Probabilistic source location of explosion quakes at Stromboli volcano estimated with double array data

Mario La Rocca a,*, Gilberto Saccorotti a, Edoardo Del Pezzo a, Jesus Ibanez b

a I.N.G.V. Osservatorio Vesuviano, Via Diocleziano 328, 80124 Naples, Italy
b Istituto Andaluz de Geofísica, Universidad de Granada, Campus Universitario de Cartuja, Granada, Spain

Received 21 May 2002; accepted 22 September 2003

Abstract

Data from two dense arrays of short-period seismometers are used to retrieve source locations of the explosion quakes at Stromboli volcano. Slowness vectors estimated at both arrays with the zero-lag cross-correlation technique constitute the experimental data set. A probabilistic approach based on a grid search spanning the volcano interior is used to calculate the probability of the source location. Results depict a shallow source, located beneath the crater area, at depths not greater than 500 m below the surface. Results are slightly different from, but comparable to, those obtained in a companion experiment carried out in the same time period using a broad-band seismometer network, which show a source shifted some hundreds of meters northwest of the crater area. The method is revealed to be effective and useful for future studies having the purpose of real-time tracking of the explosion quakes and tremor.

Keywords: Stromboli; array; explosion quake; source location

1. Introduction

Stromboli Volcano is a small island located in the Tyrrhenian Sea, Southern Italy (38°47′N, 15°13′E). The volcanic edifice rises about 3000 m from the seafloor, reaching 927 m above sea level. Its present volcanic activity is characterized by mild, intermittent jets of gases laden with molten lava fragments which burst in short eruptions lasting 5–15 s and occurring at a typical rate of 3–10 events per hour. The active vents are located about 750 m above sea level, at the top of a sector graben named Sciara del Fuoco (Fig. 1). Seismic activity at Stromboli is dominated by spindle-shaped signals associated with the summit explosions, superimposed on a background of sustained volcanic tremor. Typical tremor spectra span the 1–5-Hz frequency range, while explosion quakes depict a more complex spectrum, which broadens over the 0.1–10-Hz frequency interval (Neuberg et al., 1994; Langer and Falsaperla, 1996; Chouet et al., 1999).

Over the past 20 years, many efforts have been aimed at elucidating the wavefield properties of the seismicity associated with the Strombolian activity. Among others, we make particular reference to the work by Chouet et al. (1997), as it...
describes the details of the frequency–wavenumber spectrum of tremor and explosion quakes using data from a small-aperture array of short-period seismometers deployed on the northern flank of the volcano during 1992. Chouet et al. (1997) used these results to quantify the importance of source and path effects in the observed wavefield, and modeled the explosive source as a gas-piston mechanism operating in the upper part of a quasi-vertical crack-like conduit. At the end of their study, Chouet et al. (1997) left some open questions which are summarized by two main points. The first is related to the quantification of the physical mechanism driving the fluid transport which is at the basis of the generation of tremor and explosion quakes. The second regards the spatial extent of the zone in which the high-frequency energy of tremor and explosion quakes is generated. To answer the first question a further experiment was carried out in 1997, during which a temporary broad-band network was operated at Stromboli (Chouet et al., 1999). These broad-band data show that the very-long-period components of the explosive signals are associated with the action of two stationary sources repeatedly activated in time, with a source centroid offset 300 m beneath and 300 m northwest of the active vents. From analysis of the source–time functions of these signals, Chouet et al. (1999) proposed a source process characterized by a three-step mechanism. The first force system would be associated with the pressurization of the crack-like conduit,
in response to the ascent of a gas slug. The second force system would be representative of a depressurization of the conduit in response to a mass withdrawal caused by the emission of magma during the explosions, and the third would be associated with the repressurization of the conduit as a consequence of magma re-filling.

To address the second question regarding the location and space extent of the high-frequency source, Saccorotti et al. (1998) developed a Bayesian approach to retrieve the probability of source location from inversion of the array data collected during the 1992 array experiment. This inversion was essentially based on the back-tracing of the wave vectors observed at a single array, and thus the results obtained by Saccorotti et al. (1998) were completely unresolved along the source array direction. At the end of their paper, Saccorotti et al. (1998) extended the method to the case of a double array deployment, and used numerical simulations to demonstrate that a second array located on the west flank of the volcano was needed to properly locate the high-frequency source at Stromboli.

