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1-D *P*-velocity Models of Mt. Vesuvius Volcano from the Inversion of TomoVes96 First Arrival Time Data

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Abstract—We applied a revised version of the 1-D τ–p inversion method to first *P*-arrival times from the active seismic experiment performed at Mt. Vesuvius (southern Italy) in 1996 (TomoVes96 Project). The main objective of this work is to obtain 1-D velocity models of Mt. Somma-Vesuvius volcano complex and surrounding area. Moreover we show that combining the 1-D information we provide a reliable 2-D initial model for perturbative tomographic inversions. Seismic and geological surveys suggest the presence of a refractor associated with the contrast between carbonate basement and volcanic/alluvial sediments; synthetic simulations, using a realistic topography and carbonate top morphology, allowed us to study the effect of topography on the retrieved velocity models and to check that the 1-D τ–p method can also approximately retrieve the refractor depth and velocity contrast. We analysed data from 14 on-land shots recorded at stations deployed along the in-profile direction. We grouped the obtained models in three subsets according to the geology of the sampling area: Models for carbonate outcrop area, models for the Campanian Plain surrounding the volcano edifice and models for Mt. Somma-Vesuvius volcano complex. The found 1-D *P*-velocity models show important vertical and lateral variations. Very low velocities (1.5–2.5 km/s) are observed in the upper 200–500 m thick shallow layer. At greater depths (3 km is the maximum investigated depth) *P* velocities increase to values in the range of 4–6 km/s which are related to the presence of the carbonatic basement. Finally we interpolated the 1-D models to demonstrate an example of misfit for a 2-D interpolated model whose residuals are confined in a narrow band around zero.

Introduction

The Somma-Vesuvius complex is a strato-volcano located in the southern sector of the Campanian Plain (southern Italy), a periTyrrhenian graben bordered by Mesozoic carbonate outcrops and formed during the Plio-Pleistocene extensional tectonic which involved the southern Apenninic chain. The Vesuvius morphology is characterised by a volcanic cone (Gran Cono) built within the older Somma caldera. The volcano structure, built during the last 25,000 years, is the result of long periods of prevalently effusive activity alternating with highly explosive eruptions (Plinian). Basic lavas are associated with the first type of eruption while pumice-fall and pyroclastic flow deposits belong to the second one.

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Structural models of the volcano are based on surface geology and on scarce geophysical data. A seismic reflection survey carried out in 1973 in the Bay of Naples indicated a strong reflector interpreted as the top of the carbonate basement underlying the volcano (FINETTI and MORELLI, 1974). A deep borehole drilled on the southern slope of the volcano (Trecase well) reached the carbonate basement at a depth of 1.665 km below sea level (BALDUCCI *et al.*, 1985).

Further details on carbonate basement structure have been generated during the last 10–15 years by analysis of gravimetric data (CASSANO and LA TORRE, 1987), seismic reflection profiles on land (BRUNO *et al.*, 1998) and seismic reflection surveys in the Gulf of Naples (MILIA *et al.*, 1998).

To better understand the internal structure of Mt. Vesuvius, seismic experiments were carried out during the last five years. The Tomography of Mt. Vesuvius (TomoVes) Project began in 1994 with a single profile to determine the feasibility of a three-dimensional tomography of the volcano, which is located in a densely populated and noisy area (ZOLLO *et al.*, 1996). In 1996 a three-dimensional field experiment using explosive sources onland and air gun sources offshore was implemented.

A preliminary 2-D tomographic image of the volcano, achieved by the analysis of data collected during the feasibility experiment, showed a strong lateral and vertical P-velocity variation in the first 3 km of the volcano structure (ZOLLO *et al.*, 1998). This result necessitates a reliable initial model for 2-D/3-D tomographic inversions.

The objective of this paper is to obtain detailed 1-D velocity models of the shallow structure that will be combined to construct starting models for a tomographic study. We applied a revised τ–p inversion method based on a nonlinear approach to a set of first P-wave arrival times from TomoVes 94 and 96 experiments. To date velocity measurements on Mt. Vesuvius rocks are not available, therefore lithological interpretations are based on the laboratory measurements of P and S velocities on volcanic rocks from other areas (ZAMORA *et al.*, 1994; BERNARD, 1999).

Data Collection and Analysis

The data analysed in this study were collected during the recent active seismic experiment TomoVes96 (GASPARINI *et al.*, 1998). Fourteen on-land shots have been generated along four profiles, 24 to 40 km long, and recorded at a dense seismic network. The acquisition layout was multi-2-D, with profiles crossing at the Vesuvius crater and extending to the carbonate outcrops which border the Campanian Plain (Fig. 1). For each shot the recording stations were deployed along the in-profile direction and the quasi-orthogonal one in order to have both 2-D and approximate 3-D ray sampling. The explosive source had variable charge size from

