CHAPTER SIX

CONCLUSIONS AND FURTHER WORK

6.1 Methods

Two methods were used to determine the source parameters corresponding to VLP phases at Stromboli. First a simple technique was derived based on extrapolation of decay laws back to the magma/rock boundary. Separate models for a point source and a line source were derived. The second technique (the k-ω method) was more complex (and is presumed to be more accurate). It is based on an analytic solution of the wave equation in the k_z-ω domain. The problem is broken down into an infinite sum of plane waves travelling in all directions at all phase speeds which allows any source to be represented. Moving sources were modelled in addition to point sources and line sources by simulating the perturbations in pressure and shear stress at the conduit wall.

Both of these methods accounted for near-field effects. Problems that were ignored were heterogeneity, attenuation, tilt and complex conduit geometries. For the very-long-period phases studied at Stromboli, heterogeneity and attenuation do not matter since the seismic stations were only a few percent of a wavelength distant from the source. Tilt may be a problem for horizontal components but at worst it may double the signal ampltiudes observed [*Forbriger and Wielandt*, 1997]. Neither method was designed to deal with conduits that are not vertically oriented cylinders.

6.2 Application to VLP phases at Stromboli

The results of amplitude modelling applied to VLP phases recorded at Stromboli in 1995 [Fig. 6.1] suggest that the source of the VLP phases is either:

(i) a point source ~350 m beneath the vents which displaces a volume of ~600 m^3 [Fig. 6.1a],

(ii) a line source extending from the vents to a depth of 300 m [Fig. 6.1b], or

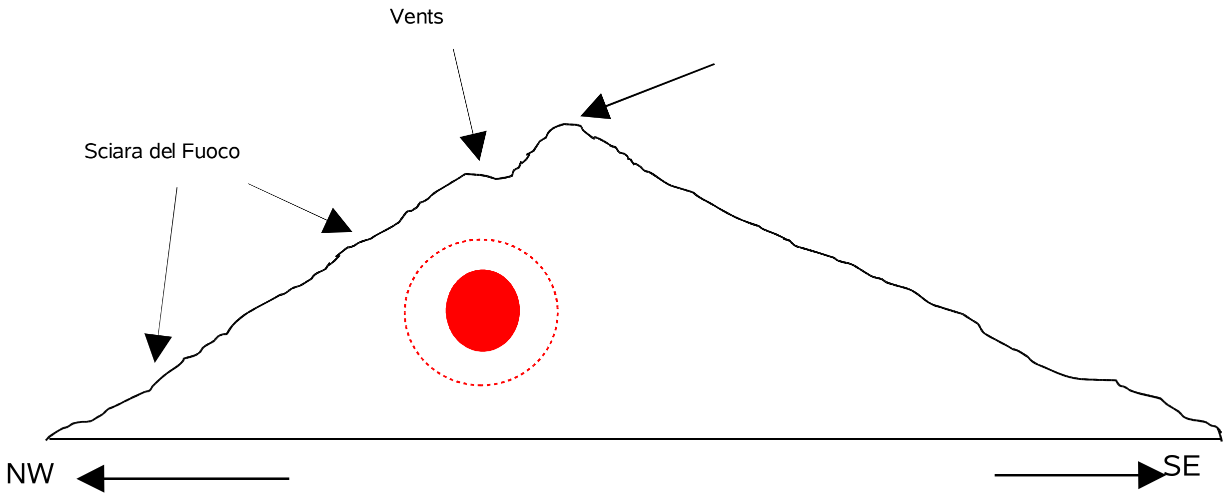
(iii) a line source extending from the vents to a depth of 700 m [Fig. 6.1c].

It is likely that the point source represents a spherical magma chamber centred on a depth of 350 m, and that line sources represent uniform pressure changes throughout cylindrical conduits. Assuming the maximum possible pressure change is 20 MPa the magma chamber must have a radius of at least 30 m, or the conduit must have an initial radius of at least 4 m. A better constraint is radial stress, which in Chapter 3 we assumed couldn't exceed 10^{-4} even for these transient signals. In that case it was found that the conduit radius must be at least 10 m, and the magma chamber radius must be at least 100 m.