Following the above indications, a further experiment was carried out during September 1997. The experiment consisted in the deployment of two seismic arrays of short-period seismometers located on the north and west flanks of Stromboli Island, at Labronzo and Ginostra sites respectively (Fig. 1). The details of the experiment deployment, a description of the instrument response and other technical information can be found online (Del Pezzo and Stromboli Working Group, 1998, http://www.ov.ingv.it). Here we report only the main information.

Ginostra array included 15 vertical-component and three three-component Mark Products L4C seismometers, which have a natural frequency of 1 Hz. Labronzo array included 26 vertical-component and two three-component Mark Products L15B seismometers with a natural frequency of 4.5 Hz. Electronic extension allowed these sensors to operate at a higher frequency range.

2. The experiment

In September 1997 two small-aperture, short-period seismic arrays were installed on the north and west flanks of Stromboli Island, at Labronzo and Ginostra sites respectively (Fig. 1). The details of the experiment deployment, a description of the instrument response and other technical information can be found online (Del Pezzo and Stromboli Working Group, 1998, http://www.ov.ingv.it). Here we report only the main information.

Ginostra array included 15 vertical-component and three three-component Mark Products L4C seismometers, which have a natural frequency of 1 Hz. Labronzo array included 26 vertical-component and two three-component Mark Products L15B seismometers with a natural frequency of 4.5 Hz. Electronic extension allowed these sensors to operate at a higher frequency range.

Fig. 2. Amplitude and phase response of the seismometers used at Labronzo (Mark L15B with electronic extension of the bandwidth), continuous line, and at Ginostra (Mark L4C), dashed line. The response includes the anti-alias filter with cut-off frequency at 49.1 Hz and the A/D converter.
to achieve a response curve almost flat down to the frequency of 1 Hz, comparable to the response of the sensors used at Ginostra, as shown in Fig. 2. The two arrays were set up in trigger mode (STA/LTA criterion), and were operated over five consecutive days. Recording at the two arrays was achieved using PC-based data loggers, each acquiring eight data channels at 200 samples per second with a dynamic range of 16 bits. Time synchronization was achieved using GPS timing at each data logger, reaching a precision higher than the sampling interval. Since the different

![Seismograms and spectra](Image)

Fig. 3. Vertical-component seismograms for five explosion quakes recorded at stations E1 and A0 of the Ginostra (left) and Labronzo (right) arrays, respectively. The two bottom panels show the average spectra. The lower amplitude at Labronzo for frequencies lower than 3 Hz, due to the different amplitude response of the sensors used in that array, is evident.
data loggers operated independently, some of the triggers were partial. In this work we use a set of 32 explosion quakes recorded at all the data loggers of both arrays.

2.1. Seismic signatures of the explosion quakes recorded at Stromboli

The characteristics of the explosion quakes recorded during the experiment closely resemble those extensively reported by many previous studies (see references in Section 1). The explosive signals observed at frequencies greater than 0.5 Hz have emergent onset and are spindle-shaped, with a duration which rarely exceeds 1 min. The first 2–3 s of the signal are dominated by frequencies between 0.5 and 2 Hz, while the successive phases show a broader frequency content (Fig. 3). An acoustic-generated phase (Ripepe et al., 1996), delayed by about 5 s with respect to the signal’s onset, depicts an even broader frequency content which may exceed 10 Hz. In some cases the first onset, although never sharp, is visible despite the high-amplitude background tremor whose predominant frequency is similar to that of the signal.

For the first 2 s of signal, particle motion patterns evidence a short window characterized by polarization oriented toward the craters, while the following larger-amplitude phases depict transverse motion. An example is shown in the lowermost left panel of Fig. 4, where the longitudinal motion is characterized by low amplitude and a duration of a few tenths of a second, and is rapidly followed by transverse motion having a larger amplitude. The low apparent velocity deduced from the value of slowness around the first onset of the event (always lower than 2 km/s, see

![Graphical representation](image)