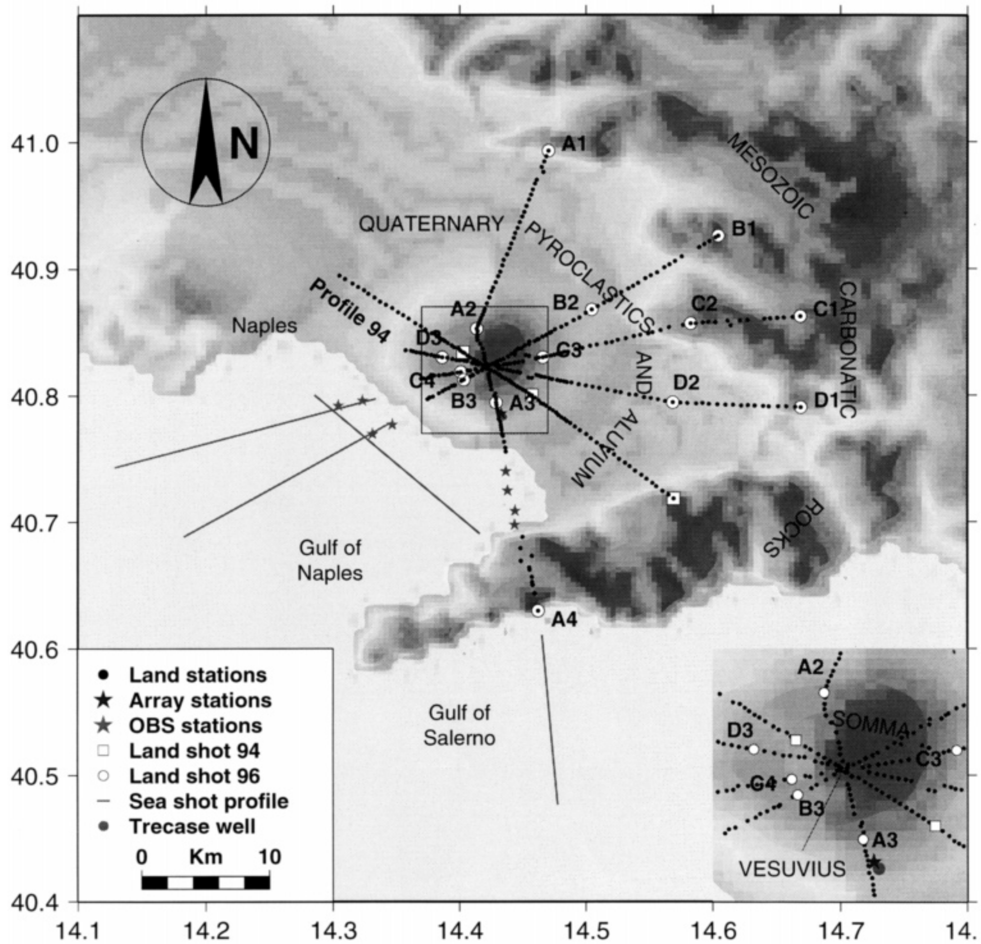

Figure 1
Map of the TomoVes active seismic experiments.

250 kg to 800 kg of gel allocated in 50-m deep holes. Six shots were carried out along the volcano slopes, two in the Campanian Plain and the remaining six on the surrounding carbonatic outcrops.

Along each profile the station spacing was variable from 250 m across Mt. Vesuvius and 500 m outside. GPS and levelling measurements provided shot and receiver positions with an accuracy of 5 m and 10 m for the horizontal and vertical coordinates, respectively. The seismic network was composed of 140 digital seismographs, 120 of them equipped with three-component short-period sensors and those remaining with single vertical components. The sampling frequency of the digital acquisition system was also variable, ranging from 5 ms to 8 ms, depending on the seismograph type.

The first arrival times were picked on unfiltered records arranged in common shot gathers. Where the signal-to-noise ratio appeared very low (typically at large distances from the source and/or near densely urbanised areas) we applied a bandpass Butterworth zero-phase shift filter with frequency limits depending on the noise spectrum. We performed two-time readings ($t1$ and $t2$) which define a window where the true arrival is expected to be. Times $t2$ and $t1$ are respectively the upper and lower limits for the first P-arrival time. For clear and unnoisy phase arrivals, $t1$ and $t2$ may coincide. In case of unclear and noisy phase arrivals only time $t2$ can be estimated (in this case $t2$ corresponds to the most energetic arrivals and it is therefore the estimated upper limit for the arrival time of P phases). The measured travel-time curves (times $t1$ and $t2$) are shown in Figure 2 for each profile and the appropriate shots. Examples of seismic record section are shown in Figure 3. Generally the shots located on carbonatic outcrops provided the best quality records, showing a satisfactory signal-to-noise ratio even at elongated distances. This is related to the larger source size and the lower energy absorption of carbonatic rocks relative to volcanic and alluvium materials. Medium quality records and shorter recording distances are associated with the smaller shots located in the densely urbanised Campanian Plain and on the volcano edifice.

A sharp slope decrease on travel-time curves is shown in Figure 3. It suggests the presence of a probable head-wave arrival. This phase is likely to be generated at the interface between the volcanic/alluvium deposits filling the Campanian Plain and the carbonatic formation.

The Inversion Method and its Application to Synthetic Examples

The 1-D τ–p Equation

In order to model the first arrival times from the TomoVes96 experiment, we preferred to move from the (X, T) to the (τ, p) domain. The inversion in (τ, p) domain is a method extensively used in 1-D problems and it was applied to define 1-D P-velocity models in a complex volcanic structure like the Jasper seamount (Hammer *et al.*, 1994).

The reason for which we adopted the τ–p transformation is the relatively smooth behaviour of τ–p with respect to travel times curves, which instead appear very irregular due to the stronger influence of reading errors, topographic effects and medium heterogeneities. This makes the inversion process more stable and robust. We applied a revised version of the 1-D τ–p method which assumes P-wave velocity to be a monotonic, increasing function of depth in a laterally homogeneous medium (De Matteis *et al.*, 1997). We preferred to model the observed first arrival time curves, using the continuous velocity approximation, since the latter leads to acceptable models for further tomographic studies because the retrieved model is