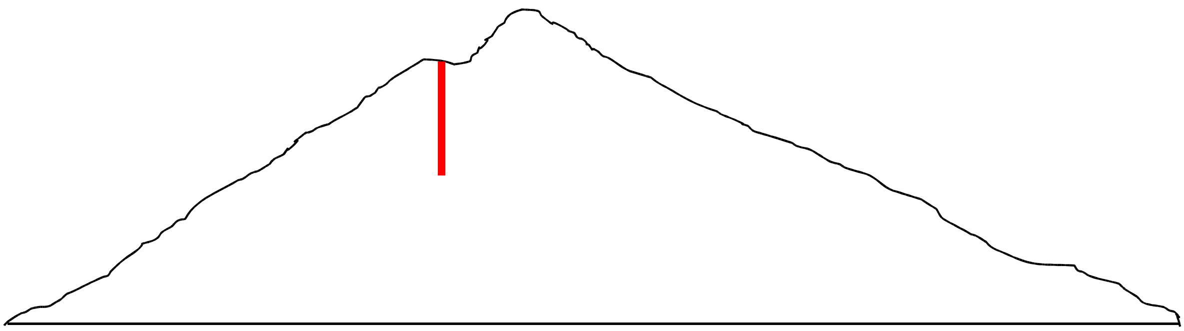
For a magma chamber with a radius of 100 m, a pressure change of 0.5 MPa produces signals with the same amplitude as the observed VLP signals. A pressure change of 0.5 MPa was also computed for a 100-m-radius spherical magma chamber using the simpler technique derived in Chapter 3 for VLP phases recorded in 1992.

For a conduit of length 700 m and radius 10 m, decay modelling modelling required a pressure change of 0.5 MPa to produce the VLP phases recorded in 1992. For k-ω modelling, a pressure change of 4 MPa is required to produce 1995 VLP phases, a factor of 8 larger. If we assume a radial strain of 10^{-4} corresponds to a pressure change of 0.5 MPa for both modelling methods, then the conduit radius would have to be 30 m [Fig. 5.10] which seems unrealistic.

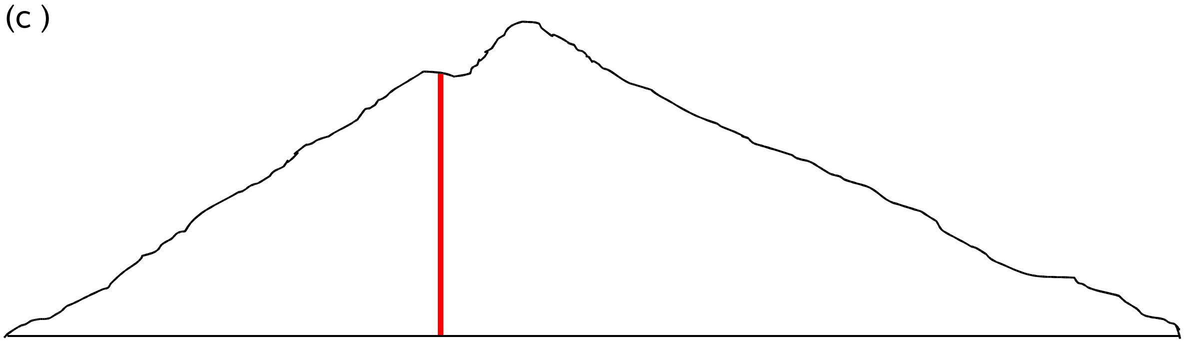
There is no evidence to support such a large conduit radius, and this favours the model given in Fig. 6.1a, which obviously must be connected to the vents by a narrow cylindrical conduit. This second order model is shown in Fig. 6.2a. However, the modelling in this thesis assumed the VLP seismic source (and therefore the chamber/conduit) to lie directly beneath the vents. *Chouet et al.* [1999], using semblance analysis [*Ohminato et al.,* 1998], found the source to lie at the same depth, but also 300 m to the north-west [Fig. 6.2b], directly beneath the Sciara del Fuoco.



