Chronology and complex volcanic processes during the 2002–2003 flank eruption at Stromboli volcano (Italy) reconstructed from direct observations and surveys with a handheld thermal camera

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Received 6 April 2004; revised 12 November 2004; accepted 22 November 2004; published 2 February 2005.

[1] Effusive activity at Stromboli is uncommon, and the 2002–2003 flank eruption gave us the opportunity to observe and analyze a number of complex volcanic processes. In particular, the use of a handheld thermal camera during the eruption allowed us to monitor the volcano even in difficult weather and operating conditions. Regular helicopter-borne surveys with the thermal camera throughout the eruption have significantly improved (1) mapping of active lava flows; (2) detection of new cracks, landslide scars, and obstructions forming within and on the flanks of active craters; (3) observation of active lava flow field features, such as location of new vents, tube systems, tumuli, and hornitos; (4) identification of active vent migration along the Sciara del Fuoco; (5) monitoring of crater’s inner morphology and maximum temperature, revealing magma level changes within the feeding conduit; and (6) detection of lava flow field endogenous growth. Additionally, a new system developed by A. J. L. Harris and others has been applied to our thermal data, allowing daily calculation of effusion rate. These observations give us new insights on the mechanisms controlling the volcanic system.


1. Introduction

[2] The typical activity of Stromboli volcano (Aeolian Islands, Italy) is characterized by regular explosions that send small volumes of lava 50–100 m above the craters every 15–20 min; this type of behavior has become synonymous with the volcano. Lava effusion is also observed on Stromboli; prior to 2002 the last such event initiated on 6 December 1985 and lasted for 141 days. During that eruption, lava flows extended down the Sciara del Fuoco (SDF), a collapse scar located on the N flank of the volcano (Figure 1a), and emplaced a ~6 × 10^6 m^3 lava flow field [De Fino et al., 1988] at a typical effusion rate of ~0.3 m^3 s^-1 [Harris et al., 2000a].

[3] On 28 December 2002, a new effusive eruption started at Stromboli volcano that continued until 22 July 2003. This 2002–2003 eruption was heralded by an increase in the intensity of strombolian activity and a heightened magma level within the main conduit, both of which increased from May 2002 onward. The increasing magma level culminated in an overflow from the N rim of the central crater (Crater 2, Figure 2) during November and resulted in the emplacement of a lava flow, a few tens of meters in length, on the upper SDF. Explosions increased in both intensity and frequency during early December, where eruptive activity was particularly intense at Crater 1 (Figure 2). On 28 December the height of ejecta emitted during strombolian events reached 200 m above Crater 1. In addition, the wide spread of ejecta emitted during explosions, as observed in images provided by the INGV-CT web-cam located at Il Pizzo (available at http://www.ct.ingv.it) (Figure 1a), suggested that the level of the magma surface was very close to the crater rim. This activity climaxed during the afternoon of 28 December with the opening of a NE-SW trending eruptive fissure on the SDF (Figure 1b). Fissures opening in the same location, and with the same orientation, had previously fed the 1975 and 1985 north flank effusive
eruptions of Stromboli [Capaldi et al., 1978; De Fino et al., 1988].

[4] Although the value of ground-based (handheld or tripod-mounted) thermal imaging and thermal infrared thermometers in volcanology has long been recognized [e.g., Decker and Peck, 1967; Shimozuru, 1971; Birnie, 1973], only recently has the routine use of ground-based thermal measurements become widely reported. In recent years, ground-based thermal infrared measurements of active volcanic features have been used to achieve the following: (1) recognize magma movements within the central conduits, detect the upward movement of shallow feeder dikes and track eruptive activity [Calvari et al., 2003; Harris et al., 2003]; (2) measure the thermal and rheological properties of active basaltic lava flows, lava tubes and silicic (block) lava flows [Flynn and Mounigis-Mark, 1994; Hon et al., 1994; Harris et al., 2002; Kauahikaua et al., 2003; Wright and Flynn, 2003; Andronico et al., 2004]; (3) analyze the evolution of fumarole fields [Harris and Maciejewski, 2000], active eruption plumes [McGimsey et al., 1999; Dehn et al., 2002; Kaneko et al., 2002; Matsushima et al., 2003], strombolian activity and persistent degassing [Harris et al., 1996; Ripepe et al., 2004]; (4) obtain effusion rate for

Figure 1. (a) Topographic map of Stromboli (north at the lower base) showing in orange the 30 December 2002 landslide scars. (b) Map of the area NW of the Sciara del Fuoco, showing the eruptive fissure on the NE flank of Crater 1, the first lava flow field (30 December 2002 to 15 February 2003) fed from the 500 m elevation vent, and the second (15 February to 22 July 2003) lava flow field from the 670 m vent. (c) Photo from NW of the Sciara del Fuoco, showing the summit craters, the location of effusive activity during 2003, and the Spiaggia dei Gabbiani beach. The red square indicates the area represented in Figure 4.

Figure 2. Photo taken on 28 September 2003 showing the summit craters of Stromboli.
active lava flows [Harris et al., 2005]; (5) detect potential failure planes on recently formed cinder cones [Calvari and Pinkerton, 2004] and fractures developing just before flank collapse at active volcanoes [Bonaccorso et al., 2003]; and (6) analyze active lava lakes [Flynn et al., 1993; Oppenheimer and Yirgu, 2002].

During the 2002–2003 eruption of Stromboli, regular (daily) monitoring with a handheld (and sometimes tripod mounted) portable thermal imagers (FLIRs) proved extremely useful. This instrument was used in one of two modes: either from a helicopter hovering above and/or offshore of the flow field and/or from vantage points overlooking the flow field. In this paper we present an account of the eruption based essentially on information derived from thermal images and visual observations collected throughout the eruption. These images and observations allowed us to detect a number of complex phenomena never described before for this volcano, such as flow field inflation, lava tube formation, apparent vent migration, and alternation between lava flows and debris. Additionally, thermal image-based constraints of the lava effusion rate and the mechanisms governing the magma dynamics within the conduit are presented. We demonstrate that regular monitoring using a handheld thermal imager allows both improved quantification and greater understanding of eruptive phenomena.

2. Chronology of the Eruption

At 1830 LT on 28 December 2002, a 300-m-long eruptive fissure, trending NE-SW, opened on the NE flank of Crater 1 (Figure 1b) of Stromboli volcano. The fissure extended from the NE rim of Crater 1, at an elevation of ~750 m a.s.l., down to 600 m elevation. Debris from the shattered flank of the crater mixed with a fast-moving lava flow that spilled from the crater, mixing with loose material excavated from the steep slope of the SDF. This formed a hot avalanche that flowed down the SDF, following the eastern margin of the 1985 lava flow field, and reached the sea at the Spiaggia dei Gabbiani (Figure 1c). On this short beach, the hot avalanche formed a ~4-m-thick, reddish deposit, composed of subrounded lava clasts and a very fine-grained ash matrix. This deposit was soon covered by ‘a’a lava flows that poured out from the distal end of the fissure. These flows formed two branches on the SDF that extended down the eastern and western margins of the 1985 lava flow field. Local observers reported that within 30 min of onset the two flows had reached the sea, about 1.7 km distant from the source (Figure 1c), giving a time-averaged velocity of ~0.9 m s⁻¹.

2.1. Period of 28 to 30 December 2002: Single Flow Units

The two lava flows erupted from the NE fissure on 28 December 2002 entered the sea along a ~400 m wide front, partially covering the Spiaggia dei Gabbiani beach (Figure 3). However, on steeper slopes higher on the SDF, flow widths were much narrower. The morphology of these flows was typical of early stage flows observed during the 1984 eruption of Mauna Loa [Lipman and Banks, 1987], where lava channel development was inhibited by the high effusion rate, and the flow moved as an unconfined sheet. Given an estimated mean flow thickness of 1 m, the eruption rate for these early stage flows was extremely high at ~280 m³ s⁻¹.

After a two-hour pause in lava effusion, a new flow issued from the distal end of the fissure. This flow was diverted westward by the previously emplaced flows so that it extended down the middle section of the SDF. This flow also entered the sea where it continued to cover the Spiaggia dei Gabbiani beach (Figure 3). The steeper, upper sections of this flow were characterized by ‘a’a morphologies, within which a narrow channel developed. Distally, the flow comprised a thick apron of flowing lava mixed with solid (entrained and rafted) debris. A helicopter-borne thermal survey carried out on the morning of 29 December at 1130 LT (all times are local) allowed us to map the, by then, inactive 28 December lava flows, where surface temperatures had already fallen to below 200°C.

Local residents reported that eruptive activity resumed on 29 December 2002, at ~1630 LT, and a new thermal survey on 30 December at 1130 LT allowed us to distinguish five thermal features: (1) the inactive lava flow of 28 December; (2) the inactive lava flows and vents from 29 December; (3) active and slowly spreading lava flows of 30 December; (4) a new vent at ~570 m elevation feeding a short lava flow; and (5) the opening of hot cracks along the SDF.

