.
mass fed by the feeder plume. The plume front is typically thermally and morphologically distinct from the feeder plume in these cases.

3.3 Equipment and measurement setup

A thorough description of the basic methods and setup of the experiment are described in Chapter 2, and a short summary of that methodology is provided here. The FLIR measures longwave radiation (7.5-13 \text{ \mu m}), producing a calibrated thermal image of 240 x 320 pixels with a field of view of 18° x 24° (instantaneous field of view is 1.3 milliradians), and is capable of measuring temperatures from −40 to 1200 °C. At the observation distance in this study (~450 m), the pixel size is ~60 cm and field of view is 145 m x 190 m. Several models were used in the original experiment (Chapter 2), but only data from the ThermaCAM™ S40 model were used in this particular study. The S40 model acquires at 30 Hz via a notebook computer, and thus provides characterization of the plume dynamics at an extremely fine timescale.

The data used in this study originate from the June-July 2004 field season at Stromboli, which covered both the SW and NE craters from the Rocette site, approximately 450 m from the NE crater (Fig. 3.1). The vertical field of view of the FLIR was 145 or 190 m, depending on camera orientation (horizontal or vertical). The SW crater was partially obscured in this setup, so quantitative eruption characteristics could only be measured for the NE crater. In addition, during June-July 2004 eruptions from SW crater were consistently sluggish Type 2 eruptions that did not lend themselves
well to my analysis methods (i.e. there was no well-defined plume front). Therefore no SW crater plumes were subjected to rigorous analysis. Of the 240 eruptions imaged in 2004, 146 were from the NE crater, but 53 of these were Type 1 eruptions (which lack a robust ash plume), leaving 93 eruptions with a conspicuous ash plume. Of these plumes, many involved multiple simultaneous or oblique bursts which produced a plume front too complicated for simple analysis. Also, many were very small in size and dissipated in a matter of a few seconds. Therefore, I concentrated my quantitative analyses on 25 significantly-sized NE crater eruptions which exhibited a relatively simple, well-defined plume front. All of these selected eruptions were from the NE-1 sub-crater.

3.4 Methods of analysis

3.4.1 Plume front tracking

Most of the quantitative methods I outline here are applied to the plume front, which forms the leading portion of the rising plume (Fig. 3.3), as opposed to the entire plume. This was necessary because the plume front is the only coherent portion of the plume, allowing it to be tracked through the course of the eruption and throughout the evolution of the plume. During the analysis the remainder of the plume - i.e. the feeder plume - was thus continuously emerging from the vent at its base while being incorporated into the plume front at its top.

The basis for the quantitative analyses are plots of the plume front dimensions and temperatures with time/height. This was performed using the ThermaCAM™ Researcher
software. The extracted parameters include the position of the leading edge (top) of the plume front ($h_l$) and plume front center ($h_c$), the radius ($r$) and bulk temperature ($T_{bulk}$). The dimensions were determined by eye through pixel coordinate measurements, followed by conversion to meters using the known pixel size, calculated from $2 \times \text{[distance to target \tan(1.3 \text{ m illiradians}^2)]}$. The height values were then corrected for minor vertical distortions (<3%) due to camera viewing angle. Velocity and acceleration were measured by fitting a fifth-order polynomial to the height data (with respect to time, which was recorded by the camera for each image) and taking the first and second derivatives. $T_{bulk}$ was determined by placing a measurement circle over much of the plume front and taking the average temperature within that circle. The volume of the plume front was determined by approximating it as an ellipsoid with two horizontal radii ($r_h$) of $r$ and a vertical radius ($r_v$) of $h_l-h_c$ (Fig. 3.3). The radius of the feeder plume ($r_f$) was also measured, just below the bottom of the plume front.

In the gas-thrust region where velocities were high (>15 m s$^{-1}$) the measurements were made every 3-5 frames (or every 0.1-0.17 s), and in the convective region where velocities were low (generally <10 m s$^{-1}$) measurements were made every 10-30 frames (or every 0.3-1.0 s), depending on velocity and plume front evolution. Tracks were made for each of the 25 selected eruptions throughout their entire lifetime in the field of view, so that the above parameters were each recorded at a total of 1162 points in time (i.e. an average of 46 points for each plume).
3.4.2 Measuring plume temperature

The S40 camera measures radiance, which is then internally converted to temperature using the Planck function (see Harris et al., 1997 for an example of the Planck approach). Radiance of the target \( R_{\text{obj}} \) is calculated by the camera using the following formula (ThermaCAM™ S40 user’s manual):

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

The variables \( R_{\text{tot}}, R_{\text{refl}} \) and \( R_{\text{atm}} \) represent the total radiance collected by the camera, radiance reflected by the target and atmospheric radiance, respectively. The variables \( \varepsilon \) and \( \tau \) denote emissivity of the target and transmissivity of the atmosphere, respectively. I use the 8-14 \( \mu \)m emissivity of basalt (0.95) from Salisbury and D’Aria (1992). The transmissivity of the atmosphere is calculated with the camera’s built-in LOWTRAN atmospheric correction routine (Selby et al., 1978), where the input parameters include distance, ambient temperature and relative humidity. These latter two parameters were measured periodically during each acquisition run.

I limited my plume temperature results to those periods when I determined that the plume was opaque, as discussed in section 3.4.4. When the plume becomes translucent, apparent plume temperatures are not equivalent to actual plume temperature, due to the mixing of plume radiance with background sky (or far-crater wall at the base of the plume) radiance.
I assume efficient mixing in the plume front so that the observed temperature on the outer surface of the plume is representative of the bulk temperature. Inspection of the FLIR imagery showed turbulent mixing to be common. This, according to Turner (1979), should be a prevalent feature in plume fronts. Supporting this assumption of a reasonably well-mixed plume front, the standard deviation in temperature throughout the plume front surface was generally within 10% of the mean Celsius temperature. Further uncertainties in temperature measurement result from errors in emissivity and atmospheric correction (the latter controlled by the inputs of ambient temperature and relative humidity). These were all varied over plausible ranges to estimate the effect on resultant temperature. The combined uncertainty was garnered by treating the independent errors from 1) inherent temperature variability, 2) emissivity, and 3) atmospheric correction as vectors, and taking the square root of the sum of their squares. This resulted in an uncertainty of <12% of the Celsius temperature. Also note that temperatures above \( \sim 550 \, ^\circ C \) were not measurable, as this is the saturation temperature of the mid-gain setting used on the FLIR during my observations (Chapter 1).

3.4.3 Ejecta mass estimates

I constrained the mass of ejecta (gas and ash) in the plume fronts using a heat conservation approach as follows. I assume that the plume exits the vent as a mixture of ash and gas at their respective initial temperatures. As the mixture ascends, it distributes heat into the entrained air and the combined mixture drops in temperature. Because
radiative heat loss is negligible (section 3.6.2), the total amount of heat in the plume will be conserved throughout this process. At any point in time, therefore, the total above-background heat in the combined mass of air, gas and ash \( Q_{\text{final}} \) should equal the heat originally supplied by the mass of gas and ash at their initial temperatures \( Q_{\text{initial}} \). In this case, the equations are thus:

\[
Q_{\text{final}} = Q_{\text{initial}}
\]

\[
[m_{\text{air}}c_{\text{air}} + m_{\text{ash}}c_{\text{ash}} + m_{\text{gas}}c_{\text{gas}}] \Delta T_{\text{bulk}} = m_{\text{ash}}c_{\text{ash}}\Delta T_{\text{ash}} + m_{\text{gas}}c_{\text{gas}}\Delta T_{\text{gas}}
\]

where \( m_{\text{ash}}, m_{\text{air}}, m_{\text{gas}} \) are the respective mass values for ash, air and gas, and \( c_{\text{ash}}, c_{\text{air}}, c_{\text{gas}} \) are the respective specific heat values for the same parameters. The bulk temperature at any point in time is \( \Delta T_{\text{bulk}} \) while \( \Delta T_{\text{ash}} \) and \( \Delta T_{\text{gas}} \) are the respective initial temperatures of the ash and gas ejecta components.

A similar method was used by Wilson and Self (1980) and Sparks and Wilson (1982). However these approaches used mass (determined from plume dynamics) to estimate temperature, i.e. the reverse of my approach. Wilson and Self (1980) and Sparks and Wilson (1982), analyzing vulcanian and plinian eruptions, respectively, used the inherent mass fraction of exsolved gas from the magma (generally a few percent) to constrain the total mass values for volcanic gas and particles. Strombolian ejecta is unique from that of vulcanian or plinian eruptions in that the solid particles are entirely incidental due to deep gas decoupling from the parent magma (Chouet et al., 1974;
Blackburn et al., 1976; Vergniolle and Brandeis, 1989), and thus the relative amounts of volcanic gas and particles can potentially extend over any reasonable range.

Due to the uncertainty in gas mass fraction, \( n \), I solve for the full potential range of possible gas fractions, i.e. over end-member gas fractions from zero to one (\( n=0-1 \)). At \( n=0 \), all of the ejecta mass comprises solid particles, while at \( n=1 \) all is volcanic gas. Both of these end members are unrealistic, as some gas must be present from the slug (i.e. \( n \) must be greater than 0), and some ash is present to make the plume opaque (i.e. \( n \) must be less than 1). An additional constraint may be imposed because \( n \) values less than ~0.15 will be, initially, denser than air. Type 2a eruptions, which exhibited initial momentum, could potentially be initially denser than air, but Type 2b eruptions, which had no initial momentum, must have had \( n \) values greater than 0.15. To apply consistent constraints on all eruptions for the purposes of comparing one another, however, I apply the \( n=0-1 \) range to all eruptions.

The other uncertainty in my calculation is the initial temperature of the gas and ash. Because gas overpressures are modest in Strombolian eruptions (several bars, from Ripepe and Marchetti, 2002) I assume cooling due to adiabatic decompression is small. Thus, a magmatic temperature is appropriate (1000 °C). The initial temperature of the solid particles is more poorly constrained, as the ash is likely sourced from loose material sitting atop the magma column (Chapter 2). In this case, the temperature could be any value between magmatic temperature (~1000 °C) and ambient, depending on the thermal gradient in the loose overfill. A FLIR image taken of the vent at SW crater in May 2002
showed the vent was covered or filled with material which had a surface temperature of \(~300\ {^\circ}C\). This temperature would represent the ‘skin’ temperature of the brittle layer resulting from radiative and convective heat losses, with much higher temperatures expected just beneath the surface. As a rough constraint on the possible ash temperatures, therefore, I adopt the range of 500-1000 \(^{\circ}C\). These high temperatures are conceptually plausible because the active vents are the outlet for Stromboli’s combined 35-200 kg s\(^{-1}\) of passive degassing (Allard et al., 1994), which would stream directly through any overfill. In addition, such vent temperatures were commonly retrieved using hand-held thermal infrared thermometer (spot radiometer) data for open vents during 1994-2002 (A. Harris, unpublished field measurements).

### 3.4.4 Determining opacity

Accurate plume temperatures can be measured only when the plume is opaque, and I determined two ways to judge opacity of the plumes in FLIR images. First, the temperature of the plume compared to ambient temperature can indicate translucence. Because of its high transmissivity, the apparent temperature of the atmosphere (looking into space) will be significantly lower (generally \(<0\ {^\circ}C\)) than actual temperatures for the atmosphere \((\sim20-30\ {^\circ}C\ \text{in June 2004})\). If the apparent temperature of the plume drops below that of ambient for atmosphere, it is a clear indication that the plume is translucent and beginning to pass background sky radiation.
A plume may become translucent, however, before its apparent temperature drops below that of ambient. A more useful tool for judging plume opacity is to examine the rate of change in apparent heat content in the plume front. Total heat content is calculated using a constant gas mass fraction \((n)\), for simplicity, and plotted versus volume (Fig. 3.4). Because the feeder plume supplies new material into the plume front at a generally steady rate (Turner, 1962), the total heat content should show a roughly steady increase (Fig. 3.4). At the point that the apparent heat content ceases to increase, begins to drop (Fig. 3.4), or deviates strongly from a linear trend, I can reason that the apparent temperature is affected by the transmittance of sky radiation through the plume. After this point, apparent temperatures, and their derived parameters, are no longer representative of actual plume temperature.

3.5 Results

3.5.1 Rise rates

The rise of Type 2 plumes has been discussed in Chapter 2 (shown here in Fig. 3.5). Type 2a eruptions exhibited gas-thrust phases (initial velocities >15 m s\(^{-1}\)) which decelerated into convective velocities (<15 m s\(^{-1}\)) by height of ~100 m. Type 2b eruptions, on the other hand, showed roughly constant convective velocities throughout their observed lifespan. Time-averaged convective velocities for 42 Type 2b plumes ranged between 1.4 and 10.9 m s\(^{-1}\), with a mean value of 4.5 m s\(^{-1}\) (st. dev.: 2.1 m s\(^{-1}\)).
Figure 3.4. Testing the opacity of plume fronts. Opaque plumes (top) show a clear increase in apparent heat content with plume front volume. When the apparent heat of a plume front declines (bottom), it is a sign that the plume front has become translucent.
Figure 3.5. Height, velocity and acceleration for Strombolian ash plumes. Data from NE-1 crater in June-July 2004, separated by eruption type. Original height values are shown, while velocity and acceleration were produced from smoothed height data (fifth-order polynomial).
3.5.2 Plume morphology

Plume forms were controlled by the velocity differential between the plume front and the feeder plume. Type 2a plumes, with their high initial velocities (>15 m s⁻¹; Fig. 3.5), began as jets in which the front of the jet was largely coupled to the main body of the jet. Superimposed on this, however, was turbulent shearing along the edges which continuously sloughed off small-scale (~3-5 m wide) entrainment vortices. Upon deceleration of the jet (transitional gas-thrust to buoyant velocities, 10-15 m s⁻¹), a clear velocity differential developed between the plume front and main plume (now the feeder plume), creating a starting plume morphology. When fully buoyant behavior was established (<10 m s⁻¹), there was a pronounced differential leading to the development of the vigorous ring vortex which marked a rooted thermal. Type 2b plumes often emerged above the crater rim as amorphous masses which organized themselves directly into rooted thermals once the velocity differential was established. Because the ascent rate was slow for Type 2b plumes, the velocity differential could develop just at a short distance above the vent.

I found an effective way of quantifying the plume form using the ratio of feeder plume radius to plume front radius (Fig. 3.6). It is instructive to compare Figure 3.6 with Figure 3.5 in order to link velocities with their corresponding heights. During high-velocity gas-thrust phases the plume front comprised the tapered head of the gas jet, and extended downwards, with little interruption, to the feeder plume. The ratio of the feeder plume radius to that of the plume front was 0.8-1.2 (Fig. 3.6) for jet morphologies. At the onset of buoyant phases the jet would transition into a starting plume, with a feeder:front
Figure 3.6. Quantifying plume form. Plume forms can nominally be quantified by the ratio of feeder plume radius and plume front radius. Higher velocities exhibited a jet form, intermediate (buoyant or near-buoyant velocities) produced a starting plume, and fully buoyant plumes were rooted thermals.
ratio generally 0.75-1.0. In fully buoyant phases, either during the entirety of Type 2b plumes or the later stages of Type 2a plumes, the plume front could grow significantly larger in radius than the feeder plume. The emergence of very low feeder:front ratios (<0.75) was coincident with the establishment of the large ring vortices which formed rooted thermals (Fig. 3.2). Similar ring vortices were remarked on by Kieffer and Sturtevant (1984). Not all starting plumes transitioned into rooted thermals in my field of view. Some rose out of the field of view while still starting plumes, while small plumes tended to dissipate too rapidly to establish the ring vortex of a rooted thermal.

3.5.3 Spreading rates

Turbulent plume studies have shown consistently that plumes spread following a roughly linear trend with height (Batchelor, 1954; Turner, 1962), and the spreading rates I measured agree with this on a first order basis (Fig. 3.7). On closer inspection, I observed that plume fronts at greater heights appeared to be spreading at a faster rate (with respect to height) than at lower points, because spreading rate was different for gas-thrust and buoyant regimes (Table 3.1). For the 25 eruptions examined, I used a threshold velocity of 15 m s\(^{-1}\) (Chapter 2) to divide plume front spreading trends into gas-thrust and buoyant regimes (Fig. 3.8). The horizontal plume front radius \(r_h\) at any height \(h\) can be calculated from the relation \(r_h = (drh/dh)h + r_{hi}\), where \(r_{hi}\) is initial horizontal radius. In gas-thrust regions, \(drh/dh\) varied from 0.06-0.22 (mean: 0.13) and \(r_{hi}\) varied from -4.9 to 6.6 m (mean: 3.3 m). Negative initial radii indicate virtual plume sources above the vent, and simply mark a departure from the typical linear spreading
Figure 3.7. Dimensions of the plume front. Data from 25 eruptions in June-July 2004.
Table 3.1. Spreading rates of 25 strombolian plumes

<table>
<thead>
<tr>
<th>1) Plume front</th>
<th>Min.</th>
<th>Max.</th>
<th>Mean</th>
<th>Stdv.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Gas thrust, $r_h$</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Rate $(dr_h/dh)$</td>
<td>0.06</td>
<td>0.22</td>
<td>0.13</td>
<td>0.05</td>
</tr>
<tr>
<td>Half angle, $^\circ$</td>
<td>4</td>
<td>13</td>
<td>7</td>
<td>3</td>
</tr>
<tr>
<td>Lateral velocity, m s$^{-1}$</td>
<td>1.6</td>
<td>3.7</td>
<td>2.8</td>
<td>0.7</td>
</tr>
<tr>
<td>Buoyant, $r_h$</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Rate $(dr_h/dh)$</td>
<td>0.1</td>
<td>0.41</td>
<td>0.24</td>
<td>0.07</td>
</tr>
<tr>
<td>Half angle, $^\circ$</td>
<td>6</td>
<td>22</td>
<td>14</td>
<td>4</td>
</tr>
<tr>
<td>Lateral velocity, m s$^{-1}$</td>
<td>1.1</td>
<td>3.4</td>
<td>1.8</td>
<td>0.6</td>
</tr>
<tr>
<td>Vertical radius, $r_v$</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Rate $(dr_v/dh)$</td>
<td>0.03</td>
<td>0.18</td>
<td>0.09</td>
<td>0.04</td>
</tr>
<tr>
<td>2) Feeder plume</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Rate $(dr_v/dh)$</td>
<td>0.02</td>
<td>0.21</td>
<td>0.12</td>
<td>0.04</td>
</tr>
<tr>
<td>Half angle, $^\circ$</td>
<td>1</td>
<td>12</td>
<td>7</td>
<td>2</td>
</tr>
<tr>
<td>Lateral velocity, m s$^{-1}$</td>
<td>0.2</td>
<td>2.2</td>
<td>1.0</td>
<td>0.6</td>
</tr>
</tbody>
</table>
Figure 3.8. Spreading rates of the plume front for gas-thrust and buoyant regimes. Spreading half-angles during buoyant regimes were approximately twice those of gas-thrust regimes.
trend. The gas-thrust rates equate to spreading half-angles of 4–13° (mean: 7°) and lateral
growth velocities of 1.6–3.7 m s⁻¹ (mean: 2.8 m s⁻¹). In buoyant regions, \( \frac{dr}{dh} \) varied
from 0.10–0.41 (mean: 0.24) and \( r_h \) varied from -11.7 to 6.2 m (mean: -0.6 m). These
rates equate to spreading half angles of 6–22° (mean: 14°) and lateral growth velocities of
1.1–3.4 m s⁻¹ (mean: 1.8 m s⁻¹). Note that even though the spreading angle is much lower
in gas-thrust regimes, the higher upward velocity leads to greater lateral velocities
compared to buoyant regimes.

This range of spreading angles is in good agreement with the laboratory results of
Turner (1957, 1962). In those studies, starting plumes exhibited spreading rates of 0.15–
0.21 (half angles of 10.2°±1.7) while discrete thermals had values of ~0.30 (17°). My
results (0.06–0.41) cover the full range of starting plume and thermal spreading rates,
supporting the interpretation that plumes were transitioning between multiple plume
forms during the period I observed. My results also extend slightly lower (to 0.12) than
the lowest laboratory starting plume spreading rate (0.15), possibly reflecting the
presence of a jet spreading rate.

The rooted thermal form that I describe has not been studied in the laboratory and
therefore lacks a published air entrainment coefficient. In buoyant regimes, where the
rooted thermal form was common, the mean spreading rate was 0.24, and up to 0.41,
which overlaps with the spreading rate of a discrete thermal (0.30). Therefore, I assume
that the entrainment coefficient of a rooted thermal is roughly similar to that of a discrete
thermal. The chief difference is that a discrete thermal entrains only air (and thereby
maintains a constant above-background heat content) while the rooted thermal entrains both air and feeder plume, increasing its above-background heat content.

The majority of my spreading rates were in the range 0.10-0.30 (Table 3.1), and only one of my spreading rates (0.41) overlaps with those of plinian plumes observed at Soufrière volcano in 1979 (0.39-0.43) and measured by Sparks and Wilson (1982). Those authors attribute their higher spreading rates, in part, to decreasing atmospheric density with height over the ~10 km range through which their plumes extended. This should not be a significant factor over the 150 m vertical ascent range measured here.

The vertical radius of the plume front ($r_v$, Fig. 3.7) also increased linearly with height, but at approximately half the rate at which the horizontal radius grew, $dr_v/dh$ having a range of 0.03-0.18 (mean: 0.09). The aspect ratio of the plume front ($r_h/r_v$) generally ranged from 0.75 to 2.5, increasing steadily with height (Fig. 3.7). The plume front volume changed at a higher rate, naturally (Fig. 3.7).