(b)



(c)



(a) Pizzo Sopra la Fossa

Figure 6.1: First-order models indicated by results in this thesis, against a NW-SE cross-section through the vents. Only the part above sea-level is shown. (a) a point source centred on a depth of 350 m, indicating a pressure variations in a magma chamber with a radius of at least 100 m, possibly much greater. (b) a line source with a length of 300 m, indicating pressure variations in a conduit with a radius of at least 10 m. (c) a line source of length (at least) 700 m, again indicating a conduit with a radius of at least 10 m.

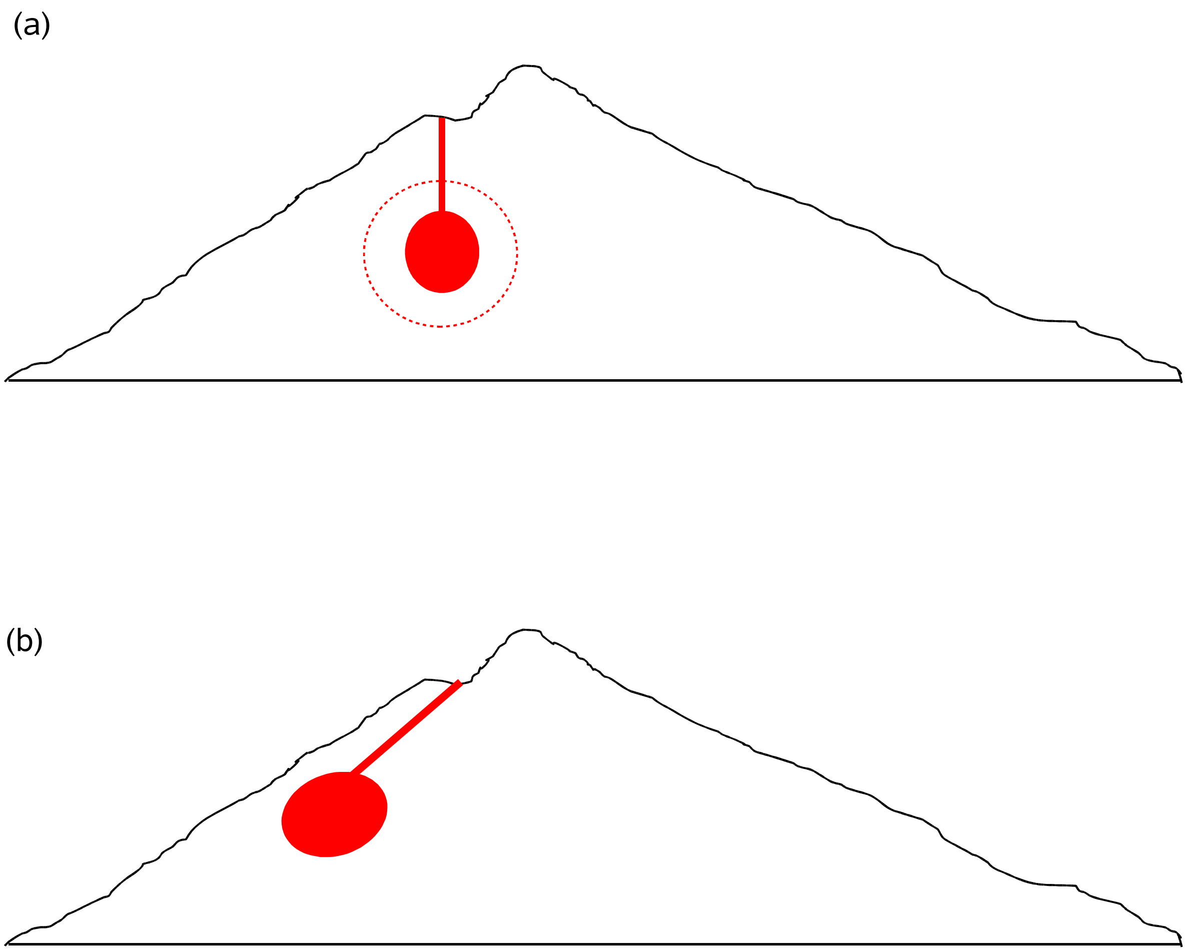


Figure 6.2: (a) The most probable model indicated by the results in this thesis. A magma chamber centred on a depth of 350 m beneath the vents, connected to the surface by a narrow conduit. Radial strains suggest the chamber radius must be at least 100 m, but may be much larger. (b) The methods used in this thesis assumed the VLP seismic source lay directly beneath the vents, but the results are in general agreement with those of *Chouet et al.* [1999], for which a model is shown here.

Given the model in Fig. 6.2a, it is interesting to think more deeply about how vent 1 eruptions may occur. A schematic model is given in Fig. 6.3. The pressure in this magmatic system probably increases prior to eruptions due to rising bubbles (advective overpressure). When the cap rock fails, a decompression wave travels down through the conduit, eventually reaching the chamber which acts as a strong seismic source because of its large radius.

To explain why the eruption isn't observed before the strong contractive seismic pulse it is only necessary to assume the base of the cap rock may be as much as 100 m beneath the vents. The eruption would be observed 1 or 2 seconds after the cap fails, which is ample time for the decompression wave to reach the magma chamber, and for the seismic pulse to be generated. Variations in depth of the cap rock or in the bubble density in the conduit would cause the relative times of the eruption and seismic pulse to vary.

