Seismic Signals Associated with Landslides and with a Tsunami at Stromboli Volcano, Italy

by Mario La Rocca, Danilo Galluzzo, Gilberto Saccorotti, Stefano Tinti, Giovanni B. Cimini, and Edoardo Del Pezzo

Abstract  In this article, we analyze the seismic signals produced by two landslides that occurred at the Stromboli volcano on 30 December 2002, recorded by both broadband and short-period seismic stations located in the 2.5–22-km distance range from the source. For both landslides, the characteristics of the low-frequency seismograms indicate a complex time history in the release of seismic energy. The first landslide occurred over the submerged part of the northwest sector of the volcano and had associated a large-amplitude, low-frequency pulse representative of the abrupt detachment of a large mass. Lower amplitude phases in the following 3 minutes possibly indicate minor detachment events. The highest amplitude, low-frequency signals are well described by a single-force source model. The second mass-failure episode is also characterized by a complex source and can be interpreted as a multiple event, with a less abrupt onset and at least four detachments occurring during 4–5 minutes and producing low-frequency signals.

Synthetic seismograms generated by a shallow single force located in the submerged area of Sciara del Fuoco and directed upslope, fit well the first low-frequency seismic pulse recorded at Stromboli and Panarea by three-component stations. From this simulation, we estimated the force exerted by the first mass failure. The estimate of the volume through two different procedures, gives values in the range of 1.0–1.5 million m$^3$ and about 14 million m$^3$, respectively.

The landslides, which involved both the submarine and the subaerial northwest flank of the volcano, produced a tsunami that struck the coast of Stromboli Island and in a few minutes reached the other islands of the Aeolian Archipelago. Three broadband seismic stations installed on land about 100 m from the coastline at Panarea Island, located 20 km southwest of Stromboli, recorded very long period seismic signals produced by the tsunami waves. Analysis of these signals gives invaluable information on the spectral content and propagation properties of tsunami waves and on their interaction with the ground at a short distance from the coast. Synthetic tsunami waves, obtained by a landslide source model and taking into account the bathymetry of the sea surrounding Stromboli and Panarea Islands, fit the observed phenomena and the experimental data very well.

Introduction

Stromboli is the northernmost island of the Aeolian volcanic arc in the southern Tyrrenian Sea, Italy. The typical continuous explosive activity of the volcano increased during the last months of 2002, and on 28 December, an effusive eruption started from a new fracture which opened along the northwest flank of the northeast crater, in the sector graben named Sciara del Fuoco (Fig. 1). This flank of the island has been affected in the past 13 k.y. by several huge-volume landslides and sector collapse episodes (Tibaldi, 2001) that probably produced large tsunami waves (Tinti et al., 1999, 2000, and 2003a). The effusive eruption started at 18:20 local time and in less than 1 hour reached the sea, flowing along the northern side of Sciara del Fuoco (for reference, see on-line daily reports at www.ct.ingv.it/Stromboli2002/Main.htm). The effusive eruption was accompanied by ash emission and by an increase in the amplitude of the volcanic tremor. Later, the lava flow split in three branches, and one of them filled a natural terrace at an elevation of about 600 m a.s.l.

On 30 December 2002, two landslides separated in time
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Figure 1. Map of Stromboli and Panarea islands with the location of seismic stations operating at the time of the landslides. Stromboli is also shown in the foreground with the topography, the Sciara del Fuoco sector culminating in the crater area, and the location of the landslide scar extending both below and above sea level. The arrow in the top left inset indicates the location of Stromboli volcano in the southern Tyrrhenian Sea, and the arrow in the submarine scar area represents the horizontal projection of the single-force source adopted to estimate the landslide volume. Full symbols represent broadband stations (the three at Panarea) and the intermediate band station (SX15), empty triangles indicate the short-period seismic stations. The origin of the reference system is lat. = 38° 47′ N, long. = 15° 11′ E.

by about 7 minutes occurred on the northern flank of the Sciara del Fuoco, in the area covered by the new lava flow (Bonaccorso et al., 2003). Shortly later, a tsunami caused severe damage along the coast of the island and reached the other islands of the Aeolian archipelago. At the same time, a dense cloud of dust and ash arose above the island and was blown east-northeast by the wind, affecting the village of San Vincenzo in Stromboli with a fallout lasting many hours. Fortunately, most of the houses along the coast were uninhabited during the winter season, and the few residents were able to escape, alerted by the rumbling sound produced by the incoming tsunami. Only a few people were injured. One landslide was observed directly by at least two persons: one of them (M. Pompilio, a volcanologist of the INGV-CT, Italy) took a short movie that turned out to be valuable for later analysis.

