Abstract

We applied a type of depth migration for converted seismic phases to active and passive seismic data sets from the Campi Flegrei volcanic region in southern Italy. The migration method is based on the diffraction summation migration technique. Travel times to grid points in a volume were calculated in smooth P and S-velocity models and trace energy near the calculated converted phase time was stacked over multiple sources at one receiver. Weighting factors based on Snell’s Law at an arbitrarily oriented local interface were applied to better focus trace energy. PP reflection images from the active data set provide increased detail to images of the caldera rim from other studies. The migrated images also show features near 2-3 km in depth beneath Pozzuoli city, which may be associated with an over-pressured gas volume, as suggested by other geophysical investigations. Possible deeper features near 4 km depth may be related to the presence of the carbonate basement or may image a previously undetected feature, such as a small body of strongly thermometamorphosed volcanic rock. The current passive earthquake data set from the 1984 ground uplift episode was not well suited to the converted phases analysis due to narrow P-S windows and high noise levels in the traces. However, two stations provide confirmation and extension of imaged features in the active data.
1. Introduction

The Campi Flegrei caldera is an active volcanic system lying west of Vesuvius volcano and the city of Naples in a densely populated area of southern Italy (Figure 1). It last erupted in 1538 (Rosi and Sbrana, 1987) but seismic, fumarolic, and ground deformation continue to the present day (De Natale et al., 1991; Avallone et al., 1999; Chiodini et al., 2001; Troise et al., 2007). In fact, ground deformation is more dramatic here than at any other volcano on Earth. The background long-term ground deformation at Campi Flegrei is subsidence at a rate of about 1.5-1.7 cm/yr (Troise et al., 2007), however, the 1538 eruption was preceded by 5-8 m of uplift (Dvorak and Gasparini, 1991; Dvorak and Mastrolorenzo, 1991). After the eruption subsidence resumed until the recent uplift period began in 1969. Since then there have been two episodes of rapid uplift with peak rates of 1 m/yr in 1969-1972 and in 1982-1984 (Barberi et al., 1984). Both were accompanied by increased microseismicity, though this increase above background levels was much stronger in 1982-1984. Ground uplift ultimately reached a maximum of 1.8 m recorded in the town of Pozzuoli, part of which was evacuated (Barberi et al., 1984). Fortunately, no eruption occurred and by January 1985 seismic activity had dropped off and slow subsidence again began, although the area still remains above the pre-1982 level. As of November 2004, uplift has again started, though at a considerably slower rate than in 1982-1984 (Troise et al., 2007), accompanied by swarms of microearthquakes at depths between 0.5 and 4 km (Saccorotti et al., 2007). These recent uplift episodes are thought to result from input of magmatic fluids from a shallow magma chamber into shallower aquifers (Troise et al., 2007).

The threat of future eruptions and/or destructive ground deformation has prompted a number of active and passive seismic studies in the Campi Flegrei region. Background seismicity at Campi Flegrei is very low with only a few events of magnitude >1 recorded per year, except when there is an uplift episode at which time thousands of such events can occur over a short period (Judenherc and Zollo, 2004). Using earthquake data from the 1982-1984 uplift episode, Aster and Meyer (1988) identified a low Vp, high Vp/Vs ratio volume centered near the city of Pozzuoli which was later associated with a low Qp anomaly (de Lorenzo et al., 2001). Based on modeling of P-S conversions in teleseismic data, Ferrucci et al. (1992) proposed the presence of a magma body at 4-5 km depth at the center of the caldera. De Natale et al. (1995) analyzed focal mechanisms of the 1982-1984 microearthquakes and interpreted them as characterizing a ring-shaped normal fault system inside the caldera. This model as well as several later models (e.g., Wohletz et al., 1999; Troise et al., 2003) assumed
the presence of a magma chamber like that proposed by Ferrucci et al. (1992), however, recent seismic tomography studies have found no evidence of a magma chamber within the upper ~6 km beneath the caldera (Zollo et al., 2003; Judenherc and Zollo, 2004; Vanorio et al., 2005; Chiarabba and Metelli, 2006; Battaglia et al., 2008). What they have found are remnants of the buried caldera rim both in the Bay of Pozzuoli at 1-3 km depth, as well as on the landward side of the caldera at 2 km depth. Bruno (2004) found evidence of several volcanic vents located along the inferred buried caldera rim spanning the Bay of Pozzuoli using legacy seismic reflection, gravity, and magnetic data. Evidence of over-pressured gas-bearing formations at a depth of about 2-4 km has been detected as a low $V_p/V_s$ anomaly beneath the caldera and the city of Pozzuoli (Chiarabba et al., 2006; Vanorio et al., 2005; Chiarabba and Moretti, 2006). A seismic reflection study using three-component ocean bottom seismometer data has confirmed the presence of a supercritical fluid-bearing formation at 3 km depth as well as low velocity zone 1 km thick at 7.5 km depth associated with a mid-crust, partial melting zone (Zollo et al., 2008). This magma sill lies at a depth similar to the proposed depth of the magma sill beneath Mt. Vesuvius (Auger et al., 2001) suggesting the possibility that the two volcanic centers are fed by a single magma reservoir (Zollo et al., 2008).