Fig. 4. Top panel shows 5 s of the three seismograms recorded by station E1 (vertical, radial and transverse) filtered in the 0.5–1.5-Hz band for the beginning of event 2570110. Particle motions of the signal contained in the shadowed window are plotted in the lower panels. In the first one, where vertical motion is plotted against radial motion, we can see the very short longitudinal motion characterized by high incidence angle (evidenced by the dashed line), followed by a higher-amplitude transverse motion. The last two particle motion plots (vertical–transverse and transverse–radial) indicate that both SV and SH components are present in the wavefield.
Section 2.2) indicates a shallow source. Particle motion at the signal onset depicts the predominance of transverse waves with the highest amplitude on the vertical component (Fig. 4, left bottom panel). A pure P wave phase is not observable in the frequency band of analysis because the S–P time would be always smaller than the wave period for the investigated frequencies, due to the short array–source distance. In synthesis, particle motion analysis for the explosion quakes recorded at Stromboli shows a predominance of transverse motion, and a clear S–P time cannot be observed. The origin of transverse motion can be attributed to a propagation effect associated with conversion at the interfaces along the ray path. However, near-field observations (Chouet et al., 1997) seem to indicate that most of the transverse wave is radiated directly by the source.

After the first 2–3 s of signal, the decrease in the apparent velocity indicates a dominance of surface waves. Thus we can conclude that the explosion quakes analyzed in the present paper show most body-wave energy (longitudinal and transversal) in the first 2 s after the first onset, and that most of this energy is associated with transverse (SV, SH) motion. These observations are consistent with others reported in several previous works (see e.g. Neuberg et al., 1994; Wassermann, 1997; Ereditato and Luongo, 1997; Chouet et al., 1997).

2.2. Array analysis

Explosion quakes recorded by all the stations composing the two arrays with a good signal to noise ratio have been analyzed using the zero-lag cross-correlation technique (ZLCC) (see Del Pezzo et al., 1997; Saccorotti and Del Pezzo, 2000) in order to estimate slowness and back-azimuth of the well correlated phases. The ZLCC is a time-domain technique, which is more adequate for short-duration seismic phases, allowing for an exact tailoring of the window length to fit the dominant period of the signal. The analysis was carried out over two frequency bands (0.5–1.5 Hz, 1.5–2.5 Hz) containing most of the energy which characterizes the signal onset, as shown in Fig. 3, where the power spectra averaged over five events at stations E1 and A0 (respectively of Ginostra and Labronzo arrays) are shown as an example. Spectra were computed over a 40-s-long window starting a few seconds before signal onset. After filtering the signals in the two frequency bands, the ZLCC technique was applied to successive windows having lengths of 1.5 s and 1.0 s for the 0.5–1.5-Hz and 1.5–2.5-Hz frequency bands, respectively. These window lengths were determined after trying several values and observing the dependence of the results. We chose the shorter window length which guarantees stable results and is greater than the wavelength corresponding to the middle of the frequency band. The window of analysis was then shifted along the array signals with 90% overlap.

A source–array distance of the order of one wavelength for the lower-frequency band analyzed would suggest the use of a ‘spherical wave array method’, as in Wassermann (1997), Almendros et al. (1999) or Wassermann and Ohrnberger (2001), to estimate slowness and back-azimuth of the wavefield. However, given the array dimensions, the travel time differences at the stations between plane and spherical waves for a source located in the crater area would be smaller than the sampling rate, and hence negligible in our analysis. For this reason we applied the conventional ZLCC method under the plane wave approximation.

The array analysis applied to data from vertical-component sensors is evidently appropriate to investigate the nature of seismic wave packets having considerable amplitudes in the vertical component of the ground motion. Therefore, only P, SV and Rayleigh waves are observable, while Love and SH waves are excluded, due to their horizontal predominant motion. In Figs. 5 and 6 an example of results given by the ZLCC technique applied to event 2570110 is shown. The analysis was applied to both Ginostra and Labronzo arrays, and results are represented by full and open circles respectively. The first phases of the event, between 12 and 13 s in the current time scale, are characterized by slowness values around 0.6–0.7 s/km. The corresponding apparent velocity, on the order of 1.5–1.6 km/s, is compatible
with S waves impinging with a large incidence angle.