A subaerial detachment scar extending from the shoreline to about 500 m a.s.l., striking N40° W (Fig. 1) was observed during an aerial survey conducted a few hours later. Successive observations indicate that this scar was produced by two distinct mass failures, both associated with the second multiple landslide. The picture in Figure 2, taken a few days later, shows the subaerial landslide scar seen from the north. A submarine survey conducted some days later showed that the mass-failure scars extended below sea level (Chiocci et al., 2003; Tommasi et al., 2004). This was ex-
The event of 30 December attracted the attention of the scientific community and during the following weeks and months, many geophysical and volcanological investigations were conducted. Some preliminary analyses of the seismic signature of the 30 December landslides have been recently published by Bonaccorso et al. (2003) and Pino et al. (2004). In this article, we analyze the low-frequency seismic signals associated with the landslides to characterize the corresponding source processes. Comparison with synthetic signals generated by a shallow, single-force source allows estimating the force involved in the main slide. Then, using two different methods, we infer the volumes associated with the first submarine detachment and discuss the reliability of these two different results in the light of available geologic and geomorphological information. Very long period seismic signals recorded on land at Panarea Island have been interpreted as produced by the impact of tsunami waves on the coast.

Seismic Signals Associated with Landslides

The seismic signals produced by the events of Stromboli on 30 December were recorded by seven seismic stations installed on Stromboli and Panarea Islands (Fig. 1). Among them, one station equipped with a three-component 5-sec seismometer (SX15, Cimini et al., 2004), was operating at the INGV observatory, located about 2.5 km east of the place where the landslides occurred. Two short-period stations of the permanent seismic monitoring network of INGV-CT were installed respectively, at Ginostra (STR, southwest of the landslides, three components) and Punta Lena (PL1, south of the sources, vertical component only). Four seismic stations were installed at Panarea Island, about 20 km southwest of Stromboli. Three of them, PCAB, PAN, and PELB (Fig. 1), were equipped with three-component broadband seismometers (response from 60 sec to 50 Hz). These stations had been installed in early December 2002 by the INGV-OV mobile seismic network with the specific goal of monitoring the anomalous degassing offshore Panarea. The three broadband stations installed on Panarea Island, on land about 100 m from the coastline, provided the best data available about the tsunami.

The landslides produced seismic signals characterized by high amplitude and very emergent onset, irregular envelope, frequency content in the band 0.1–5 Hz, and a duration of many minutes (Fig. 3). The time of origin of the landslides cannot be defined with precision because the seismograms do not contain any sharp pulse. However, the maximum amplitude is seen on the seismograms at around 12:15 UTC for the first event and 12:23 UTC for the second (Fig. 3). The time of origin of the landslides cannot be defined with precision because the seismograms do not contain any sharp pulse. However, the maximum amplitude is seen on the seismograms at around 12:15 UTC for the first event and 12:23 UTC for the second (Fig. 3). The time of origin of the landslides cannot be defined with precision because the seismograms do not contain any sharp pulse. However, the maximum amplitude is seen on the seismograms at around 12:15 UTC for the first event and 12:23 UTC for the second (Fig. 3). The time of origin of the landslides cannot be defined with precision because the seismograms do not contain any sharp pulse. However, the maximum amplitude is seen on the seismograms at around 12:15 UTC for the first event and 12:23 UTC for the second (Fig. 3).
The spectral content of these signals is rich in low frequencies, with high-amplitude peaks spanning the 0.01–5 Hz frequency band, depending on the site and kind of seismometer (Fig. 4). From the spectrogram of the seismic signals recorded by broadband stations at Panarea and by station SX15 at Stromboli, a well-defined, low-frequency (0.1 Hz) wave packet is clearly visible at the beginning of both landslide-generated signals (Fig. 5). For the first event, this phase starts at 12:14:54 at station SX15 (Fig. 3 and Fig. 6), and the signal amplitude exceeds the average noise amplitude from 12:14:15. During this 40-sec time window, the seismic signal increases gradually without an important change in frequency content. This effect is clearly visible at the three stations installed on Stromboli, but it is not observed at Panarea.
area. Then a low-frequency signal, recognized in the low-pass-filtered seismograms, starts with the highest amplitude on the horizontal components.

The seismic signals recorded at Panarea are naturally low-pass filtered due to the much greater source–station distance (21 km for PCAB against about 2.8 km for SX15, Fig. 1). Thus, the very emergent onset, characterized by frequencies in the band 0.5–5 Hz, is not so evident as described before for SX15. On the contrary, a low-frequency phase (around 0.1 Hz) starting at 12:15:00 constitutes the landslide seismic signal onset (Figs. 3 and 7). In this case, the highest amplitude is also observed on the horizontal components, and the coherence of this signal among the three broadband stations is very high. The estimated 7–8 sec difference in travel time of this \( S \)-phase between SX15 and PCAB is compatible with the shear-wave velocity of about 3 km/sec that characterizes the shallow crust in the area around Stromboli and Panarea Islands (De Luca et al., 1997; Neri et al., 2002). Array analysis applied to seismograms from the three broadband stations gives an apparent velocity in the range of 2.5–3.5 km/sec for this low-frequency phase, as described in detail later.