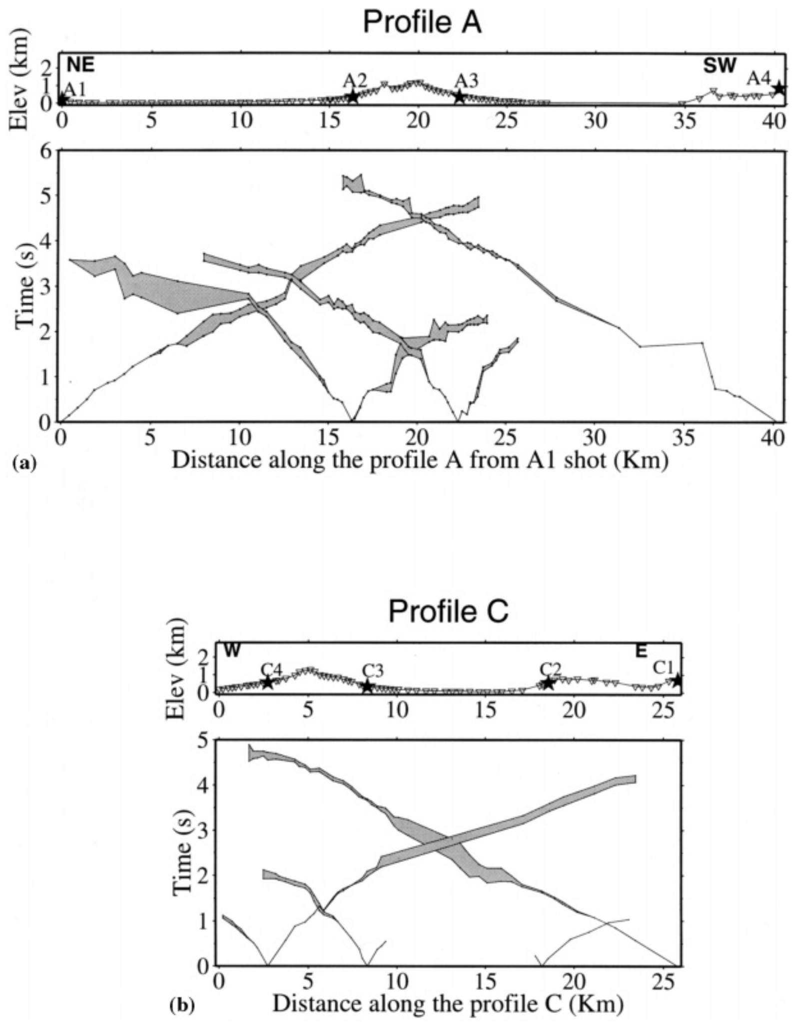

Figure 2
Travel-time curves along *A* profile (a) and *C* profile (b) for each shot. The topographic profiles and the shot locations are shown in the top of the panel.

smooth. We note that, given a first arrival data set showing sharp slope changes, one can always find a discontinuous or continuous model which equivalently explains the observations, due to the non-uniqueness of the first arrival inversion problem (AKI and RICHARDS, 1980). Under these conditions the function $\tau(p)$ can be written:

$$\tau(p) = 2\int_{p}^{u_{\max}} \frac{uz(u)}{\sqrt{u^2 - p^2}}\,du$$

where $u = v^{-1}$ is the slowness, and the unknown function $z(u)$ gives the depth vs. slowness values. We approximate the function $z(u)$ by a polynomial series:

$$z(u) = \sum_{i=-n}^{n} a_i u^i.$$

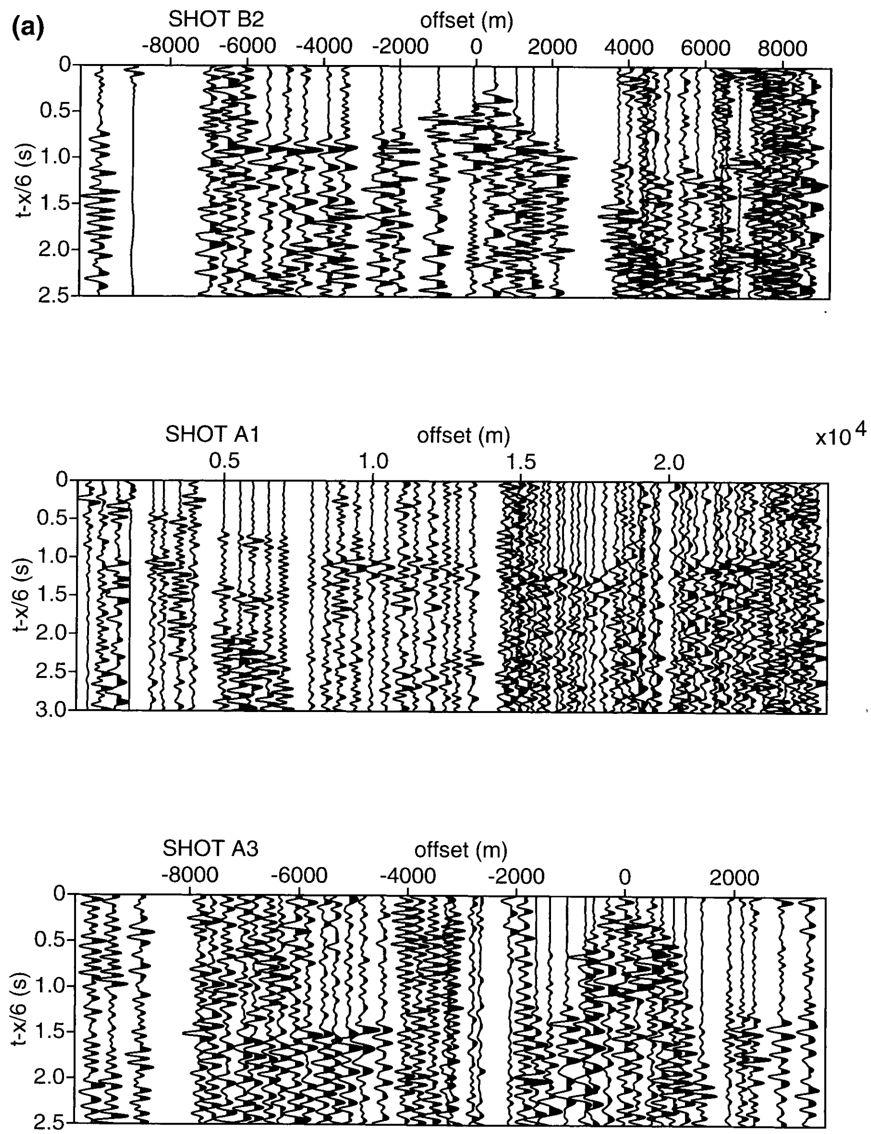

Figure 3.