The slightly increasing trend of slownesses for increasing lapse times allows the identification of surface waves (Rayleigh) on the basis of the following consideration. For a given frequency, the ratios between compressional and Rayleigh wave velocity and between shear and Rayleigh wave velocities are on the order of 2 and 1.2, respectively. Then, if the beginning of the explosion quake were composed of a pure P wave, we should observe an apparent velocity value at least

Fig. 5. Results of the analysis with the ZLCC technique of explosion quake 2570110 for the 0.5–1.5-Hz frequency band. From top to bottom, the first and second panels show the seismograms recorded by station A0 (Labronzo array) and E1 (Ginostra array), respectively. The shaded area includes the portion of signal analyzed using 1.5-s windows sliding along the seismograms with 90% overlap. For the unfiltered signals see Fig. 1, first seismograms at the top. The following three plots show correlation coefficient, slowness and back-azimuth of the wavefield versus time. Results from Ginostra and Labronzo arrays are represented by full and open circles respectively, whose abscissa is the middle window time.
double or greater (depending on the incidence angle of the P wave) with respect to the following surface wave velocity. In contrast, as shown in Figs. 5 and 6, we observe that the decrease of apparent velocity (or the increase in the apparent slowness) is more compatible with an S-to-Rayleigh wave change.

As shown in Fig. 3, the average spectra computed at Ginostra stations depict more energy in the 1–2-Hz band than those computed for the

---

Fig. 6. Results of ZLCC analysis for the same explosion quake as in Fig. 5, but relative to the 1.5–2.5-Hz frequency band. In this case the sliding window adopted for the analysis has a duration of 1 s (200 samples). Results from Labronzo array (open circles) show an evident jump at 11.8 s, when a correlated wavefield coming from the crater area (the explosion quake signal) emerges from a poorly correlated (0.4) surface wave (deduced by the high slowness value) propagating from a different direction (230°).
same set of events at Labronzo, due to the different sensor response. As the starting phases of the explosion quakes are characterized by a predominant frequency lower than 2 Hz, the first arrival is consequently much better visible at Ginostra than at Labronzo, although it is never sharp. For the cases in which the onset time is too emergent to be visually measured, we fix the onset of the signal as the time at which the correlation values begin to increase showing contemporaneously a stable back-azimuth with values close to the array crater direction at both sites.

Once the start time is fixed, we select three windows of 300 samples, with 50% overlap for the 0.5–1.5-Hz frequency band and five windows of 200 samples with 50% overlap for the 1.5–2.5-Hz frequency band. For example, event 2570110 is considered to start at 11.5 s in the current time scale (corresponding to 01:10:34.5, see Figs. 5 and 6) with an uncertainty of some tenths of a second. For the location procedure, we use the slowness vector components and correlation coefficients calculated with the ZLCC technique for windows centered at 12.1, 12.85, 13.6 s (1 Hz) and 12.0, 12.5, 13.0, 13.5, 14.0 s (2 Hz). The difference in travel times at the two arrays, computed for sources in the crater area, is on the order of 0.1 s, as illustrated in the following sections. Therefore, the two synchronized 1-s-long signal windows at arrays Labronzo and Ginostra will contain at least 90% of signal coming from a common source. For the 1.5-s-duration signals filtered in the 0.5–1.5-Hz band, this percentage increases.

For the 32 events analyzed we selected 96 windows (three for each event) for the 0.5–1.5-Hz band, and 160 windows (five for each event) for the 1.5–2.5-Hz band. Histograms in Figs. 7 and 8 show the distributions of slowness components, back-azimuth and correlation coefficients obtained using the ZLCC analysis in these windows.

The statistics of these data are reported in Table 1. It is evident that data for the Ginostra array are better correlated, particularly at lower fre-

![Fig. 7. Histograms of experimental results obtained by the array analysis applied to data filtered in the 0.5–1.5-Hz frequency band. Three windows with 50% overlap were selected for any of the 32 events analyzed. The two distributions shown in each plot are distinguishable through the gray color and bold lines when they overlap each other.](image-url)
frequencies. Each distribution obtained for the Ginostra array is characterized by a standard deviation smaller than that of the distribution corresponding to the Labronzo array.

3. Location of the seismo-volcanic source

On the basis of the observations discussed in Section 2, we develop our location procedure making the following simplifying assumptions:

(1) The slowness data used for the source location analysis are mainly representative of S waves;

(2) Each individual slowness measurement is representative of a distinct source. Consequently, separate locations are calculated for each synchronous pair of slowness data obtained at the two arrays.