Particle motion of the initial low-frequency phase, in the band 0.05–0.2 Hz, has been studied accurately for the high dynamic range stations SX15 and PCAB. On the scale of our analysis, seismic signals at periods around 10 sec are not affected by the Earth’s heterogeneities, but particle motion can be significantly modified by the interaction of the wavefront with the free surface and the topography (Neuberg and Pointer, 2000; Hellweg, 2003). However, given the low number of stations available, we do not use particle motion to locate the source, so we have not applied any correction to account for topographic effects. Since the frequency response of SX15 is flat only above 0.2 Hz (5-sec free period), the seismograms from this station have been corrected for the sensor transfer function. For both stations, particle motion is predominantly horizontal, highly rectilinear, and oriented northwest–southeast, with the first pulse toward the southeast, as shown in Figs. 6 and 7. This is perfectly compatible with a single-force source model, the force due to
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Figure 5. Spectrograms stacked over the three components of stations SX15 and PCAB for the first landslide. Data from SX15 have been corrected for the sensor response, and for both stations, data have been integrated to obtain the ground displacement before the spectrogram computation. The low-frequency pulse around 0.1 Hz is very well defined around 900 sec at SX15 and between 900 and 940 sec at PCAB.

The recoil of the ground when a large mass starts sliding down. The single-force model has been adopted to interpret seismic signals produced by landslides by several authors, such as Kanamori and Given (1982), Kanamori et al. (1984), Eissler and Kanamori (1987), Kawakatsu (1989), Okal (1990), Dahlen (1993), Pelinovsky and Poplavsky (1996), and Ma et al. (1999).

From the landslide scar shape and direction, we expect the force acting at the source to be oriented S50° E ± 5°. Then the angle between the source force and the SX15 station direction is about 45°. This means that SX15 is located in the direction of the minimum energy radiation pattern and the motion has both longitudinal and transverse components (Aki and Richards, 1980; Lay and Wallace, 1995). Since the source–station distance is no more than 2.8 km and we are analyzing data in the 0.05–0.2 Hz band, the S–P time is negligible with respect to the wave period. Then the composition of these two motions results in an almost horizontal and rectilinear motion oriented northwest–southeast. Hence, the experimental result, shown in Fig. 6, is in good agreement with the theoretical model.

Station PCAB, located on Panarea Island (Fig. 1), forms an angle $\phi$ of about 72° with respect to the source force direction. The wave radiated in that direction by a single-force source is mostly transversal (the $S$-wave and the $P$-wave amplitudes proportional, respectively, to $\sin \phi$ and to $\cos \phi$ (Aki and Richards, 1980). The particle motion obtained by PCAB data, shown in Fig. 7, agrees with this model; it is mostly horizontal and oriented northwest–southeast. These considerations indicate that the low-frequency seismic signal observed around 12:15:00 UTC is produced by a large mass which starts sliding down almost simultaneously along the submerged slope of Sciara del Fuoco. The precursor higher frequency emergent signal, characterizing the seismograms recorded at Stromboli by stations close to the source, can be interpreted as produced by crack opening and small blocks sliding and rolling down before the major mass detaches. Similar considerations can be invoked to explain the long coda of the seismograms.

The low-frequency signal envelope has a quite sharp onset at both stations and reaches the highest amplitude in two periods of the signal (Fig. 8). Several lower amplitude but highly visible wave packets are present in the low-frequency band (0.05–0.2 Hz) after the beginning of the first landslide. They are identified at both stations around 12:15:28, 12:16:02, 12:16:41, and 12:17:52.

The second landslide was certainly more complex than the first. It contains many low-frequency phases, some of which are different from the first case described before. Seismograms filtered over the same frequency band (0.05–0.2 Hz) show...
Figure 6. Top: Three components of the ground displacement observed at SX15 at the beginning of the landslides. These signals were obtained through deconvolution of the seismometer response and integration of the original seismograms. Data have been also corrected for the acquisition system characteristics and filtered in the band 0.05–0.2 Hz. Bottom: Three-particle motion of the low-frequency signals shown above, relative to the time window 12:14:20–12:15:10, demonstrate that the low-frequency ground shaking was predominantly horizontal and characterized by high rectilinearity. These features and the polarization direction are in good agreement with the wavefield produced by a shallow, single-force source.

Hz, see Fig. 8) contain at least five distinct phases in the time interval ranging from 12:22:40 to 12:28:05. At 12:22:40, a well-defined phase similar to the first event is observed on the horizontal components of SX15, but it cannot be recognized on the seismograms recorded at Panarea. At about 12:23:10, a second low-frequency phase, with vertical amplitude similar to the horizontal ones, is recorded by all the stations. It is followed by other similar phases for about 40 sec. These characteristics of the seismic signals may indicate that the second landslide was generated by a more complex source mechanism, radiating energy in several steps, sometimes with a pattern different from that of the first event. Further pulses around 12:28:00 could be interpreted as a third landslide as well.