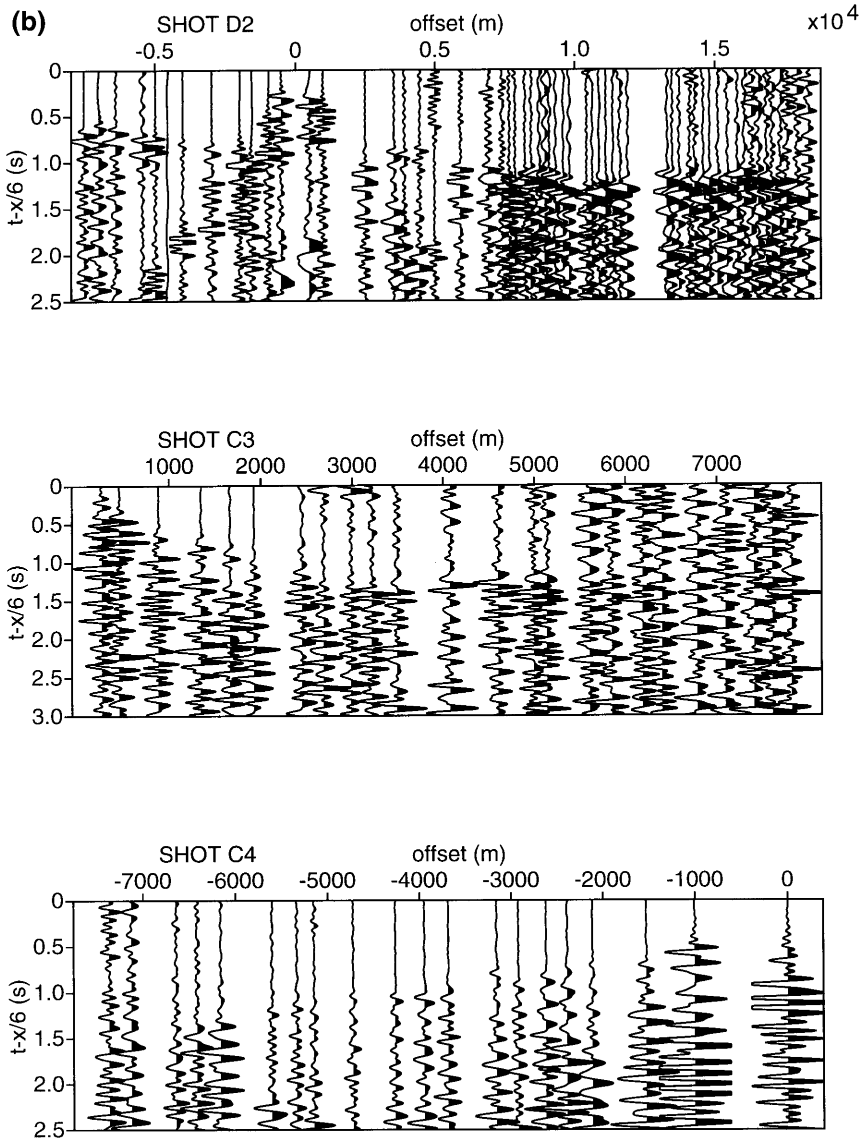

Figure 3
Examples of vertical component recordings. Data are reduced ($v_p = 6$ km/s), bandpass filtered and Automatic Gain Control windowed. (a) Shots B2, A1, A3, (b) Shots D2, C3, C4.

The coefficients of the polynomial function are the unknown parameters of the inverse problem which is solved by a nonlinear optimisation method based on the search for the minimum of a L-square norm.

Measurements of τ–p Curves: Bessonova (1974) vs. Radon Method

The accurate estimate of $\tau(p)$ curves is very important. We adopted two different approaches. The first one was proposed by BESSONOVA *et al.* (1974) and is based on data transformation from the T–X to the τ–X space. For each investigated p the value $\tau(p)$ is measured at the maximum of the corresponding $\tau(X)$ curve (Fig. 4b). In the case of Mt. Vesuvius, $T(X)$ curves showed a complex pattern which sometimes made difficult the use of this technique (Fig. 2).

In such cases we preferred to follow the second approach based on the evaluation of the Radon transform of synthetic seismic sections, obtained by resampling the first arrival time curves and replacing the arrival time readings by pulse-like signals. The method has been successfully checked on synthetic examples using simple to very complex arrival time curves computed from known 1-D velocity profiles. The Radon transform has been computed by using the Seismic

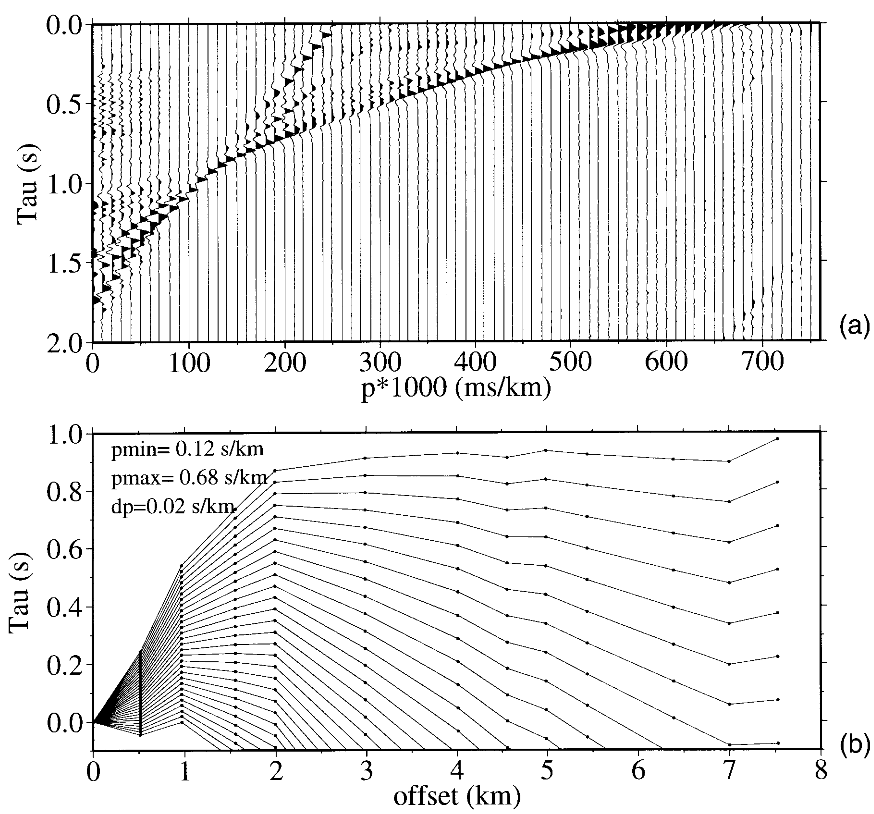

Figure 4

(a) Seismic section in the τ–p domain obtained by applying the Radon transform to the T–X record section relative to shot D-2. (b) $\tau(X)$ curves computed for different p-parameter values. The τ values for each p are determined by the extreme points of $\tau(X)$ curves (Bessonova method).