While the first four features are shown in Figure 3, the hot cracks are illustrated in Figure 4. Less than two hours later, at 1415 LT, the fractures forming along the SDF merged, causing failure of two large portions of this already unstable flank of the volcano [Bonaccorso et al., 2003]. The volumes of the two failures were estimated respectively at ~5 × 10⁵ m³ and ~6 × 10⁶ m³ [Bonaccorso et al., 2003], with a submarine portion of ~20 × 10⁶ m³ [Pino et al., 2004]. The subaerial portion of the landslide was recorded by video, which revealed sea entry at velocities in excess of 100 km h⁻¹ (M. Pompilio, Eruzione Stromboli 2002–2003: Cronologia dell’eruzione, localizzazione e migrazione delle bocche eruttive, INGV-CT internal report, 2003). The landslides triggered two tsunami’s that caused extensive damage along the east coast of the island. On Stromboli the tsunamis were observed as a sea regression followed by two waves of several meters in height that entered the villages of Stromboli and Ginostra (Figure 1). Tsunami inundation zones extended over 100 m inland, causing damage to buildings and boats, and slightly injuring a few people. Large waves were also recorded as far away as Milazzo on the northern coast of Sicily, ~60 km south of Stromboli. Previous tsunamis at Stromboli occurred in 1930, 1944 and 1954 [Barberi et al., 1993], and have been associated with paroxysmal explosive activity, landslides on the SDF, or pyroclastic flows entering the sea, but never before with lava flow emplacement. Several minor landslides occurred after the major event, involving decreasing volumes of rock with time [Pino et al., 2004].
December 2002 (M. Pompilio, internal report, 2003). It initially extended from between the cracks opening on the SDF, and showed an increase in emission rate around two hours later. Subsequently, a new effusive vent formed at ~500 m elevation, within the largest landslide scar. This new vent fed a lava flow that reached the sea causing phreatic explosions at the shoreline, with wet ash clouds rising from the sea and falling on the village of Stromboli, ~3 km away. Ash fallout was also caused by rock fall activity from the collapse scar, which significantly decreased the following day but continued for several days at a lower intensity during several minor collapses.

Explosive activity ceased at the summit craters of Stromboli following the start of the flank eruption. Persistent effusive activity became concentrated within the 30 December landslide scar on the SDF (Figure 1), resulting in a perched, compound lava flow field that formed during January and the first half of February. This flow field was located in the middle of the SDF and was fed by the 500-m elevation vent. Lava flows were frequently covered by collapse debris from failures at the boundaries of the landslide scar, which also caused widening and upslope migration of the scar. Accumulation of debris on the active lava flows added to the insulating effects of crust formation,
thus promoting tube growth and the establishment of tube-fed ephemeral vents within the flow field [e.g., Calvari and Pinkerton, 1998].

[13] During the first few weeks of this eruptive phase, an apparent down-slope migration of the effusive vents was observed. However, comparison between photos and thermal images collected every day, and often several times a day, allowed us to recognize that the apparent active vent migration was due to the hot interior of the lava flow emerging from below the landslide debris. This usually occurred at marked breaks in slope gradient. Accumulation of a debris fan at the shoreline caused frequent changes in direction of the lava flow front. This resulted in a significant increase in the width and thickness of the lava flow field at the foot of the SDF.

[14] The morphology of the lava flow field was strongly influenced by the steep (∼35°–45°), planar slopes of the SDF. The lava flow field was characterized by ‘a’a surface texture, where the shape, size and features of the flow field, as well as the style of activity, evolved with time and as a function of the effusion rate. During peaks in effusion rate, single (nonbraided) ‘a’a lava flows entered the sea, causing phreatic explosions at the flow front and accumulation of an apron of lava blocks at the foot of the SDF (Figures 5a and 5b). Decreases in effusion rate produced flow bifurcation. Such low effusion rate flows did not reach the sea, but instead spread out, forming a fan in the middle of the SDF (Figure 5c). Further decreases in effusion rate caused tube growth and skylight opening along the flow path (Figure 5d).

[15] Effusive activity from the 500 m vent continued, although at a gradually decreasing rate, until 15 February 2003, when this vent completely shut down. The final shape of this flow field was triangular, with the apex at the 500-m elevation source vent and a ∼ 300 m wide base at the shoreline (Figure 1b). Given that this flow field completely filled the largest of the 30 December 2002 landslide scars, its volume is probably comparable to that of the largest landslide. Therefore, considering a minimum volume of $6 \times 10^6$ m$^3$ of vesiculated lava for this lava flow field, and a duration of 19 days for its emplacement, we obtain a time-averaged eruption rate of approximately $3.7$ m$^3$ s$^{-1}$. Assuming a vesicularity of $\sim 22 \pm 12\%$ [Harris et al., 2000a], this gives a bulk rock eruption rate of $2.9 \pm 0.4$ m$^3$ s$^{-1}$.

2.3. Period of 15 February to 22 July 2003: Lava Flow Field From the 670-m Elevation Vent

[16] On several occasions during January and February the effusion rate from the 500 m vent decreased. This was accompanied by effusion from a second vent that opened at about 670 m elevation (Figures 3 and 4). Slow, short lava flows extended from this vent for periods lasting from a few hours to a few days. Such activity occurred on 30 December 2002, on 9, 23, 28, 29, 30 and 31 January 2003, and on 1, 2, and 3 February 2003. These effusive events built a small lava flow field on the upper east section of the SDF. After this intermittent activity, the 670 m elevation vent became stable on 15 February, when more persistent effusion began to feed a compound lava flow field extending down the eastern side of the SDF (Figure 1b). Structures within this flow field were strongly influenced by the underlying slope, which contained a relatively flat bench around the 600 m elevation (at the NE foot of Crater 1) and a steeper zone (∼40°) below the ∼570 m elevation (Figures 1 and 3).

[17] During the second half of February, effusive activity was relatively weak. Lava flows from the 670 m vent did
not reach the sea, and formed relatively short, single unit flows fed by low effusion rates. Oscillations in the effusion rate were common, with changes occurring every few hours to days. The very low effusion rate was thus, essentially, feeding a series of cooling-limited lava flow units [Guest et al., 1987]. These flows resulted in the construction of a lava shield at the head of the flow field, on which three tumuli grew during February and March. Below the break in slope at the ∼570 m elevation, lava flow front failure was common. This resulted in the detachment of blocks and clinker from the flow front. These rolled down slope, triggering the movement of unconsolidated coarse clasts as well as finer grained material along the slope. Thus, in turn, fed rock falls, grain flows and ash clouds that rose above the SDF.

From the beginning of March, several ephemeral lava streams formed within the flow field, only a few of which entered the sea. In addition, several ephemeral vents opened between the 600 and 450 m elevations, feeding small lava flows that developed into a braided tube system. During March, a fourth tumulus grew in the middle lower portion of the lava flow field, suggesting lengthening of the tube system [e.g., Calvari and Pinkerton, 1998] and indicating tube-fed inflation in the medial section of the flow field.

After mid-March no single flows extended to the sea. Flows continued to form tubes in the upper portion of the flow field, with ‘a’a breakouts extending just a few tens of meters from tube exits. Lava flows from such breakouts often coalesced and braided again when encountering small obstacles, where superimposition of numerous, short lava flows caused a progressive thickening on the upper lava flow field. This added to the construction of the distal shield on the ∼600 m elevation flat zone.

In summary, during March, we observed two effusive patterns: (1) the formation of numerous short lava flows that were typically fed by 1–2 main (master) streams at low effusion rates; such periods of reduced effusion rates favored lava tube formation; (2) the development of a single, apparently higher effusion rate flows, that typically extended for greater distances than the lower effusion rate flows; only these, more organized flows, would be capable of extending to the sea.

2.4. The 5 April 2003 Paroxysm

While lava was still erupting from the 670-m elevation vent on the upper SDF, on 5 April the summit craters of the volcano, inactive since the start of the flank eruption, gave rise to an extremely powerful paroxysmal event. This event was the strongest recorded at Stromboli after the 1930 paroxysm [Rittmann, 1931; Barberi et al., 1993], and is described in detail by S. Calvari et al. (The 5 April 2003 vulcanian paroxysmal explosion at Stromboli volcano (Italy) from field observations and thermal data, submitted.
Journal of Volcanology and Geothermal Research, 2004, hereinafter referred to as Calvari et al., submitted manuscript, 2004). It is worth noting that the paroxysm followed a day of heavy, continuous rain, which probably saturated the thick carpet of debris that was obstructing the summit craters. We were fortunate that the paroxysm occurred while two of us (SC and LL) were flying near the summit craters during a routine thermal survey. This allowed us to collect both photos and thermal images before, during and after the paroxysm (Figures 6a and 6b; Calvari et al., submitted manuscript, 2004).