It must be pointed out that assumption 1 gives a lower bound to the source depth. In fact, if the S waves were generated by conversion at an interface along the path, the source location would be shallower than in the case of direct S wave. Since

<table>
<thead>
<tr>
<th>Array</th>
<th>Labronzo</th>
<th>Ginostra</th>
</tr>
</thead>
<tbody>
<tr>
<td>Frequency</td>
<td>1 Hz</td>
<td>2 Hz</td>
</tr>
<tr>
<td>$S_X$</td>
<td>$0.20 \pm 0.14$</td>
<td>$0.15 \pm 0.23$</td>
</tr>
<tr>
<td>$S_Y$</td>
<td>$0.64 \pm 0.14$</td>
<td>$0.67 \pm 0.23$</td>
</tr>
<tr>
<td>Slowness</td>
<td>$0.69 \pm 0.11$</td>
<td>$0.72 \pm 0.23$</td>
</tr>
<tr>
<td>Back-azimuth</td>
<td>$198^\circ \pm 14^\circ$</td>
<td>$193^\circ \pm 19^\circ$</td>
</tr>
<tr>
<td>Correlation</td>
<td>0.84</td>
<td>0.74</td>
</tr>
</tbody>
</table>
there is no evidence of strong velocity discontinu-
ties below the array sites (Chouet et al., 1998;
Petrosino et al., 2002), we assume that the signal
windows used for source location are mainly rep-
resentative of the direct S waves radiated by the
primary (magmatic) source.

3.1. Method

In our procedure the source location is repre-
sented by a probability distribution computed
over a regular, three-dimensional cartesian grid
encompassing almost the whole volume of the is-
land. The grid dimensions are $4 \times 4 \times 2$ km along
the $X$, $Y$ and $Z$ directions, respectively, where the
$X$-axis is oriented east–west, the $Y$-axis is oriented
north–south and the $Z$-axis is oriented normal to
the Earth’s surface and points upward. The origin
of the grid, with reference to the local coordinate
system of Fig. 1, is set at the point (0.4, $-0.8,$
$-1$), and the spacing among adjacent grid nodes
is 0.025 km in all directions.

The procedure starts from assuming a velocity
model (described in Section 3.2), which is used to
calculate the travel times from the arrays to all the
nodes of the grid. For this calculation, the actual
array elements are replaced by five grid points
surrounding the array center. This procedure is
justified by the fact that we observed a greater
numerical stability when the travel times were cal-
culated for sources located exactly on a grid node.

Travel times are computed using the finite dif-
ference code developed by Podvin and Lecomte
(1991), and adapted to the 3-D case by Lomax
et al. (2000). For both arrays the travel times
associated with each grid node are then converted
to slowness vectors using a conventional plane-
wave fitting. At the end of this procedure each
grid node has associated the two theoretical hor-
izontal slowness vectors which would be observed
at the two arrays if the source were located ex-
actly at that node.

In the case that the errors in the estimate of the
two components of either the theoretical or ob-
served slowness vectors have a Gaussian distribu-
tion, the misfit between the predicted and ob-
served slownesses is written as:

$$P_A(x,y,z) \propto \exp\{-0.5[S_A^P - S_A^O]^T \text{Cov}(S_A^P) + \text{Cov}(S_A^O)]^{-1} [S_A^P - S_A^O]\} \quad (1)$$

where $S_A^P$ and $S_A^O$ are the predicted and ob-
served slowness vectors respectively, relative to
the $i$th measurement made at array $A$. A com-
prehensive assessment of the errors in the measure-
ment of slowness was made by Saccorotti and Del
Pezzo (2000), who estimated the complete proba-
bility function of the vector slowness $P(S)$ for six
events recorded at Ginostra array. We repeated
the procedure of Saccorotti and Del Pezzo (2000), and calculated the function $P(S)$ at the
Labronzo array for the same data set. Results
from this procedure show that:

1. the magnitude of the errors in the estimate
   of slowness vector is roughly the same for the
two components of horizontal slowness;
2. these uncertainties are constant for the dif-
   ferent events;
3. the uncertainties are the same at both ar-
   rays.