The radius of the feeder plume ($r_f$) also grew linearly with height, with approximately the same rate as that calculated for the plume fronts in gas-thrust regimes. Growth rates ($dr_f/dh$) ranged from 0.02 to 0.21, with a mean of 0.12. These equate to spreading angles of 1-12° (mean: 7°) and lateral growth velocities of 0.2-2.2 m s⁻¹ (mean: 1.0 m s⁻¹). The similarity between spreading rates for feeder plumes, plume front jets and starting plumes is consistent with the fact that they all entrain air through a large number of small, disorganized eddies, without the benefit of the large ring vortex of a thermal.
3.5.4 Plume front temperatures

The first step in measuring real plume temperature is determining opacity, i.e. the plume must be opaque if we are to extract actual plume temperature in a straightforward manner. Based upon the opacity test developed in section 3.4.4, seven of the plumes became translucent (Fig. 3.9) within the 130 m high field of view. Translucence was more common and pronounced among the June 07 eruptions than other observation days. Only three plume fronts dropped below ambient temperature (i.e. reached extreme translucence), and these all occurred on June 07. At least four eruptions out of a total of eight measured on that day became translucent to some degree (two additional eruptions were possibly translucent). Two eruptions out of seven became translucent on 11+13 June, but these were two low-velocity plumes which remained in the field of view for a substantial amount of time. Comparing translucence between days is somewhat troublesome as it is a function of overall plume scale (e.g. very small plumes always became translucent within the field of view, often immediately above the vent). Though I limited my selection of 25 eruptions to relatively large plumes, some minor bias in plume scale may still exist among the days.

For opaque plumes fronts, the bulk temperature asymptotically approached the background (ambient) temperature through time (Fig. 3.9). Figure 3.9a indicates the uncertainty (±12%) for a sample eruption. Mean plume front temperatures for the 25 eruptions were greater than ~300 °C at the crater rim, with several initially exceeding the 550 °C saturation temperature of the camera’s medium gain setting (Fig. 3.9b). Visual
Figure 3.9. Bulk temperature of the plume front. A) Example track (eruption 94) with error bars showing +/- one standard deviation, B) All 25 temperature tracks vs. height, and C) the same vs. time. Dotted portions show periods where the plume was likely translucent, and therefore temperature values are unreliable.
extrapolation of the cooling trends (Fig. 3.9) suggests that these temperatures were
somewhere between 600 and 700 °C at the crater rim. Most plume fronts dropped below
100 °C within the first 130 meters of ascent (Fig. 3.9b), which equates to the first ~5-10
seconds of the plume life (Fig. 3.9c). The mean cooling rate was 3-4 °C m\(^{-1}\) over the first
100 meters. Note that the cooling rate (per unit height) during the first 100 m was
roughly independent of initial temperature (Fig. 3.9).

Plume front temperature variations were not randomly distributed among
observation days. Figure 3.10 shows plume front temperatures at the 50 m height, with
error bars for temperature uncertainty. The 8 eruptions from June 07 show plume
temperatures tightly clustered between 100 and 250 °C, while plumes on June 11+13 had
a capacity for greater temperatures (200-500 °C).

3.5.5 Ejecta mass in plume fronts

Results of the ejecta mass approach for the 25 tracked eruptions are shown in
Figure 3.11. The ejecta mass estimation method was only applied to opaque periods.
The end-member solutions for \( n \), the gas mass fraction of total ejecta, range from 0
(ejecta is entirely ash) to 1 (ejecta is entirely gas). Derived ejecta mass for the plume
front increases steadily with height, a result of the continuous supply of mass from the
feeder plume. For \( n=0 \) (ejecta is entirely ash), ejecta mass in the plume front began
between \( 10^2-10^3 \) kg and increased to \( 10^3-10^4 \) kg by 150 m height. For \( n=1 \) (i.e. ejecta is
entirely volcanic gas), ejecta mass began between \( 10^1 \) and \( 10^2 \) kg, and increased to \( 10^2-
Figure 3.10. Mean temperature of the plume fronts at 50 m height. Error bars show uncertainty. Note plumes on June 11+13 attained higher temperatures than those on June 07.
Figure 3.11. Plume front mass values. Results constrained with the heat conservation approach for end-member gas mass fractions ($n$). At $n=0$, all of the ejecta mass (and all above-background heat) is assumed to be from ash. At $n=1$, the entirety is attributed to volcanic gas.
$10^3$ kg by 150 m height. As ejecta comprises a minor portion of the plume volume at any significant height above the vent, the air mass solution is similar between the two end-member solutions ($\pm 1$-6%). Air mass began between $10^1$ and $10^3$ kg and increased to $10^4$-$10^5$ kg by 150 m height. Results are shown in Table 3.2, at 130 m height, for those plumes which did not become translucent in the field of view.

Low plume temperatures were observed on June 07 and high temperatures were more common on June 11+13 and, therefore, it was of interest to quantify any ejecta mass difference which could explain this temperature disparity (Fig. 3.12). Solutions for $n=1$ extend to $\sim 1000$ kg for eruptions on June 07, and up to $\sim 2000$ kg for those on June 11+13, at $\sim 150$ m height. For $n=0$, eruptions on June 07 extend up to $\sim 4000$ kg while those on June 11+13 extend to $\sim 8000$ kg. For a given gas fraction, that is, the eruptions on June 11+13 contained more mass. The uncertainty in $n$, however, does not allow a definite comparison between mass values on June 07 and June 11+13 because the end-member solutions extend over a range greater than the relative difference between these days.

3.5.6 Miscellaneous observations

Sedimentation was visible in the FLIR images in two forms. First, very fine particle sedimentation appeared as a faint curtain of elevated temperatures below overhanging sections of the plume, moving in concert with the plume. Sedimentation curtains, or veils, have been described for larger-scale eruptions by Houghton et al. (2004) and Clarke et al. (2002). Second, coarse sedimentation appeared as discrete
Table 3.2. Plume characteristics at 130 m height

<table>
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<tr>
<th>Erupt. Date</th>
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<th></th>
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<td>Ejecta mass, kg $n=0$ (all ash)</td>
<td>Ejecta mass, kg $n=1$ (all volc. gas)</td>
<td>Air mass, kg</td>
<td>Front volume fraction of whole</td>
<td>Ejecta mass, kg $n=0$ (all ash)</td>
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<td>54</td>
<td>2527</td>
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<td>10859</td>
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Figure 3.12. Ejecta mass in the plume front through time. Results based on heat conservation approach, for end-member solutions of gas mass fraction, $n$. 
elevated pixels, resembling ballistic particles except that their paths were traced directly
down from the plume rather than as a parabolic arc originating from the vent (Fig. 3.13a).
The nature and intensity of sedimentation varied among the observation days. For
instance, significant sedimentation was not visible for any plumes on June 07, while on
June 11+13 sedimentation (both coarse and fine) was common for large plumes. Very
small plumes on June 11+13 did not exhibit any visible sedimentation. On June 19,
sedimentation was most pronounced and plumes exhibited prolonged periods of fine
sedimentation, sometimes to the point of obscuring the feeder plume.

Helical vortex motion was observed in the feeder plume on several occasions
(Fig. 3.13b), and could be cyclonic (counter-clockwise from above) or anticyclonic.
Helical motion was not always clear, and plumes were deemed definite or probable
examples. The helical motion likely results from the rotation of an originally horizontal
entrainment rotational axis towards a vertical axis (Maxworthy, 1973), and this could
result from lateral variations in buoyancy. Helical motions were only observed during
buoyant velocity regimes. Because these eruptions are short-lived and small (and
unaffected by Coriolis forces), they are much like dust-devils and exhibit a rough 50/50
ratio of cyclonic/anticyclonic spins (Idso, 1974). Such an even ratio is close to what was
observed: of definite and probable helical plumes I recorded 4 cyclonic examples and 5-6
anticyclonic examples. Three of the five definite helical plumes were on June 13, with
the other two on June 19. No definite helical eruptions, and just one probable example,
occurred on June 07 and 27.
Figure 3.13. Miscellaneous plume observations. A) Coarse sedimentation (dotted arrow) from a plume detached from the vent and drifting east (solid arrow). B) Example of cyclonic helical motion in the feeder plume. C) A portion of the plume front in the process of sloughing off (dotted arrow) while the remainder of the plume front continues rising (solid arrow).
During the course of the plume front growth, minor portions at the edge of the plume front would occasionally decelerate sufficiently to slough off and detach from the plume front. Sloughing was caused by two mechanisms. First, during gas-thrust phases the turbulent shearing along the edges of the plume front would tear off portions of the plume circulating along the plume front margin. Second, during buoyant phases colder portions of the plume front would fall behind the main plume front (Fig. 3.13c), representing portions of the plume front which could not maintain the buoyancy of the remainder. For instance, the portion sloughing off of the plume front of the rooted thermal in Figure 3.13c has a significantly lower temperature (70 °C) than the remainder of the plume front (130 °C) which rose at a higher rate. Thus, occasional thermal heterogeneity could arise in plume fronts, but this was more often limited to small portions. The detached slough would then follow one of two trends. Orphaned plume slough would detach completely and be left behind by the plume front, usually drifting into the dissipating remnants of the feeder plume column. Alternatively, recycled plume slough would be immediately caught by the ring vortex at the base of the plume front and reincorporated into the plume front.

3.6 Assumptions

3.6.1 Sources of radiance

During opaque periods I assume that all radiance originates from the solid particles in the plume. Calculations with the MODTRAN4 radiative transfer program
(Berk et al., 1999) on mixtures of water vapor suggest that radiance from the gas phase will be low. Varying the concentration of water vapor in hypothetical gas plumes, MODTRAN4 estimated that the transmissivity was 0.94 for the worst-case scenario of 0.2 kg m\(^{-3}\) water vapor at 1000 °C (i.e. a gas parcel leaving the vent). In a gaseous phase, emissivity is the complement of transmissivity, and thus the emissivity of 0.06 can be compared to that of basalt at 0.95 (Salisbury and D’Aria, 1992). In an ash plume image, there is concern that gas parcels between the particles and the observer could contribute excess radiance and corrupt temperature measurements. But because the emissivity and transmissivity are complements, a parcel of gas (at the same temperature as the particle) will contribute no net radiance. In a situation where the gas temperature is higher or lower than the particle, the low relative emissivity of the gas would ensure a minor effect on the apparent particle temperature.

One additional component that could compromise my heat budget approach to ejecta mass estimation is water aerosol, which can exhibit significant emission at these wavelengths (Carlon, 1971) as well as liberate latent heat. Water aerosol typically develops at a relative humidity >60% (Carlon, 1971), but the relative humidity of gas exiting the vent (at, say, 0.2 kg m\(^{-3}\)) is extremely low (<1%) due to the high vapor saturation density of high temperatures. Thus, I would only expect aerosol formation in the plume after humid air has been entrained. Only on June 07 was the ambient relative humidity greater than 60%, (61% at the start of the acquisition), but this dropped to 19% within 2.5 hours as the air temperature rose, suggesting the relative humidity was below
60% for most of the acquisition that day. On other acquisition days relative humidity values were consistently below 40%. Based upon several gas flux simulations, my calculations suggest that the volcanic gas will not appreciably raise the relative humidity of the mixture upon entrainment (due to the counterbalancing of heat and vapor saturation density). Therefore, I can discount water aerosol as a radiant contributor or influence on thermal budgets during my particular observation days.

3.6.2 Heat loss from the plume

What heat transfer mechanism controls the temperature of these plumes on the timescales that I observed them (<30 s)? The plume temperature can decrease primarily by either radiative heat loss to the atmosphere or entrainment of air at ambient temperature; adiabatic cooling of the plume due to rising in a stratified atmosphere is assumed negligible as I am examining only a vertical path length of ~150 m.

Cooling due to air entrainment can be shown as the dominant factor by comparing observed cooling rates to modeled cooling rates calculated assuming radiative cooling. Observed cooling rates were measured from the 25 tracked eruptions using the first-derivative of the smoothed temperature trends (Fig. 3.14) and ranged between -50 and -200 °C s⁻¹, with a few outliers showing semi-prolonged positive changes. Increases in temperature occurred from rapid injections of hot feeder plume material into a colder plume front, always in the initial stages just above the vent, or from colder obscuring material moving out of the way.
Figure 3.14. Observed cooling rates for the 25 tracked plume fronts. These are significantly higher than the radiative-only models, which predicts a maximum bulk temperature change of only 0.3 °C s⁻¹, indicating that radiative heat losses are negligible.
For comparison, I modeled hypothetical plumes with the time-varying dimensions and temperatures of the 25 tracked plumes, a range of densities from 0.2-1.0 kg m\(^{-3}\), and emissivity of unity (as an overestimate). These were cooled radiatively over the same time interval for comparison with the observed cooling rates. I modeled radiation from only the outer surface, as these plumes were determined to be opaque for the periods shown (section 3.4.4). The radiation-only model gave a maximum cooling rate of just -0.3 °C s\(^{-1}\) (highest for lower densities) while the observed cooling rates were significantly greater (up to 200 °C s\(^{-1}\)). I also took the plume model of Wilson and Walker (1987), which assumes cooling only by air entrainment, and inserted a radiative cooling routine. The effect on the results obtained using the cooling-by-air-ingestion-only model was negligible. Therefore I assume radiative heat transfer is insignificant compared to that resulting from the mixing of cold entrained air.

3.6.3 Thermal equilibrium

The gas phase (air and volcanic gas) comprises a large portion of mass in the plume front (Fig. 3.11), yet the FLIR is only measuring radiance from the ash particles (section 3.6.1). To gain a representative temperature of the gas phase, therefore, the particles must be in thermal equilibrium with the gas. Woods and Bursik (1991) estimated the timescale for complete thermal equilibrium to be 100\(d^2\), in which \(d\) is the diameter of the particle in cm. Based on terminal velocities buoyant phases (<10 m s\(^{-1}\)), for instance, will support a maximum particle size of 3-6 mm (depending on gas/air
concentration and temperature), which takes 9-36 s to equilibrate completely. This is longer than the residence time of most plumes in the field of view, suggesting that the largest clasts in these plumes will not attain complete thermal equilibrium on the timescale of my observations.

Terminal velocities, however, relate only the theoretical maximum particle size while the majority of particles in the plume will be significantly smaller. For a particle 1 mm in diameter, thermal equilibrium takes 1 sec according to the Woods and Bursik (1991) relation. I assume that the majority of particles are smaller than 1 mm, and therefore are capable of transferring heat on a timescale of a second. Supporting this, Wilson and Self (1980) collected ash particles for three plumes which, according to their descriptions, were very similar in scale and velocity to the plumes measured here. For their plumes the largest particles they collected ranged from 200 to 900 microns in diameter.

3.7 Discussion

3.7.1 Are strombolian plumes analogous to plinian plumes?

It is important to recognize which aspects of strombolian plumes are analogous to much larger-scale plinian plumes and which are not. Besides scale and velocity, the greatest differences between strombolian plumes and plinian plumes are the 1) gas mass fraction and 2) initial temperature of the particles, both of which are related to 3) the incidental nature of strombolian particles. Plinian ejecta is thermally and mechanically
coupled upon leaving the vent (Wilson, 1980) and thus would have similar initial temperatures among ejecta components (Sparks and Wilson, 1982) as well as a well-constrained gas mass fraction (<0.05, Sparks and Wilson, 1976). As I have shown, due to the incidental nature of strombolian particles, a wide range of initial particle temperatures and gas mass fractions are possible. Strombolian plumes, at least in some instances, have gas mass fractions >0.15 (section 3.7.6), which is greater than plinian values (<0.05). Strombolian plumes, in other words, can be much less dense than plinian plumes upon leaving the vent. This means that the ejecta mass values I measured will not be representative of plinian values even after scaling the plume dimensions. For the same reasons, the temperatures of strombolian plumes (Fig. 3.9) will not be representative of plinian plumes.

Nevertheless, there are several significant aspects of the strombolian plumes I measured that are analogous to plinian plumes, albeit at a much smaller scale. First, I have shown how strombolian explosions eject particles which are distinctly coupled (ash plume) and decoupled (ballistic particles) from the gas phase (Chapter 2). Second, the two-part model of gas thrust-convective phases of the ash plume, more commonly ascribed to plinian eruptions, accurately characterizes strombolian plumes as well (Fig. 3.5, see also Blackburn et al., 1976; Sparks and Wilson, 1976). The buoyant sloughing I observed (section 3.5.6; Fig. 3.13c) may be thought of as the buoyant equivalent of partial plume collapse to the extent that the sloughing indicates thermal heterogeneities in the plume. In these strombolian plumes, the sloughed material was buoyant in the first
place, but in vulcanian and plinian plumes the density variations of heterogeneous plume portions can bound the density of air, controlling which parts collapse.

3.7.2 Plume front parameters

The quantitative parameters presented here depict the properties of the plume front, a result of the difficulties of tracking temperatures and dimensions of the highly transient feeder plume. To provide a rough idea of the ejecta mass in the entire plume, the volume of the plume front can be compared to the total volume of the entire plume (Fig. 3.15; Table 3.2). This was achieved through a simple approximation of the plume front as an ellipsoid (as discussed) and the feeder plume as a column with the radius $r_f$ and height of $h_f-(h_t-h_c)$. Plume fronts generally comprised 20-50% of the entire plume volume, indicating that mass values can be multiplied by 2-5 for a first-order estimation of whole plume ejecta mass. Table 3.2 shows mass values at 130 m height (the top of the field of view) for the plume fronts, along with estimates for whole plume mass. Whole plume mass estimates ranged from 3220 to 30330 kg (mean: 10455 kg) for $n=0$, and from 920 to 7690 kg (mean: 2760 kg) for $n=1$.

3.7.3 Transient air entrainment regimes

Strombolian plumes exhibit different plume forms (jets, starting plumes, rooted thermals) which I have quantified with the feeder to plume front radius ratio (Fig. 3.6). A particular plume can transition from a jet (gas-thrust), to a starting plume (transitional), to a rooted thermal form (fully buoyant) over the course of seconds to a few tens of seconds.
Figure 3.15. Extrapolating plume front mass values to the entire plume. Plume fronts generally formed 0.2-0.5 of the whole plume, indicating that plume front mass values can be multiplied by 2-5 to roughly estimate whole plume mass values.
Each plume form has a unique air entrainment coefficient (jets=0.065, thermals=0.25), indicating that the plumes I studied changed their air entrainment coefficient fairly rapidly. Higher entrainment coefficients in buoyant phases (due to the rooted thermal form) would mean greater spreading rates, and this is shown by Figure 3.8 where the mean buoyant spreading rate (0.22, or 13°) is almost twice that in the gas-thrust region (0.12, or 7°).

Changing air entrainment dynamics are of paramount importance in the study of pyroclastic flow generation, as the air influx rate has a large control on buoyancy potential. Previous volcanic plume models essentially assume a constant (Wilson and Walker, 1987; Woods, 1988; Glaze and Baloga, 1996) air entrainment coefficient, equivalent to assuming a constant spreading rate (Batchelor, 1954; Turner, 1962). This is completely reasonable for the large, steady-state plinian columns for which the models are intended. My results show, however, that small-scale (tens to hundreds of meter scale) plumes have transient air entrainment dynamics, indicating that existing plume models must be modified to characterize smaller-scale eruptive phenomena (e.g. strombolian and vulcanian styles).

3.7.4 Changes in daily plume temperatures

What caused the greater capacity for high temperature plumes on June 11+13, compared to a low temperature day such as June 07? Two lines of evidence from the FLIR data support an increase in particle content as being a primary source for the increased heat. This evidence includes: 1) higher tendency for translucence on June 07
coincident with low plume temperatures (Fig. 3.10) and 2) the more pronounced visible sedimentation on June 11+13 (Fig. 3.13) coincident with higher temperatures (Fig. 3.10).

In Chapter 2, I discussed how ash in Strombolian plumes is likely sourced from loose material which has accumulated in a backfill layer on top of the magma column. I here speculate that the particle concentration is linked to the available mass of fine particles in this backfill layer, which may change from day to day. On June 07, just 1-2 days into the Type 2 phase at NE-1 crater, backfill levels were likely minor. Given continued Type 2 eruptions, and particularly an increase in the frequency of lower intensity eruptions, over the next four days, the backfill thickness may have increased. By June 11+13, the gas plume would have then been traveling through a greater thickness of backfill, encouraging greater entrainment of fine particles and increasing the mass and heat content of the plumes.

If particle content controlled these variations in plume temperature it would demonstrate the ability of accidental material, as all strombolian particles are essentially incidental (Chapter 2), to drastically alter the mass and heat content of a plume. Bulk plume front temperatures at 50 m varied over a range of 350 °C (Fig. 3.10). In other scenarios, such as a vulcanian eruption, any accidental material could strongly effect the heat and mass content of a plume. In addition to initial momentum and air entrainment rates, the balance of accidental mass and accidental heat could ultimately control the collapse potential of a plume. The main problem, as I have shown, is that quantifying the accidental particle content in a plume is inherently difficult as it depends upon incidental circumstances and not the inherent gas solubility.
3.7.5 Strombolian gas mass fraction

As discussed, the uncertainty in gas mass fraction for strombolian plumes arises from the decoupled nature of the gas from its parent magma, and this uncertainty imposes a major limitation on my ability to extract explicit mass values from the thermal imagery. The gas mass fraction has been estimated for Strombolian ejecta in two previous studies. First, Chouet et al. (1974) measured ballistic particle velocities and inferred the gas exit velocity. From a known vent radius and assumed gas density, they calculated total gas flux and compared it to particle mass measured in their photographs. They determined that the gas:lava mass ratio was 2 and 16 (or \( n=0.67 \) and 0.94) for the two eruptions they studied. This value is irrelevant for my study, however, as they were observing Type 1 eruptions, which contain no significant ash plume. Their ratio, therefore, represents the relative gas mass to ballistic particle mass.

Second, Blackburn et al. (1976) estimated the gas:lava mass ratio by taking the bulk density (inferred from rise speed) of ash plumes and reconciling this with the original adiabatic decompression of the explosion. They estimated an average gas:lava mass ratio of 0.25 (or \( n=0.2 \)). Their study, however, suffered from the fundamental error of disregarding air entrainment. Thus, they extracted the gas mass fraction from plumes which they assumed were entirely volcanic gas and particles but were likely, for the most part, air (see Fig. 3.11). This explains why they estimated an average gas mass fraction of 0.2, as this is approaching the gas mass fraction which is equivalent to the density of air. Their results, therefore, are not representative of any real gas mass fraction, and we
are left with no value in the literature to characterize gas mass fraction in strombolian plumes.