A few minutes before the paroxysm, lava was flowing down the SDF from three vents. A diffuse gas plume was emanating from the summit craters, which suddenly became thicker and denser, apparently due to an increase in the emission of steam, and its apparent temperature increased from 10°C to ~17°C. Next, a small quantity of reddish (probably lithic) ash issued from Crater 1, bringing the apparent temperature to ~40°C. Seconds later a darker plume of juvenile material was emitted from Crater 1. This had an apparent temperature of up to 150°C. Emission of juvenile material was accompanied by a cauliflower-shaped jet that rapidly rose above the crater. About 2–3 s later Crater 3 emitted a jet of juvenile material with a minimum temperature of 240°C. The two jets (from Crater’s 1 and 3) then coalesced, widening and rapidly spreading upward. An extremely powerful blast accompanied the merging of the two jets, pushing the helicopter away from the crater and increasing its velocity by ~30 knots. Oxygen isotopic ratios obtained from freshly ejected pyroclasts suggested that a shallow aquifer was the source of water in the clasts (F. Sortino, personal communication, 2004).

Observations after the 5 April paroxysm revealed that the lava flow field from the ~670 m vent had been completely covered by a carpet of light brown pyroclastics (golden pumice) that was ejected from Crater 1 during the initial phase of the event. The volume of ejecta (vesiculated material) was ~140,000 m³ [Andronico and Coltelli, 2003] which, considering a typical vesicularity of 40% (R. A. Corsaro et al., Attivita di monitoraggio petrologico a Stromboli nel 2003, INGV-CT internal report, 2004, available at http://www.ct.ingv.it/Ufvg/, 2004), gives a dense rock equivalent (DRE) of 84,000 m³. This volume falls in the range for typical paroxysmal eruptions at Stromboli [Bertagnini et al., 2003].

Alternating pulses of black and red ash rose, mainly from Crater 3, for about 30 minutes after the main blast. The upper part of the volcano above the 700 m elevation was completely covered by a carpet of pyroclastics (golden pumice) that was ejected from Crater 1 during the initial phase of the event. The maximum thickness of this deposit, estimated on the basis of comparison between helicopter photos recorded immediately before and after the explosion, was ~10 m. The temperature on the surface of the pyroclastic deposit, measured ~1 minute after emplacement, was between 250°C and 300°C. A thick, diffuse steam cloud rose from the debris carpet, probably caused by vaporization of the wet pyroclastic deposit above the hot lava flows (Figure 6b). Oxygen isotopic ratios obtained from freshly ejected pyroclasts suggested that a shallow aquifer was the source of water in the clasts (F. Sortino, personal communication, 2004).
Within a few minutes of the paroxysm ending, lava flows were observed as still active on the SDF at the ~600 m elevation, emerging through the debris carpet, along a line marking the break in slope within the upper flow field.

### 2.5. Post 5 April Effusive Activity

Another strong explosion was recorded on 10 April by the INGV seismic network. This event corresponded with a significant increase in temperature on the floor of the summit craters (Figure 7) and an increase in the effusion rate (compare with Figure 15). After the 5 April explosion, some high-pressure degassing was heard from the upper flow field. This suggested that gas-rich magma was being erupted. This inference is also consistent with sulphur deposits that formed on the upper lava flow field and, after 18 April, with the development of spattering hornitos.

Following the 5 April paroxysm, lava effusion continued at a generally low rate, forming numerous, short lava flows. Clusters of ephemeral vents developed at 550 and 590 m elevation and were fed by a number of tumuli that developed approximately along the line of the NE trending fracture from the foot of Crater 1. One hornito was gradually built from a source of particularly strong degassing along this line. The lava pile thus continued to thicken in the upper lava flow field, due to both endogenous (inflation) and exogenous (lava flow) activity. In addition a fracture developed at the break in slope, probably as a consequence of the increasing weight of the lava pile and the steep slope onto which it was extending.

In early May a second hornito grew beside the first one at ~650 m elevation. The lava flow field became increasingly complex, with frequent formation of ephemeral, tube-fed, vents that were active for just one to two days. On 28 May a significant increase in spattering was observed at the main hornito. In the following days several flows departed from this hornito, in concomitance with a reduction of spattering. However, on 31 May the effusive activity, sustained from the hornito-based vents, was accompanied by almost continuous spattering from these vents. At the same time a number of ephemeral vents opened at the hornito foot producing radial lava flows that further contributed to the vertical growth of the upper flow field lava shield. The two hornitos had, by that time, attained an estimated height of ~20 m.

At this time, effusion rates were low but variable. Thus flow lengths were also relatively short, but also variable [Harris et al., 2005]. On 1 June individual lava flows, fed at a relatively low effusion rate, extended no more than 300 m down the SDF. A decline in effusion rate was noted during 6–7 June, which resulted in a reduction in flow length to 150 m. At this point, spattering at the main hornito ceased, but emission of bluish gases persisted. This trend continued until 15 June, when a tube-fed lava flow emerged from ephemeral vents located between 500 and 590 m elevation. This represented the beginning of a final, waning phase such that after 22 June only very short flows were active from ephemeral vents in the upper lava flow field and along the main lava tube. At the same time degassing from the hornitos declined. An up-flow-field migration of effusive vents was accompanied by the limited effusion of short-lived lava flows such that, after 10 July, lava flows had become confined to the upper lava field within the shield that formed on the proximal bench.

A total cessation of effusive activity occurred between 21 and 22 July. The total volume of vesiculated lava erupted during this phase of the eruption was estimated at ~7 × 10^6 m^3. Considering an emplacement time of 156 days, this gives an eruption rate of ~0.5 m^3 s^-1. This equates to dense rock values of 5.5 ± 0.8 × 10^6 m^3 and 0.4 ± 0.1 m^3 s^-1.

### 2.6. March to September 2003: Explosive Activity Resumes at the Summit Craters

Explosive activity from the summit craters ceased on 28 December 2002, in coincidence with the start of the flank eruption. At this time, withdrawal of magma from shallow levels caused the crater floors to collapse, transforming the previously three-crater system into an elongate trench,
within which the locations of the three craters could still be determined.

[11] Temperatures recorded on the floor of the trench showed a significant decline after 28 December. However, the shift in effusive activity from the 500 m vent to the 670 m vent on 15 February 2003 was accompanied by a steady increase in the floor temperature measured for all three craters. Throughout March 2003 the first signs of fresh scoria emissions were observed, along with ash emissions from Crater 1. A juvenile component within the ash emissions also became evident after 27 March (L. Miraglia et al., Caratterizzazione delle ceneri emesse dallo Stromboli nella seconda metà di Marzo 2003, INGV-CT internal report, available at http://www.ct.ingv.it, 2003).

[12] The 5 April event caused further excavation of the crater trench. After this excavation the summit vents appeared to be effectively plugged, with ash explosions occurring approximately once per week. This frequency increased with time, concentrating mostly at Crater 3 until early June, when activity also began to intensify at Crater 1. Around the 20 June a resumption of strombolian explosions was observed; activity became focused at one main vent within Crater 1 and three vents inside Crater 3. Over the following months a steady increase in activity produced syncrater scoria cones, whose rate of construction reflected the accelerating trend in strombolian event frequency. Explosive activity significantly increased during mid-September, and the rate of construction of the scoria cones increased both within Craters 1 and 3, such that the tricrater morphology that had existed prior to the start of the effusive eruption became reestablished.

3. Thermal Surveys

[33] Helicopter-borne thermal surveys of the active flow field and summit craters proved extremely useful during this flank eruption, allowing us to better constrain the chronology given above and to observe a number of complex volcanic phenomena never described before. Additionally, we were able to quantify a number of eruptive processes using the thermal images collected during the eruption.

[34] We used a Forward Looking Infrared Radiometer (FLIR) (model TM695) to collect thermal images. The FLIR instrument has been used intermittently at Stromboli since 2001 (J. Dehn et al., Handheld infrared imaging of strombolian eruptions, submitted to Bulletin of Volcanology, 2004), and was employed during June 2003 to calculate daily effusion rates [Harris et al., 2005]. The FLIR consists of an uncooled microbolometer that detects emitted radiation in the 7.5 to 13 μm wave band. The instrument is capable of acquiring a 320 × 240 pixel image every 2 seconds, and the 24° × 18° field of view equates to a 0.075° × 0.12° field of view per pixel. Over a typical 1–2 km long line-of-sight, this equates to pixel diameters of 1.3–2.6 m. Internal calibration, together with an atmospheric correction that uses input parameters such as distance to target, ambient temperature, humidity and emissivity, allows the FLIR to calculate realistic digital source temperatures. Ambient temperature and air humidity are measured every time at the start of the thermal survey at appropriate elevations, emissivity for lava flows is taken 0.95, and these values inserted in the thermal camera analysis software before recording of thermal images starts. Precision of the instrument is ±2%, and thermal sensitivity is <0.08°C at 30°C.