Many efforts have been aimed at obtaining an accurate location of the sources involved in the landslides, with the main goal of locating the two major events. Unfortunately, available data are not sufficient to achieve this purpose, since the closest stations, STR and PL1, equipped with short-period sensors, are not useful for analyzing the signals around 0.1 Hz associated with the main mass failures. Moreover, the seismogram of PL1 and the north–south component of STR were clipped in the second landslide due to the low dynamic range of the acquisition system. Therefore, neither correction for sensor response nor signal filtering can be applied. At higher frequency, in the band 0.2–0.8 Hz, the seismograms contain more pulses, but the identification of the same phase at different stations is more difficult or impossible when three-component data are not available.

Array and Polarization Analysis of the Low-Frequency Seismic Wavefield

The three broadband stations on Panarea Island, arranged in a triangular configuration, are far enough from the source to allow the application of array techniques to study the kinematic properties of the wavefield. Thus we applied the zero-lag cross-correlation (ZLCC) method (Del Pezzo et al., 1997) to the seismograms filtered between 0.05 and 0.2 Hz, to obtain correlation, apparent velocity, and backazimuth versus time. We also performed polarization analysis in the time domain, in the same frequency band, using the covariance matrix method (Jurkevics, 1988). This analysis is based on the evaluation of eigenvectors and eigenvalues of the “array stacked covariance matrix” obtained by three-component seismograms shifted in time by the value cor-
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Figure 7. Top: Three components of the ground displacement at Panarea (station PCAB) in the 0.05–0.2 Hz frequency band. These signals come from integration of the original seismograms and correction for the data logger characteristics; the deconvolution of the sensor response has not been applied because the corner frequency is much lower than the frequency analyzed. Even the signal-to-noise ratio is much lower than that observed at SX15 (Fig. 6) due to the greater source–station distance and the lower sea–station distance; the characteristics of the wavefield evidenced by the particle motions (bottom) are similar to those observed at Stromboli.

responding to the highest correlation value found in the ZLCC analysis (La Rocca et al., 2001). The eigenvector associated with the highest eigenvalue and the three eigenvalues are used to compute the incidence, azimuth, and rectilinearity of particle motion. The polarization incidence is the angle between the first eigenvector and the vertical axis, and the polarization azimuth (or backazimuth) is the angle between the projection on the horizontal plane of this vector and north, measured clockwise. The rectilinearity is computed by the formula $RL = 1 - (\lambda_2 + \lambda_3)/2\lambda_1$, where $\lambda_1 \geq \lambda_2 \geq \lambda_3$ are the three eigenvalues of the covariance matrix.

Both ZLCC and polarization analyses were carried out using a 30-sec-long window with 90% overlap sliding along the signals. Horizontal components were rotated clockwise by 33° to handle directly with radial and transverse components. The ZLCC analysis was performed three times, once for each component of the ground velocity. Even the array was composed of only three stations, and its configuration was elongated in the radial direction; the analysis carried out using the three components separately gives similar results. This is not surprising because seismic waves produced by landslides have vertical, radial, and transverse motion, as seen in the low-frequency seismograms, as expected from the source characteristics discussed before and from the source-array direction. Moreover, the seismic wavelength corresponding to the analyzed frequency band is much larger than the distances among the three stations, so the plane wave approximation holds, and the correlation between each pair of seismograms recorded at the three sites is very high.

Figure 9 shows the results obtained from ZLCC analysis of the transverse components (which are characterized by the highest amplitude) and polarization analysis. The envelope at the top of the figure is the same as in Figure 8 for station PCAB. The results relative to windows characterized by cross-correlation greater than 0.9 are shown by bold symbols to emphasize the most correlated phases of the wavefield. Many of the low-frequency phases marked by arrows in Figure 8 are characterized by very high correlation which, in some cases, reaches values as high as 0.99. For the first landslide, the correlation is higher than 0.95 for more than 1 minute, the apparent velocity is in the range of 2.5–4 km/sec, the backazimuth in the range of 10–50°, the polarization azimuth is stable around 300°, and rectilinearity and polarization incidence are both high. All these values indicate that the low-frequency signal is composed mostly of shear waves produced by a shallow source located in the direction of Stromboli Island. The wide backazimuth range is not surprising because the array, composed of only three stations...
and elongated in the radial direction, has low resolving capability for directions of arrival pointing toward Stromboli. The signal between 12:17:20 and 12:18:30 is also very well correlated and the results of ZLCC and polarization analysis are the same as those of the first landslide. These results are compatible with the hypothesis of several important mass failures occurring over 3 minutes following the first landslide. On the contrary, the correlation of phases between 12:16:30 and 12:17:20, also marked by arrows in Figure 8, does not exceed 0.9. Two well-correlated phases appear at 12:19:00 and around 12:21:25, but their apparent velocity, backazimuth, and polarization parameters are different from the values corresponding to those of the first landslide, particularly for the second window (12:21:10–12:21:40). The signals marked at 12:22:40 and 12:23:13 in Figure 8, which are clearly visible on the seismograms recorded at Stromboli but not evident at Panarea stations, do not exceed our correlation threshold at the array. We interpret these phases as produced by two mass failures characterized by a different radiation pattern, so they are well recorded at Stromboli but not at Panarea stations.