In September of 2001, an active marine seismic experiment called SERAPIS (Seismic Reflection/Refraction Acquisition Project for Imaging complex volcanic Structures) was carried out in the Bay of Naples and Bay of Pozzuoli in order to investigate the caldera structure and its feeding system (Zollo et al., 2003). During this experiment, ~5000 shots were produced by the vessel Le Nadir of Ifremer and recorded on an array of ocean bottom seismometers (OBS) and land stations. A subset of the SERAPIS active data set was combined with earthquake sources from the 1982-1984 uplift episode that occurred in the Campi Flegrei caldera to produce high-resolution 3-D tomographic P- and S-velocity models of the caldera (Battaglia et al., 2008). These models confirm the presence of the caldera rim imaged as a high P-velocity anomaly at about 1-1.5 km depth in the southern part of the Bay of Pozzuoli. A very high $V_p/V_s$ anomaly was found beneath the city of Pozzuoli at ~1 km depth, and a low $V_p/V_s$ anomaly was imaged at about 4 km depth below much of the caldera.

In this paper we investigate the subsurface structure of the Campi Flegrei caldera by applying a type of depth migration for converted phases to active as well as passive data sets for the Campi Flegrei area. This migration method has been previously applied to passive data sets in the Gulf of Corinth region (Latorre, 2004) and the Molise region in southern Italy (Latorre et al., 2008). The active and passive parts of the Battaglia et al. (2008) and...
SERAPIS data set were analyzed separately. The active data set provided good coverage across the caldera for both PP and PS reflected phases. However, the acquisition geometry of the passive data set was not well suited to the converted phases analysis presented here. Most stations lay almost directly above the earthquake hypocenters resulting in short P-S time windows. Although some possible converted phases could be detected in the traces, the small P-S windows and energetic S-arrivals limited the coverage to small areas around each station thus resulting in a very poor image. Therefore, we use the passive data only to confirm some of the features we find in the images from the active data set.

The converted phases migration method is well suited for regions with complex velocity structure like Campi Flegrei. Tomographic studies show “smooth” velocity variations obtained by only inverting P- and/or S-first arrival times (Aster and Meyer, 1988; Zollo et al., 2003; Judenherc and Zollo, 2004; Vanorio et al., 2005; Chiarabba and Moretti, 2006; Battaglia et al., 2008). By migrating converted reflected waves in depth we image impedance contrasts (high frequency structures) that provide increased detail of some subsurface structures and thereby complement as well as confirm the tomographic images. Although for large data sets like the one used in this study the computation cost is not trivial, it is yet less than that required by more sophisticated methods such as full-waveform tomography techniques (e.g., Ravaut et al., 2004). Moreover, the full-waveform tomography requires a seismic folding difficult to reach for each focal point of the medium, providing room for a simpler investigation such as the one we suggest. Sources of error associated with this method (e.g., effect of velocity model, grid size, and attenuation) are discussed in detail in section 3.

2. Data

We investigated two data sets from the Campi Flegrei region: an active marine seismic data set consisting of OBS and land station recordings of airgun sources, and a passive data set consisting of land station (temporary and permanent) recordings of microearthquake sources. The active seismic data used here is a subset of the SERAPIS experiment performed in September of 2001. Approximately 5000 shots were produced by an array of synchronized airguns; 12 16-liter airguns were used offshore and 6 in the Bay of Pozzuoli and near the shore. Receivers included 70 three-component OBS, 66 three-component land stations, and 18 vertical-component land stations. The OBS were equipped with 4.5 Hz
sensors and continuous recording devices, while almost all on-land stations were equipped with 1 Hz sensors. Within the Campi Flegrei area, shots were spaced 125 m apart although some lines were re-shot to attain a shot spacing close to 63 m. The SERAPIS experiment thus resulted in a density of coverage that approaches commercial exploration surveys and allows us to stack seismic traces. This makes the data set well suited to investigations of geologic hazards in the Campi Flegrei area.

For the converted phases analysis, we selected data from a subset of the SERAPIS data used by Battaglia et al. (2008) in a recent 3-D seismic tomography study of Pozzuoli Bay. We chose three-component stations that have at least 300 shots with P-wave first arrival picks. Thus, our active data set includes: 1528 shots, 33 OBS (41778 source-receiver pairs), and 13 3-component land stations (7484 source-receiver pairs). Selected stations and shots are shown in Figure 2.

We can identify as many as three coherent secondary phases at some receivers by visual inspection of traces (Figure 3). OBS receivers recorded a strong phase arriving just after or with the first arrival, as shown in selected traces for OBS O12 (phase A, Figure 3). Phase A is noticeable on all three components with higher frequency content in the horizontal components. We treat this phase as a PP reflection from a shallow interface. A second and in some cases third secondary phase can be discerned later in the traces at some receivers (phases B and C, Figure 3). These arrivals are visible in all three components but are often stronger on horizontal components in un-normalized traces. These later phases sometimes appear to have higher frequency content than the phase A, but the complexity of the signal makes interpretation difficult. We treat the phases B and C as being either PP or PS reflected phases from deeper interfaces. Records from land stations are more difficult to interpret due to higher noise levels but most stations show at least one coherent secondary phase in one or more component records.