On the basis of these observations, the matrix
of errors associated with the estimate of predicted
and observed slowness vectors may thus be written
as $\text{Cov}(S_A^P) + \text{Cov}(S_A^O) = \sigma^2 I$, where $\sigma$
includes the uncertainties in the estimate of pre-
dicted and observed slowness vector components.
These values are taken equal to 0.01 s/km and
0.05 s/km for the experimental and theoretical er-
rors, respectively (Saccorotti et al., 1998; Sacco-
rotti and Del Pezzo, 2000).

Eq. 1 thus expresses the probability for the
source to be located at the node $(x,y,z)$ in terms
of the magnitude of the difference between the
theoretical and experimental slowness vectors at
a given array. As the slowness measurements at
the two arrays are independent, the compound
probability for that node is:

$$P_i(x,y,z) \propto P_{Gi}(x,y,z)P_{Li}(x,y,z) \quad (2)$$

where the subscripts G and L indicate the Ginoc-
tra and Labronzo arrays, respectively. Finally, the
overall probability for the source to be located at
node $(x,y,z)$ is given by stacking Eq. 2 over the
$N$ slowness measurements:
\[ P^{\text{Tot}}(x, y, z) = \sum_{i=1}^{N} (C_{L_i} + C_{G_i}) P_i(x, y, z) \]  

(3)

where \( N \) is the number of time windows (three or five, depending on the frequency of analysis) times the number of analyzed explosion quakes (32). In Eq. 3, the weights \( c_L \) and \( c_G \) correspond to the array-averaged correlation coefficients derived for individual windows of analysis at the Labronzo and Ginostra arrays, respectively.

The function \( P^{\text{Tot}}(x, y, z) \) is integrated over the 3-D gridded volume described above, and normalized so that:

\[ \int_V P^{\text{Tot}}(x, y, z) \, dx \, dy \, dz = 1 \]  

(4)

Fig. 9. Results of the array analysis applied to the synthetic seismograms used to simulate two sources characterized by different positions and times. The first pulse is relative to a source located at \( X = 2.6, Y = 1.0, Z = 0.4 \), while the second one is produced 2 s later by a source located at \( X = 2.8, Y = 0.5, Z = -0.2 \).
3.2. Velocity models

Recently, several studies have been aimed at determining the shallow velocity model in the northern and western flanks of Stromboli Island (e.g. Chouet et al., 1998; Petrosino et al., 1999, 2002). However, these studies report the velocity structure only for the shallower 300 m below the Labronzo and Ginostra sites, whereas the distribution of seismic wave velocities throughout the whole volcanic edifice remains unknown. Therefore, for our location procedure, we tried several reasonable velocity models compatible with the available information. Each model consists of a function describing the variation of velocity with depth and is applied to the whole grid by setting as zero depth the topographical surface of the island. To reduce the horizontal velocity gradient, the topography has been smoothed and the velocity gradient made dependent on the topographical elevation, in order to have the same velocity at a depth of 1 km below sea level, as can be seen in the vertical cross-sections shown in Figs. 11–14. Other kinds of model, such as a single layer over a half space (which in the present case means one cone shell over a cone with the axis in the crater area, to simulate the volcano edifice), were also adopted but rejected because they produce unrealistic source clustering at the interface and depth dependence on the velocity.

Fig. 10. Location of the synthetic sources using the data presented in Fig. 9. In these cross-sections the true source positions are indicated by crosses. The second maximum in the upper left plot and the two in the upper middle plot are cross-sections of the probability function not passing through the synthetic source positions, so they represent the probability function in the neighborhood of the sources.
3.3. Test with synthetics

To test the method, we used a synthetic signal given by:

\[
S(r, t) = A_{\tau} \left( t - t_0 + \Delta t_r \right)^b e^{-k(t-t_0+\Delta t_r)} \sin(2\pi f_0(t-t_0+\Delta t_r)) + A_{\tau} \left( t - t^* + \Delta t'_{r} \right)^b e^{-k(t-t^*+\Delta t'_{r})} \sin(2\pi f_0(t-t_0+\Delta t'_r))
\]

defined for the time \( t > 0 \) and for the distance \( r \), measured along the ray path from the source to the receiver. The values assigned to the constants are \( \tau = 0.005 \text{ s}, t_0 = 0 \text{ s}, t^* = 2 \text{ s}, b = 0.7, k = 2 \) and \( f_0 = 1.0 \text{ Hz} \). \( t^* - t_0 \) is the difference between the origin times of two sources located at two different grid nodes, indicated by the crosses in Fig. 10. Eq. 5 represents a wave packet which originates from two point sources. The signature of the synthetic signals, filtered in the 0.5–1.5-Hz band, are shown in the upper panels of Fig. 9 for two receivers located at Ginostra and Labronzo sites respectively. The propagation of the synthetic signal from the source point to the arrays is simulated by shifting the signals of the corresponding travel time, given the velocity structure. Both the arrays in this test were composed of five elements, as described previously for the estimate of the