Unix code (COHEN and STOCKWELL, 1994) (Fig. 4a). In order to avoid spatial aliasing, the Radon transform has been applied to linearly interpolated travel-time data, resampled at a finer distance interval. We are aware that the interpolating process is somewhat arbitrary and that this procedure cannot provide details regarding the velocity models which are smaller than the true average station spacing.

Inversion with Synthetic Data

Since the travel-time curves (for distant shots) indicate a clear evidence for a head wave probably generated by the carbonatic top discontinuity, we checked whether the 1-D τ–p method was able to retrieve the refractor depth and velocity contrast. The synthetic simulation using a realistic topography allowed us to study the effect of topography on the retrieved velocity models.

This analysis was performed using synthetic travel-time curves computed in a 2-D preliminary reference Vesuvius model (PRVM) which was built based on active seismic, surface geology and gravity data (CASSANO and LA TORRE, 1987; FUSI, 1996; BRUNO *et al.*, 1998; FERLITO, 1998). The model is a 1-layer over a half-space structure which includes the earth's surface and the carbonate top morphology along a profile A. We assume a linear velocity variation with depth in the upper layer, $V = (1.7 + 1.2z)$ km/s, where z is measured from the earth's topography. A P velocity of 5.6 km/s has been assumed for the carbonatic formation (Fig. 5a).

The first arrival times have been computed in the 2-D propagation medium with topography using a finite difference solution of the eikonal equation (PODVIN and LECOMTE, 1991). We did not apply the topographic correction moving from the (T, X) to (τ, p) plane for $S1$ and $S2$ shots; in fact the dominant topographic effect is observed for arrivals at stations located on the volcano slope corresponding to the head wave generated at the top of the carbonatic formation. In the τ–p plane, the head-wave arrivals collapse to the point which represents the minimum observable p. The topographic effect will therefore only be relevant for the estimate of velocity at the maximum investigated depth. However, as in the actual cases, the head wave is already observed the nearest stations where the topographic effect is negligible. These results suggested that we can neglect the effect of topography for shots located at the extreme points of all profiles and in the Campanian Plain.

The inferred velocity models for the cases ($S1S2$) and ($S2S1$) are shown in Figures 5b and 5c. Velocities in the upper layer are well constrained by the τ–p 1-D inversion. This part of the velocity depth curves corresponds to the area investigated only by turning waves. At higher depths $v(z)$ curves demonstrate a flexure which is associated with the abrupt velocity variation. The comparison between the theoretical and retrieved curves illustrates that the depth of the flexure matches well the discontinuity depth in a region under the shot. The maximum apparent velocities of the two conjugate profiles (7 and 4.9 km/s) provide an average value (5.9 km/s) which nears the real velocity (5.6 km/s).

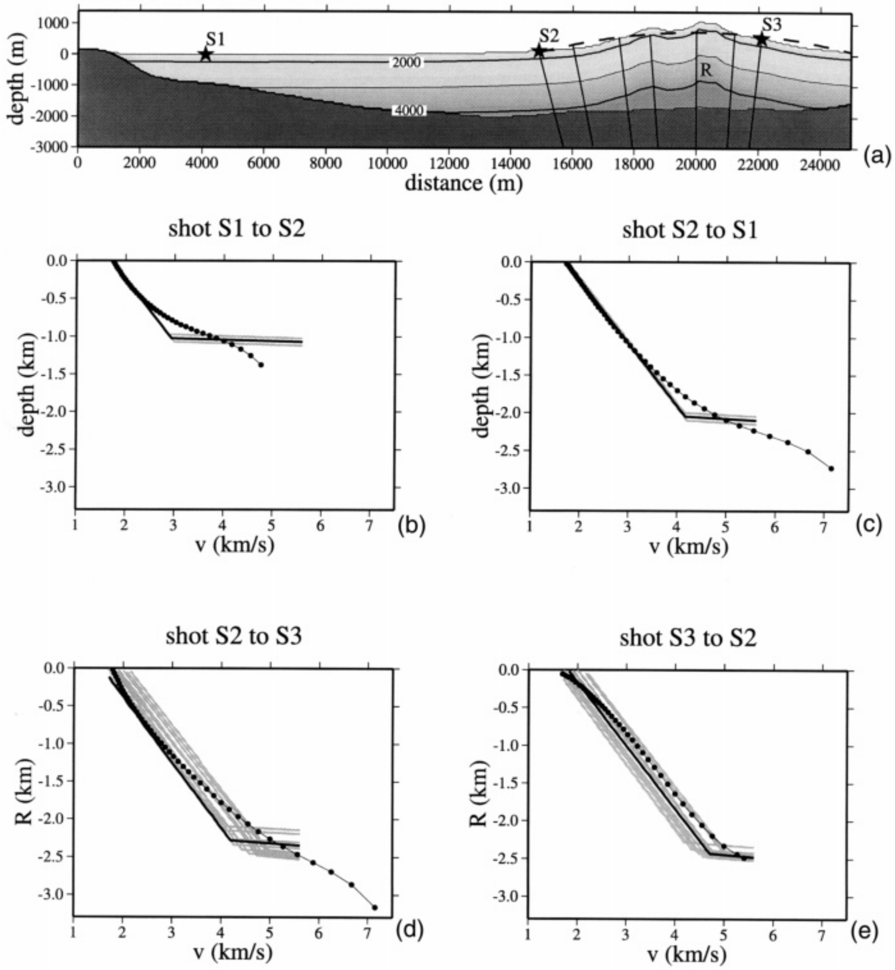

Figure 5
(a) For the synthetic example the topography and the top of the carbonatic basement are relative to the *A* profile. (b) and (c) Velocity models obtained for the two conjugate profiles *S*1 towards *S*2 and *S*2 towards *S*1 (dotted lines) are compared with three 1-D velocity models (grey continuous lines) extracted from the 2-D synthetic model. (d) and (e) For the shots on the volcanic edifice, *S*2 and *S*3, velocity is recovered as a function of orthogonal distance from the surface topography. The retrieved velocity models (dotted lines) are compared to different "true" radial velocity profiles (grey continuous lines). The distance is measured from the arc of the circle approximating the volcano topography. The "true" velocity models at a distance of 1 km from the shots are represented with a continuous line.