We also examine passive data selected from that used by Battaglia et al. (2008) in their tomography study in order to complement the migration results from the active data. During the 1982-1984 uplift episode, the University of Wisconsin undertook a field experiment deploying a temporary network of 21 three-component digital short-period seismometers (Aster and Meyer, 1988). As part of the SERAPIS project, data from this experiment was compiled and re-picked. Battaglia et al. (2008) chose earthquakes with azimuth gaps in station coverage <180°, root mean square (RMS) of arrival time residuals <0.5 s, at least 3 S-picks, at least 4 P-picks, and a vertical location error <1km (determined by hypo71). In addition, we only considered stations with at least 90 events.
3. Methodology

The migration imaging technique is based on the diffraction summation method (e.g., Yilmaz, 2001). The target volume is broken into a regular 3-D grid. Each grid point is considered to represent a local arbitrarily oriented interface (all having a common orientation) where seismic reflection and transmission is assumed to occur. We then sum weighted trace energy along complex quasi-hyperbolic paths, whose shapes are determined by the velocity structure of the medium, and map the summed energy to the corresponding grid points in space in order to build the image. Only kinematic and geometrical aspects of the migration procedure are considered; we do not take geometric spreading or wavelet-shaping factors into account.

The background velocity model is assumed to be smooth enough that ray theory can be applied throughout the volume resulting in travel times that are consistent with arrival times in the trace data. Into this smooth background we introduce sharp oriented interfaces at grid points where we assume reflection and transmission of waves occurs. Alternatively, the grid points can be considered as point scatterers by not assuming an interface orientation at each grid point. The method works with all kinds of reflected waves with the assumption (common to all migration techniques) that they are primary reflections.

We analyze the data in receiver gathers. For each receiver, our migration procedure consists of three steps after data preparation: (1) Calculate travel times from each source and receiver to each image point, (2) Determine weighting coefficients based on Snell’s Law, and (3) Stack converted energy at each image point.

3.1 Travel time computation

Travel times are computed for each source-grid point and each grid point-receiver couple in the heterogeneous but smooth 3-D P- and S-wave velocity models as illustrated in Figure 4(a). At each grid point first arrival P- and S-waves are taken into account. First we use the finite difference solution of the eikonal equation in a fine grid to construct P and S first arrival travel time fields in the entire medium (Podvin and Lecomte, 1991). Then, we recompute more accurate travel times along the rays by tracing backwards from each source and each receiver to each grid point and then integrating the slowness field along the rays (Latorre et al., 2004).
3.2 Determine Snell’s Law weights

We consider a local oriented interface at each grid point and examine the rays hitting this surface for each source-receiver pair. The interface orientation is arbitrary and is the same for all grid points. The Snell’s Law weighting factor is calculated as the cosine of the angle $\theta$ between the computed grid point-receiver ray direction and the ray direction predicted by Snell’s Law (see Figure 4b). This weight is equal to one when Snell’s Law is perfectly satisfied and decreases towards zero as the calculated rays move farther away from the condition predicted by Snell’s Law. Reflections and transmissions are considered separately so that source-receiver pairs with the wrong ray geometry at the grid point (i.e., reflection-like geometry when we are considering transmissions and vice versa) are given a weight of zero. Then, only source-grid point-receiver paths with $\cos(\theta) \geq 0.5$ are considered. In addition, ray paths that resemble “upside down reflections” (i.e., rays reflecting from the underside of an interface) can also be down-weighted. Applying these weights to the traces focuses energy and mitigates the limited folding by smearing out “smiles” in the image.

3.3 Stack converted phase energy

For one receiver, we stack trace energy at each grid point from all sources with ray paths that satisfy the Snell’s Law requirement ($\cos(\theta) \geq 0.5$). Traces are first filtered, muted (if desired), and then normalized. Single component or three-component normalization can be performed. The maximum trace amplitude is selected from within a narrow window $t_{\text{win}}$ around the calculated converted phase time and Snell’s Law weight coefficients are applied (again, if desired). Note that if the Snell’s Law weights are not applied the image points are considered as point diffractors.

Weighted seismic energy is stacked over multiple sources by summing the amplitude squared. The final image for one receiver is then normalized by the maximum stacked trace energy in the model. In the images, each grid point is represented by a pixel whose color indicates the relative amount of focused seismic energy. High relative values relate to high energy focusing and can be interpreted as the most likely boundaries where observed phases may be generated.

To combine images from multiple stations we simply sum the energy at each grid point and divide the sum by the number of stations contributing a non-zero value to the sum. The stacked energy grid for each station is normalized by the maximum value in the grid before the sum over multiple stations is performed. Summing a large number of station...
images tends to blur out the image, therefore we select stations using a trial and error method while referring to individual station images to obtain the best image for each case.

Attenuation is an important factor, especially in volcanic areas such as Campi Flegrei. In our method, we do not take into account amplitude variations because we only consider kinematic aspects of the migration procedure. We stack seismic energy (amplitude squared) and normalize traces after appropriate muting before stacking. In this way we eliminate effects of source-receiver distance, restoring weak signals that may be due to attenuation, but we also do not allow for cancellation of wavelet phases along the migration ellipses. We thus perform a preserved amplitude migration where we preserve amplitude ratios along each trace (or each set of 3-component traces when 3-component normalization is used) rather than a true amplitude migration. Although this is a simplification, it allows us to ignore effects of unknown phase conversions at the water/seafloor boundary and mitigates poor knowledge of the source excitation and instrument calibration errors. In the future, it may be possible to analyze true amplitudes by including source and receiver response corrections and geometrical spreading calculations in the procedure. However, we find that useful results can still be obtained with our current kinematic migration scheme.