The second large landslide, which occurred after 12:23, produced very well correlated signals at the array for more than 2 minutes between 12:23:40 and 12:26:00. However, only in correspondence with the highest signal amplitude, for about 1 minute around 12:24, do apparent velocity, backazimuth, and polarization parameters assume values corresponding to those expected. The last two, very short, well-
correlated phases visible around 12:28 and 12:29 have characteristics similar to those of the main landslide events.

Polarization analysis has also been applied to data recorded by station SX15 in the same frequency band, and the results are shown in Figure 10. In this case, bold symbols represent the results relative to time windows for which the envelope stacking shown at the top is greater than an arbitrary threshold which selects all the phases marked by arrows in Figure 8. Again, we see that both major landslide events are characterized by very stable values of the polarization angle, whereas the phase around 12:21:25 shows a completely different pattern.

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**Estimate of the Landslide Volume**

The qualitative agreement of observed particle motions with expected results led us to compare data from the first landslide with synthetic signals associated with a shallow single-force source. For this purpose, we generated synthetic waveforms at both sites of the stations described before and compared them with the observed low-frequency seismograms. The aim of this analysis is to estimate the magnitude of the force exerted by the sliding mass from which we can infer the landslide volume. The volume has been estimated using two different approaches: first, we apply the formula

![Figure 10. The results of polarization analysis for station SX15. The frequency band, length, and overlap of the sliding window are the same as those described for Figure 9. Bold symbols show the phases with large amplitude, as shown by the signal envelope plotted in the first panel. This selection includes all the phases marked by arrows in Figure 8.](image-url)
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\[ F = \alpha^2 MD \text{ (Dahlen, 1993)} \]

and then we consider a simple model consisting of a body sliding down a tilted plane at constant acceleration.

Synthetic seismograms have been obtained using the wavenumber integration method implemented in Computer Programs in Seismology software by R. B. Herrmann, version 3.25 (2002). This software evaluates Green’s functions for any seismic sources, including a single force, given a propagation medium model and specifying the source–time function. Then, an appropriate linear combination of Green’s functions, taking into account the source magnitude, orientation, and position with respect to the receiver points, gives the three components of the ground velocity at the two station sites.

The velocity model at Stromboli Island is known in detail only for the shallower few hundreds meters (Petrosino et al., 2002). Therefore, we adopted a simple multilayer model to simulate a constant velocity gradient in the upper 8 km (Fig. 11), with velocity values compatible with the results of seismic tomography studies carried out for the south Tyrrhenian Sea region (De Luca et al., 1997; Neri et al., 2002).

For our simulation, we consider a single-point force applied 50 m below the sea bottom in the center of the submarine scar (Fig. 1). The direction is 140° from north, the main scar direction is observed both below and above sea level (Tommasi et al., 2003), and the tilt is 25° upward, consistent with the slope of the sea bottom. Synthetic seismograms were generated using different source–time functions and changing their duration over a wide range of time. For each source–time function, the normalized cross-correlation between synthetic and observed seismograms was computed for any components of both stations SX15 and PCAB. Though the fit of horizontal components is very good, particularly at SX15 (correlation up to 0.93), the synthetic vertical motions do not fit the observed data. Since most of the energy is observed on the horizontal components, we considered only the latter to estimate the source force magnitude from the best fit. We believe that the misfit between synthetic and observed vertical motions indicates that the source was more complex than the simple model adopted, and maybe our velocity model is quite different from reality. Moreover, the vertical component is also the most affected by topography, due to the P–SV conversion at the free interface, but this interaction was not considered in our analysis.

The total momentum of the whole process must be zero, hence the time integral of the source–time function must vanish. Our first two source–time functions consist of a sine wave and a sine wave tapered with a Hanning window (see Table 1 and Fig. 12). The latter function gives a better fit of the observed seismograms. However, we believe that these symmetrical source–time functions do not satisfactorily represent the submarine landslide because the deceleration process is expected to be much slower than the accelerating stage. Moreover, in a submarine landslide, the friction along the sliding plane during the deceleration stage is much smaller than that in the acceleration due to the water that lubricates the contact and reduces the effective density through buoyancy. These characteristics of landslide dynamics have been pointed out in many cases (Hasegawa and Kanamori, 1987; Okal, 1990; Dahlen, 1993). For this reason, we also tested a source–time function with a nonzero integral to simulate the signals corresponding only to a positive acceleration process. This function, shown in Figure 12, is the same as that used by Kanamori and Given (1982) to simulate the signals produced by the 1980 eruption of Mt. St. Helens.