There are two possibilities for estimating the gas mass fraction in ejecta from my imagery. First, the observed plume front volume growth rates can be reconciled with predicted rates of air entrainment to estimate volcanic gas influx into the plume front. Comparing this gas influx to the total heat influx can then garner the contribution from solid particles, which can be assumed to have no significant volume (Wilson and Self, 1980). This approach was attempted, but the uncertainty in air entrainment rates (due to rapidly changing plume forms) was so great that no insightful results could be obtained.

The second approach to estimating gas mass fraction is described in the Appendix, and involves taking a FLIR snapshot of the plume in a mature state (e.g. approaching the top of the field of view) to maximize thermal equilibrium and large scale mixing and then estimating the total heat content of the entire plume, both feeder plume and plume front. The plume velocity at the crater rim is estimated for each point in time using the 30 Hz data, from the eruption start to the time of the snapshot. This continuous velocity is used to constrain total gas output, and the gas heat contribution is then compared to the total heat content measured in the snapshot. The difference in total heat would then be attributed to solid particles. This method was applied to seven eruptions whose plume fronts remained opaque and whose feeder plumes appeared opaque at the time of the snapshot. The results, for ash temperatures of 500 and 1000 °C, range between 0.08 and 0.76, corresponding to gas:lava mass ratios between 0.09 and 3.2. The
average minimum gas mass fraction was 0.22 while the average maximum value was 0.48 (corresponding to gas:lava mass ratios of 0.28 and 0.92, respectively). If these values are trusted, they indicate that Strombolian ash plumes (i.e. Type 2 eruptions) generally involve more solid mass than volcanic gas mass. This contrasts with the Type 1 eruptions studied by Chouet et al. (1974), which lacked an ash plume and contained only coarse ballistic particles and volcanic gas, where the gas:lava mass ratios were 2 and 16. Type 2 eruptions, therefore, appear to involve more solid mass than Type 1 eruptions.

3.7.6 Strombolian plume density and collapse potential

Sparks and Wilson (1976), followed by Woods (1988), state that bulk densities in an ash plume will always be greater than that of air during the gas thrust phase. This is certainly true for plinian columns which normally have initial gas mass fractions <0.05 (Sparks and Wilson, 1976), and thus, initial bulk densities at least a few times greater than that of air. As I have mentioned, strombolian ash plumes are unique in that the entirety of solid particles is essentially incidental, and thus any initial gas mass fraction is possible. In fact, the Type 2b eruptions analyzed in Chapter 2, which have negligible initial momentum, must have had initial gas mass fractions >0.15 in order to be buoyant at the outset. Note that, using my gas mass fraction estimation technique, nearly all the explicit solutions for $n$ (Table 3.3) are greater than the critical $n$ required for initial buoyancy (~0.15). Five of the seven eruptions in Table 3.3 were gas-thrust eruptions (i.e.
Table 3.3. Gas mass fraction estimates

<table>
<thead>
<tr>
<th>Eruption</th>
<th>$n_{min}$</th>
<th>$n_{max}$</th>
<th>ratio$_{min}$*</th>
<th>ratio$_{max}$*</th>
</tr>
</thead>
<tbody>
<tr>
<td>13</td>
<td>0.25</td>
<td>0.56</td>
<td>0.34</td>
<td>1.27</td>
</tr>
<tr>
<td>36</td>
<td>0.14</td>
<td>0.32</td>
<td>0.16</td>
<td>0.48</td>
</tr>
<tr>
<td>68</td>
<td>0.31</td>
<td>0.67</td>
<td>0.44</td>
<td>2.00</td>
</tr>
<tr>
<td>72</td>
<td>0.08</td>
<td>0.20</td>
<td>0.09</td>
<td>0.25</td>
</tr>
<tr>
<td>103</td>
<td>0.26</td>
<td>0.58</td>
<td>0.35</td>
<td>1.36</td>
</tr>
<tr>
<td>207</td>
<td>0.36</td>
<td>0.76</td>
<td>0.56</td>
<td>3.15</td>
</tr>
<tr>
<td>227</td>
<td>0.12</td>
<td>0.29</td>
<td>0.14</td>
<td>0.41</td>
</tr>
</tbody>
</table>

* mass ratio of volcanic gas to lava
Type 2a), indicating that plumes at Stromboli are probably buoyant to begin with, regardless of initial momentum.

Again, this demonstrates how strombolian plumes are unique from plinian plumes, translating into differing controls on ballistic particle behavior. In plinian eruptions, the plume is initially very dense (>3.4 kg m$^{-3}$, assuming $n<0.05$, H$_2$O at 1000 °C and 1 bar), and gas velocities are often greater than 300 m s$^{-1}$ (Cioni et al., 2000). This exerts a substantial force on the particulate phase, requiring particles to be large (dm-scale, B. Houghton, personal communication, 2005) and dense to break this control and travel independently (i.e. ballistically). In strombolian plumes, however, the low plume densities (as mentioned, likely <1 kg m$^{-3}$) and low velocities (generally 20-50 m s$^{-1}$, e.g. Chapter 2; Chouet et al., 1974) allow much smaller (cm-scale; Chouet et al., 1974) and less dense (1000-1500 kg m$^{-3}$; McGetchin et al., 1974; N. Lautze, personal communication, 2005) particles to decouple and follow largely independent trajectories.

Also with regard to bulk density, it is worth noting that of the 171 total ash plumes in my 2004 dataset (considering all craters), none collapsed or had portions which collapsed. This disinclination for collapse is likely the result of two factors. First, the potential for high gas mass fractions due to deep gas decoupling enables initial buoyancy, as discussed, without the reliance on entrained air to lower bulk density. Second, the solid particles are sourced from material which has a continuous hot passive gas flux streaming through it, maintaining high temperatures of the particles. This would minimize the heat transferred from the gas to the particles, enabling higher rates of gas expansion and lower bulk densities. Assuming instantaneous heat transfer for simplicity,
the critical $n$ for initial buoyancy is 0.14 for particles with an initial particle temperature of 1000 °C (with the gas also at 1000 °C), while the critical $n$ for initial buoyancy is 0.26 for particles at 20 °C (with gas again at 1000 °C).

Eruptions occurring at other volcanoes with decoupled gas flow (i.e. high erupted gas mass fractions) and continuous passive degassing (i.e. higher temperatures of accidental particles) should exhibit a similarly low likelihood for collapse – regardless of initial momentum. Conversely, the lack of these conditions (i.e. low gas mass fractions or cold entrained particles) would encourage high density plumes and the possibility of collapse dependant, ultimately, upon initial momentum and air entrainment rates.

3.8 Summary and conclusions

In this study I have shown how FLIR video can convey the varying thermodynamic characteristics of small-scale ash plumes. A selection of 25 ash rich eruptions at Stromboli volcano, acquired in June and July 2004, was analyzed. The rise history of strombolian ash plumes was different for Type 2a and Type 2b eruptions, due to different initial velocities (Type 2a: >15 m s$^{-1}$; Type 2b: <15 m s$^{-1}$). These velocities are linked to the velocity differential between the plume front and the feeder plume. This, in turn, controlled the plume form, which included jets, starting plumes and rooted thermals. Jets had lower spreading rates (mean spreading half-angle of 7°) and therefore lower entrainment rates, while rooted thermals (with their vigorous ring vortex) spread at higher rates (mean spreading half-angle of 14°), indicating greater entrainment rates.
Modeling of transient plume dynamics at this scale must account for these rapidly changing entrainment regimes. Plume front temperatures began at values extending from 300 °C to slightly higher than 550 °C, and dropped at a mean rate of 3-4 °C m⁻¹ in the first 100 m of vertical ascent.

Different days exhibited different ranges of plume front temperatures. June 11+13, for instance, was generally a period of higher temperature plumes, while June 07 was limited to lower temperature plumes. This may have been controlled by the amount of backfill in the conduit and, thus, the mass of fine particles that the plume could entrain.

Extracting ejecta mass from the thermal imagery is troublesome, mainly due to the uncertainty in gas mass fraction. As all particles in strombolian plumes are incidental, a wide range of gas mass fractions is possible. Ejecta mass values in the plume front had a mean value of 3150 kg (for \( n=0 \)) to 840 kg (for \( n=1 \)). By a simple extension, this corresponded to between 10,460 kg (\( n=0 \)) to 2760 kg (\( n=1 \)) for the whole plume. Previous attempts at estimating gas mass fraction are either not applicable to ash plumes, or incorrect. My attempt at constraining the gas mass fraction resulted in values generally between 0.2 and 0.5, which is significantly larger than those for a plinian eruptions (<0.05). This enables strombolian plumes to have an initial bulk density lower than that of air, reducing the likelihood of collapse. In addition to the quantitative analyses, I observed sedimentation, helical spin, and plume sloughing (related to temperature variations) in the imagery.
The FLIR was limited mainly by its inability to provide a direct constraint on the gas mass fraction, the major uncertainty in the study of heat and mass content. The field of view (24°) was another limitation, and may have prevented observing the development of discrete thermals at higher altitudes. Wider field of view lenses (45°) are available, and may improve plume visualization.

In conclusion, the FLIR offers a tool for fine timescale analyses of plume dynamics and thermal properties. Applying the techniques described here to other types of plumes (e.g. Vulcanian) could help constrain rapidly changing thermodynamic behaviors which, up until now, have only been observed in the visible spectrum. The FLIR should also be useful to analyze the circumstances of plume collapse, or subsequently, the thermodynamics of a developing pyroclastic flow.

3.9 Appendix: Constraining gas mass fraction of ejecta

I propose the following method for constraining the gas to ash mass ratio in a plume. My approach consists of three main steps: 1) estimating gas flux by measuring continuous gas velocity near the vent, 2) estimating the instantaneous above-background heat for the entire plume (not just the plume front), and then 3) reconciling these values in terms of heat content.

1) Estimating gas flux: The gas velocity was measured at the crater rim in the 30 Hz data using a cross correlation routine. A small window (~17 m wide) was placed at the crater rim, and the image data in the window were matched with features in a successive image. The window displacement and time duration were then used to
calculate velocity. The precision is rather coarse (~6 m s^{-1}) due to the pixel size of the FLIR, but values agree well with instantaneous measurements performed manually. The velocity was multiplied by a reasonable gas density (0.2 kg m^{-3}) and area (assuming a cylindrical plume = \pi r^2) using the jet radius (r) measured at the crater rim (as an overestimate). The actual vent radius is on order of 1-2 m (Chouet et al., 1974; and my observations based on several exceptionally high-velocity jet eruptions), however, the jet will be wider above the vent due to adiabatic decompression.

2) Estimating total heat in plume: This was performed by taking an image where the plume was either a) close to the top of the field of view or b) near its final moment of opacity. The gas phase will consist of a mixture of both air and volcanic gas. As a necessary simplification, I assumed the bulk density of the plume was approximately that of air at the observed temperatures, and that the specific heat was roughly that of air as well (1000 J kg^{-1} K^{-1}). Ballistic pixels were removed from the image with a morphological image processing routine (Chapter 4). The volume of the plume was measured by assuming axial symmetry of the plume, so that the plume is a stack of disks one pixel thick with the observed radii in the image. The average temperature was measured for each disk, and density was garnered with the ideal gas law. Multiplying the volume by density gave mass, and multiplying this by specific heat gave heat. The heat in each disk in the plume was summed to obtain total heat content of the plume in the snapshot.
3) *Reconciling the heat budget:* The integrated gas flux from the eruption start to the time of the snapshot used in step 2 was calculated from the gas velocity results (step 1), and its heat contribution to the total heat (step 2) was subtracted, leaving heat \( Q_{\text{ash}} \) that was contributed by the solid phase. From this, ash mass \( m_{\text{ash}} \) was calculated using

\[
Q_{\text{ash}} = m_{\text{ash}} c_{\text{ash}} \Delta T,
\]

where \( c_{\text{ash}} \) is the ash specific heat and \( \Delta T \) is the initial temperature of the ash above background (500-1000 °C). A rough time-averaged gas mass fraction \( n \) can then be constrained.
Chapter 4

Automated analysis of small volcanic explosions using thermal (FLIR) imagery

4.1 Introduction

The activity at Stromboli volcano, Italy, is normally characterized by small, intermittent explosive events from the summit craters at intervals of ~10 minutes with ejecta heights <300 m. This typical activity is punctuated or interrupted by effusive events, which have a mean repose period of 3.7 years and generally last a few months (Barberi et al., 1993). The last four effusive events (1976, 1986, 1993, 2003) at Stromboli were preceded by increases in the size and frequency of the summit explosive activity (Capaldi et al., 1978; Smithsonian Institution, 1986; Carniel et al., 1996; Calvari et al., 2005). While lava flows themselves have not historically posed a danger to tourists or island inhabitants, the longer effusive phases represent a prolonged destabilization of the shallow conduit which has sometimes been concurrent with paroxysmal eruptions (producing pyroclastic flows and widespread ballistic fallout) and tsunamis (Barberi et al., 1993). The 2002-2003 effusive phase involved partial collapse of the lower flanks resulting in a tsunami which damaged shorefront property at Stromboli village and injured several people as well as a violent paroxysmal eruption on 05 April 2003 with block fallout on Ginostra village and (Bonaccorso et al., 2003; Calvari et al., 2005). This
effusive phase, the best monitored yet (Calvari et al., 2005; Ripepe et al., 2004; 2005),
was preceded by a significant ramping-up of explosive levels in the six months preceding
the onset of lava flow activity (Calvari et al., 2005). Also, the collimation of explosions
markedly decreased in the hours before the first lava flows appeared (Calvari et al.,
2005). This suggests that monitoring the normal explosive behavior at Stromboli could
be a predictive tool for anticipating anomalous eruptive phases and the associated hazards
for tourists and residents.

The normal explosive behavior at Stromboli has been monitored by permanent
arrays of seismometers, infrasonic microphones, and spot radiometers which give a good
indication of the intensity of the eruptions (Ripepe and Marchetti, 2002; Ripepe et al.,
2002; Ripepe et al., 2004). These data cannot, however, characterize the appearance of
the eruptions, which can convey additional information on the nature of the shallow
conduit. For instance, collimation may indicate the depth of the magma free-surface, and
eruption style (ash-rich or ballistic-rich) may hold clues as to conduit fill levels in the
upper parts of the magma column (Chapters 2+3). Also, eruption images are simply the
most straightforward data from which to calculate ejecta mass (a valuable parameter in
itself) and involve less potential error than assuming proxy values from seismic or
infrasonic data.

In this study, I present an automated method to extract eruption characteristics
from Forward Looking Infrared Radiometer (FLIR) imagery. The output of the
algorithm is three-fold. First, it characterizes the style of the eruption (ash-dominated,
ballistic-dominated, or mixed) based upon the classification scheme presented in Chapter
2. Second, the collimation of the ballistic ejecta is estimated. Third, it calculates explicit mass values for the ballistic and plume components. The algorithm was test-run on a dataset of 125 eruptions collected at Stromboli during June-July, 2004.

4.2 Background

4.2.1 Stromboli volcano

Stromboli volcano lies in the Aeolian Islands north of Sicily, in the Tyrrhenian Sea. Its notably frequent eruptive activity, consisting of ~10 eruptions each hour (Harris and Stevenson, 1997), has existed for the last ~1500 years (Rosi et al., 2000). The eruptions originate from a 300 m long crater terrace, composed of three main craters, at an elevation of ~800 m above sea level (Fig. 4.1). Ejecta are composed of incandescent lapilli and bomb size ballistic material as well as finer particles which can comprise a dark, convecting plume (Chapter 3). Ballistic ejecta are commonly thrown to <300 m above the active vents, and typically fall within 100 m of the source. In Chapter 2 I performed a study of 344 eruptions from 2001-2004 using a FLIR camera, and classified two major eruption styles (Fig. 4.2). While Type 1 eruptions were dominated by ballistic particles and lacked a significant ash plume, Type 2 eruptions were dominated by an optically-thick ash plume. The latter were subdivided into Type 2a eruptions, which had a gas thrust phase (velocities >15 m s⁻¹) and typically significant ballistic particles, and Type 2b eruptions, which exhibited only buoyant velocities (<15 m s⁻¹) and usually lacked a significant amount of ballistic particles.
Figure 4.1. View of the western flank of Stromboli volcano, Italy. The active crater terrace is ~800 m above sea level. The Rocette location (ROC), site to the current continuously operating FLIR camera, is ~450 m from the active craters. An ash-rich eruption can be seen emanating from NE crater. The view from ROC can be seen in Fig. 4.13.
Figure 4.2. Eruption styles and their FFT equivalent. Left: FLIR images of primary eruption styles at Stromboli. Right: Corresponding 2D discrete FFT images (edges are high spatial frequency, center is low frequency). In the ballistic-rich Type 1 eruption (high spatial frequencies), higher amplitudes are distributed away from the center of the FFT image. In the ash-rich Type 2b eruption (low spatial frequencies), higher amplitudes are concentrated towards the FFT image center (see Fig. 4.5). The FFT images are computed after subtracting out the crater region and thresholding at a modest level, leaving just airborne ejecta.
4.2.2 FLIR instrument and measurement setup

In Chapter 2 I used a FLIR camera, which is a thermal video camera utilizing an uncooled microbolometer array sensitive to longwave (7.5-13 \( \mu \text{m} \)) radiation. Longwave imagery is better suited to show Strombolian ejecta than visible (e.g. Chouet et al., 1974; Blackburn et al., 1976) or near-infrared imagery (e.g. Ripepe et al., 1993), as described in Chapter 2. FLIR cameras had been used solely in campaign style at Stromboli until December 2003, when a permanent thermal camera (Omega Micron) was installed at the Rocette site, ~450 m from the crater terrace (Fig. 4.1). This paper presents an algorithm to calculate Strombolian particle mass in such a setup. The method, however, relies on temperature-calibrated imagery, which is not offered by the analog model (Omega Micron) currently in place. Therefore, I apply the method to 125 eruptions captured in 2002-2004, in campaign style, by a temperature-calibrated camera, i.e. a ThermaCAM™ S40 manufactured by FLIR Systems, Inc. I anticipate that such temperature-calibrated FLIR cameras will eventually be installed in a suitable location. One is already operating from the lower flanks having been installed by the Istituto Nazionale di Volcanologia (INGV). However, this camera is too distant to provide images suitable for my analysis, as proved by collecting images with my own ThermaCAM™ S40 during September 2003. These images were collected from the 450 m elevation and ~1.5 km distant from the vents, i.e. roughly the same location as the INGV installation. The images proved to
be of insufficient spatial resolution to apply the techniques and methodologies given in this thesis.

The ThermaCAM™ S40 produce an image of 240 x 320 pixels, covering a field of view (FOV) of 18° x 24°, giving an individual pixel FOV of 1.3 milliradians. The data used in this study originate from Rocette (ROC) station (450 m distance), in which the vertical FOV is ~140 m and the pixel size ~60 cm. The pixels yield a pixel-integrated temperature, following a correction for emissivity (0.95 for basalt, Salisbury and D’Aria, 1992) and atmospheric attenuation. Atmospheric correction is performed by the camera using a built-in LOWTRAN routine, relying on user-input distance, ambient temperature and relative humidity. The latter two were measured periodically in the field during each acquisition session. More details on the instrumentation and acquisition setup can be found in Chapters 1 and 2.

4.2.3 Eruptive activity during June-July 2004

The algorithm is applied here to 125 eruptions from NE crater in summer 2004. These eruptions were collected between June 05 and July 25, 2004, at 30 Hz from ROC station (Fig. 4.1). All originate from a single vent (NE-1, Fig. 4.1; see also Fig. 2.2 in Chapter 2). The June-July 2004 period at NE-1, in general, represented a moderate level of activity with ~6 eruptions per hour and most eruptions casting ballistic ejecta to <200 m above the vents. Eruption style, however, varied by observation day. June 05, as well as July 07-18, exhibited Type 1 eruptions. Eruptions on June 07-27, as well as on July 20+25, were Type 2 style (Chapter 2).
4.2.4 The spot radiometer

The FLIR-derived mass estimates from this algorithm were compared to results from a spot radiometer. These radiometers have been recording continuously at Stromboli since May 2002 (Harris et al., 2005). They have a single 15° FOV, are sensitive to radiation in the 8-12 micron range, and are housed in weather-proof cases. At a distance of 450 m the FOV is ~120 m in diameter, which is approximately the width (i.e. short axis) of the crater terrace. Data are acquired continuously at 54 Hz, and the instruments remained operational throughout the 28 December 2002 and 05 April 2003 paroxysms (Harris et al., 2005). The radiometers offer two distinct advantages over the FLIR. First, they have more continuous data coverage than either the Omega Micron thermal camera, which can acquire up to 30 Hz but due to archiving limitations data are saved at only 1 Hz, or the campaign-style FLIR. Second, the 15° radiometer system costs about $1,150 USD (Harris et al., 2005), compared with FLIR cameras which generally cost between $15,000 (for the Omega Micron) and $50,000 (for a FLIR S40). I was interested, therefore, in exploring how well the 15° radiometer response correlated with eruption mass calculated by the FLIR, an assessment that would allow me to judge whether the spot radiometer could act as an inexpensive substitute for eruption mass calculation and tracking.
4.3 Methods for automated analysis

The algorithm was created in Matlab. An outline of the program is as follows:

1) Preprocessing,

2) Identification of eruption images,

3) Isolation of airborne ejecta,

4) Determination of eruption style,

5) Isolation of plume and ballistic components,

6) Estimation of instantaneous mass from the plume and ballistic images,

7) Estimation of ballistic collimation.