3.4 Application of Converted Phases Migration Method to Active Data

It is well known that the effectiveness of migration techniques depends on the velocity model. In our migration model, errors in the velocity model result in bad estimation of both converted wave travel times and Snell’s Law coefficients with, consequently, defocusing of converted wave energy. The velocity model from Battaglia et al. (2008) has been obtained using an expanded version of our data set and, more importantly, the same algorithm for travel time computation. It is thus the most appropriate velocity model for our study as it is kinematically consistent with our data. All other velocity models of the area (Aster and Meyer, 1988; Judenherc and Zollo, 2004; Vanorio et al., 2005; Chiarabba and Moretti, 2006) are not adequate for our study because they have been obtained using a different part of our data set (either active or passive data) and/or different travel time algorithms.

In the first step of our method, we calculate first arrival travel times for P- and S-waves in the 3-D tomographic models of Battaglia et al. (2008). The target volume dimensions are 16 x 16 km horizontally, extending from zero to 7 km in depth; horizontal area of coverage is shown in Figure 2 by the bold box. Grid points are spaced 500 m in the X
and Y directions, and 200 m in the Z direction. Airguns such as those used in the SERAPIS experiment produce peak energy around 7 to 10 Hz (e.g. Kramer et al., 1968). For an average P-wave velocity of 3500 m/s in the upper 3 km of the model, the dominant wavelength of our sources in this part of the model is ~350-500 m. Because the number of receivers and sources constituted a large data set for analysis by our method with significant processing times, we used a horizontal grid size that is on the large side at 500 m. Our vertical grid spacing of 200 meters should allow us to define depths of features in our images without significant aliasing effects. We compared migrated images at this grid spacing to images obtained for a smaller grid volume (X-Y-Z = 10 by 9 by 6 km) with grid spacing of 250 m in X and Y, and 200 m in Z. In Figure 5, the images show the same focusing of energy at the same locations in the model, although the image is blurred slightly when the grid size is larger.

We apply Snell’s Law weighting coefficients for a horizontal local interface to the grid points as it is the simplest geometry and we do not assume that we know a priori the orientation of subsurface reflectors. For a more detailed discussion of the sensitivity of the resulting image to the orientation of the local interfaces see Latorre et al. (2008). We note here that the assumption of horizontal interfaces is a good compromise when the orientation of subsurface features is not known in advance. As for classical migration methods, we do not expect to be able to image steeply dipping interfaces (greater than 45° dip). Despite the relatively large number of sources and receivers in our data set, we find that treating grid points as point scatterers does not result in a clear image and that the Snell’s Law weights are needed to focus the image (Figure 6). Note that in the image for a single station in Figure 6 we see nearly concentric elliptical bands of energy in the image without the Snell’s Law weights. This is due to migration ellipses from multiple sources overlapping over broad surfaces and is a result of our acquisition geometry. When the Snell’s Law weights are applied we remove much of this bias and greatly improve the focusing of the image. In the summed image for multiple stations in Figure 6 we see brighter energy in the image without the weights applied. This is because the images are normalized by the greatest energy in the entire model space. When the weights are applied the summed image shows the greatest focusing of energy in a different part of the model than is shown in the selected slice. When applying the Snell’s Law weights we also remove energy resulting from “upside down” reflections. This kind of energy can be seen in Figures 5 and 6 as the yellow image points just below the receivers. In these images, the additional reduction in the Snell’s Law weights for rays with upside down reflection geometry has not been applied.
Traces for all OBS were filtered using a zero phase shift Butterworth band filter with a corner frequencies of 5 and 35 Hz. Some land stations were noisier than others, so band pass corner frequencies were chosen individually for each land station (range of lower corner: 1-10 Hz, range of upper corner: 20-40 Hz).

Latorre et al. (2008) showed that the choice of the trace window length $t_{\text{win}}$ can influence the final image. Thus, we investigated different values of $t_{\text{win}}$ in order to determine the optimal choice for our data set. Sample traces showing the trace window around the calculated converted phase time are shown in Figure 7. A very small $t_{\text{win}}$ of $\pm 0.004$ s corresponds to taking a window over $\pm dt$ for the OBS data, where $dt$ is the trace sample spacing (land stations had greater $dt$ of 0.008 or 0.01 s). For $t_{\text{win}} = \pm 0.004$ the sample trace shows that we have missed the maximum amplitude of the phase. However, for $t_{\text{win}}$ of $\pm 0.05$ and $\pm 0.1$ s we see that we have captured the maximum amplitude of the phase in our window.

The use of a small window length implies the ideal condition in which computed travel times are perfectly consistent with background velocity models and source and receiver locations. In reality, this is never the case. In the velocity model used in our study (Battaglia et al., 2008), residual times have an RMS of 0.07 s, which indicates some discrepancy between the model and P- and S-arrival times. Therefore, we consider $t_{\text{win}} \leq \pm 0.07$ s suitable for taking into account discrepancies of the data with respect to the velocity models. Based on this, we choose $t_{\text{win}} = \pm 0.05$ s for our analysis.