We evaluated Green’s functions associated with each of the three source–time functions, then from these data, we computed the three components of ground velocity, taking into account the source direction and receiver position. Synthetic seismograms are band-pass filtered over the same frequency band used for experimental data, and then integrated to obtain the ground displacement. The force magnitude is computed from the average ratio between the observed and synthetic seismogram amplitude, and the fit quality is estimated by normalized cross-correlation between each pair of seismograms, as shown in Figure 12.

The following remarks are needed for correct interpretation of these results and a successive estimate of the landslide volume:

1. The data set useful for the analysis of the source is constituted by only four three-component stations, three of which were located so close to each other (the three broadband stations at Panarea) that they contain, in practice, the same information. Therefore we used only SX15 and PCAB, but the azimuth and distance coverage of the source are quite poor.
The method we are using to estimate the force, and hence the volume, was applied with very good results to landslides much larger than that of Stromboli, such as the Mt. St. Helens event of 18 May 1980 (Kanamori and Given, 1982; Kanamori et al., 1984), the Mantaro landslide (Peru, 24 April 1974) (Kawakatsu, 1989), and the Kala-pana earthquake (Hawaii, 29 November 1975) (Eissler and Kanamori, 1987; Kawakatsu, 1989). In these cases, the landslides were so large (more than 1 km$^3$) as to excite very long period seismic waves recorded all over the world.

The simple relation $F = \frac{\omega^2}{H^{1.5}} \times D^2 M$, where $M$ is the sliding mass and $D$ is the displacement of the center of mass, is very efficient in the description of large Table 1

<table>
<thead>
<tr>
<th>Source Force Function</th>
<th>Displacement</th>
<th>$T$</th>
<th>$D$</th>
<th>$F$</th>
<th>Volume (m$^3$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>$MA\sin(\omega t)$</td>
<td>$\frac{AT^2}{2\pi}$</td>
<td>12 sec</td>
<td>54 m</td>
<td>$4 \cdot 10^{10}$N</td>
<td>$1.0 \cdot 10^6$</td>
</tr>
<tr>
<td>$\frac{4\sqrt{3}}{9} MA\sin(\omega t)[1 - \cos(\omega t)]$</td>
<td>$\frac{\sqrt{3}AT^2}{6\pi}$</td>
<td>16 sec</td>
<td>54 m</td>
<td>$4 \cdot 10^{10}$N</td>
<td>$1.0 \cdot 10^6$</td>
</tr>
<tr>
<td>$\frac{MA}{2}[1 - \cos(\omega t)]$</td>
<td>$\frac{AT^2}{4}$</td>
<td>10 sec</td>
<td>61 m</td>
<td>$7 \cdot 10^{10}$N</td>
<td>$1.5 \cdot 10^6$</td>
</tr>
</tbody>
</table>

The force $F$ is computed by the ratio between synthetic and observed ground displacement, and the frequency is assumed to be $\omega = \frac{2\pi}{T} = 0.628$ rad/sec, corresponding to the 10-sec low-frequency signal. The acceleration is $A = 2.36 \text{ m/sec}^2$, computed by the formula $A = g(\sin\theta - \mu\cos\theta)$ with a slope $\theta = 25^\circ$ and a coefficient of friction $\mu = 0.2$. Mass $M$ is computed assuming a density of 2 g/cm$^3$ and neglecting buoyancy.

Figure 12. Top: The three source–time functions used to compute the synthetic waveforms. Functions a and b describe a complete sliding motion until the final rest position; function c describes only the acceleration process. The analytical functions and related displacement, force estimate, and corresponding volume are summarized in Table 1. The bottom plots show the normalized cross-correlation between synthetics and observed low-frequency seismograms computed for stations SX15 and PCAB. Diamonds represent the average correlation value between horizontal components relative to source a. Open circles and crosses show the analogous results corresponding to source functions b and c, respectively. The source durations for the final estimate of force magnitude through synthetic/observed ratios have been chosen, looking at the maxima of these data, 12 sec, 16 sec and 10 sec for functions a, b, and c, respectively.
events. On the contrary, the first Stromboli landslide, which is the object of this analysis, as well as the second one, were much smaller, involving a total volume of the order of 20 million cubic meters. Furthermore, each landslide did not produce seismic signals with periods longer than 15 sec, though this could be apparently in strong contrast with the event duration which, though not known exactly, must have been about 1 minute, given the runout distance and the slope of the volcano flank.

4. The single-force source model is an approximation of the real process, as shown by Dahlen (1993). In the present analysis, we neglect the quadrupole terms proportional to $\beta^2$, $\beta$ being the shear wave velocity in the sliding material, that is very low in a highly fractured, water-filled sliding mass. Also, we neglect any mass variation during the acceleration process. This is a reasonable assumption considering that the area of detachment is of the order of $100 \times 100$ m$^2$, thus we expect the propagation time of the fracture to be much smaller than the dominant wave period taken into account in this analysis.