4.3.1 Preprocessing steps

Images must be corrected for emissivity, distance-to-target, ambient temperature and relative humidity using the ThermaCAM™ software. The images are then exported from the native ThermaCAM™ format to a Matlab data structure, containing the image array (of Kelvin temperature values) along with several accessory variables holding the date, time, and image settings (e.g. emissivity and relative humidity).

4.3.2 Identification of eruption images

Any data covering a significant amount of time (e.g. hours) will consist mainly of non-eruption images, briefly interrupted by short (~15 s) periods of eruptive activity. Flagging those images in an image sequence which display a Strombolian eruption can be achieved by examining the standard deviation within a running difference-image. The
algorithm runs through the total number of images in a given period, subtracting the previous image from the current image to produce a running difference-image. Two sequential images during non-eruptive periods tend to have similar temperatures, and thus yield low standard deviations. However, the injection of material during an eruption causes a change in the image properties between each image, such that two sequential images become significantly different. Thus the standard deviation values of all the difference-images throughout an observation period are checked for spikes, with values for eruption difference-images standing out clearly from those for non-eruption difference-images (Fig. 4.3). The explicit flagging is most easily accomplished by simply using a standard deviation threshold above the median value for all difference-images (a value of 1-2 Kelvin has proven to be reliable), or by finding local maxima in the standard deviation and tracing the standard deviation plot around these peaks backward and forwards to the median to determine the start and end of each eruptive event. The maximum temperature, as opposed to standard deviation, of each difference-image can also be used for detection, however this typically fails to flag images later in the eruptive sequence which are dominated by cooled material.

4.3.3 Isolation of airborne ejecta

In order for the airborne ejecta to be isolated, ‘background’ regions (areas with no ejecta) must first be identified and removed. This approach is unique to each camera location and viewing geometry. As these test data were acquired from the Rocette station (Fig. 4.1), and there is now an operational FLIR (Omega Micron) at the same spot, I
Figure 4.3. Identifying eruptive events with the standard deviation. The standard deviation is calculated for the running difference-image (previous image subtracted from the current image), with eruptive images standing out clearly from non-eruptive images.
present an example approach suited to this viewing geometry. The steps, illustrated in Figure 4.4, are as follows:

1) Using one non-eruptive image, the typical range of atmospheric temperatures and crater terrace temperatures are determined and used to threshold the image at a reasonable value. A value of 25 °C was generally effective for June-July 2004 (Fig 4.4b), so that all pixel temperatures above 25 °C are classified as ground (crater terrace), and all below are sky. It is in this ‘sky’ region that ejecta will be later searched for.

2) A binary “opening” morphological operator, followed by a “closing” operator, are used to remove noise and smooth jagged features (Gonzalez et al., 2003).

3) An edge-detection routine is run (Gonzalez et al., 2003), creating a line showing the upper rim of the crater, which separates the ground from the atmosphere.

4) The line coordinates are used as a lookup table to denote the highest extent of crater pixels. No ejecta are to be considered beneath this line.

5) In subsequent eruptive images, ejecta can be isolated using a simple threshold (e.g. 25 °C) for the region above the crater rim masking line. This is possible because the apparent sky temperature in the FLIR imagery (looking into space) is generally <0 °C.

4.3.4 Determining eruption style

The eruption style (Type 1, 2a, or 2b) can be determined quickly using the 2D discrete fast Fourier transform (FFT) of an image (see Gonzalez et al., 2003, for FFT use
Figure 4.4. Image processing to isolate ejecta components. The original image (A). The regions of warm ground pixels are identified from the last non-eruptive image (B) using a simple threshold, in order to isolate airborne eruptive material. Morphological operations then separate this ejecta image into ash plume (C) and ballistic (D) component images.
in image processing). The FFT of the eruption image characterizes the range of spatial frequencies present, and thus is useful for roughly judging the relative amounts of plume and ballistic particles (Fig. 4.5). Ballistic pixels are spatially discrete due to their small sizes (sub-pixel) and high dispersal, and fall in the high frequency domain. Ash plume pixels, however, exhibit lower frequencies because of their greater spatial continuity. In the 2D FFT image, therefore, the relative amplitudes of data in the high frequency and low frequency domains will be significantly different for Type 1, 2a and 2b eruptions (Fig. 4.2).

Before the FFT is applied, the most recent non-eruptive image is subtracted from the eruptive image. This temporary difference image is then thresholded in order to reduce the image to airborne ejecta. For the purposes of the FFT, this approach to isolating airborne ejecta is adequate, and simpler than the isolating routine presented in the next section. Once the FFT has been used to characterize the abundance of frequencies present in the image, I ratio the mean squared amplitude in the high frequency domain to that in the low frequency domain, using a simple threshold to divide the two regions. This produces a single number to characterize relative ballistic vs. ash abundance in each eruption image. Eruptions consist of multiple images, and I found it best to select the maximum ratio value among images which fell in a period 6-12 seconds after the eruption start. This time period maximized the dispersal, and detection, of ballistic ejecta.

A challenge was choosing the ideal threshold between high and low frequency domains. I used 125 eruptions from the June-July 2004 dataset as a test sample, and
Figure 4.5. Using the FFT to isolate plume and ballistic particles. The 2D Fast Fourier Transform (FFT) characterized the range of spatial frequencies present in the ejecta images. Ash plume pixels were found to fall in the low frequency part of the spectrum, while ballistic pixels fell in the high frequency portion.
varied the threshold over the course of multiple FFT runs. I then used multiple discriminant analysis on these sets of results to find the most effective threshold for determining eruption style, producing the most accurate discriminant functions by comparing predicted eruption styles to those defined qualitatively by eye.

Spatial frequencies in any image range from the DC frequency (by row: 240\(^{-1}\) cycles pixel\(^{-1}\), and by column: 320\(^{-1}\) cycles pixel\(^{-1}\), as a FLIR image is 240 x 320 pixels) to the Nyquist frequency (2\(^{-1}\) cycles pixel\(^{-1}\)). The ideal threshold was determined to lie at 20% of the Nyquist frequency, but this is likely unique for this viewing geometry (i.e. 450 m distance). This threshold produced discriminant functions which had a 67% success rate in distinguishing Type 1, 2a and 2b eruption styles, and an 81% success rate in distinguishing Type 1 and 2 styles.

4.3.5 Isolation of plume and ballistic components

While the FFT is useful for style discrimination, it is not entirely effective at explicitly isolating plume and ballistic components. For instance, I originally attempted to reconstruct ballistic-only images using the high frequency FFT data alone, but found this produced images which contained ‘ballistic’ pixels on the fringe of the ash plume, where sharp contrasts (and high frequencies) exist. Instead, I found that morphological operators produced a more accurate separation of plume and ballistic pixels (Fig. 4.4). I used an “opening” filter (Gonzalez et al., 2003) to remove ballistic pixels and produce an image showing only the ash plume. The difference between the original image and the ash plume image is used to identify ballistic pixels (using a modest threshold) and make a
ballistic-mask. The mask is then applied to the original image to garner ballistic pixels with their original, unadulterated temperature values. Once the plume and ballistic component images are created, individual approaches are then taken to calculate explicit mass values.

4.3.6 Estimation of total mass from plume and ballistic images

4.3.6.1 Plume mass calculation

In Chapter 3 I detailed a heat conservation approach to calculating the ejecta mass in the plume front, which forms the circulating head of the plume and is trailed and supplied by the feeder plume. In Chapter 3, I applied this mass estimation method to the plume front only, as it was the part of the plume most amenable to manual tracking and measurement. The other benefit of the plume front for mass estimation is that it is likely the most well-mixed portion of the plume, due to constant circulation that can bring hot material from the inner portions of the plume to the exterior (Turner, 1962) on a relatively short timescale.

In the algorithm presented here, total mass in the plume must be measured, and I must therefore justify complete thermal mixing in the feeder plume if temperature measurements are to be made there. The typical model for a plume involves a hot, high velocity inner core surrounded by an annulus of cold (due to air entrainment), low velocity (due to turbulent shear) material, and a limited amount of observations supported this model for the feeder plume. During high velocity onsets, and during the first ~20 m of plume ascent, inner portions of the plume occasionally appeared to be rising at the
faster rate than material at the exterior. These inner portions could move to the exterior, and had higher temperatures than the remainder of the plume. These occurrences, however, were limited in both time and space, and during the majority of the rise of the feeder plume the plume exterior was composed of entrainment vortices that were on the same scale as the radius of the feeder plume itself. These appeared to draw material from the interior to the exterior on a small timescale, and thus support the idea that the feeder plume, like the plume front, is reasonably well-mixed for most of its lifespan.

The first step is to measure the volume of the entire plume in one pixel height slices, assuming axial symmetry over that small height increment (Fig. 4.6). Next, the bulk temperature ($\Delta T_{\text{bulk}}$) is taken as the average FLIR temperature over that increment. To determine the ejecta mass in each one-pixel plume slice, the heat budget equation from Chapter 3 is applied:

$$Q_{\text{final}} = Q_{\text{initial}}$$  \hspace{1cm} 4.1

$$[m_{\text{ash}}c_{\text{ash}} + m_{\text{air}}c_{\text{air}} + m_{\text{gas}}c_{\text{gas}}]\Delta T_{\text{bulk}} = m_{\text{ash}}c_{\text{ash}}\Delta T_{\text{ash}} + m_{\text{gas}}c_{\text{gas}}\Delta T_{\text{gas}}$$  \hspace{1cm} 4.2

where, for the air, ash and gas components of the plume, $m$ is the respective mass value, $c$ is the respective specific heat value, and $\Delta T$ is the temperature above background ($\Delta T_{\text{ash}}$ and $\Delta T_{\text{gas}}$ denote initial temperatures of ash and gas, respectively). I solve for two end-member solutions for the gas mass fraction ($n$), where the heat is either attributed entirely to gas ($n=1$) or ash ($n=0$). I assume $\Delta T_{\text{gas}} = 1000 \, ^\circ C$ and $\Delta T_{\text{ash}} = 500-1000 \, ^\circ C$ (Chapter
Figure 4.6. Plume volume calculation. The approach assumes axial symmetry, using a stack of circular disks (1 pixel thick) fit to the shape of the cloud.
3). This method disregards radiative heat loss, which I have already shown to be insignificant (Chapter 3).

4.3.6.2 Ballistic mass calculation

This method uses a two-component radiance budget to estimate the size, and then the mass, of ballistic particles in the ballistic component image. Each pixel is ~60 cm in dimension, at 450 m distance, which is likely much larger than the vast majority of ballistic particles. I thus assume all ballistic particles are sub-pixel. For instance, Chouet et al. (1974) measured the mean particle size for two eruptions at Stromboli to be 2.4 cm, with a maximum size of 32 cm. Ripepe et al. (1993) found that nearly all particles were <30 cm in their analysis of infrared photos of Strombolian explosions.

For each ballistic pixel, extraction of the a two-component radiance budget entails calculating: 1) the pixel-integrated radiance ($R_{\text{int}}$), 2) the background radiance ($R_{\text{bg}}$), taken from the ash plume component image, and 3) the theoretical ballistic radiance ($R_{\text{ball}}$). For $R_{\text{ball}}$ the ballistic particles are assumed to be 500-1000 °C, as 500 °C is the lower limit of incandescence (Macdonald, 1972) and incandescence is a common characteristic of these explosions. The fractional area ($p$) of the ballistic particles in the pixel can now be calculated with:

$$R_{\text{int}} = pR_{\text{ball}} + (1-p)R_{\text{bg}}$$
All radiance values are adjusted for the FLIR detector response (Chapter 1). The ballistic particle area is calculated from the fractional area ($p$) and the total pixel area. Assuming a spherical particle to calculate volume, along with a reasonable ballistic density (1100 kg m$^{-3}$, N. Lautze, personal communication, 2005), a range of ballistic mass values (for ballistic temperatures of 500-1000 °C) can be calculated for each ballistic pixel in the ballistic component image.

4.3.7 Collimation

Collimation is the degree to which ballistic ejecta paths are parallel, and this parameter may be important for judging the relative height of magma in the conduit. My method simply measures the distribution of ballistic pixels among the vertical image columns (Fig. 4.7). The ballistic image is scanned, and the column number of each particle is stored in an array. The standard deviation of this array (converted from pixels to meters) is a proxy for the lateral dispersal of ballistic ejecta. Lower standard deviation values (i.e. lower dispersal) indicate increased collimation.

4.3.8 Instantaneous mass estimation

As the images are only 1 s apart, many images will contain ash and ballistic particles included in previous and subsequent images, and therefore a simple summing of image mass values will produce over-estimates. Following Ripepe et al. (1993) I address this problem by using the maximum mass value for ash and ballistics during each eruption. Maximum ash mass tends to occur when the plume is nearing the top of the
Figure 4.7. Estimating the collimation of ballistic ejecta. The algorithm measures the lateral (horizontal) standard deviation of ballistic pixels in the ballistic component image. A) Very low collimation eruption from 05 June 2004 (see plot). B) Very high collimation eruption from 14 July 2004. At the bottom is the graph of mean collimation proxy values for 90 eruptions from NE-1 crater in 2004.
field of view, and the maximum ballistic mass occurs when the ballistic fountain is at its apex. The maximum mass will generally not, however, account for the complete mass ejected during the eruptive event, and these results must be considered 'instantaneous' mass values as opposed to total, integrated, erupted mass. Any mass erupted after the maximum mass value will be unaccounted for, and the magnitude of this portion is estimated in the results section.

The instantaneous mass is, nevertheless, a highly desirable value to calculate for several practical reasons. First, instantaneous mass is more amenable for extraction from the simple, low-bandwidth (i.e. 1 Hz) algorithm presented here. In order to account for additional mass following the maximum mass value, some knowledge of the residence time of plume and ballistics must be gauged from the imagery. This may be possible with higher frame rate data (e.g. 30 Hz, see Appendix in Chapter 3 for such a method), but it is difficult or impossible with 1 Hz data. Second, instantaneous mass from the FLIR imagery is the most straightforward result to compare to the amplitude of the spot radiometer signal recorded during a strombolian event, which is a component of this study.

4.4 Results I: June-July 2004 eruptions

The algorithm was test run on 125 eruptions from NE-1 crater during June-July 2004. The original 30 Hz data were sub-sampled to 1 Hz to emulate an operational data stream.
4.4.1 Collimation

Based upon a comparison with visual impressions of the eruptions, the collimation routine proved effective at characterizing the lateral spread of ballistic ejecta (Fig. 4.7). Eruptions on June 05 appeared poorly-collimated (and thus, had high values of lateral dispersal; mean: $\sim 33$ m), while eruptions on July 14 were well-collimated (low values of lateral dispersal; mean: $\sim 11$ m).

4.4.2 Eruption styles from FFT ratio values

The overall distribution of FFT ratio values is shown in Figure 4.8 for each of the primary eruption styles. Type 1 eruptions had a range of 0.50-0.97 (mean: 0.67), while Type 2 eruptions had a range of 0.38-0.67 (mean: 0.50). Type 2a eruptions had a range of 0.41-0.67 (mean: 0.54) and 2b eruptions had a range of 0.38-0.57 (mean: 0.45). Because NE-1 crater showed a progression of eruption styles during June-July 2004 (Chapter 2), the FFT ratio values likewise changed in time (Fig. 4.9). Ratio values during the Type 2 periods were generally low, correctly indicating high ash contents. Those during Type 1 were high, indicating ballistic dominance. One notable departure was the high ratio value on July 25, showing high ballistic proportions in a Type 2 phase. As discussed in Chapter 2, July 25 exhibited ballistic-rich Type 2 eruptions which appeared to be transitioning into Type 1 activity.
Figure 4.8. FFT ratio results for different eruption styles. Higher ratios (i.e. higher relative amounts of ballistic pixels vs. ash plume pixels) corresponded to Type 1 eruptions, compared to low values for Type 2 eruptions.
Figure 4.9. FFT ratio values and initial velocity through time. Daily average FFT ratio values for June-July 2004, at NE-I crater.
4.4.3 Mass values

Plume mass was calculated for each of the end-member scenarios, these being: 1) gas mass fraction \((n)\) of the ash plume extending from 0 to 1, and 2) the initial ash temperature extending from 500 to 1000 °C. Ballistic mass was also calculated for ballistic temperatures extending from 500 to 1000 °C. The total range of uncertainty in plume mass values can be expressed by a mean value with a ±60% uncertainty, while the mean ballistic mass results have an uncertainty of ±70%.

Mean plume mass values (Fig. 4.10; Table 4.1) extended from 0 to 40,800 kg (overall mean: 3100 kg). Type 2a eruptions had plume mass values ranging from 120 to 40,800 (overall mean: 7100 kg), while Type 2b had values ranging from 2 to 10,850 kg (overall mean: 880 kg). Type 1 eruptions could register appreciable plume masses (up to 4000 kg) due to partial radiance from the plume or ballistic particles misclassified into the plume component image. I disregard these values, however, as the plume is translucent and accurate temperature measurements cannot be made – thus, plume mass results in these cases would be inaccurate. As I explain in the discussion, these results depict only ‘observable’ mass.

The mean ballistic mass values (Fig. 4.10; Table 4.1) extended from 0 to 6230 kg, (overall mean: 500 kg). Mean Type 1 ballistic values had a range of 1-6230 kg (overall mean: 470 kg), and Type 2 had a range of 0-6200 kg (overall mean: 520 kg). Type 2a
Figure 4.10. Mass results by eruption style and ejecta type. Results for 118 NE-1 crater eruptions in June-July 2004. Note that ballistic mass estimates have an uncertainty of +/- 70%, while the plume mass values have an uncertainty of +/- 60%. Also note that because gas mass fraction is unknown, it is not possible to differentiate the heat source (gas or ash) in the plume.
Table 4.1. June-July 2004 ejecta mass results (kg)*

<table>
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<th>Style</th>
<th>#</th>
<th>Ballistics</th>
<th>Plume</th>
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<tr>
<td>Overall</td>
<td>118</td>
<td>500 (0-6230)</td>
<td>3100 (0-40,800)</td>
</tr>
<tr>
<td>Type 1</td>
<td>41</td>
<td>470 (1-6230)</td>
<td>-</td>
</tr>
<tr>
<td>Type 2</td>
<td>77</td>
<td>520 (0-6200)</td>
<td>3100 (0-40,800)</td>
</tr>
<tr>
<td>Type 2a</td>
<td>43</td>
<td>950 (3-6200)</td>
<td>7100 (120-40,800)</td>
</tr>
<tr>
<td>Type 2b</td>
<td>34</td>
<td>30 (0-320)</td>
<td>880 (2-10,850)</td>
</tr>
</tbody>
</table>

* shown as mean (min-max)
eruptions had mean ballistic mass values of 3-6200 kg (overall mean: 950 kg) and Type 2b had a range of 0-320 kg (overall mean: 30 kg).

These ballistic mass results, however, must be considered minimum estimates of the actual ballistic mass due to two reasons. First, the ballistic detection method (i.e. the morphological operator) relies upon the contrast between ballistic pixels and their surroundings. I noticed that several Type 1 eruptions of relatively similar magnitude (judged visually) had markedly different ballistic mass values (in one case, by up to 8 times) due to differences in the number of ballistic pixels detected. Controlling the number of detected ballistic pixels was the degree to which faint plume (within which the ballistics were contained) had obscured or dampened the ballistic contrast. In other words, any faint plume between the sensor and the ballistic could act as a low-pass filter and reduce the effectiveness of the morphological operator (a high-pass filter). In this case, therefore, the ballistic particles would be under-detected. Faint plume is also typically low temperature, and can thus lower the apparent temperature of any ballistic pixels which are detected through the plume, decreasing the size and mass of ballistic results. Second, in the case of Type 2 eruptions, the opaque plume will simply block any ballistic particles residing within or behind it. In general, due to these factors of obfuscation, as well as the inherent limitation of coarse pixel size, the FLIR as used at this range is not ideally suited for total mass measurements of ballistic particles.

For those ballistic pixels which were detected, however, the radiance budget approach produced plausible ballistic dimensions (Fig. 4.11). The results do not depict the entirety of ballistic sizes, as the morphological operator has an inherent detection
Figure 4.11. Ballistic particle sizes and mass values. Particle size distributions for end-member assumed particle temperatures, using the radiance budget approach. Results are shown for all eruption images for all 125 eruptions - hence, a particular particle may appear more than once. Note that the number of particles is lower for the T=500 °C solution because a limited number of ballistic pixels had pixel integrated temperatures greater than 500 °C and, thus, no solution was possible. Mass values were calculated using the dimensions and assuming a spherical shape, and density of 1100 kg m$^3$. 
threshold. The smallest particle detected with the $T_{\text{particle}}=500$ °C solution was 4.3 cm, while the smallest for the $T_{\text{particle}}=1000$ °C solution was 2.6 cm. The mean ballistic sizes, for detectable particles, are 9.2 cm for the $T_{\text{particle}}=500$ °C solution and 5.3 cm for the $T_{\text{particle}}=1000$ °C solution. Figure 4.11 shows that the particle distribution is highly skewed towards smaller sizes, probably reflecting a large number of particles smaller than the detection threshold which are not accounted for. Chouet et al. (1974), using a technique with a much lower spatial detection threshold (2.5 mm), measured mean particles sizes of 2.2 and 2.5 cm, with 99% of the particles <6 cm and the largest particle being 32 cm. Ripepe et al. (1993) used a technique with a 10 cm detection threshold, and showed that nearly all particles were <30 cm. Their distribution is very similar to the $T_{\text{particle}}=500$ °C solution within the range that these two datasets overlap. My results agree with Chouet et al. (1974) in that the mean particle size is <10 cm. Furthermore, these results agree with both Chouet et al. (1974) and Ripepe et al. (1993) that virtually all particles erupted in Strombolian eruptions are <30 cm in size, and the majority fall in the region <15 cm.