After the converted phases migration analysis has been completed for each station we combine individual station images to obtain a summed image including results from as many stations as possible. Our procedure for choosing which station images to sum together is based strictly on a trial and error method with qualitative image assessment done “by eye”. A great number of different combinations of station images was tried until a set was found that showed the most “real” features (features near the center of the model volume where rays were densest and which were surrounded by areas of markedly weaker focused energy) and the least edge artifacts. This time consuming trial and error method was necessary because there was no progressive improvement in image quality as the number of stations used to construct the image increased even though we normalized each station image before summing. Some individual stations had unexpected effects on the summed image causing an overall decrease in image contrast or loss of “real” features imaged by several stations in favor of a bright feature imaged by only one station. In addition, some station images had very bright edge artifacts located just inside the region of non-zero energy focusing. Station
O30 in Figure 6 with Snell’s Law weights applied illustrates a mild case of these artifacts in
the two orange boxes at the southernmost edge of the central imaged region (non-gray
region). Thus, each summed image shown here is the final choice selected after looking at
tens of possible station combinations. In the future, it may be possible to augment a more
automated and quantitative method of image quality assessment that will speed this process
by using image analysis tools.

4. Results and Discussion

The acquisition geometry for the active experiment with sources above the ground
surface (just below the water surface) is likely to result in strong reflection phases and weak
or unnoticeable transmitted phases from rays that turn at depth and travel back upwards
through the medium. Thus, we focus our investigation on PP and PS reflected phases. We
consider the phase A (Figure 3) as a PP reflection and analyze it applying no mute to the
traces. We then focus on the phases B and C by applying a mute to remove the portion of the
trace up to and including phase A (i.e., amplitude before the green line in Figure 3 is zeroed).

4.1 Results for phase A

Using a selection of stations to produce a summed image for PP reflections without
muting traces, we are able to image what has been interpreted in other studies as part of the
caldera rim in the southern part of the Bay of Pozzuoli (near Y = -1 km in Figure 8). This
feature is captured in the P- and S-velocity models of Battaglia et al. (2008) used in our
study. In their models it appears as a high velocity feature at depths of 1-1.5 km. Our X-Y
and X-Z images in Figure 8 show PP energy focusing over depths of 1.2 to 2.0 km in the
same location as the high velocity features found in tomography models. Vertical X-Z slices
(bottom of Figure 8) show that we can image this feature well; there is a distinct drop in
focusing below ~2 km above the broad deeper feature.

4.2 Results for phases B and C

PP reflection images for individual stations when phase A was muted out showed
features at a variety of depths, but most stations showed focused energy near 2.5 km and/or 4
km depth. We selected 23 stations to produce the summed image in Figure 9. The shallow
feature near 2.5 km depth is most clear in the northern part of the Bay of Pozzuoli at Y = 2
km (highlighted with an arrow in Figure 9). Further south at Y = 1 km, a deeper feature is visible farther to the west near X = -1.5 km. Near the mouth of the bay at Y = -1.5 km, two shallow features at the east and west sides of the bay can be discerned. The summed image of PS reflections from a set of 25 selected stations is shown in Figure 10. PS reflection images showed focusing of energy near 2 km depth in many individual station images. Some stations showed energy focused deeper, near 3.5 to 5 km depth. The summed image (Figure 10) primarily shows energy focused near 2 km depth across most of the caldera in the Bay of Pozzuoli.

There is some overlap between areas of focused energy in the PP and PS reflection images, although the brightest areas in the PS image tend to lie at slightly shallower depth compared to the PP image. However, a visual comparison of Figures 9 and 10 suggests that the later secondary phases B and (possibly) C are likely to be PP reflections. The focusing of energy in the image considering PP reflections is considerably less blurred and shows more localized features compared to the case considering PS reflections. It is possible that the airgun sources (purely a pressure wave in the water) tend to produce very weak PS converted waves in the subsurface resulting in PP reflections being the dominant signal in the traces.