For *S*2 and *S*3 we followed the approach suggested in De Matteis *et al.* (1997) where the volcano edifice has been assumed spherically concentric; this is a better approximation than a flat layer case for the Mt. Vesuvius geometry. In order to take into account the spherically concentric variation of velocity, we can correct the velocity estimates obtained by the 1-D inversion method using a stretched, Cartesian frame. This is done by applying the transformation formulas (AKI and RICHARDS, 1980):

$$z = r_{\otimes} \ln\left(\frac{r_{\otimes}}{r}\right)$$

$$v(z) = v(r)\frac{r_{\otimes}}{r}$$

where r is the radial distance from the centre of the selected circle of a radius $r_{\otimes}$ which is approximately 15 km for Mt. Vesuvius.

The obtained 1-D velocity model is an average model of the region near the shot as shown by comparison with theoretical curves in Figures 5d and 5e. In order to see the effect of noise on the inversion we added an error equal to ± 0.05 s to τ values. The results of the inversions show that the two retrieved curves limit a region around the velocity model without error and that the extreme parts of the curves are less resolved.

Inversion of TomoVes96 First Arrival Time Data

Two different τ–p curves are computed for shots located inside the seismic line, relative to the left and right side shot gathers (Fig. 6). The inversion method has been applied to τ–p curves inferred separately for $t1$ and $t2$ arrival times. The final $v(z)$ curves are represented as the envelope of the models obtained using $t1$ and $t2$ (the upper and lower limits for the first P-arrival time). The total number of retrieved 1-D models for Mt. Vesuvius area is 20, corresponding to 14 shot gathers. The obtained models are grouped in three subsets according to the geology of the area. They are: the models belonging to the carbonate outcrops which border the southeastern part of the Campanian Plain, the models for the plain surrounding the volcano edifice and the models for the Mt. Somma and Vesuvius volcano complex. The derived 1-D models relative to areas characterised by similar geological conditions are shown in Figures 7 and 8.

Models for the Carbonate Outcrop Areas

At shallow depths the $v(z)$ curves present highly variable P velocities (from 2.5 km/s at $C2$ site to 4.3 km/s at $C1$ site) and different gradients (Fig. 7a). The low velocities at sites $C2$ and $B1$ are probably linked to the occurrence of a cover of pyroclastic deposits. Velocities in carbonatic rocks at shallow depth range 3.5–4.3 km/s which is consistent with expected values in fractured near-surface carbonate rocks. P-velocities increase up to 5.5–6.0 km/s with depth. These values are compatible with those inferred by FINETTI and MORELLI (1974) for Mesozoic carbonates which form the basement of the Gulf of Naples.

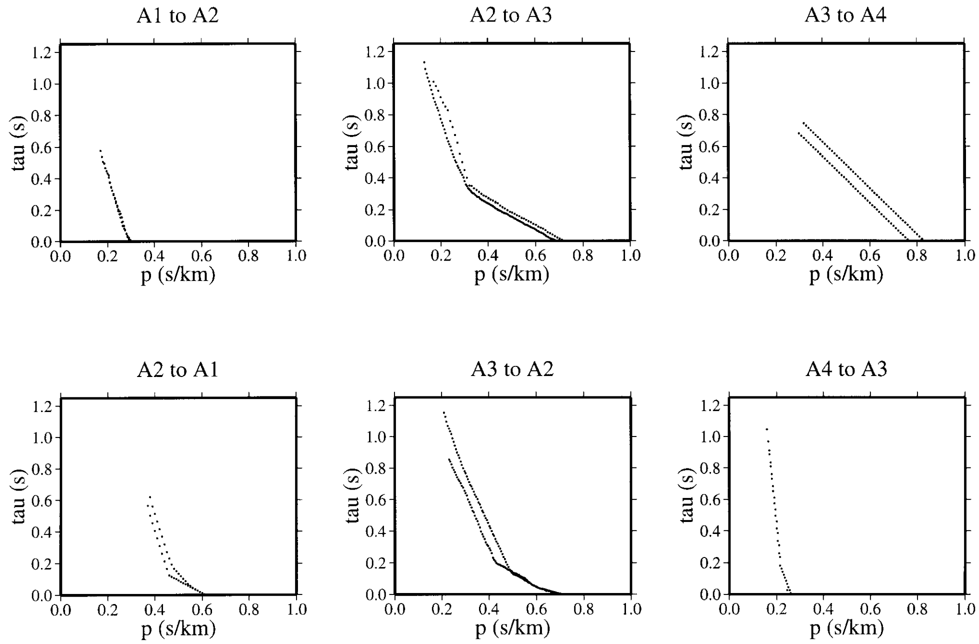

Figure 6
Example of τ–p curves computed for shots located along A profile.

Models for the Campanian Plain Surrounding the Volcano Edifice

All but one ($A2$ to $A1$) of the analysed profiles show a striking similarity both in absolute velocities and gradients (Fig. 7b). Near-surface velocities are very low (around 1.5 km/s) and increase regularly to 3.5 km/s at about 1–1.3 km depth. The very low velocities are consistent with the presence of a thick shallow water layer in unconsolidated sediments and pyroclastics as found in several boreholes in the Campanian Plain (BELLUCCI 1994; DI VITO *et al.*, 1999). The high gradients can be related to the compaction of these sediments with depth. Only the $A2$ to $A1$ curve located north of Vesuvius presents higher near-surface velocities (around 2 km/s) and smaller gradients with velocity reaching 2.5–2.7 km/s at about 1.5 km depth. This indicates the occurrence of shallow volcanic material (Somma lavas) buried under the northern foot of Mt. Vesuvius (BELLUCCI, 1998). North of Mt. Somma-Vesuvius edifice the 1-D model reaches a maximum depth of 1.3 km and P velocity of 2.7 km/s, and does not evidence the carbonate top discontinuity.