[35] The FLIR 695 is equipped with an extra lens to reach the maximum temperature of 1500°C, and can record data in one of three temperature ranges: −40°C to 120°C, 0°C to 500°C, and 350°C to 1500°C. These ranges actually have a greater overlap than indicated by the FLIR manufacturer. We thus find that the low temperature range is capable of recording temperatures of up to 240°C, and the middle range can extend up to 850°C, although some loss in accuracy might arise outside the range stated by the manufacturer. In collecting images, it is thus important to select a temperature range appropriate to the feature being targeted, as well as the line-of-sight distance. A careful balance is necessary to allow detection of low temperature anomalies while avoiding saturation over higher temperature features. Effective pixel size increases with distance, and the resulting increase of temperature heterogeneity within each pixel causes a decrease in pixel-integrated temperatures over greater line-of-sight distances. Over 1–2 km, we generally used the low-temperature range for detection of low-temperature features such as fractures and ground cracks, taking care to avoid solar reflection and direct sunlight. Solar radiation can cause an increase in the temperature of ambient surfaces by several tens of degrees Celsius, thus swamping subtle thermal anomalies [Calvari and Pinkerton, 2004]. Thus, early morning flights were preferred to avoid the peak of solar heating. The middle temperature range was generally used for observing explosive and effusive activity, and the high range was used only for imaging particularly active lava flows at close range.

[36] While the radiances measured with the thermal camera are accurate to ~2% (FLIR), translating quantitative radiance to quantitative temperature is not without challenges. There are three main sources of uncertainty; the emissivity of the radiation source, topography/viewing geometry and atmospheric absorption (from gas and aerosols). Basaltic lavas have a high and relatively stable emissivity, 0.95–0.98 [Buongiorno et al., 2002]. The FLIR images used for quantitative measurements were obtained using specially selected images that have direct, overhead or orthogonal views of the target thermal anomalies. We can therefore estimate a 5% error due to geometrical effects. Finally, the effect of gas and aerosol absorption on broadband FLIR imagery is potentially very important. Recent work (G. Sawyer and M. R. Burton, Atmospheric and volcanic absorption effects on handheld thermal infrared imagery, manuscript in preparation, 2005) demonstrates that for path lengths of less than 500 m the effect of infrared absorption from volcanic SO2 can be significant for FLIR data, while water vapor has a lesser effect. The lava flows within SDF have already lost most of their SO2 through prior degassing, and therefore cloud-free images can be considered relatively unaffected by volcanic absorptions. We estimate an error of 3% for the effect of water vapor in these images. Data collected looking into the craters themselves are much more strongly affected, because of the high concentrations of SO2 present in the persistent gas emissions from the volcano. Errors on these images are therefore higher, between 10–20%. Overall
therefore, we estimate an error of 6% for temperatures derived from lava flow images and 11–20% for temperatures from crater images.

[37] FLIR data were collected mainly as part of regular helicopter-borne surveys that occurred periodically on at least a daily basis between 28 December 2002 and 15 September 2003. After June 2003, the same flight path was followed on each day where observation points were fixed using a handheld GPS. Repetition in flight paths allowed a good precision in distance estimates; a detailed description of the data collection approach is available from A. J. L. Harris et al. (Calculation of lava effusion rates at Stromboli Volcano using FLIR data, Italian Civil Protection internal report, Centro Operativo Avanzato di Stromboli (San Vincenzo, Italy), June 2003). For measurements prior to June, a distance correction (essential to obtain pixel size) was applied on the basis of the observed dimensions of the summit craters and distances between stable, ground control points. These control points were located during field surveys with a Leica range finder and/or obtained from topographic maps and preexisting digital elevation models.

3.1. Summit Crater Floor Temperatures

[38] The maximum temperature obtained from images of the floor of the summit craters showed interesting changes with time. A total of 376 thermal images recorded during the eruption permitted good views of the crater floor, allowing us to plot floor temperature variation with time (Figure 7).

[39] Prior to the eruption, during January and March 2002, temperatures were low and mostly below 300°C. Temperatures increased in the months prior to the flank eruption. The onset of effusive activity on 28 December 2002 caused a sudden decrease in the maximum temperature recorded at the summit craters, to below 300°C. This temperature drop is consistent with collapse of the crater floor, termination of explosive activity, and reduction of the magma level within the conduit. The temperature drop would have been a logical consequence of tapping of the central conduit by the flank vents.

[40] In the months after 15 February 2003, the temperature at all three of the summit craters gradually increased, showing a particularly marked increase at Crater 1 during June. During this phase of the eruption the lava flow field from the 500 m vent closed down, and the 670 m vent became stable, suggesting that the closure of the lower vent caused an increase in magma level within the feeder conduit at this time. Floor temperatures peaked during July, following the termination of lava effusion onto the SDF. This also correlated with a marked increase in explosive activity.

[41] Changes in the peak temperature recorded for the crater floors, although uncorrected for field of view [Harris et al., 1997a; Dehn et al., 2002] and recorded at different distances, thus appear to be a function of both magma level within the feeder conduit and activity within the summit craters. Increasing temperatures at the summit craters occurred when effusion at the lava flow field decreased, suggesting a shallow connection between the lateral effusive vents and the main conduit feeding the summit craters and a constant magma supply rate, such that decreased tapping of the main magma column caused the free-surface level in the conduit to increase [Harris et al., 2005].

3.2. Temperatures Obtained for the 28 December 2002 Hot Avalanche

[42] The hot avalanche emplaced on the Spiaggia dei Gabbiani on 28 December (Figure 3) was visible in thermal images acquired on 29 December, when it showed a maximum temperature of 368°C (Figure 8a). The obvious expected change with time after emplacement would be a gradual decline due to cooling of the exposed surface. However, after a period of declining temperature, peaks in temperature were observed on 2, 8, and 9 January, as well as on 6 and 14 February. These points are circled 1 to 6 on Figure 8a. Of these, one apparent increase in temperature (the two measurements marked by circles 1 and 2, Figure 8a) was due to differences in the line-of-sight distance, being lower for the first image. However, all other temperature variations are apparently real. It is worth noting that front of the hot avalanche emplaced along the beach (Figure 8b) was repeatedly collapsing due to erosion by wave action. This caused regular exposure of the inner, hotter, portion of the avalanche, as clearly visible by comparison between Figures 8c and 8d. Due probably to the large amount of fine-grained matrix comprising the deposit, which decreases the amount of open spaces and increases cohesion, the inner temperature remained above 300°C for at least 45 days. After this date the deposit became buried by debris from the lava flow fronts.

[43] The range of maximum temperatures that we found for this deposit (250° to 350°C, Figure 8a) fits within the range of temperature of the deposits obtained with the thermal remanent magnetization technique on different types of pyroclastic density currents from the A.D. 79 eruption of Vesuvius [Cioni et al., 2004]. It is worth noting that the measurements from Cioni et al. [2004] record the temperature of the deposit and not the emplacement temperature. Our values are also within the range of temperatures derived for pyroclastic flows at Montserrat, measured within 72 hours of deposit emplacement [Calder et al., 1999]. However, they fall only within the lower end of the emplacement temperature range obtained for pyroclastic flow deposits spread at Mt. St. Helens during 18 May–17 October 1980 [Banks and Hoblitt, 1981].

[44] Peaks 3 to 6 (Figure 8a) relate to secondary, reworked but still hot deposits, emplaced local collapse of the primary deposit, rather than by primary volcanic processes, 1.5 months after emplacement. Such exposures made this an opportunity to track the cooling rate of both the surface and the interior of a pyroclastic deposit in the field with direct measurements, Banks and Hoblitt [1981] being an excellent, but rare, second example. Although the exact timing of a front collapse on 14 February is not known, as is not known the amount of rainfall, the maximum temperature recorded on that day indicates a maximum cooling rate of 0.66°C/day for the inner portion of the deposit during the first 45 days of its emplacement. Conversely, the cooling rate for the surface is significantly higher at 37.7°C/day.