Bearing these considerations in mind, our simulation is aimed at finding a simple source mechanism which gives a good qualitative fit to the first low-frequency seismic phase. The distance traveled by the center of mass is computed by double integration of the acceleration function, which is proportional to the source force. The peak acceleration $A$ has been computed considering a body in motion over a plane with slope $\theta$ and friction coefficient $\mu$. Assuming $\theta = 25^\circ$ and $\mu = 0.2$ (Brodsky et al., 2003), the acceleration given by $A = g(\sin\theta - \mu \cos\theta)$ results in 2.36 m/sec$^2$. The relation between the source–time function and distance $D$, as well the results obtained, are indicated in Table 1. The duration $T$ of the source function is that corresponding to the highest correlation at both stations between synthetics and observed signals. The force $F$ which gives synthetic horizontal component seismograms of amplitude equivalent to the observed signals, averaged on SX15 and PCAB, is also indicated. Mass and then the volume have been computed by the formula $F = \omega^2 MD$ assuming $\omega = 0.1 \cdot 2\pi$ rad/sec, and a rock density of 2 g/cm$^3$. The volume of the first large detachment is $10^6$ cubic meters for the first two source functions and $1.5 \cdot 10^6$ m$^3$ for the third source. These results, even relative only to the first mass failure, are not consistent with the submarine landslide volume of $13 \cdot 10^6$ cubic meters estimated by bathymetric and photogrammetric data collected before and after the landslides (Tommasi et al., 2004). The estimated center of mass displacement $D$, around 55 m for the two source–time functions describing acceleration and deceleration, is also much smaller than the observed value which is about 300 m. On the contrary, the displacement of 61 m covered only during the acceleration stage, obtained from the third source–time function (Fig. 13 and Table 1), is much more reasonable. However, the corresponding volume is still too

Figure 13. Comparison of low-frequency signals recorded at SX15 and PCAB (thin continuous line) with synthetic seismograms corresponding to the source–time functions b (dashed line) and c (thick gray line). The source–time functions of the synthetic signals are shown in Figure 12; the direction of the source force is toward the southeast (140° from north) and upslope (25°).
small, even if it is greater than that in the first two cases. Assuming that the procedure applied before is inadequate to study small landslides like those of Stromboli, we follow a different approach to obtain another estimate of the volume. The acceleration of a body in motion on a plane with slope \( \theta \) is given by \( a = g(\sin \theta - \mu \cos \theta) \), where \( \mu \) is the coefficient of friction. The force exerted on a sloping plane by a body of mass \( M \) is obviously \( F = Ma \), and the corresponding volume is given by

\[
V = \frac{F}{g(\sin \theta - \mu \cos \theta) \rho}.
\]

With the same numerical values used above for \( \theta \), \( \rho \), and \( \mu \), and \( F = 7 \times 10^{10} \) N, this formula gives a volume of \( 14.8 \times 10^6 \) m\(^3\). This second model, much simpler than the first, produces a volume one order of magnitude greater than the previous and much closer to the expected value between \( 10^6 \) and \( 14 \times 10^6 \) m\(^3\).

The uncertainty associated with our estimates of the volume is not easily computed. The parameter which affects the volume estimate more is the coefficient of friction \( \mu \). Changing \( \mu \) in the range of 0.1–0.3, the acceleration \( A \) along the plane with slope \( \theta = 25^\circ \) varies between 3.3 and 1.5 m/sec\(^2\). This corresponds to an uncertainty of 37% on the displacement \( D \) and hence on the volume. The problem is further complicated by the presence of water, which can much reduce the friction. The coefficient of friction for dry landslides varies in the range of 0.2–0.6 (Brodsky et al., 2003), and it can be one order of magnitude lower in submarine landslides (Tinti et al., 2000). When the submarine slide starts, it is reasonable to assume a high value of friction, neglecting the presence of water at the interface. However, during motion, the highly fractured sliding mass is expected to become more similar to fluid flow than a solid body sliding over another solid. In this hypothesis, the friction with the sea bottom is much smaller than the slide onset and gives a very low contribution to the generation of seismic waves. Therefore, we believe that the low-frequency seismic signals analyzed before correspond only to landslide detachment and are produced by the force drop between the preslide equilibrium state and the sliding motion.