The term 'cap rock' is used throughout this thesis for emphasis, but it is more realistic to think of a skin which forms (due to cooling) at the top of the magmatic column, where the magma is exposed to the air. Cooler magma has a higher viscosity and yield strength and therefore pressure can build beneath it. It is unlikely that there is sufficient cooling for this skin to solidify since the interval between eruptions at vent 1 can be as little as 3 minutes (and may be less). It would be interesting to investigate whether eruptions that occur after relatively short repose periods differ in their eruptive style from those following much longer eruption periods.

This model of vent 1 eruptions is consistent with thermal and degassing budgets, as discussed in Section 2.4. Eruptive activity at vent 3 is different because this vent remains open between eruptions, so the eruptions are probably due to bursting slugs at the air-magma interface [e.g. *Jaupart and Vergniolle*, 1988]. What cannot be explained is why vent 3 remains open while vent 1 becomes 'capped'. It seems unlikely that advective overpressure would be an efficient pressure building mechanism if some vents are open.

The k-ω modelling results can only be considered preliminary since (i) only 4 stations were used, (ii) only six VLP signals were analysed and (iii) the effects of a free surface may significantly alter the results. One process that can be completely disregarded though is the Bernoulli process, which at best can only produce pressure changes of the order of a few kPa at Stromboli. Both the modelling techniques used in this thesis support this conclusion.

Although the source of VLP phases at Stromboli was found to be either a point source or a line source, it is likely that at some point in the future seismic phases associated with rising magma will be identified, since the k-ω method has shown that they should be micrometres or even millimetres in size

6.3 Further work

The modelling techniques applied to VLP phases at Stromboli show that the source is either a point source or a line source. Modelling also demonstrated that for these types of sources, the waveforms recorded at each station (assuming a radially isotropic model) match the source function (this is not generally true for all source types, though it is often assumed). So given that we now know the source function, we need to explain it. In order to do this it will be necessary to model various two-phase fluid-flow processes, such as changes in density, velocity and viscosity that occur as bubbles rise and coalesce and as magma rises. If the pressure and shear stress perturbations at the conduit wall can be calculated, seismograms can be computed using the k-ω method.