These ballistic dimensions can be translated to mass values by assuming a spherical shape and a reasonable density (1100 kg m$^{-3}$, N. Lautze, personal communication, 2005) (Fig. 4.11). For $T_{\text{particle}}=500$ °C, the majority of particles are less than 1 kg, with a median value of 300 grams. For $T_{\text{particle}}=1000$ °C the majority are less than 500 grams, with a median value of 60 grams. Again, these values are limited by the detection threshold mentioned earlier.
4.4.4 Relation of instantaneous to total mass

For the 125 eruptions analyzed, the times of maximum instantaneous mass were compared to the total eruption duration. For the plume mass, the maximum mass value occurred, on average, at 52% of the total eruption duration (st. dev.: 22%). For the ballistic mass, the maximum mass value occurred at 40% of the total eruption duration (st. dev.: 21%). In general, however, the mass flux of an eruption is not uniform through time, but instead exhibits a rapid waxing followed by a prolonged waning (Fig. 4.12, see also Ripepe et al., 1993).

A better judge of instantaneous mass relative to total mass would be based upon the trend of mass flux through time, which could account for the higher mass flux at the onset of eruptions. In the Appendix of Chapter 3, a method was presented which demonstrates how near-vent velocities can be tracked throughout the course of the eruption. The method was applied to the 25 eruptions analyzed in Chapter 3. The velocities were calculated by measuring the motion of the plume at the crater rim. Although this does not precisely quantify ballistic behavior, ballistic velocity was observed to follow the same approximate trends as the plume velocity, as both are ultimately tied to the gas exit pressure at the vent at any given moment. Thus, the velocity trend would offer a reasonable proxy for mass flux of plume and ballistics (Fig. 4.12).

For the 25 eruptions the instantaneous plume measurement occurred, on average, at the moment of 60% of total mass output (st. dev.: 18%). For most eruptions, this occurred when the plume was near the top of the field of view, indicating that the plumes
Figure 4.12. Plume velocity at the crater rim. Results for E009, using an automated cross-correlation window technique.
at that moment contained, on average, 60% of the total mass erupted. The 60±18% value should be kept in mind when placing the instantaneous values in their proper context. It is more difficult to quantify the role of instantaneous measurements for ballistic masses, as they generally have smaller, and more variable, residence times in the field of view than the plume. For instance, low velocity particles fall out of the field of view rapidly, while high velocity particles can easily extend above the field of view during their ascent to reappear later during their descent. I note, however, two issues with the measurement of the ballistic phase. First, I show that the ballistic mass value is an underestimate to begin with, as it is unlikely that all particles are being detected (see above). Second, the ballistic mass will be a minor component of total mass, as discussed below.

4.5 Results II: Comparison with the spot radiometer

The FLIR-derived mass results were compared to radiometer amplitudes for the same June-July 2004 eruptions from NE-1 crater. The radiometer was situated very close to the FLIR (within 50 m) (Fig. 4.13), and the field of view of the radiometer (~120 m, diameter) was approximately equivalent to that of the FLIR (~140 m in its vertical dimension), allowing a direct comparison. Radiometer amplitudes were measured as the difference between the peak of the signal during an eruption and the background signal between eruptions.

As discussed, Type 1 mass results from the FLIR are only minimum estimates because the plume is not observable by the FLIR and ballistic particles can go undetected in many circumstances. In Type 2 eruptions, the plume (which comprises the majority of
Figure 4.13. Mass estimates by spot radiometer. Top: The spot radiometer (bottom right) in a waterproof Pelican case, about the size of a shoebox, pointed at NE crater. Bottom: The relation between radiometer amplitude and total observable mass determined from FLIR imagery for Type 2 eruptions. Red data points (and fitted power-law) depict solutions for gas mass fraction ($n$) equal to 0, while the blue data points and line show that for $n=1$. 

$n=1$: Mass = $59.3 \times T_{\text{rad}}^{1.55}$

$n=0$: Mass = $217 \times T_{\text{rad}}^{1.64}$
the ejecta mass) is observable and mass values can be considered reliable estimates of total instantaneous mass, within the uncertainty of parameters such as the gas mass fraction. I limit my comparison, therefore, to Type 2 eruptions (68 total). Results are shown for a range of gas mass fractions \((n=0-1)\) (Fig. 4.13). Power-law fits to these upper and lower bounds had \(R^2\) values of 0.92. For \(n=1\), the explicit relation is \(M=59.3k^{1.55}\), where \(M\) is total mass (kg) and \(k\) is radiometer amplitude (°C). For \(n=0\), the relation is \(M=217k^{1.64}\). Notice that the correlation reflects a power law, with an exponent of \(~1.6\), stemming from the fact that the radiometer signal reflects a vertical cross-section of the ash plume and ballistic particles (giving a 2 dimensional view). Therefore, the actual volume and mass of the ejecta are one dimension greater (3 dimensional).

4.6 Discussion

4.6.1 Limitations of the algorithm

As mentioned, there are important limitations to the data that preclude a simple interpretation of ejecta mass results. My conceptual model for Strombolian eruptions involves two main components: ballistic particles and plume. The plume is composed of volcanic gas and, in some cases, very fine particles (ash). During Type 1 eruptions the plume is essentially invisible in the infrared due to its high transmissivity (Chapter 3), and therefore only ballistic particle mass extractions can be trusted in these cases. In Type 2 eruptions, the gas heats the entrained ash particles, which are visible in the infrared, allowing measurement of plume heat content and constraint of gas and ash.
mass. Thus, during Type 2 phases we have constraints on all ejecta components: ash, volcanic gas and ballistic particles. The ejecta mass results show ‘observable’ mass, therefore, which is different for the two eruption styles. This underlines the importance of the FFT ratio to establish eruption style in order to assess the meaning of the extracted mass values.

Another observational limitation is the irregular capability of the algorithm to detect ballistic particles. In some cases, faint plume can mute thermal contrasts and stymie the morphological operator responsible for ballistic detection. In other cases, the opaque plume can completely block the thermal signature of ballistic particles. Thus, the method does not reliably detect ballistic particles and any ballistic mass values must be considered minimum estimates. In Type 1 eruptions, where the ballistic particles comprise the entirety of observable mass, the total mass values are, therefore, minimum values. In Type 2 eruptions the ballistic particles, most likely, comprise a minor portion of the total mass. For instance, Chouet et al. (1974) estimated that volcanic gas was responsible for 67 and 97% of the total mass erupted in their two eruptions (both Type 1). The relative mass of ballistics is further subdued in Type 2 eruptions with the addition of ash mass. Thus, regardless of any error in ballistic mass, the mass results for Type 2 eruptions will be a good approximation of total instantaneous mass.

Also, the fact that the largest mass value was measured results in the mass values being representative only of maximum ‘instantaneous’ mass (i.e. the mass present in a scene at any one time), which is often not equivalent to total mass ejected during an individual eruption. As described, assuming the plume comprises the majority of total
mass in Type 2 eruptions, the instantaneous mass was shown to be approximately 60±18% of the total mass ejected during an eruptive event. Any attempt to quantify total erupted mass more precisely would require high frame rate data and more sophisticated routines for tracking ballistic particles and portions of the plume.

In addition to these observational limitations in the FLIR data, the observed mass values cover a wide range due to uncertainty in 1) the gas mass fraction \( n \) of the plume, 2) the initial temperature of ash and 3) the time-varying temperature of ballistics in flight. First, the gas mass fraction is unknown and variable for strombolian eruptions, as all particles in strombolian explosions are essentially incidental, and thus any gas mass fraction is possible (Chapter 2). The bursting gas slug, which causes a strombolian eruption, is completely decoupled from the parent magma and any solid particles are those incidentally picked up and accelerated by the exploding gas. This contrasts with plinian eruptions, where the gas is coupled to its parent magma and the inherent exsolved mass of volatiles can be used to estimate \( n \). Second, in the plume mass calculation, any heat contributed to the ash-sized particles depends upon their initial temperature. As these particles are likely sourced from loose material atop the magma column, they are not necessarily at magmatic temperature (i.e. \( \sim 1000 ^\circ C \)). Following Chapter 3, I assume a range of temperatures from 500 to 1000 °C. Finally, the ballistic temperature will vary throughout flight. If the particle is sourced from the fluid magma column (Chouet et al., 1974; Blackburn et al., 1976), it will begin its ascent at magmatic temperatures but drop rapidly due to radiation and forced convection. As many particles are incandescent throughout their journey, and the lowermost incandescent temperature is \( \sim 500 ^\circ C \), I
assume a range of 500 to 1000 °C. As mentioned, the lack of knowledge on these three parameters produces a ±60% uncertainty in plume mass values, and ±70% uncertainty in ballistic mass values. This ballistic mass uncertainty is overshadowed, however, by the detection limitations described earlier.

Considering these limitations, this automated method is able to characterize Type 2 mass far more effectively than Type 1 mass. In Type 1 eruptions, only a portion of the total instantaneous mass is observable, and any mass results are minimum estimates. In Type 2 eruptions, the majority of mass is observable.

4.6.2 June-July 2004 FFT ratio and mass values

It is interesting how the FFT ratio values follow a similar trend to that of the initial velocities calculated in Chapter 2 (Fig. 4.9). As discussed in Chapter 2, this is likely due to the accidental nature of ash-sized particles in the Type 2 eruptions, which tend to lower velocities. Thus lower velocities will be associated with lower ballistic:ash ratios. One exception to this relation is July 14, whose high velocity eruptions showed only a moderate FFT ratio. Eruptions on this day were highly collimated (see Fig. 4.7b), and ballistic particles were visually difficult to separate from the faint ash plume because both components were so laterally confined. Thus, the FFT ratio values are a function of both relative ballistic abundance and, to a lesser degree, collimation.

The differences between mass ejected during Type 2a and 2b eruptions are sufficiently large that they exceed the uncertainty in my method, and we can say with
confidence that Type 2b eruptions emit significantly less plume and ballistic ejecta than Type 2a eruptions. Type 2a eruptions are defined as those having initial plume velocities >15 m s\(^{-1}\), while Type 2b eruptions have initial plume velocities below this threshold. Type 2b eruptions, therefore, erupt ejecta with very little initial momentum (Chapter 2), suggesting that that the gas slug driving the explosion has minor initial overpressure. The lower mass values of Type 2b eruptions thus seem to be caused simply by a deficiency in source energy.

These mass values generally agree with the range of previous Strombolian ejecta mass estimates, in part because the range of previous results is also very large (Table 4.2). It is important to understand, however, that previous mass values incorporate different vents, eruption styles and ejecta types. For instance, based upon their descriptions, Chouet et al. (1974) and Ripepe et al. (1993) measured Type 1 eruptions while Blackburn et al. (1976) measured Type 2 eruptions. Thus, the results from the latter study reflect plume mass while Ripepe et al. (1993) measured ballistic mass. Chouet et al. (1974) measured the ballistic mass directly and inferred gas mass from the photographs. In general, these studies agree that ejecta mass values for eruptions at Stromboli can extend up to a maximum of ~50,000 kg, and have a mean value in the vicinity of several hundred to several thousand kilograms.

Several previous studies have used ejecta mass values to calculate the magma budget for Stromboli (McGetchin and Chouet, 1979; Allard et al., 1994; Harris and Stevenson, 1997). It is important to recognize that the mass values of Type 2 eruptions
<table>
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<td>3</td>
<td>-</td>
<td>37 (16-62)</td>
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<tr>
<td>Ripepe et al., 1993&lt;sup&gt;2&lt;/sup&gt;</td>
<td>6</td>
<td>2100 (200-4200)</td>
<td>-</td>
</tr>
<tr>
<td>This study&lt;sup&gt;3&lt;/sup&gt;</td>
<td>118</td>
<td>500 (0-6230)</td>
<td>3100 (0-40,800)</td>
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</tbody>
</table>

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<td>180 (120-240)</td>
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<td>Blackburn et al., 1976&lt;sup&gt;1&lt;/sup&gt;</td>
<td>5</td>
<td>-</td>
<td>1444 (496-3040)</td>
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<td>4</td>
<td>12,225 (2000-31,000)</td>
<td>-</td>
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*values in kg shown as mean (min-max)

<sup>1</sup> Type 2 eruptions only

<sup>2</sup> Type 1 eruptions only

<sup>3</sup> Both Type 1 and 2 eruptions
represent, at least in part, recycled material (Chapter 2) and therefore do not relate to a total magma budget in a straightforward manner.

4.6.3 Radiometer relations

It is not surprising that the radiometer amplitude correlates well with the FLIR-derived mass results, because they are both directly linked to the integrated 8-13 μm radiance emitted by the ejecta. The radiometer signal is essentially a reduction of the two dimensional FLIR image to one dimension and the correlation, therefore, does not serve as a strong corroboration of the FLIR-derived mass measurements. In practice, it is impossible to corroborate these mass values as no other independent means of measuring ejecta mass are available.

It must also be appreciated that the radiometer will suffer from the same shortcomings as the FLIR in producing inconsistent mass values for Type 1 and Type 2 eruptions, due to limitations in observable mass. Unlike the FLIR, the radiometer cannot discriminate the style of eruption, limiting the usefulness of the radiometer mass results if eruption style is unknown.

4.6.4 Operational monitoring

As already mentioned, previous effusive phases at Stromboli volcano have been preceded by intense explosive activity with very poor collimation, suggestive of a high gas flux into the shallow system coupled with magma very near the level of the crater
terrace. This algorithm permits me to make estimates on eruptive style, ejecta mass, and ballistic collimation. In an operational environment, all of these parameters would prove useful for understanding the ongoing state of eruptive activity. Essential to interpreting and using the results for higher order analysis and modeling, however, is appreciation of the large uncertainty in my methods (60% for plume mass, and 70% for the ballistic mass minimum estimate).

Overall, the method is best suited to ash-rich eruptions (Type 2), whose plumes are more amenable to total instantaneous mass estimates. In the proper setting, the method could prove useful for constraining rough ejecta mass values and detecting trends in mass, collimation and eruption style through time. Anomalous changes in these parameters may be a precursor to more hazardous eruptive activity. Detectable changes in ejecta mass, however, would need to be considerable due to the large uncertainties in my mass estimation approach.

4.7 Conclusions

The method outlined here is intended to estimate eruption collimation, eruption style, and ejecta mass for small explosive eruptions, and I envision this algorithm applied to an operational stream of FLIR data at a persistently active volcano. Although applied to strombolian eruptions here, other small-scale explosive eruptions could be monitored in such a way. A major limitation of the method is the large uncertainty in ejecta mass results. This is due to observational limitations and lack of knowledge of the gas mass.
fraction, ballistic temperature and initial temperature of the ash particles in the plume. These are thus parameters that require further constraint using field measurements.

These results could be improved with different instrumentation. Instantaneous mass could more closely resemble total erupted mass if the field of view was increased (such as with a 45° lens, sold by FLIR Systems). A larger lens, however, would increase the pixel size, which is the other main limitation. The pixel size must be greatly reduced (to at least a scale of 1-2 cm) to begin to make accurate ballistic mass measurements. This is impossible with a larger field of view, and thus, dual cameras might be considered for improved mass estimates during ash-rich eruptions. But any FLIR configuration will still be unable to detect the transmissive gas released during ash-poor eruptions, indicating that another instrument could be necessary to provide such information. Instruments that detect SO₂ (such as a FLYSPEC, see Horton et al., 2002) may aid in measuring this gas output.

The other limitation of the algorithm is the unreliable ejecta mass values estimated during Type 1 phases. At Stromboli, Type 1 activity is generally more common than Type 2 activity during normal activity, based on anecdotal evidence (M. Ripepe and A. Harris, personal communication). This imposes a limit on the usefulness of the ejecta mass approach (and, likewise, derived mass from the radiometer) for Stromboli. At other volcanoes, however, the mass estimation approach outlined here may prove useful when 1) ash-rich eruptions are the most common activity style and 2) the gas mass fraction is known with more certainty.
Chapter 5
Conclusions

5.1 Summary

In this dissertation I investigated the utility of handheld thermal infrared cameras for understanding eruptions at Stromboli volcano (Italy), and by extension, strombolian eruptions in general. The type of camera I used, called a FLIR (for Forward Looking Infrared Radiometer) is emerging as a popular tool for research and monitoring of active volcanoes. For example, these cameras have played a prominent role in operation monitoring of recent activity at Kilauea (Kauahikaua et al., 2003), Mt. St. Helens (Schneider et al. 2004), and Stromboli (Calvari et al., 2005) volcanoes. Current interest has led to the proposal of a special session at the Fall 2005 meeting of the American Geophysical Union to focus on the capabilities and application of FLIR data in research and monitoring scenarios (J. Dehn, personal communication, 2005).

The first part of my work began with a simple classification scheme to conceptually deconvolve the myriad explosive behaviors that Stromboli can exhibit. This resulted in the definition of a Type 1 vs. Type 2 dichotomy, in which Type 1 eruptions are ballistic-dominated and Type 2 are ash-dominated. Extending this further, Type 2 eruptions exhibiting a gas-thrust phase and abundant ballistics were considered Type 2a, and those with only buoyant velocities and few ballistics were classified as Type 2b. Type 1 plumes had velocities with a maximum value of 101 m s\(^{-1}\), and a mean of 34 m s\(^{-1}\).
Likewise, Type 2a eruptions exhibited gas thrust velocities (max 38 m s\(^{-1}\), mean 31 m s\(^{-1}\)), while Type 2b eruptions rose at buoyant velocities (max 11 m s\(^{-1}\), mean 7 m s\(^{-1}\)).

Type 1 activity can be considered the most common activity style at Stromboli, and occurs when the magma free surface is unobstructed so that bubble bursting can throw molten clots of bubble wall freely into the atmosphere. In some cases, however, the conduit can be obstructed by loose debris. This provides finer particles by secondary fragmentation (milling/grinding during passage of the explosion) that are entrained into otherwise normal eruptions to create ash-rich plumes (and, therefore, Type 2 eruptions). Type 2a styles are caused by explosions of high overpressure bubbles, while Type 2b bubble overpressures are very low. The style of eruption (Type 1 vs. 2) for a given vent was generally maintained over a duration of days to weeks, indicating that backfill can have a residence on that timescale.

Strombolian plumes were then investigated in detail, by fine timescale tracking of the rise rates, dimensions and bulk temperatures using 30 Hz data. Plumes transitioned through a number of different plume morphologies (jets, starting plumes, rooted thermals) on a timescale of seconds. Type 2a eruptions exhibited gas-thrust phases (initial velocities >15 m s\(^{-1}\)) which decelerated into convective velocities (<15 m s\(^{-1}\)) usually by a height of ~100 m. Type 2b eruptions, on the other hand, showed roughly constant convective velocities throughout their observed lifespan. Time-averaged convective velocities for 42 Type 2b plumes ranged between 1.4 and 10.9 m s\(^{-1}\), with a mean value of 4.5 m s\(^{-1}\). Type 2a plumes, with their high initial velocities began as jets in which the front of the jet was largely coupled to the main body of the jet. Upon jet
deceleration, a clear velocity differential developed between the plume front and main plume (now the feeder plume), creating a starting plume morphology. When fully buoyant behavior was established (<10 m s\(^{-1}\)), there was a pronounced differential leading to the development of the vigorous ring vortex which marked a rooted thermal. Type 2b plumes often emerged above the crater rim as an amorphous mass which organized itself directly into a rooted thermal once the velocity differential was established.

The rapid evolution of plume morphology indicated that air entrainment rates were likewise changing at high rates. Highly variable air entrainment coefficients were shown implicitly by measuring the lateral spreading rate of the plumes, agreeing well with previous laboratory results. Spreading rates (i.e. radius vs. height) ranged from 0.1 to 0.4, representing jet, starting plume and rooted thermal rates. I also measured plume temperatures with the FLIR, one of the first times this has been done from the ground. Plume temperatures began between 300 and 600 °C at the crater rim, and dropped to <150 °C within the first 100 m of ascent. Measuring actual plume temperatures can only be done on opaque plumes. Therefore I developed a simple method to discriminate opaque from translucent plumes. The method entailed tracking the total heat content in a plume front versus volume, where sudden drops are clear indictors of translucence. Extracted plume temperatures can then be used to estimate ejecta content in the plumes using a heat conservation approach. Unfortunately, the ejecta estimates are sensitive to the gas mass fraction, which is a poorly constrained parameter in Strombolian plumes because of the incidental nature of the particles. This imposes a fundamental weakness in
understanding Strombolian plume thermodynamics, and introduces an uncertainty of \( \approx 60 \% \) into my mass extraction results.

The final part of the dissertation took these ideas and integrated them into a model for an operational algorithm that processes FLIR data for small volcanic explosions, such as strombolian eruptions. The automated method determines eruption style (Type 1 vs. 2), measures collimation of ballistic ejecta and attempts to constrain mass values, within the uncertainty of the gas mass fraction. The ballistic detection scheme was judged less reliable than the plume measurement approach, and thus Type 1 mass values could only be considered minimum values. Also, the fact that the mass measurement represents only mass within the field of view at a particular time means that mass results do not directly represent the total mass ejected during an eruption. Extracted ash masses had a range of 120 – 40,800 kg for Type 2a eruptions and 2 – 10,850 kg for Type 2b eruptions, with ballistic masses in the range 1-6230 for Type 1 eruptions.

In the case of Type 2 eruptions there was a simple relationship between mass and the amplitude recorded during the eruptions by the much simpler to use and more spot-radiometer. Although the spot radiometer lacks the information content that a FLIR image contains, it is much less expensive. The thermal amplitude of the radiometer signal \((k)\) can be related to FLIR-measured mass \((M)\) by \(M = ak^b\). For a gas mass fraction of 0, \(a=217\) and \(b=1.64\), while for a gas mass fraction of 1, \(a=59.3\) and \(b=1.55\). This indicates that, in the case of ash-rich eruptions (Type 2), spot radiometers can be used to judge the instantaneous ejecta mass in eruption plumes within \(\pm 60\%\). At Stromboli, or
other volcanoes where eruptions are predominantly Type 1, using spot radiometers alone to judge reliable mass is not feasible or recommended.