4.3 Discussion

The converted phases migration method can be considered as an intermediate step between 3D tomographic inversion of first P- and S-arrivals and full waveform inversion. In our migration approach, travel time computation is performed using 3D velocity models that have been previously computed by Battaglia et al. (2008). This is a step up in complexity from many migration techniques that are applied using 1D velocity models (e.g. Louie et al., 2002), but it is necessary considering that we are working with wide-angle data in an area with complex upper crustal structure. The result is an increase in the required computing time, but also an increase in the accuracy of the calculated travel times. Using wide-angle data does limit the fold at any one image point compared to multichannel seismic profiling, however, deploying an industry-style multi-streamer swath for 3D imaging (typically about 4.5 km long and ~400 m wide) in a small bay like the Bay of Pozzuoli would simply not be possible. The converted phases migration method also benefits from the flexibility of being able to use both active and passive seismic data. Though at the moment only local data can be used, in the future it may be possible to extend the method to be used with regional and teleseismic data.
In addition to requiring compatible 3D P- and S-velocity models, our migration technique is subject to some of the same kinds of limitations as other seismic methods. The acquisition geometry is always an important factor and determines the areas that can be adequately lit by the seismic rays. In active seismic experiments somewhat more control can be exercised over the placement of sources and receivers as well as the source strength, although in heavily urbanized areas this may not be the case. When active sources are used at the surface, the depth that can be imaged is limited by the depth of penetration of the sources. In the passive source case, the best chance for imaging will come from transmitted converted waves traveling upward from the sources to the receivers; the depth of the hypocenters will thus control the depth of imaging. Noise is also a limiting factor, particularly when working with weaker secondary signals. Many of the land stations at Campi Flegrei were too noisy to permit the identification of any reflected phases, thereby limiting the extent of coverage of the images. We note, however, that it is not necessary to strictly identify the secondary phase(s) in order to apply this migration method. When a phase can be detected in the traces but not positively identified as, say, a PP reflection or a PS reflection, then as we have done in this study, imaging can be attempted for each expected phase and the quality of energy focusing in the images can be compared to determine the most likely phase type. At this time, the comparison of images must be done in a qualitative “by eye” method but a future improvement could include automated image quality analysis, which would also aid in choosing groups of receiver images to sum together.

Where there are strong, positively identified secondary phases (in active or passive three-component seismograms) a clear image can be obtained by this method that provides a useful complement to smooth velocity models from tomography (Latorre et al., 2008) with a smaller computational effort compared to full waveform inversion. In the Campi Flegrei case, we see that even when the secondary phases are not as sharp and easy to identify we can still obtain a useful image that can confirm the presence of features detected by other means (Bruno, 2004; Judenherc and Zollo, 2004; Vanorio et al. 2005) and provide new details.

The migrated images of PP reflections contribute considerable detail to the caldera rim structure based on gravity, well logs and tomographic inversions of first arrival times. We image the caldera rim at two depths. When considering phase A we can image part of the rim at a shallow depth of ~1.2-2 km spanning the mouth of the Bay of Pozzuoli (Figure 8). When phase A is muted out we can image a deeper portion of the caldera rim around 2.4-2.6 km depth (Figures 9, 11). Tomographic images do not capture these two distinct
horizons; they image sections of a ring-shaped region of high velocities over a depth range of 
~0.8-3 km (Zollo et al., 2003; Judenherc and Zollo, 2004; Vanorio et al., 2005; Chiarabba and Moretti, 2006; Battaglia et al., 2008). Gravity studies provide maps of the aerial extent 
of the dense volcanic and metavolcanic materials that make up the caldera rim (AGIP, 1987; 
Florio et al., 1999; Zollo et al., 2003; Judenherc and Zollo, 2004), and the lack of depth 
resolution can be mitigated by tying gravity models into well log data (e.g., AGIP, 1987; see 
Figure 1). However, the resulting models are necessarily simplified and well data that can 
control the depth of the layers is sparse. The migrated image in Figure 8 shows that the 
interface defining the upper portion of the caldera rim lies at a shallower depth in the center 
of the bay opening (~1.2 km) compared to the edges towards Miseno (~1.6 km) and Nisida 
(~2.2 km). This interface could correspond to that between the tuffs and hydrothermalized 
lavas proposed by AGIP (1987) based on gravity and well logs (Figure 1). Similarly, the 
lower reflector of the caldera rim that we image when phase A is muted out (Figures 9, 11) 
could correspond to the interface between the hydrothermalized lavas and deeper 
thermometamorphic rocks (Figure 1). The shallower features along the caldera rim could 
also be related to volcanic vents detected by Bruno (2004). Bruno (2004) detected four 
features in multichannel seismic profiles interpreted to be volcanic vents that formed when 
the caldera floor collapsed downward in a piston-like action and caused lava to be squeezed 
out along the ring faults around the caldera rim. Three of these events can be correlated to 
the interfaces we image in Figure 8. Bruno’s (2004) vent β, the shallowest (depths in two 
way travel time, TWT), lies towards the center of the mouth of the Bay of Pozzuoli, although 
~1.5 km south of the feature we image in Figure 8. Vent δ lies just east of Miseno at a 
slightly greater TWT and vent α lies just to the southwest of Nisida at about the same TWT. 
The TWT to the tops of these vents in Bruno’s (2004) study can be converted to approximate 
depths considering an average seismic velocity in the upper 1.5 km of Battaglia et al.’s 
(2008) P-velocity model (3600 m/s) and taking into account the water layer. This yields 
approximate depths of ~1.1 km for vent β and ~1.7 km for vents α and δ which agree 
reasonably well with the depths of the features imaged in Figure 8.