3.3. Detection of Cracks and Fractures

[45] Helicopter-borne thermal surveys allowed us to recognize the opening and growth of a number of fractures in
which hot gases were transported. These occurred in three main locations: (1) along the SDF, one hour before the 30 December flank failure [Bonaccorso et al., 2003]; (2) around the summit craters; and (3) on the lava flow field fed by the 670 m vent. These steps caused a characteristic kink aspect of the lava flows extending from the 550 m vent on 30 December (Figure 3b). In fact, the lava flows erupted from this vent followed the step-like morphology before spreading down the slope as a straight stream. These lava flows thus enhanced the thermal contrast between the cool SDF surface and the developing cracks (Figure 3b). The cracks were the source of ash emission (Figure 4), where emission was not pulsating as would be the case of pressurized gas emission, but was rather mild and continuous. This sug-

**Figure 8.** (a) Graph showing changes with time of maximum temperature recorded at the exposed surface of the hot avalanche emplaced on the Spiaggia dei Gabbiani on 28 December 2002. (b) Photo of the avalanche on 29 December 2002. (c) Thermal image of the avalanche on 29 December and (d) on 14 February 2003. The green line in Figures 8c and 8d for comparison indicates the same portion of the deposit before and after erosion.
gested that the ash output was caused by rock fragmentation due to friction during rock sliding rather than by explosions accompanied by degassing [Bonaccorso et al., 2003; Andronico et al., Ash characterization (December 30, 2002), INGV-CT internal report, available at http://www.ct.ingv.it/Stromboli2002/Main.htm, 2002]. A lava flow passively filled the opening cracks, pouring out in a well-fed flow immediately after the failure. A thermal survey carried out on the morning of 30 December 2002 revealed propagation of fractures from ~500 m elevation down slope, and extending from the west to the east boundary of the landslide scar.

[47] During January 2003, improved weather conditions allowed the first thermal survey of the crater area to be completed. This showed that the three summit craters had coalesced to form a single, trench-like depression with a NE-SW orientation. The crater zone was surrounded by arcuate fractures open to the north (Figures 9a–9d), suggesting an initial northward (seaward) movement of the upper part of the feeding system. A structural survey carried out a few days later, however, indicated that these fractures were consistent with near-vertical sinking of the summit zone due to magma drainage from the main conduit through the 500 m vent (G. Lanzafame and M. Neri, Eruzione Stromboli 2002–2003: Fratture e cinematismi osservabili lungo la Scia研究生 - Rapporto Preliminare, INGV-CT internal report UFVG2004/012, available at http://www.ct.ingv.it, 2003).

[48] A third set of fractures developed on the upper lava flow field emplaced from the 670 m vent. These fractures were concentrated at the break in slope at the northern edge of the proximal bench and developed during early May and August 2003. Cracks that opened in May resulted from inflation of the upper lava flow field, whereas those opening in August probably reflected the instability of the inactive lava flow field, that was contracting due to cooling and was sliding northward on the steep slopes of the SDF.

3.4. Lava Flow Structures

[49] A number of interesting structures characterized these lava flows, some structures being the result of emplacement on the steep and unconsolidated slopes of the SDF. Lava flow-front failure, landslides, and rock falls were very common during early January, building talus fans at the foot of the SDF and covering the surface of the lava flows with a carpet of debris for most of their length. Debris accumulating on the surface of lava flows aided in insulating the flow core, increasing unit thickness and enhancing tube formation. On 4 January, first-order (tube-fed) ephemeral vents [Calvari and Pinkerton, 1998] developed along the flow field between the 500 and 350 m elevations. Several further ephemeral vents were observed in the following days along the SDF. Initially these vents caused concern because, if primary, they indicated a down slope propagation of the feeder dike. Thanks to the thermal images, we could reconstruct the path of the still active lava flow below the debris and inside lava tubes, revealing that neither dike injection nor dike propagation were occurring after 28 December 2002 along the SDF.
Instead they were the result of vent migration by tube extension.

### 3.4.1. Talus Fans

The initial lava flow field from the 500 m vent was emplaced on the steep collapse scar excavated by the 30 December landslide on the SDF. The slope was constant between the 500 and ~350 m elevations at ~45 degrees. At lower elevations, this steep slope caused ‘a’a flow fronts to become unstable so that they failed to form a coarse, loose debris pile or talus mound at the foot of the slope. Additional material accumulated as a result of minor, but common, landslides and rock falls from the walls of the collapse scar. These continued, steadily widening the landslide scar and causing the headwall to creep up slope, until the end of the eruption. Gradual piling up, and thus up-slope extension, of this talus mound caused the local slope at the base of the SDF to gradually decrease with time [e.g., Borgia et al., 1983; Calvari et al., 1994]. The surface of this distal talus zone was comprised mostly by fine-grained material in the upper part, and coarse blocks across the lower section. Such talus fans, as we describe, influenced subsequent lava flow paths.

### 3.4.2. Excavated Debris Levees

Lava flows extending from the 500 m vent on 2 January 2003 contained lateral levees (Figure 10), unlike any of the levee types defined by Sparks et al. [1976] or Lipman and Banks [1987], which were largely composed of fine-grained ash and scoria from the SDF that was excavated downward and pushed laterally by the lava flows. Excavated debris levees improved channel formation, and were distinguished from normal levees for their lower temperature.

### 3.4.3. Sea Entry

Lava initially entered the sea as a single flow unit contained within initial levees [Sparks et al., 1976]. On 10 January 2003, we observed a moustache-shaped zone of heated water extending offshore (Figures 11a and 11b), similar in form to anomalies observed in airborne thermal data for Kilauea by Realmuto et al. [1992]. In our case, we suggest that accumulation of talus at the foot of the SDF obstructed the lava flow, diverting it in two opposite directions on entering the water (to the right and left of the obstacle) thus leaving a corresponding, bifurcated warm signature on the sea surface (Figure 11b).

Thermal images of the sea entry acquired during late December 2002 thus revealed well-defined zones of warm water, ~300 m wide and extending ~50 m offshore (Figure 11c), where water depth is ~50 m. The temperature contrast of 9°C between the warm, lava heated water and the surrounding zone of cooler, ambient water allows us to estimate the lava volume necessary to obtain this contrast [Harris et al., 1998]. We used water and lava specific heat capacities of 2100 J/kg K and 1225 J/kg K, with the assumption that the lava cools from 1200 to 300 K during water heating and an estimated volume of heated water of $3.5 \times 10^6$ m$^3$ to obtain a cooled lava volume of $2.3 \times 10^4$ m$^3$.

### 3.4.4. Vent Migration

Migration between the 500 and 670 m elevation vents occurred during changes in effusion rate. Whenever output rate from the 500 m vent significantly decreased, the upper vent erupted short-lived lava flows lasting a few hours to a few days. We thus suggest that the upper vent operated as a pressure valve, where decreases in output from
the lower elevation vent caused the system to pressurize so that the magma level within the feeder conduit increased to allow supply to the upper vent.

A third vent opened between these two primary vents, at the 550 m elevation, on 24 January, followed by a small collapse at the 580 m elevation on the SDF during 25 January. This vent erupted lava for just 4 days. These three active vents were either located within the deepest part of the landslide scar, as was the case for the 500 m elevation vent, or along the fractures that were delimiting the step-faults on the upper SDF (as with the 550 and 670 m vents). We thus suggest that these vents represented the passive fingering of magma from the central conduit and NE fissure, rather than an active dike intrusion. In fact, a dike intrusion would have caused a progressive migration of effusive vents that was excluded by our thermal measurements. The clear position of the active vents has been detected each time with the aid of the thermal camera, allowing us to distinguish the hot spot in between the cool debris.

3.4.5. Lava Flow Form

The form of the lava flow field fed by the 500 m vent was related to effusion rate. High effusion rates resulted in lava flow to the coast as a single unit flow (Figure 5b). However, moderate effusion rates fed multiple, cooling-limited flows [Guest et al., 1987] that were active across the middle section of the SDF (Figure 5c). Low effusion rates resulted in tube growth, with short lava flows emerging from vents or skylights along the tubes (Figure 5d). Tube extension during such phases resulted in down slope migration of first-order (tube-fed) ephemeral vents.

Emplacement and structure of the upper lava flow field generated by the 670 m vent was strongly influenced by the underlying slope. For this purpose, we distinguish four characteristic zones: (1) a proximal, low gradient zone extending between the 670 and 600 m elevations, where slope was between 0° and 20°; (2) a medial, intermediate gradient zone extending between the 600 and 580 m elevations with slopes between 20° and 30°; (3) a distal, high gradient zone which included most of the SDF below the 580 m elevation where slope was greater than 30°; and (4) a low gradient toe below the 30 m elevation, corresponding to the Spiaggia dei Gabbiani beach (Figure 1), where slope was between 5° and 15°.