5.2 Implications of results

First and foremost, these results illustrate how strombolian activity encompasses a much broader range of behaviors than has been commonly attributed to this style of eruption (e.g. Chouet et al., 1974; Vergniolle and Mangan, 2000). This was partly due to the small number of eruptions that had been imaged up until now (fewer than 30) (Chouet et al., 1974; Blackburn et al., 1976; Ripepe et al., 1993). Although ash-rich (Type 2) strombolian eruptions have been noted in the literature (e.g. Blackburn et al., 1976), their behaviors have not been fully articulated. My work serves to reinforce that the paradigm of strombolian behavior ought to be broadened to consider these ash-rich (Type 2) eruptions, which can be considered a modified form of normal strombolian activity (Type 1). Type 2 eruptions represent the same underlying mechanism (bubble bursting) as the more typical Type 1 eruptions, but in a slightly ‘messier’ conduit. Appreciating this is imperative to describing small-scale eruption styles in the field, as Type 2 strombolian behavior could be confused with vulcanian activity, which is not thought to be related to bubble bursting (Morrissey and Mastin, 2000).

Strombolian ash plumes are unique in that fine scale fragmentation is not effected by anomalously high explosive energy. Instead, fine particles are produced through secondary fragmentation mechanisms – i.e. mechanical breakage of originally larger, but brittle, clasts. This runs counter to the common volcanologic maxim that fine scale
fragmentation is necessarily the result of high explosive energy (Walker, 1973), indicating that some prehistoric fine deposits at strombolian volcanoes may have originated from a secondary fragmentation mechanism. In my case, I document the persistence of ash rich strombolian emissions over prolonged periods of unspectacular, and indeed normal, levels of Strombolian activity. Not a single paroxysmal strombolian or vulcanian eruption was observed.

In my exploration of plume dynamics, the transient nature of small-scale plume entrainment regimes was uncovered. Plumes transitioned among several plume morphologies, and therefore varying entrainment rates were operating. These changed on a time scale of seconds. These rapidly evolving air entrainment regimes have ramifications on the buoyancy potential of growing plumes and, thus, their inclination to collapse and produce deadly pyroclastic flows. In understanding the collapse potential of an emergent plume, these entrainment dynamics must thus be measured and considered at a similarly fine timescale of seconds to 10ths of a second, as opposed to assuming a constant air entrainment rate as done for large scale plumes (e.g. Wilson and Walker, 1987; Glaze and Baloga, 1996).

This is one of the first studies to measure plume temperatures from the ground, and therefore some fundamental work on understanding plume temperature in infrared imagery was necessary. Of greatest importance to measuring temperature is plume opacity, as temperature values for a translucent plume are not representative of actual plume temperature. The method I outlined here for determining plume opacity should prove useful in future ground-based plume temperature measurement campaigns, as
should all methodologies, assumptions and approaches considered. Here, the FLIR is a relatively new piece of hardware in volcanology, thus new image processing and data analysis techniques capable of operating with FLIR data for dynamic volcanic events need to be defined, written, and tested. A suite of such techniques have been presented and tested to allow examination of small-scale explosive behavior with high temporal and spatial resolution thermal data obtained from the ground.

Likewise, in the last part of my study I used image processing techniques to measure strombolian eruption parameters. I introduce the Fast Fourier Transform (FFT) as a tool for discriminating ballistic-dominated from plume-dominated eruptions, as each exhibits a unique distribution of spatial frequencies that the FFT method exploits. The use of morphological operators was also found to be effective in isolating ballistic pixels explicitly. In general, there are a host of basic image processing techniques that are useful for analyzing infrared images of explosive eruptions.

The spot radiometer, though vastly inferior in information content, has several benefits over FLIR including simplicity, field robustness, and low cost. In the case of Type 2 eruptions, I found that the spot radiometer amplitude offers a proxy for ejecta mass within the uncertainty of the gas mass fraction (60%). If the gas mass fraction could be tied down, then the spot radiometer could potentially produce relatively well-constrained ejecta mass estimates of small plumes. The spot radiometer signal cannot be used to assess mass, however, if the eruption style is unknown, due to observational shortcomings for Type 1 eruptions.
5.3 Limitations

The greatest uncertainty that pervades these three studies is the gas mass fraction. Plinian eruptions, having gas and parent magma closely coupled throughout the ascent, have a well constrained gas mass fraction. Strombolian plumes, however, involve primarily incidental particles which frustrate our ability to assume a constant gas mass fraction. The lack of knowledge of gas mass fraction leads directly to large uncertainties in the total ejecta mass results (Chapter 4).

Translucence in plumes is a major hindrance on our ability to measure accurate temperatures (Chapter 3). Although I establish a tool for discriminating opaque from translucent plumes, the method can only be applied to plume fronts in its current form. Establishing opacity in the feeder plume is more troublesome, and may require more sophisticated tracking than I have presented here.

Another major limitation with FLIR imagery is the coarse pixel size. An image is only 240 x 320 pixels, or put in popular terms, approximately 0.08 Megapixels. At a distance of 450 m, the pixel size is 59 cm. This seriously degrades our ability to measure ballistic particle sizes and masses, as mean ballistic sizes are less than 10 cm at Stromboli. It also complicates tracking particles, as the pixels are sufficiently large to potentially encompass several particles at a given moment.

The mass extraction method outlined in Chapter 4, as I have already mentioned, produces values which must be interpreted based upon the eruption style. In Type 1 eruptions, only the ballistic mass can be determined as the plume mass is unobservable. Furthermore, the ballistic mass value, due to threshold limitations, is only a minimum
value. In Type 2 eruptions, ballistic mass (i.e. minimum ballistic mass) can be calculated along with plume mass, as fine particles make the plume observable. This inconsistency in results makes it imperative that eruption style be determined. For the spot radiometer, which can only produce reliable mass estimates during Type 2 phases, the eruption style must be known beforehand. At Stromboli, this is troublesome as eruption styles can vary from day to day (Chapter 2). At volcanoes which are consistently ash-rich (some equivalent of Type 2), then the radiometer results may prove useful to constrain mass.

5.4 Role of the FLIR at explosive volcanoes

The majority of previous FLIR studies have been applied to effusive volcanism or passive degassing (e.g. Bonnaccorso et al., 2003; Wright and Flynn, 2003; Calvari et al., 2005). My work entails the first major study devoted solely to understanding the FLIR in an explosive environment. I have shown that the FLIR is much better suited to viewing the range of ejecta types (ash and ballistics) than visual imagery. Visual imagery is limited by the time of day, viewing some components (ballistics) well at night, and others (ash) well during the daylight, but it cannot view both simultaneously. FLIR data, being sensitive to emitted radiance, is not hostage to the time of day in any way. This enables great flexibility in any field campaigns using the FLIR, and it provides a great strength in its potential for continuous monitoring in an operational environment. The high frame rate acquisition of the FLIR (30 frames per second) and ease of connection to modern computer systems (via Firewire) allow an added benefit of easily capturing and storing
highly dynamic explosive processes in a calibrated, fixed and standard digital format that then facilitates ease of processing and analysis.

5.5 Future work

To improve our ability to track strombolian activity with the FLIR, several objectives should be sought. Achieving these objectives is necessary if we are to further strengthen our understanding of the processes presented here. These improvements include:

- Placing the FLIR in view of the actual vent. My results were somewhat limited by the presence of the crater rim, meaning that ejecta were imaged after they had exited the vent and attained the level of the rim. In viewing the vent, one could see how the conduit is modified, such as by collapse events or material falling back into the crater. The temperature of the overfill could also be gained, which would help with the mass estimates. In addition, exact muzzle velocities and at-vent dynamic and thermal conditions can be directly extracted from such data. Naturally, such a setup would require remote operation of the camera to minimize the residence time for scientists working near the craters.

- Running the FLIR continuously for extended periods. In this way, we could examine the exact timescales over which eruption styles and dynamics evolve and persist. This would also allow consideration of the transition phases and...
the possible driving factors behind the transitions. Due to data rates involved with continuous acquisition, some routine would obviously be necessary to identify eruptive events and discard non-eruptive images, similar to that presented in Chapter 4, thus preventing hard-discs being filled rapidly.

- Using a higher magnification lens to study ballistic behavior near the vent. FLIR Systems offers telescopic lenses down to 7°. By moving the camera to 250 m (the distance of Pizzo sopra la Fossa) and using this telescopic lens the pixel size would be reduced from 59 cm (in this study) to 13 cm. This is still larger than most ballistic particles, but would nonetheless resolve individual particle motions better.

- Augment the FLIR with Doppler radar to determine particle sizes. Doppler radar has been shown to discern particle sizes (Hort et al., 2003). When coupled with the FLIR data this information could help establish the gas mass fraction.

- Augment the FLIR with conventional high-frame ratio photography. A high-speed camera would be able to discern and measure ballistic particles much better than the FLIR (Chouet et al., 1974).

- Augment the FLIR with a gas measurement device. An SO₂ instrument, such as the FLYSPEC (Horton et al., 2002) may help measure gas output in Type 1 eruptions, when the gas does not radiate significantly in the longwave infrared.
• A multispectral imager would better determine the temperature distribution within a given pixel. This could somewhat compensate for the lost information due to the large pixel size of imaging systems like the FLIR at the observation ranges we used. Current portable multispectral radiometers, however, are not imagers but spot radiometers (Flynn and Mouginis-Mark, 1992; 1994).

• Vertical stacking of multiple FLIR field of views to image entire range of ballistic heights. Many of the eruptions in this study cast ballistic particles above the top of the field of view. Using two cameras, with some degree of overlap of the field of views, one would likely see most of the highest ballistic particles.

• Full comparison of FLIR, seismicity and infrasound data. This is essential if, for individual eruptions, we are to fully model and constrain the relationship between bubble overpressure and exit velocity. As noted in Chapter 2, the previous attempt to model this relationship for strombolian eruptions failed to accurately characterize strombolian exit velocities. Further work, with this refined data, would provide powerful data for modeling and understanding the dynamics and mechanisms that drive strombolian activity.

• Focus on sedimentation. My remarks on sedimentation were very limited, and data could be collected to specifically address different sedimentation regimes. Sedimentation could then be tied to plume velocity, which is directly
visible in the FLIR imagery. This knowledge could improve models linking depositional characteristics with plume dynamics.

- A closer examination of entrainment vortices. The FLIR imagery capture the dynamics of the exterior entrainment vortices quite well, as they were inevitably related, to a degree, to the thermal contrast with adjacent material. The spatial scale and motions of these vortices for different plumes might offer insights into the turbulence and fluid viscosity of the plume.

- Modeling gas transmissivities. I briefly used MODTRAN4 to study water vapor transmissivities to understand the degree of emitted radiance. If a more rigorous approach is taken, and especially if some knowledge of particle concentration is available, it could help us understand the gas density in the translucent periods which mark all Type 1 and the dissipated phases of Type 2 plumes. The emissivity of the gas phase is, essentially, unknown and requires definition for reliable thermal mixture modeling of plumes rich in gas.

- Development of improved ejecta tracking methods. The use of the cross correlation routine I present for tracking ejecta and extracting velocities is relatively crude, and is also limited by the coarse pixel size. Improved methods, possibly utilizing telescopic lenses, might help track material at the vent (exiting plume) as well as in the air (rising plume and ballistic particles). Keeping track of all of the ejecta is essential if we are to move beyond the ‘instantaneous’ mass results I presented in Chapter 4. Knowing where the
ejecta is and where it has been since the eruption began, we can estimate total mass ejected in a given eruption.

- Improvement of the current ballistic detection scheme. The ‘opening’ morphological operator used in Chapter 4 does not detect faint ballistic pixels as well as the human eye can in the FLIR imagery, and thus the ballistic mass estimate is only a minimum of total ballistic mass. Improving the cleverness of the detection scheme could bring the ballistic mass value closer to the total value.

Thinking beyond strombolian activity, I believe this work only scratches the surface of FLIR’s potential for understanding explosive volcanism. The FLIR should prove useful for the following projects:

- Imaging vulcanian eruptions. These eruptions are generally unpredictable and can have widespread ballistic and ash fall-out, as well as pyroclastic flow, hazards. Thus remote operation of the FLIR might be necessary, and a high frame rate will be necessary (30 frames per second). This eruption style is of particular interest because of the high likelihood for column collapse and generation of pyroclastic flows, and there is no reason why all of the eruption parameters described here could not be extracted using the proposed methodologies.
- Imaging plinian columns. With a telescopic lens, and possibly remote FLIR operation, the exterior of a large plinian column could be imaged. Again, high frame rate acquisition is imperative. Although the hot, high velocity inner core would not be visible, the temperature of the column exterior would still be of interest. Temperature measurements could help quantify air entrainment rates and their spatial variations across, and along the axis of, a large plume.

- Imaging pyroclastic flows. Viewing the thermal behavior of one of the most hazardous volcanic behaviors should greatly improve our understanding of their dynamics. Again, remote operation might be necessary, along with a high frame rate. Depending on the eruptive scenario, capturing high quality data of pyroclastic flows could be quite feasible. The FLIR could be positioned, for example, on a high ridge above a valley which is known to funnel flows from, say, an unstable dome. The bulk temperature of the cloud could help constrain air entrainment rates, and viewing temperature variations across the flow could help discriminate more hazardous (low-density) surge components from the more channelized (high density) flow portion.

5.6 Concluding remarks

Previous studies have struggled to infer thermal volcanic properties from visible imagery, and the FLIR now allows these properties to be extracted directly with relative ease. Recent innovations have improved the reliability, weight, accuracy and cost of
FLIR cameras, making them increasingly attractive for independent researchers and volcano observatories. The capabilities of the FLIR mean that it is rightly receiving more and more attention across the volcanological community. Thus I envisage it rapidly becoming established as a standard, routinely used tool for both volcanologic science and monitoring.

Studies such as this help to demonstrate the ways that FLIR cameras can be employed in understanding volcanic processes. By placing temperature in a spatial context, the FLIR adds another important constraint – perhaps, the most important constraint of all – to modeling and understanding volcanic phenomena.
References


Hort M., Seyfried R., and Voge M. 2003. Radar Doppler velocimetry of volcanic eruptions: theoretical considerations and quantitative documentation of changes in


Pinkerton H., Harris A., Ball M., James M. 2004. An analysis of the development of surface textures and flow morphology on small lava flows on Kilauea using high frame


### Appendix A

**Summary of the 344 eruptions**

#### 1) 2004 Eruptions

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<th>Type</th>
<th>Plume velocity</th>
<th>Spot Rad amp, F</th>
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Appendix B
Matlab files for Chapter 4

function [maxratio, eigenval1, eigenval2, maxash1, maxash2, maxspat, ...
        maxspat2, ashconc1, ashconc2] = ejectamass
%function is a new program to estimate mass of ash and spatter of strombolian
%eruptions from a sequence of FLIR images. This is an improvement upon the
%FLIRMASS1-6.m series, which was flawed in several ways.

% Steps to the program
%---------------------
% 0) Initial ingestion of Matlab images
% 0b) Determine background region (crater terrace)
% 1) Go through series of images
%  a) Determine if viewing conditions are clear
%  b) Identify eruption images using running standard deviation
% 2) For each eruption image
%  a) Run FFT for eruption style
%  b) Isolate ash and spatter components
% 3) For the spatter image
%  a) Do radiance mixing to calculate spatter size and mass
%  b) Calculate collimation
% 4) For the ash image
%  a) IF PIT>air temp, then use heat conservation approach for ash mass
%  b) IF PIT<air temp, then use radiance mixing to calculate ash mass

cd i:\FLIRmatfiles %go to master directory with FUR images
load NE1erupts2; %this matfile has noimages and NE1erupts array
noimages=noimages2; %which eruptions not to load
masterdist=zeros(0);
maxratio=zeros(length(NE1erupts),1);
for j1=1:length(noimages)
    for j2=1:length(NE1erupts)
        if noimages(j1)==NE1erupts(j2)
            NE1erupts(j2)=0
        end
    end
end
%for each eruption you want to load
for j=1:length(NE1erupts)
if NE1erupts(j)>0
    % Step 0: Ingest Matlab images
    a=NE1erupts(j)
    if a<10
        b=['0' num2str(a)];
    elseif a<100
        b=['00' num2str(a)];
    else
        b=num2str(a);
    end
    cd(b);

    %disp('Ingesting FLIR images...')
    [Sess,imgstart,imgstop,dtime,c]=ingest2b(a);
    span=imgstop-imgstart; %min and max image number

    % Step 0b: Determine background region (crater terrace)
    [cr,cc]=craterrimb(Sess,imgstart,c,imorient(j);

    % Step 1: Go through images, find cloudy images and ID eruptions
    disp('Now finding eruption images...')
    % Steps 1a+1b: Identify eruption images using std and cloudy images
    [eruptarray,cloudy]=IDeruptb(Sess,c,imgstart,imgstop);

    % Step 2: Process eruption images
    disp('Processing eruption images...')
    fftratio=zeros(1,span+1);
    smass=zeros(1,span+1);
    amass=zeros(1,span+1);
    aconc=zeros(1,span+1);
    totalmass=zeros(1,span+1);
    count=0;
    transectsum=zeros(1,320);
    totalcount=zeros(0);
    totalval=zeros(0);
    totalspat=zeros(0);
    totalash=zeros(0);
    % for each image in this eruption
    for i=imgstart:imgstop
        if eruptarray(i)>0 %if eruption image
            count=count+1;
            a=[c num2str(i)];
            A=zeros(0);
            A=getfield(Sess,a);
            if mod(floor(((i-imgstart)/span)*100),10)==0
                disp(['num2str(floor(((i-imgstart)/span)*100))']
            end
        end
    % Step 2a: Run FFT for eruption style
        b=[c num2str(imgstart)];
        B=getfield(Sess,b);
        [%fftratio(i),style(i)]=newfft(A,B,cr,cc);
        ratio=newfft2b(A,B);
        if count<=7
            transectsum=transectsum+transect;
        end
        %figure(2)
        totalcount=[totalcount count];
totalval={[totalval ratio];
%plot(totalcount,totalval)

%rotate image if needed
if imorient(i)==1;
    A=rot90(A);
end

% Step 2b: Isolate ash and spatter components
[ashimg,spatimg]=morphsplit4b(A,ffratio,cr,cc);

% Step 3a: Estimate spatter mass using radiance mixing
[smass(i),snum(i),sdist]=spatmassb(ashimg,spatimg);
masterdist=[masterdist sdist];

% Step 3b: Estimate collimation
%colval(i)=collimate3b(spatimg);
%xxx=colval(i)
colval=zeros(0);

% Step 4a: Ash mass, if PIT>air temp
[amass(i),aconc(i)]=ashmass3b(ashimg);
xxx=amass(i);
totalspat=[totalspat smass(i)];
totalash=[totalash amass(i)];

% Step 4b: Ash mass, if PIT<air temp
% to do...low priority
end

%write max fft ratio
if length(totalval)>6
    e1=6;
    if length(totalval)>6+6
        e2=10;
    else
        e2=length(totalval);
    end
maxratio(i)=mean(totalval(e1:e2));
coltemp=zeros(0);
for j=[j1:j2
if isnan(colval(j))=0
    coltemp=[coltemp colval(j)];
end
end
eigenval1(i)=sum(coltemp)/length(coltemp);
maxash(i)=max(amass(e1:e2));
maxspat(i)=max(smass(e1:e2));
ashconc1(i)=mean(aconc(e1:e2));
maxash1(i)=maxash(i);
ashconc1(i)=ashconc(i);
elseif length(totalval)>5
    e1=5;
    if length(totalval)>5+6
        e2=9;
    else
        e2=length(totalval);
end