We detect features at ~2 km depth beneath the town of Pozzuoli in PP reflection 
images when phase A is muted out (Figure 9, 12). The reflector is identified in Figure 12a 
below the town of Pozzuoli, where the maximum of the two ground uplift episodes that 
ocurred in 1970-1972 and 1982-1984 were localized (see Figure 1). The depth of the 
reflector is located where a smooth vertical change occurs from the high \( V_p/V_s \) toward the
low $V_p/V_s$ anomaly shown in the Chiarabba and Moretti (2006) model (Figure 12b). In their study, Chiarabba and Moretti (2006) interpret the low $V_p/V_s$ anomaly located between 2 and 4 km depth as a rock volume filled by over-pressured gas. Thus, the seismic horizon indicates that the vertical velocity change is not gradual as indicated by first-arrival time tomography, but it is an impedance contrast that separates the two different bodies. However, the depth of the interface in our model is shallower than other studies suggest for the top of the caprock over the overpressured gas-saturated volume. New results from a study of seismic amplitude variation with distance confirm the presence of a gas- or water-bearing rock formation beneath the caldera in the Bay of Pozzuoli and place its top at 2.7 km depth (Zollo et al., 2008). A recent 3D inversion of gravity at Campi Flegrei indicates that the low-density caldera fill material reaches down to at least 2.4 km depth (Russo, 2007) in the Bay of Pozzuoli. 2D and 3D modeling of the ground displacement at Campi Flegrei yield best fits to the displacement pattern when the source is placed 2.5 to 3 km below the ground surface (Russo, 2007). This suggests that the caprock of the gas-bearing formation lies near the bottom of the caldera fill, possibly coinciding with the transition from caldera fill to more coherent rock at the bottom of the caldera at ~2.4-2.5 km depth. Vanorio et al. (2005) image a shallow high $V_p/V_s$ anomaly and a deeper low $V_p/V_s$ anomaly beneath Pozzuoli with the transition between them occurring over depths of ~1.5-3 km. Based on rock physics modeling, they interpret the deeper, low $V_p/V_s$ anomaly to represent overpressured gas-bearing rocks at supercritical conditions. Following Aster and Meyer (1988), the shallower high $V_p/V_s$ anomaly was interpreted as rocks containing brine likely due to steam condensation at the lower temperatures measured at those depths by AGIP (1987). These results agree with a placement of the caprock between 2-3 km depth with our model providing the shallowest estimate. We note, however, that seismic tomography methods tend to blur features in depth and thus provide an average depth for regions of anomalously low or high $V_p/V_s$. It may be that we are imaging the topmost surface of the caprock where the impedance contrast is high.

A deeper feature can be tentatively identified in our migrated images lying near 4 km depth (Figure 9). This could be related to the carbonate basement detected by tomography at approximately the same depth (Judenherc and Zollo, 2004), however we note that the Judenherc and Zollo (2004) model did not have good resolution at this depth to constrain this interpretation. It is important to note that although first arrival tomography methods are limited in the depth that can be resolved, when we work with reflected waves we can investigate much deeper structures (e.g., Zollo et al., 2008). As the carbonate basement is
believed to be continuous beneath this region, we must speculate that the impedance contrast is low and/or there is a limitation in our data set (e.g., low signal to noise ratio at these depths) which prevents us from imaging this interface in other areas of our model. A more likely alternative interpretation is that we are imaging part of the thermometamorphic complex described by AGIP (1987, Figure 1). The interface we see is certainly weaker and appears smeared in depth compared to other features in our images (Figure 9), thus it is likely that we are imaging the top of a small body of more strongly metamorphosed volcanic rock (possibly due to a local fracture which allowed magmatic liquids to reach this locale) or a small, possibly older, vent like those detected by Bruno (2004).

Figure 11 shows that the passive data are independent data that can be used to complement and constrain the results from the active data where the coverage of the two data sets lie close to each other, although the area of coverage resulting from the passive data is much smaller and the regions of coverage from each station overlap only over very small areas. When we consider PS transmitted phases generated by the earthquakes, station W04 images a bright interface at 2.4 km to the west of the S. Vito 8 well (Figure 1). As reported by Bruno (2004), deposits recovered in the well at shallow depths are mainly composed of tuffs and lavas, while all the rock samples from 2.350 km to 2.868 km (the well bottom) are composed of strongly thermometamorphosed lavas. Thus, the bright interface at 2.4 km is perfectly compatible with the well data and corresponds to the top of the thermometamorphic body, also in agreement with the AGIP model (1987, Figure 1). Unfortunately, except for station W14, other passive stations did not provide additional confirmation of the active results. Station W14 did provide an image of a feature centered at ~3 km depth just to the north of Nisida, which may correspond to part of the thermometamorphic volcanic rocks found in the S. Vito 8 well (Figure 1). Thus, we can argue that the passive data shows a continuation of this feature to the north. This improves our confidence that the features we image are real features in the subsurface.

5. Conclusions

We applied a form of converted phases migration to active seismic data from the SERAPIS project at Campi Flegrei caldera to image subsurface reflections. Focusing on PP reflected phases, we imaged features at different depths in our model. We were able to recover part of the caldera rim at two depths in the southern part of the Bay of Pozzuoli: 1.2 to 2.0 km depth when we considered an early phase in the traces, and 2.4-2.6 km when this
early phase was removed and later phases were analyzed. The shallower caldera rim feature
can be related to the transition from tuffs and loose caldera fill to denser hydrothermalized
lava. Locations and depths of the shallow features also correlate with several volcanic vents
detected in multichannel seismic data. The deeper caldera rim feature may correlate to the
transition from hydrothermalized lavas to deeper rocks that have undergone a greater degree
of thermometamorphism. Migrated images also showed focusing of trace energy at 2-3 km
beneath the town of Pozzuoli and near 4 km depth beneath the western portion of the Bay of
Pozzuoli. The shallower feature could mark the top of the caprock over a volume of
overpressured gas-saturated rock imaged by seismic tomography and seismic amplitude
variation with distance studies. The deeper feature in our images is less clear, but could be
related to the presence of the carbonate basement below the caldera. The image from passive
data constrained by well logs provides some confirmation of the deeper caldera rim feature
imaged by the active data and shows a continuation of this feature on land to the north.