**Very Long Period Seismic Signal Associated with the Tsunami**

Data recorded by the three broadband stations at Panarea constitute a rare recording on land of a tsunami. About 5 minutes after the onset of the first seismic signal produced by the first landslide, a very long period wavefield is re-

![Figure 14](image-url)

**Figure 14.** Top: Three components of the ground displacement at Panarea (station PCAB) obtained through integration of the original seismograms and filtering in the 0.005–0.03 Hz frequency band. Bottom: This very long period signal produced by tsunami waves is characterized by negligible vertical motion; the horizontal motion is polarized perpendicularly to the shoreline, incidentally, radial for this station.
corded at the station PCAB, which is located on the northern flank of Panarea, about 100 m from the beach. This very long period (VLP)-signal is characterized by a well-defined onset (bearing in mind the uncertainty due to the extremely long period), frequencies between 0.006 and 0.015 Hz (Fig. 4), and a very long coda, lasting several hours. Particle motion shows that the onset is almost completely horizontal with a highly rectilinear pattern in the radial direction (Fig. 14) that in this case incidentally coincides with the direction normal to the shoreline. For the stations PAN and PELB, the onset particle motion is always horizontal and oriented substantially perpendicularly to the shore and very different from the station-source direction. Therefore, we interpret the VLP signals recorded by the seismic station as the effect of local ground tilt produced by the tsunami waves impinging on the coast. It is well known that horizontal components of broadband seismometers are very sensitive to tilt, whereas the vertical component is not (Aki and Richards, 1980; Wie- landt and Forbriger, 1999). This explains the horizontal particle motion oriented toward the sea and is confirmed by the analysis of arrival times. The starting time of this VLP signal at the three stations is incompatible with any reasonable propagation velocity of seismic waves in the crust, but perfectly compatible with the traveling speed of tsunami waves. Simulations of the tsunami generated by the first submarine slide performed by using a deformable body assumption for the landslide dynamics and a shallow-water approximation for tsunami generation and propagation (Tinti et al., 1999) demonstrate that the travel time of the tsunami to cover the sea distance from the source area at Stromboli to North Panarea, where the station PCAB is located, is about 300 sec. In Figure 15, the tsunami water elevation snapshot computed 300 sec shows that the tsunami propagation from the source is far from being isotropic but is governed by the sea depth. The leading trough, producing a negative first arrival at the coastline, arrives at North Panarea from the north and northwest since in the northeast it is delayed by the impact against the little island of Basiluzzo about 2 km northeast of Panarea and by the shallow-depth sea surrounding Dattilo and Lisca Bianca (Fig. 1). Note also the complicated wave pattern around Stromboli, where a system of local trapped waves has formed and is traveling around the island; this which is a well known effect of the interaction of tsunamis with volcanic islands (Tinti and Vannini, 1995; Piatanesi and Tinti, 1998). In addition, the relative large delays between the onset of the signal in PCAB and the signals in PAN and in PELB, which is the furthest station, support the hypothesis that the signals are due to the tsunami waves propagating along the coasts of Panarea: east of Panarea in the region between the islands of Basiluzzo, Dattilo, and Lisca Bianca.

Figure 15. Snapshot of tsunami elevation field computed 300 sec after the submarine landslide. Only the islands of Stromboli (S) and of Panarea (P) are shown on the map; the smaller islands such as Basiluzzo and Dattilo are omitted.
(Fig. 1), the sea is very shallow, with an average depth about 50 m, corresponding to tsunami velocity in the range of 20–25 m/sec.

Discussion and Conclusions

The low-frequency seismic signals associated with the two major landslide episodes can be described by a shallow single-force source model. However, it is also evident that the landslides were characterized by a succession of multiple detachments. Available data do not allow locating the beginning of mass detachments; thus we cannot infer the occurrence sequence among the submarine slump and the landslides which produced the two scars along the Sciara del Fuoco. On the other hand, analysis of the seismic signal associated with the tsunami indicates that the submarine slump started before the subaerial landslide reached the sea.

The comparison of low-frequency signals with synthetic seismograms generated by a shallow single-force source allowed an estimate of the force exerted by the first mass failure over the detachment surface. We used this value to estimate the landslide volume applying two different techniques. The first method, applied with good results to large scale landslides by several authors [Kanamori and Given, 1982; Kawakatsu, 1989; Dahlen, 1993], in the present case, clearly underestimates the volume. This indicates that the method is not reliable for a quantitative analysis of landslide volumes on the order of $10^{5} - 10^{6}$ cubic meters, which do not excite enough long period seismic waves. On the contrary, our second estimate of the volume, based on simple motion along a tilted plane with friction, gives a volume value much higher, about 10 times the first estimate. This volume of about $10^{7}$ m$^3$ for the first large detachment, is probably an overestimate of the true value but is more compatible with the total landslide volume (about $20 - 10^{6}$ m$^3$). The low-frequency seismic signal amplitude indicates that the first landslide was the equivalent of or greater than the second one which occurred between 12:22:40 and 12:24:10 (Fig. 8). This confirms the estimated volumes of about $13 - 10^{6}$ m$^3$ for the submarine slump and about $7 - 10^{6}$ m$^3$ for the subaerial event (Tommasi et al., 2004). On the other hand, under the hypothesis that any low-frequency pulse following the first one represents a mass detachment, Pino et al. (2004), using an analysis similar to that applied here, show that the sum of all secondary mass failures is smaller than the first detachment. Thus we expect the volume of the first large mass failure to be at least one-half of the total volume. An analysis of VLP seismic signals recorded at Panarea Island, produced by tsunami waves, confirms that the first landslide occurred below the sea.

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