![Fig. 11. Horizontal and vertical cross-sections of Stromboli Island with source location results for the 0.5–1.5-Hz band. The three top panels represent horizontal cross-sections at \( Z = 0.5, 0.3 \) and 0.1 km above sea level, as indicated by bold contour lines. The thin contour lines represent the probability for source location, which is given by integration of the spatial probability density function computed on the 3-D regular grid with 0.025-km steps. The two bottom panels show vertical cross-sections of the island. In these graphs the probability function is represented by contour lines and gray color. The velocity model is also shown by contour lines of constant velocity. The labels on these lines indicates the S wave velocity in km/s.](image-url)
theoretical slowness model, and are located in the same sites where the actual Ginostra and Labronzo arrays are positioned. The technique described in Section 3.2 was finally applied to the synthetic data set.

Results are reported in the plots of Figs. 9 and 10. As shown in Fig. 10 the locations of the two synthetic sources are well recovered, and the bias is practically negligible.

3.4. Results

The spatial probability density function (Eq. 3) for the real data obtained from the array analysis for both frequency bands (0.5–1.5 and 1.5–2.5 Hz) was computed using a suite of different velocity models. Reasonable and stable results were obtained using the subset of S wave velocity models characterized by values around 2.5 km/s at a depth of 1 km below sea level. In order to show the variability of the solutions we represent in Figs. 11–14 the results obtained using two different models. In the two bottom panels of these figures the velocity models are depicted by contour lines of constant S wave velocity (indicated in km/s). The model represented in Figs. 11 and 12 (and 9, being the same as used for the test) is assumed as the reference model, since it has associated the highest maximum value of the probability function for each individual event. In contrast, models with higher values of velocity, such as those shown in Figs. 13 and 14, have associated a large number of solutions located at the free surface and are characterized by probability values lower than those obtained with slower models.

In Figs. 11–14 the isolines of equal probability are plotted on horizontal and vertical cross-sections of the island. In all these figures the three
The top panels show horizontal cross-sections of the island at different elevations (0.5, 0.3 and 0.1 km above sea level), whereas the lower panels represent two vertical cross-sections crossing the model space in the crater area. In the horizontal cross-sections the bold line represents the topographic elevation, whereas the thin contour lines bound the areas of equal probability for source location. The most external contour, labeled 1, depicts the area where the probability density function is greater than zero, and represents the zone where the source can be located with 100% probability. Since the values of the isoline labels represent the probability integral extended to the volume marked by the isolines themselves, they decrease toward the inner zone, where the probability density function reaches its maximum value. Therefore, the maximum of the probability density function (the maximum probability per unit volume) corresponds to the minimum of the integrated function.

In the cross-sections obtained for the 0.5–1.5-Hz frequency band, the zone of maximum probability for source location is associated with a well-defined volume centered below the crater position. The point corresponding to the maximum probability may be viewed as a source centroid which for the 0.5–1.5-Hz frequency band is located about 500 m below the craters.

The cross-sections obtained for the wavefield at frequency 1.5–2.5 Hz produces a more complicated pattern with several relative maxima, one of which approximately coincides with the source centroid obtained for the 0.5–1.5-Hz band. The relative maxima of the probability function with a location different from that of the low-fre-
Fig. 14. Source location of the 1.5–2.5-Hz frequency band obtained using the same model as Fig. 13.

Fig. 15. Stromboli Island seen from the north, with the representation of the source located using 0.5–1.5-Hz results and the velocity model depicted in Figs. 11 and 12. The array positions and crater area are also indicated.
frequency centroid may be interpreted assuming that the energy at 1.5–2.5 Hz is radiated partly by the same source which radiates in the 0.5–1.5-Hz band and partly by secondary sources, as for example strong near-source scatterers.