Tubes, ephemeral vents and tumuli characterized the proximal and medial zones, and were detected with the thermal measurements. Sheet flows were observed only during the early stages of the flow field emplacement, and then only in the proximal zone. Auto-brecciated lava flows and skylights were present only in the distal zone, where the high slopes favored ‘a’a lava flows formation. Well-developed lava channels were also characteristic of the medial, distal and toe zones. Levees differed in the medial and toe zones, being of the initial type [Sparks et al., 1976] in the toe zone, and of the excavated type in the medial zone. Ground morphology changed following the 5 April event, where accumulation of pyroclastic material at the foot of Crater 1 caused the

Figure 11. (a) Photo and (b) thermal image of the sea surface taken on 13 February 2003. It shows (Figure 11b) the lava flow entering the sea and forming “moustaches” due to lava blocks entering the sea being diverted by an obstacle. (c) Thermal image of a circular area with a temperature contrast of 7°C in respect to the surroundings is caused by hot lava entering the sea, 3 January 2003.
3.4.6. Development of Tubes and Tumuli

Short lava flows were emplaced several times during short-lived effusive events in January and February from the 670 m vent. These piled up close to the vent, forming a focal tumulus [Duncan et al., 2004] that became particularly well established following 29 January (tumulus A, or TA, Figure 12). At first these short flows remained confined to the distal, flat zone at the base of Crater 1 above the 590 m elevation. Low effusion periods lasting several days resulted in the formation of a system of lava tubes that developed on the upper flat zone such that the source of short flows could move further from the vent region (TA). On the upper flat zone, the master tube typically followed the NE trending line of the eruptive fissure (Figure 12). Several ephemeral vents opened along this path during January, giving rise to both a complex tube system and a compound lava flow field.

Following 15 February, when effusion from the 670 m vent became persistent, the position of the tube system became marked by a series of satellite tumuli [Duncan et al., 2004]. On 18 February, due to tube drainage during a decrease in effusion rate, several skylights opened above the master tube. At the lower end of the master tube (at the ~620 m elevation) a second tumulus (TB, Figure 12) formed due to the superimposition of several short lava flows. A third tumulus, tumulus C (TC, Figure 12), developed from 22 February onward at the break in slope at the ~600 m elevation (Figure 12).

At this time, a flow extended between tumuli TA, TB, and TC (Figure 12) where it was diverted north by the east wall of the SDF. This flow built excavated debris levees in its upper section and levees across its distal section. During this phase of the eruption, several third-order ephemeral vents also opened on the flanks and at the bases of the tumuli, with new vents opening during periods of increased effusion rate. These changes also resulted in inflation and deflation of the tumuli, with vents opening during inflation and shutting down during deflation.

A fourth tumulus (TD, Figure 12) grew after 17 March following roofing over of a lava flow erupted from tumulus B. This tumulus formed at the break in slope bounding the northern edge of the distal, flat bench, where lava could pond, inflate and pile up around tube-fed ephemeral vents. After 19 March, a significant change in the morphology of the flat, distal zone of the lava flow field was observed, as a consequence of numerous short lava flows extending from the four tumuli and filling the depressions of the lava flow field. This represents the mature stage of a lava flow field growth [Kilburn and Lopes, 1988, 1991], where no significant lengthening takes place, and further lava flow field development involves thickening of the flow field. Therefore, the complex of tumuli and flows comprised a shield-like structure at the head of the flow field.

A decrease in effusion rate following 20 March probably caused the deactivation of TC. In addition, lava drained through the tube system connecting TB and TD, causing a third-order ephemeral vent to open at the base of TD. A further, significant decrease in effusion rate, number of active vents and number of lava flows was observed between 31 March and 1 April. This was accompanied by a deflation of the upper flow field. However, this trend was reversed between 2 and 3 April when a significant inflation of the proximal lava flow field was observed. This was accompanied by an increase in temperature at TA, the emergence of a number of sheet flows from the base of TB, reactivation of TC, with a number of small lava flows pouring out from its base, and activation of a number of vents surrounding TD that fed lava flow onto the upper SDF.

The 5 April paroxysmal event covered the upper lava flow field with a carpet of lithic blocks and highly vesiculated pyroclastic material. This had a minimum thickness of ~10 m and effectively created a new, level and smooth surface across the upper lava flow field. New structures formed on top of this deposit immediately after its emplacement, where lava flows emerged from below the pyroclastic carpet a few hours after the paroxysm. These erupted from three vents that opened at the break in slope, which had shifted about 100 m down slope due to debris accumulated by the 5 April event. Only TA projected above the 5 April debris deposit, the other tumuli being covered by products of the explosive event. Emplacement of this deposit caused deactivation of the deeper tubes connecting TB with TC and TD, which were no longer fed after the paroxysm. However, a new series of tubes and tumuli grew above the 5 April deposit on the upper SDF where three new satellite tumuli (T1, T2 and T3; Figure 12) formed after 7 April from the ephemeral vents opening within the 5 April deposit at the break in slope. All of these were fed by TB.

On 7 April a thermal anomaly on the upper part of the SDF suggested a new vent was opening. This was accompanied by inflation and development of radial cracks in the vicinity of TB. On 8 April an effusive vent opened along the line of the eruptive fissure, about 20–30 m up slope from the three effusive vents of 5 April. The new vent
was located on top of TB, indicating that lava was still accumulating at this site and feeding vents down slope. In addition, a thermal anomaly at TA occurred any time that TB was inflating, indicating the presence of a (lava tube) connection between the two. On 22 April the three tumuli (T1, T2 and T3, Figure 12) merged together to form a large structure still fed by TB through a system of tubes, and adding further to the integrated, shield structure at the head of the flow field.

3.4.7. Lava Flow Field Fracturing, Inflation, and Deflation

Fractures developed within the upper flow field between 2 and 6 May 2003 (Figure 13). At the same time, acceleration of the deformation rate of the eastern section of the SDF detected by the LISA-ISPRA-University of Firenze joint research team (P. Farina, personal communication, 2003) prompted a more careful analysis of the flow field thermal images. These images indicated that the fracturing resulted from inflation (endogenous growth) of the upper section of the flow field, which continued until 6 May when two vents opened within the inflated area. The first vent opened in the morning to cause deflation of the eastern section of lava flow field, whereas the second vent opened in the afternoon to drain the central section flow field. The vent openings were accompanied by a reduction in deformation along the eastern SDF (P. Farina, personal communication 2003).

3.4.8. Lava Flow Field and Summit Crater Temperatures: A Linked System

After the end of the lava effusion, hot fractures developed in the proximal lava flow field (Figure 14), surrounding the most elevated portion of the flow field and suggesting northward movement of the shield-like pile on the upper flow field in the unbuttressed, seaward direction.

3.4.8. Lava Flow Field and Summit Crater Temperatures: A Linked System

Figure 13. Cracks opened along the upper lava flow field from the 670 m vent due to inflation and endogenous growth, recorded in May 2003. (a) Photo taken from NNW of the upper flow field on 4 May 2003. (b) Thermal image taken from the north on 4 May 2003, showing the inflated upper flow field and cracks (dashed white line) and active lava flow (yellow). The yellow flow on the right is the same as in the middle of Figure 13a. (c) Thermal image taken from the north in the morning of 6 May, showing the opening of vents causing deflation of the margins of the flow field. (d) Thermal image taken from the north in the afternoon of 6 May, showing the opening of a vent in the middle of the flow field. Active lava flows are yellow, cracks in red.
maximum temperatures recorded on the floor of the summit craters. This graph indicates a relationship and close connection between lava output from effusive vents and maximum temperature at the summit craters. In this graph, increases in crater floor temperature are associated with increases in maximum temperature extracted for flow surfaces. Following Harris et al. [2005], this correlation can be explained by a sequence of events whereby an increase in the magma free surface causes an increase in the crater temperatures. The same increase in the magma free surface results in increased effusion rates. Longer, more vigorous flows with higher channel velocities may be associated with increased crustal foundering and cracking, thus generating the higher flow field temperatures as areas of flow core are more readily exposed. Thus this provides evidence for an oscillating free surface level and effusion rates, and a hydrostatic balance between the two in a system linked to the central conduit [Harris et al., 2005].

The end of lava effusion on 22 July was followed by a marked increase in temperature at the summit craters, with

Figure 14. Photo and thermal image of the upper lava flow field recorded after the end of the eruption on 23 September 2003. View is from the east. The thermal image shows three sets of parallel hot fractures suggesting northward movement of the perched lava flow field.

Figure 15. Comparison between maximum temperature recorded at the three summit craters (CR1, CR2, and CR3) and on the surface of the active lava flow field recorded at Stromboli with the FLIR thermal camera.
strombolian activity becoming more regular. Again, this is good evidence for a shallow connection between the central feeder conduit and the flank vents. In this case, shut down of the flank vents meant that the central column was no longer tapped, allowing an increase in the free surface level and a resumption of persistent explosive activity.

4. Effusion Rates

[70] Effusion rate has been shown to influence maximum lava flow length [Walker, 1973; Pinkerton and Wilson, 1994; Calvari and Pinkerton, 1998; Kilburn, 2000] as well as the shape and extent of lava flow fields [Wadge, 1978; Kilburn and Lopes, 1988, 1991; Wadge et al., 1994; Calvari et al., 1994; Calvari and Pinkerton, 1998]. Ground-based effusion rate measurements are often difficult and limited to those periods when a lava flow field can be safely accessed [Harris et al., 2000a]. They were even more difficult to carry out during the 2002–2003 effusive eruption of Stromboli because flows were emplaced on extremely steep slopes bounded by cliffs. In addition, persistent rock fall and grain flow activity made work on the flow field hazardous. We therefore have employed a more rapid, safe and easy method to calculate lava effusion rates from satellite thermal data [Harris et al., 1997a, 1997b, 1998; Wright et al., 2001].