Models for Mt. Somma and Vesuvius Volcano Complex

The velocity models in the Mt. Somma and Vesuvius areas are shown in Figure 8. The velocity models for profiles on Mt. Vesuvius display a rather different shape

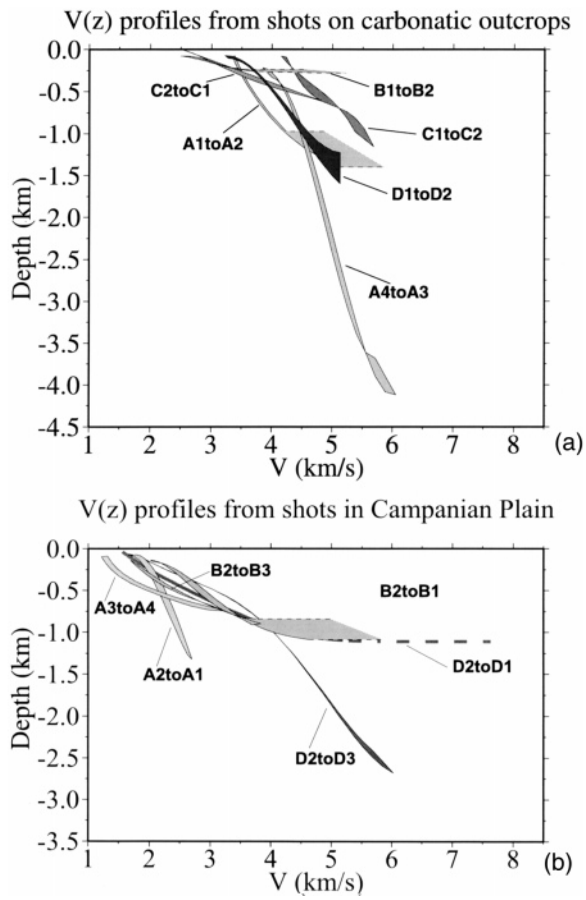

Figure 7
(a) 1-D velocity models for carbonatic outcrop areas which border the Campanian Plain. (b) 1-D velocity models for the plain surrounding the volcano edifice.

vs. depth which suggest a dominant effect of lateral heterogeneities. *P*-velocity gradients are very steep (about 3 s^{-1}) from the surface down to 0.5 km, then decrease to about 1 s^{-1} down to about 2 km. Small *P*-velocity values (1.5–2.5 km/s) observed for shallow layer (< 500 m) characterise both pyroclastic material and densely fractured and altered lavas (BERNARD, 1999).

The inferred 1-D models ($C4$ to $C3$, $A2$ to $A3$, $D3$ to $D2$) relative to the region NW of Mt. Vesuvius have *P* velocities of about 1 km/s higher than those in the SE sector. This is consistent with the presence, NW of Mt. Vesuvius of a high velocity shallow body ($V_p = 3$–4.5 km/s) whose location and dimensions match well the high velocity region detected by a previous 2-D tomography study (ZOLLO *et al.*, 1996). The measured *P* velocities (4–5 km/s) suggest that this body may represent a buried subvolcanic structure (dykes complex and/or thick compact layers). The

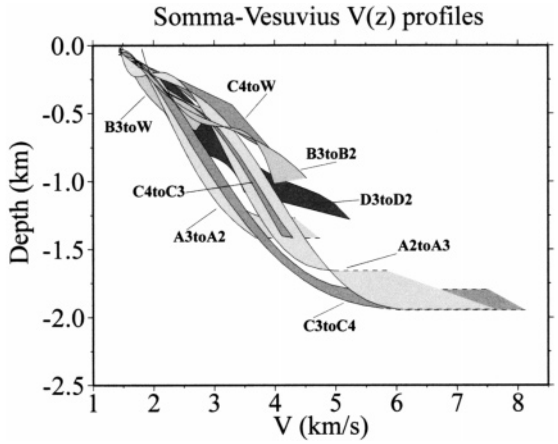

Figure 8
1-D velocity models for the Mt. Somma-Vesuvius volcano edifice.

small evidence for thermometamorphosed volcanic rocks among Mt. Vesuvius ejecta rules out the possibility that they may be a significant component of the high velocity body.

Previous geophysical studies have shown that the carbonate formation which outcrops at the SE border of the plain gently deepens toward the NE, reaching maximum depths of 3 km in the Gulf of Naples. The head-wave arrival is coherently detected at stations located outside, but also on the volcano (Fig. 3). This suggests that the seismic discontinuity is still present beneath the volcano in correspondence with the area affected by the original caldera formation and repeated eruptive episodes. Although the applied τ–p method is designed to model continuous velocity variation with depth, inversion of synthetic and observed data points out that the method is able to detect sharp velocity discontinuities whose depth can be measured from the 1-D models from the flexure of the $v(z)$ function. The depth of the carbonate top is calculated at about 1.3–1.6 km b.s.l. underneath the volcano. Outside the volcanic edifice the obtained carbonate top depth is consistent with other estimates from seismic and gravity data and matches well the interface depth of the PRVM model.

A 2-D Reference Model for Seismic Tomography

Recent tomographic studies of volcanic regions based on active seismic data have employed average 1-D flat- or concentric-layer models as starting models for first arrival time inversions (HAMMER *et al.*, 1994; AMATO *et al.*, 1996; ZOLLO *et al.*, 1996). In the present case, the found 1-D P-velocity models show an important lateral variability, moving from one section to another on the same profile. This is related to the expected complex morphology of the carbonatic basement top and to velocity variations due to changes in lithology. In this case it is very important to have a good 2-D initial model for tomographic inversions. If we combine the retrieved 1-D information we can take into account large-scale lateral velocity variations. Thus we present an example of possible interpolation techniques between the 1-D models, using a weighting procedure based on the horizontal distance from the source.

The shots inside the profiles provided two velocity models for the right and left record sections. The distance at which the 1-D models have been located along the profile have been chosen according to the maximum distance for turning waves. The PRVM model has been used to extrapolate the *P*-velocity curves down to the depth of the carbonate top. The 2-D interpolated model for profile *A* is shown in Figure 9. In order to validate the inferred 2-D model we computed the first arrival times using a revised version of the PODVIN and LECOMTE (1991) code (HERRERO and ZOLLO, 1999), based on the finite-difference solution of the eikonal equation. Moreover, to assess the quality of the derived 2-D model we compared it with a simpler PRVM, computing first arrival times by forward ray tracing. The arrival time residuals of interpolated τ–p model are confined in a band around zero smaller than one of PRVM (Fig. 10). The inadequacy of PRVM is essentially due to the discrepancy between the observed and theoretical travel times at small offsets.