The present result is fairly consistent with those obtained by other authors using different approaches. In particular Chouet et al. (1999) applied a combination of semblance and particle motion analysis to broad-band data, obtaining a source location slightly shifted toward the northwest with respect to that obtained in this study. More recently, Chouet et al. (2003) located the source centroids of two different types of explosion quakes, finding two slightly different locations, both shifted 80 m in the northwest direction from the active vents at depths of 220 and 260 m, respectively.

4. Discussion and conclusions

Using double array techniques, we have confirmed previous results about the wave composition of the explosion quakes and have obtained the location of their source. Our results concluded a multiple experiment aimed at understanding the eruptive mechanism at Stromboli Volcano.

Our approach for determining source location is essentially probabilistic, in the sense that we searched for the points of a gridded model space for which the misfit among the observed and predicted slownesses is minimized. All the procedures used in the present paper were inspired by the paper of Saccorotti et al. (1998) in which a more rigorous, but essentially similar, method was applied to the data from a single array. An approach similar to the present has been also adopted by Almendros et al. (2001a,b) who used three arrays deployed on Kilauea volcano to map the tremor source.

The essentials of the method are described by Eq. 3, which represents the probability for source location as the difference between the measured and theoretical slowness vector. The compound probability is weighted for the array-averaged correlation coefficient (3), thus weighting most the well correlated wave packets detected by the arrays. The grid search approach allows a complete mapping of the error function over the whole volume investigated, thus permitting the detection of either the principal or secondary maxima of the probability pattern.

The multiple array source location technique applied in the present paper revealed to be well suitable for transient signals with a very emergent first arrival, like the explosion quakes recorded at Stromboli at a distance of approximately 1.7 km from the source. The main problems in its application were essentially two. The first is the complete ignorance of the velocity model at depths greater than 300 m below the surface. The second is related to the presence of well correlated noise added to the explosion quake signal (the volcanic tremor), which produces uncertainty in signal synchronization at the two arrays. The problems related to the ignorance of the velocity models were partially solved generating a suite of reasonable models on the basis of what has been observed on other volcanoes. The problems related to the presence of the volcanic tremor were overcome using considerations based on the results of the array analysis applied to the pre-event signal. We used the changes in the pattern of correlation coefficient as a function of time to pick the event onset at both arrays, and were able to identify the beginning of the events utilized for source location.

Despite the above-mentioned limitations, the method gives results characterized by high resolution and by a reasonably short computing time. Its characteristics can be usefully applied to volcano monitoring, for the easy identification of possible spatial migrations of the seismo-volcanic source.

Results indicate that the volume containing the source of the explosion quakes at Stromboli volcano is located beneath the crater area, and extends to depths not greater than 700 m below the surface. The horizontal extension, estimated at 80% probability (see isolines of Figs. 11–14 and the sketch of Fig. 15), is between 0.5 and 1 km. It is also noteworthy that locations obtained at 1 and 2 Hz are different, the 2-Hz solution giving a source multiplicity with respect to that obtained at 1 Hz. In the interpretation of Chouet et al.
the low-frequency, quasi-monochromatic onset of the explosion quakes would be associated with the resonance of a shallow (about 200 m) fluid-filled crack in response to the ascent and consequent bursting of a gas slug. Thus, the results we obtained at frequency 1 Hz may be taken as representative of this phenomenon. Conversely, the multiple solutions we obtained at frequency 2 Hz may be attributed to the process of magma ejection at the free surface, or to the action of near-source scatterers. These two latter hypotheses are not alternative, and both of them may concur in producing the complex pattern depicted in Figs. 12 and 14.

Acknowledgements

Roberto Carniel and Joachim Wassermann are gratefully acknowledged for their comments that greatly improved the quality of the paper. This work was supported by G.N.V. (Gruppo Nazionale per la Vulcanologia) of Italy’s C.N.R. The Stromboli 1997 array experiment would not have been possible without the participation and technical assistance of the University of Granada group (G. Alguacil, J. Almendros, E. Carmona, M. Abril), the CSIC group (R. Ortiz and A. Garcia) and the University of Salerno group (S. Petrosino, B. Grozea, G. Maritato and M. Simini). Javier Almendros is acknowledged for useful discussions.

References