240
end
maxratio(j)=mean(totalval(1:e2));
coltemp=zeros(0);
for jji=1:e2
  if isnan(colval(jji))==0
    coltemp=[coltemp colval(jji)];
  end
end
eigenvall(j)=sum(coltemp)/length(coltemp);
maxash(j)=max(amass(1:e2));
maxspat(j)=max(smass(1:e2));
ashconc(j)=mean(aconc(1:e2));
maxashl(j)=maxash(j);
ashconc1(j)=ashconc(j);
elseif length(totalval)>4
  e1=4;
  if length(totalval)>4+6
    e2=8;
  else
    e2=length(totalval);
  end
maxratio(j)=mean(totalval(1:e2));
coltemp=zeros(0);
for jji=1:e2
  if isnan(colval(jji))==0
    coltemp=[coltemp colval(jji)];
  end
end
eigenvall(j)=sum(coltemp)/length(coltemp);
maxash(j)=max(amass(1:e2));
maxspat(j)=max(smass(1:e2));
ashconc(j)=mean(aconc(1:e2));
maxashl(j)=maxash(j);
ashconc1(j)=ashconc(j);
elseif length(totalval)>3
  e1=3;
  if length(totalval)>3+6
    e2=7;
  else
    e2=length(totalval);
  end
maxratio(j)=mean(totalval(1:e2));
coltemp=zeros(0);
for jji=1:e2
  if isnan(colval(jji))==0
    coltemp=[coltemp colval(jji)];
  end
end
eigenvall(j)=sum(coltemp)/length(coltemp);
maxash(j)=max(amass(1:e2));
maxspat(j)=max(smass(1:e2));
ashconc(j)=mean(aconc(1:e2));
maxashl(j)=maxash(j);
ashconc1(j)=ashconc(j);
else
  maxratio(j)=nan;
end
if length(totalval)>3
  maxash(j)=max(amass(3:end));
  maxspat(j)=max(smass(3:end));
ashconc2(j)=mean(aconc(3:end));
colarray=colarray colval(j);
for jji=1:e2
  if isnan(colval(jji))==0
    coltemp=[coltemp colval(jji)];
  end
end
eigenva12(j)=sum(coltemp)/length(coltemp);
else maxash(j)=0;
  maxspat(j)=0;
  maxash2(j)=maxash(j);
eigenva12(j)=0;
end
if maxspat(j)>0 & max(totalspat)>0
  maxspat(j);
totalspat;
sj(j)=find(totalspat==maxspat(j))
end
if maxash(j)>0 & max(totalash)>0
  ai(j)=find(totalash==maxash(j))
end

maxspat(j);
% colarray=zeros(0);
% if length(totalval) >= 6 & length(colval) > 0
%   % for kk=3:7
%   %   if colval(kk)>0
%   %     colarray=[colarray colval(kk)];
%   % end
% end
% if length(colarray)>0
%   eigenva1(j)=mean(colarray);
%   eigenva2(j)=min(colarray);
% end
% end
% colarray=zeros(0);
% for kk=1:length(colval)
%   if length(smass)==kk
%     if smass(kk)>50 & colval(kk)>0.0001
%       colarray=[colarray colval(kk)]
%     end
%   end
% end
% if length(colarray)>0
%   eigenva1(j)=mean(colarray);
%   eigenva2(j)=min(colarray);
% else eigenva1(j)=nan;
%   eigenva2(j)=nan;
% end

cd ..
end
%y=[maxratio;maxash;maxspat];
save masterdist masterdist
return
size(dtime);
figure
plot(dtime,fratio,'*')
figure
totalmass=amass+smass;
p=plot(totalmass)
%axis([min(dtime) min(dtime)+.0014 0 max(totalmass)*1.1]);
hold on
plot(amass,'r')
plot(smass,'r:')
y=[dtime;totalmass];
return
plot(mass(:,1),mass(:,4));
hold on
plot(mass(:,1),mass(:,5));
datetick('x',13);
function [cr,cc]=craterim(Sess,imgstart,c);
%function identifies the crater rim of stromboli in the flir image
i=imgstart;
a=[c num2str(i)];
A=getfield(Sess,a);
thresha=20+273.15; %threshold between air and crater rim
At=(A>thresha); %threshold at min crater rim temp
At2=bwmorph(At,'open'); %take out noise
At3=bwmorph(At2,'close'); %take out noise
bwl=edge(At3); %find edges
[cr,cc]=find(bw1>0); %row, col indices of crater rim

function [eruptarray,cloudy]=IDerupt(Sess,c,imgstart,imgstop)
%function identifies eruption images in an image sequence
span=imgstop-imgstart;
for i=imgstart:imgstop
a=[c num2str(i)];
A=zeros(0);
B=zeros(0);
A=getfield(Sess,a);
if mod(floor(((i-imgstart)/span)*100),10)==0
  disp([num2str(floor(((i-imgstart)/span)*100)) '%'])
end
if i==imgstart %do not subtract if first image in session
  %Sess=setfield(Sess,a,A); %keep this image the same
  runstd(i)=0;
end
if i>imgstart %subtract for all other images
  b=[c num2str(i-1)];
  %name=load(b); %load the previous image
  B=getfield(Sess,b);
  size(A);
  size(B);
  B=A-B; %subtract last image from current image
  %Use stand dev to ID eruption images
  runstd(i-imgstart)=std2(B);
runmax(i-imgstart)=max(B(i));
%and cloudy images
cloudvalue(i-imgstart)=std2(A);
end
%check to see if image is cloudy
thresh=5; %insert correct value here for threshold
if std2(A)<thresh
cloudy(i)=1;
else cloudy(i)=0;
end
end
runstd;
mean(runstd);
median(runstd(1:8));
median(runstd);
std(runstd);
eruptarray=zeros(1,imgstop);

%right now eruption check only works for single eruption sequences!!!!!!!
%will have to put in moving window check to make for full sessions
%
for i=imgstart+1:imgstop
%if runstd(i-imgstart)=(median(runstd)+(3*std(runstd(1:8)))); %if stand
%dev exceeds threshold
%find highest std eruption image:
%if runstd(i-imgstart)=(-7*(max(runstd)-median(runstd))+median(runstd));
%if stand dev exceeds threshold
if runstd(i-imgstart)==max(runstd); %if stand dev exceeds threshold
eruptarray(i)=1; %value=1 if eruption image
%find all previous eruption images
leftcheck=0;
for k=i-imgstart:-1:1
if runstd(k)=(.01*(max(runstd)-median(runstd))+median(runstd)) & ...
leftcheck==0
eruptarray(k+imgstart)=1;
k+imgstart;
elseif runstd(k)<=(.01*(max(runstd)-median(runstd))+median(runstd))
leftcheck=1;
end
end
rightcheck=0;
for k=i-imgstart:length(runstd)
if runstd(k)=(.01*(max(runstd)-median(runstd))+median(runstd)) & ...
rightcheck==0
eruptarray(k+imgstart)=1;
k+imgstart;
elseif runstd(k)<=(.01*(max(runstd)-median(runstd))+median(runstd))
rightcheck=1;
end
end
elseif eruptarray(i)=0; %value=0 if not eruption image
end
end
%count how many eruptive episodes there are in this session
ecount=0;
for i=2:length(eruptarray)-1
if eruptarray(i-1)==0 & eruptarray(i)==1
ecount=ecount+1;
end
if eruptarray(i)==1
eruptarray(i)=ecount;
end
% if eruptarray(i)==1 & eruptarray(i+1)==0
% eend=eend+1;
% end
eruptarray;
%ecount=max(estart,eend);

function [ratio]=newfft(A,B,cr,cc)
%function does the FFT to determine eruption style

% mask=ones(240,320);
% for i=1:length(cc)
% mask(cr(i):240,cc(i)=0;
% end
% A=mask.*A;
A=A-B;
threshb=20;
A=(A>threshb).*A;
% figure(2)
% imagesc(A,[0 50])

fimage=fft2(A);
fimage=ifftshift(fimage);
X=log(abs(fimage));
%X=fimage;
%figure(1)
% subplot(311)
% imagesc(X,[2 10])
% colorbar

%transect=X(120,:);
for c=1:320
 transect(c)=sum(X(: ,c);
end

a=min(min(X));
\( b=max(max(X))\);
%imagesc(X,[a b]);
a2=0;
b2=0;
\( r=size(A,1)\);
\( r0=round(.15*r)\);
\( r1=round(.35*r)\);
\( r2=round(.5*r)\);
\( r3=round(.65*r)\);
\( r4=r-r0\);
\( c=size(A,2)\);
\( c0=round(.15*c)\);
c1=round(.35*c);
c2=round(.5*c);
c3=round(.65*c);
c4=c-c0;

% high frequency domain
hfl=X(r0:r1,c0:c4);
hf1=hfl(1);
hf2=X(r1:r3,c0:c1);
hf2=hf2(1);
hf3=X(r1:r3,c3:c4);
hf3=hf3(1);
hf4=X(r3:r4,c0:c4);
hf4=hf4(1);
fhs=um([hfl;hf2;hf3;hf4]);
hfr=mean([hfl;hf2;hf3;hf4]);

% low frequency domain
fl1=X(r1:r3,c1:c3);
fl1=fl1(1);
fl2=X(r2-3:r2+3,c2-3:c2+3);
fl2=fl2(1);
flsum=um(fl1);
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function [ashing, spatimg] = morphsplit(A, ftratio, cr, cc)
% function performs an f1t splitting of eruption image into gas and spatter
ratio=ffiratio;
%se=ones(2,2);
se=[strel('disk',1,0);strel('disk',1,0);strel('disk',1,0)];
%se=[strel('pair',[1 0]);strel('pair',[1 0]);strel('pair',[1 0])];
%se=[0;1];
%se=[strel('disk',1,0);strel('disk',1,0);strel('disk',1,0)];
%se=[strel('disk',1,0);strel('disk',1,0)];
%se=[strel('disk',1,0);strel('disk',1,1,0)];
%se=[strel('disk',1,0);strel('disk',1,1,0)];
% if ratio<0.8
% se=strel('disk',1,0);
% elseif ratio<=0.8 & ratio<=0.85
% se=[strel('disk',1,0);strel('disk',1,0)];
% elseif ratio>0.85
% se=[strel('disk',1,0);strel('disk',1,0);strel('disk',1,0)];
% end
if ratio<0.8
se=strel('disk',1,0);
elself ratio>0.8
se=[strel('disk',1,0);strel('disk',1,0);strel('disk',1,0)];
end
ashcomp=imerode(A,se);
ashcomp=imdilate(ashcomp,se);
% gascomp=imopen(Subimage,se);
ashcomp=double(ashcomp);
%make spat mask
spatcomp=A-ashcomp;
thresh=10;
spatcomp=spatcomp>thresh;
%take out stuff below the crater rim
for i=1:length(cc)
    spatcomp(cr(i):size(A,1),cc(i))=0;
end
spatimg=spatcomp.*A;
ashmask=ones(size(A,1),size(A,2));
for i=1:length(cc)
    ashmask(cr(i):size(A,1),cc(i))=0;
end
ashimg=ashmask.*ashcomp;

function [smass,snum,sdist] = spatmass(ashimg,spatimg)
%function computes the two components of a flir pixel

x=linspace(7,13,100);
x2=x.*x.~6;
c1=3.742*10.~16;
c2=0.0144;

return
12.83; 13.04; 13.38; 13.88; 14.1; 14.38;...
15.08; 15.75; 15.92; 16.55; 16.76; 17.0; 17.53;...
17.67; 17.91;
ry=[0; 0.03; 0.08; 0.2; 0.43; 0.69; 0.85;...
0.91; 0.95; 0.97; 0.98; 0.96; 0.93; 0.92;...
0.93; 0.95; 0.95; 0.95; 0.93; 0.89; 0.86;...
0.75; 0.62; 0.48; 0.47; 0.46; 0.42; 0.39;...
0.37; 0.31; 0.29; 0.3; 0.31; 0.23;...
0.09; 0.08; 0.11; 0.11; 0.09; 0.01; 0;...]
mass=zeros(0);
radius=zeros(0);

for r=1:size(ashimg,1)
  for c=1:size(ashimg,2)
    if spatimg(r,c)>0
      %BOMB:
      T=1000; %Temp in Celsius
      T0=T+273.15;
      %radiance vs wavelength
      y0=(c1*(x2. A-5).I(pi*(exp(c2.1(x2.*T0)-1)));
      %response correction
      % for i=1:length(y0)
      % for j=2:length(rx)
      %  if rx(j)>y0(i) & rx(j-1)<y0(i)
      %    y0(i)=y0(i)*ry(j);
      %  end
      % end
      % end
      %integrated radiance
      z0=trapz(x2,y0);
      %BACKGROUND
      T1=ashimg(r,c); %+273.15; %BG Temp in K
      %radiance vs wavelength
      y1=(c1*(x2. A-5).I(pi*(exp(c2.1(x2.*T1)-1)));
      %response correction
      % for i=1:length(y1)
      % for j=2:length(rx)
      %  if rx(j)>y1(i) & rx(j-1)<y1(i)
      %    y1(i)=y1(i)*ry(j);
      %  end
      % end
      % end
      %integrated radiance
      z1=trapz(x2,y1);
      %PIXEL INTEGRATED
      T2=spatimg(r,c); %+273.15; %Pixel int. Temp in K
      %radiance vs wavelength
      y2=(c1*(x2. A-5).I(pi*(exp(c2.1(x2.*T2)-1)));
      %response correction
      % for i=1:length(y2)
      % for j=2:length(rx)
      %  if rx(j)>y2(i) & rx(j-1)<y2(i)
      %    y2(i)=y2(i)*ry(j);
      %  end
      % end
      % end

  end
end
%integrated radiance
z2=trapz(x2,y2);

%z0=bomb radiance
%z1=background radiance
%z2=PIT radiance
if z0>z1 & z2>z1 & z0>z2
    %PIXEL MIXING
    f2=(z0-z1)/z2 %fraction of hot component
    f2=(z2-z1)/(z0-z1);
    if f2>1
        disp(f2)
        f2=0;
    end
    pixarea=0.3426; %area of flir pixel at 450 m
    bombarea=f2*pixarea;
    newradius=sqrt(bombarea/pi);
    radius=[radius newradius];
    volume=(4/3)*pi*newradius^3;
    density=1520;
    density=1100;
    newmass=density*volume;
    mass=[mass newmass];
    kl=newmass;
else disp('Whassup?')
    %mass(i)=0;
    T0;
    T1;
    T2;
    end
end
end

smass=sum(mass);
smass=length(mass);
sdist=radius;

return

%plot(T2-273.15,radius*2)
%xlabel('Temperature (C)');
%ylabel('Bomb size (diameter)');

%pause

%plot(x,y2,y)
%xlabel('Wavelength (micrometers)');
%ylabel('Radiance (W m^-3)');

function y = collimate2(b(A))
%function calculates collimation of spatter image
rar=zeros(0);
car=zeros(0);
spstd=nan;
A=A>10;
ccount=zeros(0);
rcount=zeros(0);

for c=1:size(A,2)
    for r=1:size(A,1)
        if A(r,c)>0
            ccount=[ccount c];
        end
    end
end

for r=1:size(A,1)
    for c=1:size(A,2)
        if A(r,c)>0
            rcount=[rcount r];
        end
    end
end

if length(ccount)>50
    pmean=mean(ccount)/mean(rcount);
    spstd=std(ccount)/std(rcount);
end

y=spstd;

function [amass,aconc] = ashmass(ashimg);
%function calculates ash content in strombolian plumes

psize=0.5854;
airdense=1.05;
Tashl=1000;
Tash2=1000;
Tgas=1000;
Tash2=750;
Rc=8.314;
mwair=0.0289;
mwh2o=0.018;
cash=1100;
cair=1000;
cgas=2000;
ratio=0.25;
P=0.9e5;

%go into eruption image, isolate plume and calculate ash mass
thresh=25+273.15;
bgtemp=25;
ashimg=(ashimg>thresh).*ashimg;
dTash=Tash1-bgtemp;
dTgas=Tgas-bgtemp;

%figure(3)
%imagesc(ashimg,[280 400]);
%pause
A=ashimg;
Mass=0;
%go through image row by row, find ash plume
for r=1:size(ashimg,1)
    minc=1;
    maxc=1;

key=0;
for c=2:size(A,2)
    if ashimg(r,c)>0 & ashimg(r,c-1)==0
        minc=c;
        key=1;
    end
    if key==1 & ashimg(r,c)==0 & ashimg(r,c-1)>0
        maxc=c;
    elseif c==size(A,2) & ashimg(r,c)>0
        maxc=size(A,2);
    end
end
if (maxc-minc>1
    Tbulk(r)=mean(ashimg(r,minc:maxc));
    dTbulk(r)=Tbulk(r)-273.15-bgtemp;
    radius(r)=floor(((maxc-minc)/2)*psize;
    vol(r)=pi*(radius(r)^2)*psize;
    areaarray(r)=2*pi*radius(r)*psize;
    air dense=P./(287*Tbulk(r));
    airmass1(r)=air dense*vol(r);
    airmass2(r)=airmass1(r)*Tbulk(r)/(Tash1-Tbulk(r));
term1=((mwair*P*vol(r)*cair*dTbulk(r)/(Rc*Tbulk(r)))
term2=fcgas*dTgas;
term3=(mwair*cair*dTbulk(r))/(mwh2o);
term4=fcgas*dTbulk(r);
% airmass, n=0;
Mass1(r)=airmass1(r)*cair*dTbulk(r)./(((cair*dTash)-(cair*dTbulk(r)));
% gas mass, n=1;
Mass2(r)=term1.//(term2+term3-term4);
Mass(r)=Mass2(r);
% airmass(r)=(((P*vol(r)/(Rc*Tbulk(r)))-(ratio*airmass2(r)/mwh2o))*nwair;
% bulk dense(r)=(airmass2(r)+gasmass(r)+airmass(r))/vol(r);
end
end
amass=sum(Mass);
%amass=sum(airmass2);
aconc=0; %amass/(sum(airmass)+sum(gasmass)+amass);
Appendix C

First recorded eruption of Mount Belinda volcano (Montagu Island), South Sandwich Islands

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Abstract

The MODVOLC satellite monitoring system has revealed the first recorded eruption of Mount Belinda volcano, on Montagu Island in the remote South Sandwich Islands. Here we present some initial qualitative observations gleaned from a collection of satellite imagery covering the eruption, including MODIS, Landsat 7 ETM+, ASTER, and RADARSAT-1 data. MODVOLC thermal alerts indicate that the eruption started sometime between 12 September and 20 October, 2001, with low-intensity subaerial explosive activity from the island's summit peak, Mount Belinda. By January 2002 a small lava flow had been emplaced near the summit, and activity subsequently increased to some of the highest observed levels in August 2002. Observations from passing ships in February and March 2003 provided the first visual confirmation of the eruption. ASTER images obtained in August 2003 show that the eruption at Mount Belinda entered a new phase around this time, with fresh lava effusion into the surrounding icefield. MODIS radiance trends also suggest that the overall activity level increased significantly after July 2003. Thermal anomalies continued to be observed in MODIS imagery in early 2004, indicating a prolonged low-intensity eruption and the likely establishment of a persistent summit lava lake, similar to that observed on neighboring Saunders Island in 2001. Our new observations also indicate that lava lake activity continues on Saunders Island.
Keywords: Mount Belinda volcano, Montagu Island, Mount Michael, Saunders Island, South Sandwich Islands, satellite monitoring, ASTER

Introduction

New volcanic activity was initiated in September or October 2001 at Mount Belinda, on Montagu Island in the remote South Sandwich Islands, according to both automated and visual interpretation of thermal satellite imagery (Smithsonian Institution, 2003, 2004). This eruption is especially significant in that it is the first recorded volcanic activity on Montagu Island, and its detection was facilitated by the global MODVOLC satellite monitoring system, based at the Hawaii Institute of Geophysics and Planetology (Wright et al., 2002, 2004).

Data were subsequently gathered from a variety of sources to document the eruption at Mount Belinda (Table 1). The detection of activity and initial analysis were performed using the automated MODVOLC thermal alert system. Covering an observation period from the start of the eruption (October 2001) to May 2004, MODIS (Moderate Resolution Imaging Spectroradiometer) thermal images (1 km pixel size) showed the relative changes in heat flux that could be related to variations in activity. The MODVOLC system allowed us to quickly and easily generate heat flux time series capable of revealing such variation (Wright and Flynn, 2004). Several visible band MODIS images (250 m pixel size) showed the progression of eruptive activity at the summit. Supplementing these data, Landsat 7 Enhanced Thematic Mapper Plus (ETM+) images
Table 1. Summary of key observations for Montagu Island.

<table>
<thead>
<tr>
<th>Date</th>
<th>Source</th>
<th>Observation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sept 1992</td>
<td>Hand-held photographs</td>
<td>No signs of activity; summit ice covered</td>
</tr>
<tr>
<td>1995-1998</td>
<td>AVHRR</td>
<td>Possible low-intensity intermittent activity</td>
</tr>
<tr>
<td>6 Mar 1998</td>
<td>RADARSAT-1</td>
<td>No signs of activity</td>
</tr>
<tr>
<td>Jan 1997</td>
<td>Visual observation from neighbor island</td>
<td>No signs of activity; summit ice covered</td>
</tr>
<tr>
<td>24 Jan 2001</td>
<td>Landsat 7 ETM+</td>
<td>No signs of activity; summit ice covered</td>
</tr>
<tr>
<td>12 Sept 2001</td>
<td>ASTER</td>
<td>No signs of activity; summit ice covered</td>
</tr>
<tr>
<td>20 Oct 2001</td>
<td>MODIS (thermal)</td>
<td>First thermal anomaly</td>
</tr>
<tr>
<td>23 Dec 2001</td>
<td>ASTER</td>
<td>Elongate tephra deposit E of summit</td>
</tr>
<tr>
<td>4 Jan 2002</td>
<td>Landsat 7 ETM+</td>
<td>Ash plume, small lava flow</td>
</tr>
<tr>
<td>19 May 2002</td>
<td>MODIS (visible)</td>
<td>Tephra deposit W of summit</td>
</tr>
<tr>
<td>Aug 2002</td>
<td>MODIS (visible and thermal)</td>
<td>Apparent escalation in activity, lava or tephra deposit N of summit</td>
</tr>
<tr>
<td>Early Feb 2003</td>
<td>Photographs from HMS Leeds Castle</td>
<td>Ash plume from summit</td>
</tr>
<tr>
<td>2 Mar 2003</td>
<td>Visual observations from RRS Ernest Shackleton</td>
<td>Prominent ash plume from summit</td>
</tr>
<tr>
<td>1+17 Aug 2003</td>
<td>ASTER</td>
<td>2 km long lava/debris flow</td>
</tr>
<tr>
<td>7 Dec 2003; 9 Feb 2004</td>
<td>ASTER</td>
<td>Small plume and tephra cover SE of Mount Belinda</td>
</tr>
<tr>
<td>May 2004</td>
<td>MODIS (thermal)</td>
<td>Thermal anomalies present at time of submission</td>
</tr>
</tbody>
</table>
and Advanced Spaceborne Thermal Emission and Reflection Radiometer (ASTER) images (15-90 m pixel size) provided higher spatial-resolution snapshots of the impact of eruptive products on the surface of the ice-capped island. RADARSAT-1 synthetic aperture radar (SAR) imagery (C-band, 9-30 m pixel size) completed the suite of satellite data available for eruption tracking purposes. Finally, visual observations and photographs of the island taken from two passing ships corroborated the eruptive activity.

Background

Montagu Island and the South Sandwich Islands

The South Sandwich Islands lie between approximately 56° and 60° S latitude, and 25° and 30° W longitude, at the eastern extremity of the Scotia Sea, approximately 2000 km south-east of the Falkland Islands and about the same distance north of the Antarctic continent (Fig. 1). This young (<5 Ma) volcanic arc is a product of westward subduction of the South American plate beneath the South Sandwich plate, with lavas ranging from tholeiitic and calc-alkaline basalts to basaltic andesites, and minor occurrences of more silicic rocks (Pearce et al., 1995; Leat et al., 2003). Eruptive activity has largely been effusive in nature (LeMasurier and Thomson, 1990). Active fumaroles are, or have been, present on Leskov, Zavodovski, Candlemas, Bristol, Thule and Bellingshausen islands, and eruptions have been reported for Zavodovski, Visokoi, Saunders, and Bristol islands.
Figure 1. The South Sandwich Island archipelago, located in the Scotia Sea. The South Sandwich Trench lies approximately 100 km east, paralleling the trend of the islands.
(Holdgate, 1963; Baker, 1990). A notable exception to the typical effusive nature of the islands was the explosive submarine eruption of rhyolite from Protector Shoal (about 56 km north of Zavodovski Island) in 1962, which produced a raft of pumice that drifted as far as New Zealand (Gass et al., 1963; Coombs and Landis, 1966).