Since magma rise is a slow process (modelled as a superposition of several terms which can be modelled as moving point sources, and expanding line sources), it is more likely to be applicable to phases of even longer duration than VLP signals. It may be that the k-ω method derived in this thesis is more applicable to deformation data (from GPS or EDM) rather than broadband seismic data, though it is not clear whether the volcanic processes discussed in this thesis could generate signals large enough to be detected by the relatively insensitive deformation techniques.

The k-ω method is currently implemented in Fortran, although most of the figures were produced using Matlab. It would be very convenient if the method were coded in Matlab also, as a GUI interface could be programmed, allowing users unfamiliar with the theory to select model parameters from a sequence of menus, displaying the results automatically.

The k-ω method should be compared against other techniques, such as boundary element modelling and finite element modelling. Indeed it would be useful to review all the modelling techniques in the literature that can be used for calculating synthetic volcano-seismograms, and apply each of these to Stromboli broadband seismic data. The goal of such a study would be to identify which methods are best for which purposes. The criteria would include accuracy of results, computational speed, ease of use, variety of sources that can be used, and degree of heterogeneity the method can deal with.

The k-ω method should also be developed to deal with variations in conduit radius and shape. It is likely that at depths of a few hundred metres the conduit has a rectangular (rather than circular) cross-section, which may explain why spectral peaks for volcanic tremor and long-period events vary with azimuth at some volcanoes [*McNutt*, pers. comm.]. A rectangular source is unlikely to be applicable to VLP phases at Stromboli because the wavefield is axially (azimuthally) isotropic. The condition that the conduit be oriented vertically is not really a restriction since an angled conduit can be simulated by rotating station positions and altering pressure gradients to account for the rotation of the gravity vector.

A thorough analysis of the 1995 data has yet to be performed. The following questions haven't been resolved yet:

1. Are there significant differences between the VLP phases recorded in 1992 and 1995?

2. Is there statistically meaningful evidence for deformation phases?

Analysis of a further dataset, collected in 1996, would help to resolve these questions. In addition to cross-correlation travel-time analysis and particle motion analysis, it would be of great benefit to compute moment tensors via waveform inversion, following the technique that *Ohminato et al*. [1998] applied to very-long-period phases at Kilaeua. They were able to determine that impulsive VLP signals (uncorrelated with eruptive activity) were associated with pulses of viscous magma pushed through a constriction in a horizontal crack. This method has recently been applied to VLP phases at Stromboli [Chouet et al., 1999] and revealed the seismic source to be located 300 m north-west and below the vents. It would be interesting to see if analysis of datasets obtained by Leeds University leads to similar results.

As broadband seismometers are deployed at more volcanoes, new VLP phases are likely to be discovered and there will doubtless be a need to pull together all these observations and uncover the variety of source mechanisms that can produce such signals. Sophisticated two-phase fluid modelling coupled to the k-ω method in combination with moment tensor waveform inversion promises to be a useful way to investigate the source of VLP signals, as well as other seismic phases such as long-period events, tornillos and tremor.

Finally, it is desirable to derive a version of the k-ω modelling method which incorporates the effects of the free surface, and use this to verify the results obtained in Chapter 5. The derivation would follow that given in Chapter 4 for the infinite solid model, except that the radiation condition could not be invoked to set two of the coefficients to zero. Instead, the necessary boundary conditions would come from setting the stress at the free surface to zero. This method has been attempted, but the results were not entirely successful. Radial displacements were found to resemble those found for the infinite solid model, varying smoothly with angle of the free surface. However, vertical displacements were found to be physically meaningless. One possible explanation is that it is physically meaningless to apply boundary conditions at the conduit wall and free surface simultaneously, since the wave encounters these boundaries at different times. If this free surface problem is resolved, the method is likely to receive more interest. If it cannot be resolved then it may be necessary to resort to finite element modelling in order to check the results of Chapter 5 and link volcano seismology and magma dynamics in the future.