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Figure 1: (a) Location map of Campi Flegrei. Inset shows location of region in southern Italy. (b) Gravimetric interpretation map constrained by gravity observations and geothermal well log data (modified after AGIP, 1987). (c) 2D interpretive model of the gravimetric profile observed along the line (solid line in Figure 1b) crossing the Mofete wells, Baia,
Pozzuoli city and the Pianura quarter (modified after AGIP, 1987). The location of the S. Vito 8 well (S8) is from Bruno (2004).
Figure 2: Receiver (triangles) and shot (dots) locations for SERAPIS active data set used in the converted phases analysis. Bold box outlines target area for imaging.
Figure 3: Sample traces for station O12. Map at right shows location of the selected shots (highlighted in red). Maximum offset is approximately 3050 m. A strong early secondary phase is identified as phase A. To provide an example of how phase A appears in the traces it is highlighted in purple for a selection of traces. Two later secondary phases, phases B and C, can also be detected. Phase B is highlighted in selected traces in red, and phase C in
yellow. Green line shows time for a mute designed to remove phase A while the blue line
marks the P-pick times. Traces are band pass filtered, 5-35 Hz.
Figure 4: (a) Cartoon illustrating shot-grid point-receiver geometry for active data. Each grid point can be considered to represent a local interface with arbitrary orientation, as expressed by grid points with gray oriented interface and normal vector. All local interfaces have a common orientation. (b) Cartoon illustrating ray geometry at a single grid point for a single source-receiver pair. The angle \( \theta \) defines the difference between the ray we calculate from the tomography models and the ray predicted by Snell’s law.
Figure 5: Comparison between migrated images (vertical component) for a PP reflected phase using different grid spacing: (left) grid used in the analysis (500 by 500 by 200 meters), and (right) finer grid (250 by 250 by 200 meters) over a smaller volume in space. (Top) Sample horizontal X-Y slices. (Bottom) Sample vertical X-Z slices. Triangles show locations of stations used to make the summed images. Selected stations are: O01, O02, O04, O06, O08, O09, O12, O13, O16, O18, O22, O23, O26, O29, O30, O32, O46, BA3, BAI, NIS, OAS, PO3, STH.
Figure 6: Comparison of migrated images (vertical component) for a PP reflected phase constructed including and not including the Snell’s Law weight coefficients. (Top) Comparison of horizontal X-Y image slices for a single station, O30. (Bottom) Comparison of vertical X-Z image slices from a summed image using 23 stations (same stations as in Figure 5).
Figure 7: Sample traces for different selection window length, $t_{\text{win}}$. Traces from station O06, vertical component, with the selection window shown in gray shading centered on time zero. Traces are unfiltered. Solid line and arrow denote P-pick time at -1.0 s. Time of zero corresponds to the calculated converted phase time for a PP reflection phase at a grid point located at $X=3.5$ km, $Y=-2.5$ km, and $Z=3$ km.
Figure 8: Horizontal X-Y (top) and vertical X-Z (bottom) image slices for PP reflected phase (phase A), vertical component. Traces are not muted. Summed images are shown for stations O06, O07, O08, O09, O11, O12, O13, O14, O16, O17, HC2, NIS. Pink line in panel for $Z = 1.0$ km at upper left shows location of center X-Z slice at $Y = -1.5$ km.
Figure 9: Vertical X-Z image slices for PP reflections, vertical component. Traces have been muted to remove phase A. Color palette is the same as in Figure 8. Summed images are shown for the same stations as in Figure 5. Arrows highlight features discussed in the text.
Figure 10: Vertical X-Z image slices for PS reflections. The sum of the horizontal components is shown (component sum is the square root of the sum of the squared amplitudes at each sample point in the traces). Traces have been muted to remove phase A. Summed images are shown for stations O01, O02, O03, O04, O05, O06, O08, O09, O11, O12, O16, O17, O19, O22, O24, O25, O26, O29, O32, O54, O60, ANF, BA3, DMP, PO3.
Figure 11: Horizontal X-Y image slices from the active data (left two panels, PP reflection, phase A muted, vertical component, summed station images for same stations as Figure 8) and passive data (right panel, PS transmitted phase, sum of horizontal components for station W04). Triangles indicate station locations. Small black dots show earthquake locations at and below 2.4 km depth projected onto the horizontal plane. Color palette is the same as Figure 5. Heavy dashed line denotes inferred caldera edge from AGIP study (Figure 1).
Figure 12: (a) Horizontal and (b) vertical image slices for PP reflections, traces muted to remote phase A. Stations used to make summed image are the same as in Figure 8. Color palette is the same as Figure 5. In (a) thin line marks location of vertical profiles shown in (b). In (b), upper panel, arrows highlight features discussed in the text. Lower panel shows isolines of Vp/Vs model proposed by Chiarabba and Moretti (2006). Gray dots show earthquake locations projected onto the vertical plane.
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