[71] We used data from the advanced very high resolution radiometer (AVHRR), provided every 6 hours with a 1 km pixel size, which have previously yielded reliable effusion rates for Etna and Stromboli [Harris et al., 1997a, 1997b, 2000a; Harris and Stevenson, 1997]. In addition, we developed a new method to calculate effusion rates using the FLIR data at Stromboli during June 2003. The method is detailed by Harris et al. [2005; internal report, 2003] and summarized here.

[72] Building on the work of Pieri and Baloga [1986], this approach uses the satellite- or FLIR-image-derived heat flux ($Q_{\text{tot}}$) to estimate effusion rate (Er), using

$$E_r = Q_{\text{tot}}/\rho [C_p \Delta T + \varphi c_L]$$

(1)

Here $\rho$ and $C_p$ are lava density and specific heat capacity (2039 kg m$^{-3}$ and 955 J kg$^{-1}$ K$^{-1}$), $\Delta T$ is lava cooling (the difference between the lava temperature at the effusive vent and the temperature during its emplacement) (150–200 K), $\varphi$ is postemergence crystallization (45%), and $c_L$ is latent heat of crystallization (3.5 x 10$^8$ J kg$^{-1}$). Heat flux is calculated using the image data by estimating the radiative ($Q_{\text{rad}}$) and convective ($Q_{\text{conv}}$) heat losses ($Q_{\text{tot}} = Q_{\text{rad}} + Q_{\text{conv}}$) from

$$Q_{\text{rad}} = A \sigma T^4$$

(2)

$$Q_{\text{conv}} = A h_c (T - T_{\text{amb}})$$

(3)

in which $\sigma$, $\epsilon$, and $T_{\text{amb}}$ are the Stefan-Boltzmann constant (5.67 x 10$^{-8}$ W m$^{-2}$ K$^{-4}$), emissivity (0.95), convective heat transfer coefficient (10–100 W m$^{-2}$ K$^{-1}$) [Harris et al., 2005], and ambient air temperature (293 K), respectively. Parameters A and T are the image-derived lava flow area and surface temperature, respectively.

[73] As argued by Wright et al. [2001], because area and temperature become our main variables, we are simply calculating a reasonable slope and intercept in a relationship that relates lava effusion rate to these two variables ($E_r = mQ_{\text{tot}} - c$). For AVHRR, a range of plausible flow surface temperatures ($T = 373$–873 K) are assumed to obtain flow area [Harris et al., 1997a, 1997b]. Thus this relationship reduces further to

$$E_{r(\text{max})} = 166 A_{873}$$

(4)

$$E_{r(\text{min})} = 65 A_{873}$$

(5)

in which $A_{873}$ is the lava area (m$^2$) obtained assuming $T = 873$ K and $E_r$ is emission rate in m$^3$/s. For lava areas obtained using $T = 373$ K, the $m$ values are 6.2 and 2.5, respectively. For the FLIR data, because both temperature and area are variable [Harris et al., 2005], we obtain

$$E_{r(\text{max})} = 2 \times 10^{-9} Q_{\text{tot}}(\text{min}) - 0.016$$

(6a)

$$E_{r(\text{min})} = 1 \times 10^{-9} Q_{\text{tot}}(\text{max}) - 0.0004$$

(6b)

[74] Here we have applied this method to the entire set of thermal images collected during the eruption to provide the effusion rate time series given in Figure 16. A total of 25 AVHRR images and 64 FLIR image mosaics were suitable (i.e., cloud-free, suitable viewing geometry, etc.) for effusion rate calculations and are examined in detail by L. Lodato et al. (The morphology and evolution of the Stromboli 2002–2003 lava flow field: An example of basaltic flow field emplaced on a steep slope, submitted to Bulletin of Volcanology, 2004, hereinafter referred to as Lodato et al., submitted manuscript, 2004).

[75] The error in effusion rate estimates is typically ±40%, but is comparable to the error of field-based effusion rate measurements [e.g., Calvari et al., 2002; Harris and Neri, 2002; Sutton et al., 2003]. Field measurements of effusion rate were seldom performed, mostly due to accessibility difficulties. However, on 18 February a lava channel 2 m wide and ~1 m deep showed a surface velocity of 0.1 m s$^{-1}$, giving a mean output rate of 0.2 m$^3$/s (M. Coltelli unpublished data, 2003). This compares with AVHRR-derived effusion rates of 0.1 and 0.4 m$^3$/s on 16 and 21 February, respectively. A second field-based estimate of ~0.6 m$^3$/s was obtained on 31 May [Harris et al., 2005], and compares with a FLIR-derived value of ~0.6 m$^3$/s. A comparison between AVHRR and FLIR-derived effusion rate data for Stromboli is shown in Figure 16c. We found that fits between two same-day data points in the two data sets are generally poor. For example, an AVHRR peak value of 1.0 m$^3$/s was obtained on 7 January, whereas the highest FLIR-derived value of 1.5 m$^3$/s was obtained on 10 January. Some of this difference may be due to the different timing of the data acquisitions in respect to real variations in effusion. At this flow field, as at Etna and Mauna Loa [Lipman and Banks, 1987; Harris et al., 2000b; J. E. Bailey et al., The changing morphology of an open lava channel on Mt. Etna, submitted to Bulletin of Volcanology, 2004], we observed that effusion rate varied by a factor of ~4 over <24 hour periods where several short pulses in lava effusion can be distinguished in Figure 16a. As detailed by Harris et al. [2005] and Lodato et al. (submitted manuscript, 2004), these pulses corresponded to
the emplacement of longer flows, a relationship consistent with that observed for Hawaiian flows by Pieri and Baloga [1986]. Differences in effusion rates calculated by multiple FLIR images collected within a single day attests to such short-term variation. Variability, for example, was apparent on 10 January (1.3 to 1.5 m$^3$ s$^{-1}$), 13 February (0.5 to 0.7 m$^3$ s$^{-1}$), 13 March (0.02 to 0.1 m$^3$ s$^{-1}$), 9 April (0.2 to 0.4 m$^3$ s$^{-1}$), and 1 June (0.1 to 0.8 m$^3$ s$^{-1}$).

Such differences make point-to-point data comparisons problematic. Thus, to better assess the agreement between the AVHRR and FLIR derived effusion rates, we smoothed our data in an attempt to reduce the short term variability effects. The results from the smoothed data show a good correlation between the two data sets (Figure 16c; $E_r (AVHRR) = 0.9E_r (FLIR) + 0.1$, $R^2 = 0.8$), and indicate that the short term variability truly is a day-scale effect that overprints any longer (weeklong to monthlong) trends.

5. Discussion

[77] The 2002–2003 flank eruption of Stromboli was unique in that a huge amount of ground-based, high spatial and temporal resolution thermal data were collected before, during and after the eruptive event. A total of $\sim$100,000 FLIR images were analyzed in order to detect the features described in this paper, and in conjunction with visual observations, as well as data on seismicity, ground deformation and gas geochemistry collected routinely during the eruption, allowed us to reconstruct a detailed chronology of the eruptive events. In particular, regular thermal surveying of the summit craters and lava flows allowed us to (1) distinguish active vents, lava flows, cracks, hot debris deposits and sea entry thermal anomalies; (2) track the development of the structures on the lava flow field such as channels, tubes, tumuli and hornitos; (3) calculate effusion rates and relate them to temperatures on the floor of the summit craters and the movement of magma within the central conduit; and (4) constrain the processes responsible of the observed changes in the eruptive activity.

5.1. Feeding System

[78] It appears that initial lava effusion was not triggered by dike injection, because neither earthquakes, seismic swarms, or ground deformations that would have been associated with such an event were recorded by the INGV-CT monitoring network [Bonaccorso et al., 2003]. In addition, the NE flank of Crater 1 that failed on the first day of the eruption, was the same sector that was breached during the previous two flank eruptions (1985–86 and 1975). We thus suggest that overpressure due to increased magma levels in the central magma column was sufficient to force a failure of the crater flank, thus triggering the onset of lava flow activity.

[79] The emplacement of two lava flow fields within two different locations was rather unusual. Previously, in 1985–86 and 1975 effusive vents opened at the base of the summit cone, at $\sim$700 m a.s.l [e.g., Capaldi et al., 1978]. During the 2002–2003 eruption, the primary eruptive vents were established in two distinct zones: both very close to the central feeder conduit (Figure 17). However, the first set (the 500 and 550 m elevation vents) developed within fractures that opened within the 30 December and 24 January landslide scars, and well below the level of the active summit craters. The second set developed in a more usual location: along the eruptive fissure that extended NE from the base of Crater 1. Here, this latter set of vents developed along an identical line and elevation to the 1985–86 vents, where De Fino et al. [1988] describes a stable vent at $\sim$670 m elevation during the 1985–86 eruption, with secondary, tube-fed flows extending down to 580 m. Thus the same line of weakness was utilized by magma moving laterally from the central conduit during the 1985–86 and 2002–2003 eruptions.