Montagu Island is the largest of the South Sandwich Islands (Figs. 1, 2), measuring approximately 12 km by 10 km (note that these dimensions differ from those indicated on the only published topographical map of the island, by Holdgate and Baker (1979), which featured an incorrect scale bar). Mount Belinda, named after Belinda Kemp (Kemp and Nelson, 1931), rises to 1370 m asl and is a small summit peak situated within an extensive very gently sloping icefield that fills the largest caldera known in the island group. Mount Oceanite is a small satellite center, larger than Mount Belinda, which forms a conspicuous promontory on the south-east corner of the island (Fig. 2). Although no reliable topographic data exist for the island, examination of the high spatial-resolution imagery (ASTER and ETM+) indicates that Mount Belinda is a small summit cone on the larger shield volcano edifice represented by Montagu Island. The broad summit caldera is approximately 6 km in diameter and is entirely filled by permanent ice of uncertain depth. The outlet of the summit icefield is a tidewater glacier reaching the sea at the NW corner of the island, resulting in a temporally-variable shoreline in that area. Mount Belinda represents the youngest post-caldera eruptive center known on the island. Other likely centers, probably cinder cones in various states of degradation, are present as a small unnamed peak on the west side of the caldera, and two considerably
Figure 2. A) Map of Montagu Island, adapted from Holdgate and Baker (1979). Stippled areas show rock outcrop, the remainder is snow or ice covered. Relief is shown by form lines which should not be interpreted as fixed-interval contours. Note correction of scale from original figure. North is up. B) 12 Sept 2001, pre-eruption ASTER visible band composite image (Bands 3-2-1). Note the difference in location of Mount Belinda. The black dotted line shows the rough boundary of the apparent caldera.
degraded outcrops on the south-east outer flank facing Mount Oceanite (Fig. 2 and unpublished information). The steep-sided Mount Oceanite is capped by a 270 m diameter summit crater, which is about 100 m deep based on the interior shadow and sun elevation angle of the ETM+ image. The ASTER image (Fig. 2b) improves upon the positioning of the topographical features from the original coarse-resolution form-line map (Fig. 2a) of Holdgate and Baker (1979).

The island as a whole is about 90% ice covered, with most of the exposed rock limited to vertical sea cliffs. Because of this inaccessibility, landings have only been made on a handful of occasions, and just a few localities (Borley, Scarlett, Horsburgh, Allen and Mathias points) have ever been sampled (LeMasurier and Thomson, 1990, and unpublished information). The rocks range from basalt to basaltic andesite (49-53 wt. % SiO₂) with low amounts of Na₂O and K₂O. There was no prior record of Holocene activity on Montagu, though this is likely to reflect its remoteness, inaccessibility and extensive ice cover, which also prevent a thorough study of the island (LeMasurier and Thomson, 1990).

*Satellite imagery and the MODVOLC thermal alert system*

Because of its remote location and the persistently bad weather there have been just three formal scientific expeditions to Montagu Island, those of the RRS *Discovery* in 1930
(Kemp and Nelson, 1931), HMS *Protector* in 1962 (Holdgate, 1963), and by the British Antarctic Survey in 1997 (Leat et al., 2003). The fact that no future expeditions are being planned at the current time is not surprising: Captain James Cook, the well-traveled explorer who discovered the South Sandwich Islands in 1775, referred to them as being ‘the most horrible Coast in the World’, and the region in general as ‘a Country doomed by Nature never once to feel the warmth of the Sun’s rays, but to lie for ever buried under everlasting snow and ice.’ (Cook, 1775). That regular field studies in this region are so unlikely underlines the potential stand-alone role of satellite data for monitoring volcanic activity in the South Sandwich Islands.

Previous eruptions have been recorded in the archipelago, but ongoing volcanic activity has only recently been detected and studied in the South Sandwich Islands. These islands are sufficiently distant from population centers and shipping lanes that eruptions, if and when they do occur, can easily go unnoticed. Visual observations of the islands probably are generally limited to no more than a few days each year (LeMasurier and Thomson, 1990). Satellite data have recently provided observations of volcanic activity in the group, and offer the only practical means to monitor activity in the islands. Specifically, using Advanced Very High Resolution Radiometer (AVHRR) data, Lachlan-Cope et al. (2001) discovered and analyzed an active lava lake at the summit of Mount Michael volcano on Saunders Island, north of Montagu Island.
The satellite data used here span a wide range of spatial and temporal resolutions (Table 2). MODIS imagery is acquired on average twice daily over a given spot, and is composed of 20 thermal infrared bands at 1 km pixel size, 14 visible and infrared bands at 500 m, and two visible bands at 250 m. Of the thousands of MODIS images acquired during the study period (Oct. 2001 – May 2004) only 107 triggered the MODVOLC thermal alert system, limited either by fluctuations in activity level or, more importantly, visibility of the island due to cloud cover. Of these 107 images, all but 10 were nighttime, and thus the 250 m visible band data were useless as a complement for the majority of the thermal anomaly observations. Landsat 7 ETM+ has a much longer repeat period (16 days), with channels ranging from 15 m pixel size (panchromatic) to 60 m (thermal infrared). Again, while almost a dozen ETM+ images were acquired over Montagu Island during the eruption period, only one was sufficiently cloud-free to be of use. ASTER has infrared bands with a nominal spatial-resolution of 30 and 90 m, and visible bands at 15 m. However, its repeat period is variable since coverage must be requested and scheduled in advance. A pre-eruption, as well as five syn-eruption, ASTER images were used here. Several RADARSAT-1 SAR amplitude images (C-band; 5.6 cm), acquired in both fine (9 m pixel size) and standard (30 m pixel size) modes, improved our visualization of the island’s morphology. These data have the benefit of being unaffected by cloud cover.

Thermal anomalies in MODIS data were detected and analyzed using the automated MODVOLC satellite monitoring system (Wright et al., 2002). The alert system offers
Table 2. Sensor characteristics

<table>
<thead>
<tr>
<th>Sensor</th>
<th>Selected channels</th>
<th>Pixel size (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>MODIS</td>
<td></td>
<td></td>
</tr>
<tr>
<td>1: 0.62-0.67 μm</td>
<td></td>
<td>250</td>
</tr>
<tr>
<td>2: 0.84-0.87 μm</td>
<td></td>
<td>250</td>
</tr>
<tr>
<td>21: 3.929-3.989 μm (high gain)</td>
<td></td>
<td>1000</td>
</tr>
<tr>
<td>22: 3.929-3.989 μm (low gain)</td>
<td></td>
<td>1000</td>
</tr>
<tr>
<td>32: 11.77-12.27 μm</td>
<td></td>
<td>1000</td>
</tr>
<tr>
<td>ASTER</td>
<td></td>
<td></td>
</tr>
<tr>
<td>1: 0.52-0.60 μm</td>
<td></td>
<td>15</td>
</tr>
<tr>
<td>2: 0.63-0.69 μm</td>
<td></td>
<td>15</td>
</tr>
<tr>
<td>3: 0.76-0.86 μm</td>
<td></td>
<td>15</td>
</tr>
<tr>
<td>13: 10.25-10.95 μm</td>
<td></td>
<td>90</td>
</tr>
<tr>
<td>Landsat 7 ETM+</td>
<td></td>
<td></td>
</tr>
<tr>
<td>1: 0.45-0.52 μm</td>
<td></td>
<td>30</td>
</tr>
<tr>
<td>2: 0.52-0.60 μm</td>
<td></td>
<td>30</td>
</tr>
<tr>
<td>3: 0.63-0.69 μm</td>
<td></td>
<td>30</td>
</tr>
<tr>
<td>4: 0.76-0.90 μm</td>
<td></td>
<td>30</td>
</tr>
<tr>
<td>5: 1.55-1.75 μm</td>
<td></td>
<td>30</td>
</tr>
<tr>
<td>6a: 10.4-12.5 μm (high gain)</td>
<td></td>
<td>60</td>
</tr>
<tr>
<td>6b: 10.4-12.5 μm (low gain)</td>
<td></td>
<td>60</td>
</tr>
<tr>
<td>7: 2.08-2.35 μm</td>
<td></td>
<td>30</td>
</tr>
<tr>
<td>8: 0.50-0.90 μm</td>
<td></td>
<td>15</td>
</tr>
<tr>
<td>RADARSAT-1</td>
<td>C-band (5.6 cm): Fine mode</td>
<td>9</td>
</tr>
<tr>
<td></td>
<td>C-band (5.6 cm): Standard mode</td>
<td>30</td>
</tr>
</tbody>
</table>
global detection of high-temperature phenomena (volcanic activity, industrial hotspots and forest fires), with data displayed within 24-72 hours of image reception at http://modis.higp.hawaii.edu. The detection algorithm exploits the characteristic differences in radiance of hot (in this case volcanically active) ground surfaces compared to that of background surfaces among MODIS bands 21, 22 and 32 (see Table 2). Anomalous pixels are flagged and their locations posted on the website for analysis, but the algorithm does not yet take the role of alerting the user to new or unusual activity. Indeed, anomalies at Mt. Belinda were only discovered by analysts some months after their initial appearance, during the course of examining anomalies at nearby Mt. Michael volcano on Saunders Island.

Because the MODVOLC data are openly available on the web and the original MODIS data are free of charge from the United States Geological Survey, the costs of the imagery used here were limited to the two Landsat ETM+ (currently US $600/image), six ASTER scenes (US $55/image), and two RADARSAT-1 images (US $16/image). This kept the data cost of the entire 2.5-year satellite monitoring effort to less than US $1600.

**Chronology of activity**

Here we divide the eruption into four phases based upon MODIS thermal anomaly intensity and inferred activity style from the high spatial-resolution imagery.
There is no previous record of definitive volcanic activity on Montagu Island. Based on AVHRR images obtained in the period March 1995 to February 1998, apparent plumes and unreported single anomalous pixels were observed intermittently on images of Montagu Island and may indicate rare and sporadic (and unverified) volcanic activity prior to that now observed (unpublished observations of T Lachlan-Cope and JLS). However, during field investigations by one of us (JLS) in January 1997, Montagu Island was viewed from both Saunders and Bristol islands and was apparently inactive, with the summit region entirely clothed in snow and ice. Hand-held photographs of the island obtained in September 1992 also show the summit to be wholly inactive. Those observations suggest that the activity reported here is likely a recent phenomenon, and has probably increased relative to previous unconfirmed 1995-1998 levels.

An ASTER image acquired on 12 September 2001 constrains the start of the recent eruption as being after this date, as the summit region contains no visible deposits or thermal anomalies (Fig. 2b). Furthermore, an earlier Landsat 7 ETM+ scene, dated 24 January 2001 (not shown), also lacks any signs of activity, and indicates that the September ASTER scene likely does not contain any recent deposits that may have simply been covered by the preceding months of Antarctic winter snows. A pre-eruption RADARSAT image (5 March 1998, not shown) shows arcuate fractures in the ice cover approximately 2 km east of Mount Belinda which resemble those resulting from
subglacial volcanism (Gudmundsson et al., 1997), however their position on steep terrain and lack of recent confirmed activity suggests that these are topographic crevasses.

**Phase 1 (Oct 2001-Feb 2002): minor thermal anomalies, limited lava flow**

The first MODVOLC thermal alert on Montagu Island occurred on 20 October 2001, with a single anomalous pixel on the north side of the island, indicating that the activity started at some point in the preceding five weeks since the September ASTER acquisition. Subsequent anomalies between October 2001 and February 2002 were generally one to two pixels in size, and of low intensity relative to the subsequent phases of the eruption. Visual inspection of the Band 21 imagery revealed that these and all subsequent anomalies were located near the summit of Mount Belinda, changing slightly in position either due to satellite viewing geometry or actual migration of hot material (Fig. 3). The temporal trend of all the MODVOLC thermal alerts for the duration of our study period is featured in Figure 4, which shows the total radiant heat output throughout the course of the study period calculated from MODIS Band 21 nighttime imagery using a simple empirical relation (Kaufman et al. 1998; Wright and Flynn, 2004).

A night-time ASTER image dated 23 December 2001 (not shown here), shows the island lacking any overt signs of volcanic activity, with the exception of a subtle elongate feature extending north-east from the summit that exhibits a slightly higher radiance than the surrounding snow and ice. The lack of lateral confinement on this feature, which
Figure 3. Selected MODIS images showing thermal anomalies at Montagu Island. Band 21 is shown here. The thermal anomalies at Montagu appear to be located near the summit of Mount Belinda (see Fig. 2). Images are not georeferenced in order to maintain radiance integrity, therefore coastlines are approximate.
Figure 4. Total radiant heat output in MegaWatts at Montagu Island from the start of the eruption (Oct 2001) to the time of writing (May 2004), calculated from the MODVOLC analysis of MODIS night-time Band 21 data (Wright and Flynn, 2004). Letters denote acquisition dates of high-resolution imagery, ASTER (A), Landsat ETM+ (L), and RADARSAT (R), as well as the date of visual observations from plane and ship (V).
would be expected for a lava flow emplaced on the ice cover, suggests that this is a solar-heated tephra deposit.

Significant volcanic activity on Mount Belinda was confirmed by a Landsat 7 ETM+ image acquired on 4 January 2002 (Fig. 5a). A distinct steam/ash plume can be seen drifting south from the summit, while the entire north and east flanks are tephra covered. A large summit thermal anomaly is present in the thermal infrared Band 6 indicating a significantly heated surface, and smaller anomalies are present in Bands 5 and 7, indicating temperatures greater than about 210° C (Flynn et al., 2001). With a source at the summit anomaly and trending NE, there is an elongate high temperature feature that appears to be a short lava flow, extending ~600 m with a width of ~210 m.

**Phase 2 (Feb 2002-Nov 2002): High radiance anomalies**

The trend of the MODVOLC thermal alerts shows an apparent rise in activity level in February 2002 (Fig. 4), possibly indicating increased explosive activity at the summit. Total radiant heat output reached a local maximum on August 2, 2002, and during that week some of the largest thermal anomalies in our study period were observed (up to four pixels). The visible band MODIS images indicate that tephra was being deposited on the north flank of Belinda at this point. We can now discount the emplacement of any significant amount of lava in this period (Smithsonian Institution, 2004), as analyses of high spatial-resolution data acquired later in the eruption bear no evidence of alteration to
Figure 5. Selected high spatial-resolution images of Montagu Island. North is up. A) Landsat 7 ETM+ image from 4 Jan 2002 showing diffuse plume (P) emanating from Mount Belinda’s summit (MB) and tephra deposits on north flank. Scale bar applies to A, B and C. B) ASTER visible band composite image (Bands 3-2-1) on 7 Dec 2003, showing tephra deposits and 2003 lava flow (L2). C) RADARSAT-1 image from 30 Oct 2003 showing recent morphology, with inset (D). Arrows point to approximate summit of Mount Belinda and vent location (MB), ash plumes (P), 600 m long lava flow first observed in Jan 2002 (L1), entrenched 2 km long lava flow first observed in Aug 2003 (L2), and arcuate fractures unrelated to this eruption (F). RADARSAT image was provided by the Alaska Satellite Facility, and is copyright 2003 CSA.
the icefield. It is possible that the August maximum represents an unusually clear viewing period (as supported by the increased number of clear images), and may not reflect relative activity level.

Phase 3 (Nov 2002-Jul 2003): Low radiance anomalies

MODIS thermal anomalies between November 2002 and July 2003 were generally low to moderate in intensity (Fig. 4), and no high-resolution images were recorded during this period. At the behest of the British Antarctic Survey, observations were taken from the HMS Leeds Castle in early February 2003 providing the first photographic proof of the eruption, in which a low-level plume was shown rising above a tephra-covered icefield (Fig. 6). This remains the only ground-based photograph of the eruption in almost 3 years of monitoring by satellites, illustrating the significant difficulties in acquiring information in this remote region. One month later, a British Antarctic Survey research vessel was diverted to gather observations of the island, and although heavy fog prevented a clear view of the flanks, a prominent ash plume was observed rising from the summit on 2 March 2003.
Figure 6. Photograph of Mt. Belinda volcano in eruption in early February 2003, courtesy of the HMS Leeds Castle. View is looking to the south at the north coast of the island. Note the thin mantle of tephra covering the glaciers.
Phase 4 (Jul 2003-May 2004): High radiance anomalies, long lava flow

Anomalies were detected in MODIS imagery up until the time of writing (last on 22 March 2004), indicating that activity has persisted. Radiant heat flux reached its highest observed levels in this phase (Fig. 4), with a distinct spike on 16 October 2003.

ASTER images acquired on 1 and 17 August, 2003, show a dark sinuous lava flow extending north-east from the summit. The 1 August image (not shown here) also shows a substantial dense billowing plume, which is absent on 17 August. The latter image is the clearer of the two and shows a dark tephra deposit also directed north-eastward but slightly offset to the south of the flow, and a minor plume extending less than a km from Mount Belinda summit (Fig. 7). The lava is about 2 km in total length. It broadens from about 100 m wide close to source, to 200 m at 1.5 km, at which point it swells to a paler-colored fan-shaped flow front about 600 m in diameter with a crescent-like termination and at least one arcuate ridge-like feature. The narrow proximal section of the flow appears radiant in the thermal infrared (Band 6) on 17 August, at which point it is at least two weeks old, and surrounding shadows suggest syn-emplacement downcutting into the ice and resulting lateral confinement similar to lava flows at Westdahl volcano, Alaska, in 1991 (Dean et al., 2002). A 7 December 2003 ASTER image (Fig. 5b) and 30 October 2003 RADARSAT image (Fig. 5c,d) show this downcutting well. It is unclear whether this lava channel terminates as a lava-fed delta, entering a small meltwater lake as occurred in the 1983 eruption of Veniaminoff (Yount et al., 1985).
Figure 7. ASTER visible band composite image (Bands 3-2-1) on 17 Aug 2003, showing lava flow. The insets show (A) a visible band close-up of the lava flow and debris fan extending from the summit of Mount Belinda and (B) the equivalent view in the thermal infrared (ASTER Band 13).
The most recent ASTER image (9 February 2004, not shown) has Mount Belinda emitting only a faint plume, with thermally anomalous pixels at the vent and on the proximal sections of the still-cooling 2 km long lava flow. The distal fan-shaped portion of this flow appears to be snow covered. The lack of new lava in this image indicates that the high heat flux values observed on 16 October 2003 (Fig. 4) were not related to lava effusion.

Concurrent activity at Saunders Island lava lake

An active lava lake in the Mount Michael summit crater on Saunders Island (Fig. 1) was first reported by Lachlan-Cope et al. (2001) using AVHRR imagery from 1995-1998. The MODVOLC system has detected repeated thermal anomalies throughout 2001-2003 in the summit area (Fig. 8), indicating that the lava lake has persisted. Anomalous pixels were detected intermittently and were nearly all one to two pixels in size, consistent with the relatively small confines of the crater. At the time of submission no anomaly had been detected for ~12 months (i.e. since May 2003), however the 2001-2003 period was marked by intervals up to 7 months, indicating that detection is strongly limited by viewing conditions or activity level.
Figure 8. Total radiant heat output in MegaWatts from Mt. Michael volcano, Saunders Island, as calculated from MODVOLC thermal alerts.
Periods between observations are much longer at Michael (as much as seven months) compared with Mount Belinda, possibly due in part to their differing geometry. On Mount Belinda, material is erupted onto the volcano flanks, permitting any satellite zenith angle to view it. The maximum zenith angle for the MODVOLC thermal alert imagery analyzed at Belinda is 56.9°, which is approaching the maximum possible zenith angle, and the mean zenith angle is 20.5°. On the other hand, all of the hot material at Michael is within the ~700 m wide crater at an unknown depth, which may be variable (Lachlan-Cope et al., 2001). If sufficiently deep, only small zenith angles (i.e. closer to vertical) would permit an unobstructed view into the crater and onto the lava surface (Harris et al., 1997), similar to the situation at Shishaldin, Alaska, in 1999 (Dehn et al., 2002). The maximum zenith angle for the imagery at Michael is 28.4° and the mean is 15.2°, although the sample size is much smaller here than at Mount Belinda. The depth of the lava surface might be calculated with these values (Dehn et al., 2002), however, the 700 m diameter measurement is only a rough estimate and the actual conduit likely narrows considerably at some depth within this outer crater. Nevertheless, along with cloud cover the zenith angle is another critical factor in anomaly detection.

Anomaly detection may also be limited by the increasing pixel size at high zenith angles. For similar sensors, at extreme scan the spatial-resolution of the sensor is several times coarser than it is at nadir (Mouginis-Mark et al., 1994; Harris et al., 1997). For a small hot target, integrating its emitted radiance into these much larger areas may lead to a pixel-integrated radiance that is insufficient to trigger the MODVOLC detection scheme.
Conclusions

The South Sandwich Islands are a young active intra-oceanic volcanic arc whose remote location precludes the possibility of any comprehensive ground-based monitoring or research. As a partial remedy, satellite data offer the only realistic means of establishing the timing and evolution of the activity, as well as providing clues regarding the type, style and scale of the eruptive products. In this study, we have described the rough chronology of the first recorded eruption at Mount Belinda volcano, on Montagu Island, based almost solely upon an analysis of satellite imagery. This eruption is apparently still continuing. Its detection was made possible with the MODVOLC satellite monitoring system, a tool which will now facilitate regular tracking of volcanic activity in the South Sandwich Islands and other remote regions.

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