Figure 16. Effusion rates calculated using (a) FLIR and (b) AVHRR data. In both cases a 3-point moving average has been fitted. (c) Comparison between smoothed AVHRR and FLIR data. Smoothing has been achieved by grouping the two data sets into consecutive 10-day blocks, and the mean of each 10-day grouping is given.
5.2. Flow Morphology

Lava flow morphology changed significantly during the eruption. The first two high effusion rate flows emplaced on 28 December 2002 had a sheet-like morphology, which is typical of initial, high effusion rate, fast moving lava flows [Bertagnini et al., 1990] or early stage Hawaiian flows [Lipman and Banks, 1987], where lateral levees and channels have no time to develop [Hulme, 1974].

The third flow erupted a few hours later on the same day (28 December) it had a lower effusion rate and was characterized by a narrow lava channel across the upper, steeper part of the SDF. This fed a wider distal zone where the lava flow mixed with debris and spread onto the gentle slope of the Spiaggia dei Gabbiani (Figure 3). Here, the flow morphology was consistent with a lower effusion rate or a longer emplacement time. This flow effectively developed the more mature down-flow morphological sequence, characterized by a stable channel feeding a zone of dispersed flow, defined by Lipman and Banks [1987].

Different structures observed in the lava flows are probably related to ground morphology and effusion rate. Regular, but steep, slopes favored flow front brecciation, failure and formation of debris fans at the base of the SDF. Conversely, flatter surfaces and lower effusion rates favored establishment of tubes, tumuli and ephemeral vents. Lava tube development was particularly common in the second lava flow field that developed mostly on the relatively flat bench below Crater 1. Here ephemeral vent evolution indicated tube growth during stages of increased effusion rate, and up-slope vent migration and tumuli deactivation during decreasing effusion rate.

5.3. Vent Migration

Thermal images revealed that an apparent migration in the primary vent location was actually a result of tube extension, enhanced by the insulating effect of debris accumulating on top of the flow by continuous landslides from the margins of the 30 December landslide scar. This tube growth process served to extend the point of surface emission to secondary vents linked to the primary vents by tube systems. Real vent migration, or popping up of successive primary vents along a linear trend, did, however, occur. Daylong transitions in the primary focus of effusive activity between the 550 m and 670 m elevation vents may indicate daylong fluctuations on the magma level within the central conduit. One possible mode of achieving such fluctuations is internal: short-term changes in the supply rate from the deep source. However, an external mechanism is also possible where a declining effusion rate at the lower elevation vent and consequential backup of magma in the central conduit causes the free surface to rise and activation of the higher vent. Shut down of the lower elevation vents, and establishment of the persistent activity at the higher (670 m) vent following February 15 may be attributed to these processes – although in this case the rise in magma level was permanent, i.e., it did not drop to a lower level following February 15.

The alternation in activity between vents that ranged in elevation by over 170 m might thus have been controlled by their proximity to the central conduit, by decompression of the surrounding rock due to failure and crack opening, and to magma fingering into the open fractures. In fact, the 500 m vent opened after the 30 December landslide at the lowest point within the resulting scar. The 550 m vent opened after a small collapse that removed a 10 m thickness of rock, whereas the 670 m vent opened along the eruptive fissure crossing the NE flank of Crater 1. We thus suggest that this migration in primary vent location reflected changes in magma level within the central conduit, as shown in Figure 17a. Effusion rates and magma levels from this linked conduit system were therefore closely linked.

Self-sealing of primary vents may have been triggered by cooling and crystallization during declining
effusion rates, reducing the efficiency of the conduit as suggested at Kilauea [Kauahikaua et al., 1996]. Sealing of the 500 m conduit after February 15, as well as final shut down of the 670 m vent, thus probably resulted from a decrease in effusion rate, resulting in cooling of the magma flowing from the central conduit and a decrease in efficiency of the lateral conduits, prior to reactivation at higher elevation. Such a process could display a feedback, encouraging rapid shutdown once the process has begun.

A series of active vents that opened on the second flow field were all fed by the primary 670 m vent through a distributary system of lava tubes (Figure 17b). This primary vent became established beneath tumulus A (or TA), and fed lava onto the SDF through a master tube that also connected it to a series of secondary tumuli. Thus a series of minor secondary (or tube-related) vents, of variable location, resulted in the emplacement of numerous short flows to build a complete flow field and distal shield at the head of the flow field.

5.4. Eruption Models

Here two possible models representing the behavior of the feeder conduit are sketched in Figure 17. Model A shows conditions existing up to 15 February, when large changes in magma level within the conduit caused opening and closure of primary vents between the 500 and 670 m elevation vents. Model B represents conditions existing after 15 February when the magma level maintained a persistently higher level allowing effusion from the 670 m vent. At this point a new flow field was emplaced on the upper SDF, where all active vents had a very shallow connection to a single, primary vent through a network of lava tubes.

The eruption started with very high effusion rate from the breached NE flank of Crater 1. We estimate that the first two flows were emplaced at a rate of ~283 m$^3$ s$^{-1}$. Thereafter effusion rates fell to a much lower level of typically <1 m$^3$ s$^{-1}$. The large difference in effusion rates between the initial lava flow events and later events can be explained by the fact that the first flows were fed from direct failure of a pressurized conduit. This would have also been subject to a hydrostatic pressure exerted by magma accumulated within the ~250 m long section of conduit above the active vent. Once the central conduit had been depressurized, and the hydrostatic head drained, so the effusion rates fell to a level roughly equal to the time averaged supply from the deeper source [Harris et al., 2000a, 2005].

Thereafter, comparison between maximum temperatures recorded on the floors of the summit craters and on the surface of the lava flow fields (Figure 15), indicated a shallow connection between the central magma column and the flank vents. Backup of lava in the central column would result from declining effusion rates. This would cause magma levels to rise in the central conduit and crater temperatures to thus increase. This would increase the hydrostatic head above the flank vents, and hence increase the effusion rate once more. Such a cycle would thus end by tapping of the column during higher effusion rate periods to cause central column magma levels, and hence crater floor temperatures, to decrease once more.

At the end of the eruption, strombolian activity resumed at the summit craters and a progressive deactivation of the lava flow fields was observed, culminating in a restoration of the conditions that existed at Stromboli prior to the eruption. We believe therefore that the 2002/03 eruption of Stromboli represents a relatively minor incident in the long history of this volcano that did not significantly change the geometry of the conduit or mechanics of the system.

6. Conclusion

Regular thermal surveys during the 2002–2003 effusive eruption of Stromboli enabled quantification of eruptive parameters such as effusion rate and maximum temperature at the bottom of the summit craters, as well as qualitative tracking of lava flow features, fissures and vents. In this regard, the use of a thermal imager allowed us to obtain regular measurements of effusion rate in a situation where traditional, ground-based measurements were difficult and dangerous. Regular surveys with a thermal camera contributed to a detailed eruption chronology and understanding of the lava flow and eruption processes operating during the eruption, and provide a powerful demonstration of the use of high temporal resolution thermal images collected using simple to use, portable, handheld imagers in tracking and understanding an ongoing effusive eruption.

Acknowledgments. We wish to thank G. Bertolaso and the Italian Civil Protection for their substantial support of our activities; M. Zaia (Zaaz) and the Alpine Guides of Stromboli for their field assistance; the helicopter pilots of Air Waier and Civil Protection, whose great expertise and courage allowed us to collect a huge amount of data; colleagues of INGV from Catania, Palermo, Napoli, Roma, Pisa, and Milano, who helped during our monitoring effort, and especially A. Bertagnini, M. Coltelli, M. de Vito, S. Di Vita, P. Landi, G. Mastrolorenzo, M. Neri, M. Polacci, and M. Pompilio, who shared with us the volcanological monitoring of the volcano. Essential support from A. Bonaccorso, G. Macedonio, and E. Boschi during the entire eruption is gratefully acknowledged. F. Sortino, M. Ripepe, and the “SAR Team” from University of Florence coordinated by N. Casagli are thanked for fruitful discussions throughout the eruption, and for having shared with us their unpublished data. Thanks also to R. Wright for helping with data conversion and to G. Garfì and M. Cascone for figure handling. A. J. L. H. and M. P. were funded by NSF grant EAR-0207734. D. Pieri, V. Realmuto, and L. Mastin are acknowledged for the careful review and thoughtful suggestions on an earlier version of the manuscript.

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