The 2-D interpolated model reproduces well the low velocity basin north of Mt. Somma, where a pronounced Bouguer anomaly low is also detected (SANTACROCE, 1987). The high velocity body beneath the ancient rim of Mt. Somma caldera is located off the axis of the Mt. Vesuvius volcano edifice. The shape of this anomalous region is not well constrained and it is likely biased by the interpolation

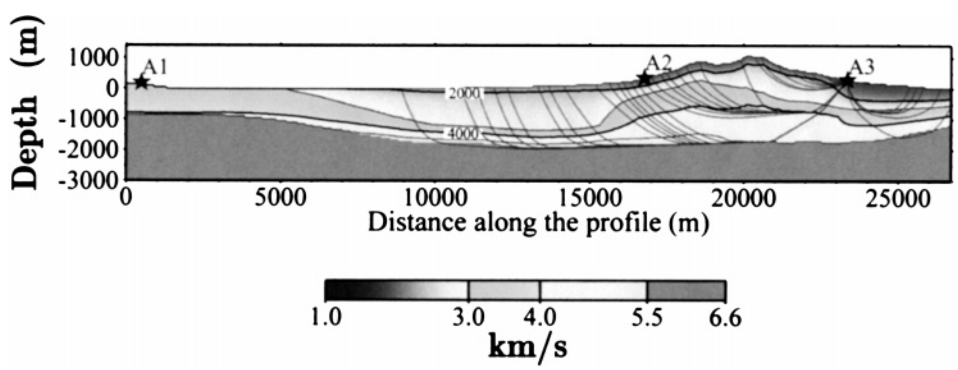

Figure 9
2-D interpolated velocity model for *A* profile. For *A*3 shot the ray trajectories are traced *a posteriori* by a back-projection technique.

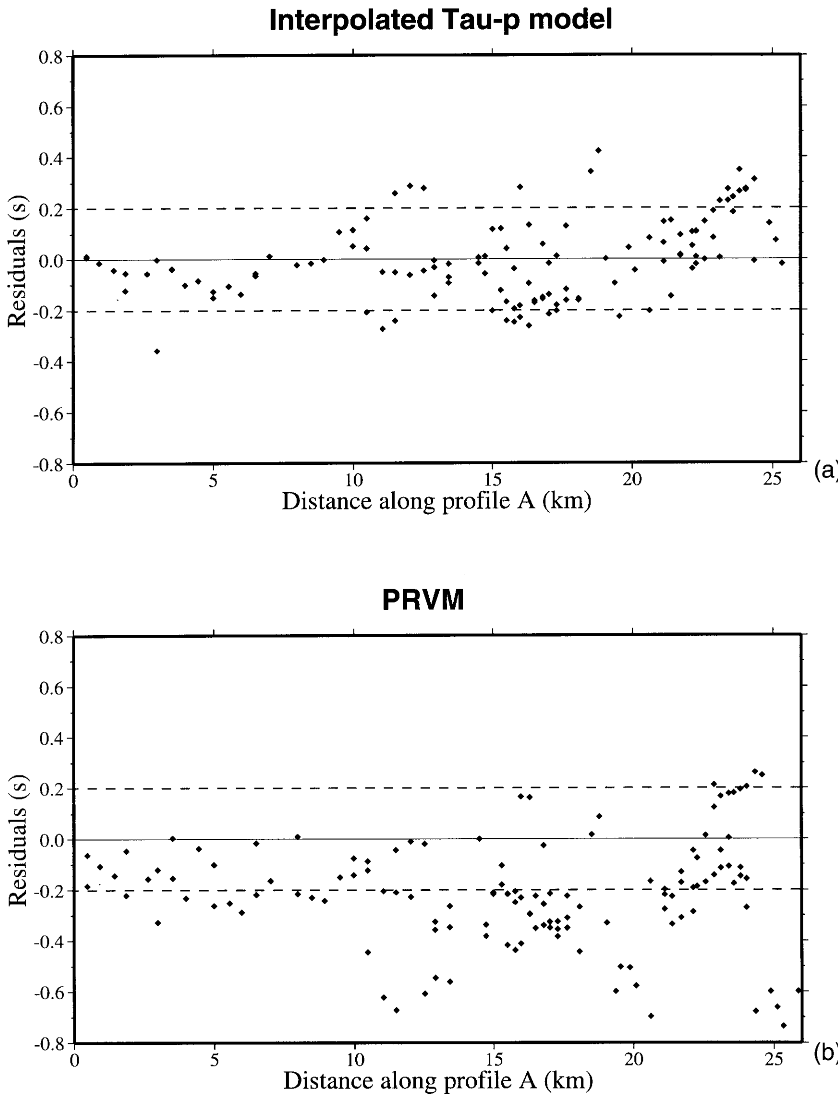

Figure 10
Arrival time residuals, computed for (a) 2-D interpolated τ–p model and (b) PRVM: residuals of the 2-D interpolated τ–p model are confined in a band around zero smaller than one of PRVM. The different distribution of PRVM residuals along the profile indicates that a simple model is not very adequate to allow for lateral velocity variations.

technique. The overall pattern of residuals of interpolated τ–p model confirm that gross features of the velocity structure are well reproduced by the interpolated model, while this is inadequate to explain small-scale details.

Conclusions

In this study we applied the revised 1-D τ–p inversion method to first arrival times from the active seismic experiment TomoVes96. The main objective of this work was to investigate the velocity vs. depth variation in the Mt. Somma-Vesuvius volcano complex and surrounding area. This information can be combined to build 2-D/3-D reference *P*-velocity models to be used for 3-D seismic tomography. The obtained results allow significant conclusions to be drawn concerning the shallow crustal structure in this volcanic district and the relationships between seismic velocities and rock lithology in volcanic environments. The results of this study can be summarised as follows:

1. Models of the shallow crust underneath the volcano show a sharp velocity variation with depth. Very low velocities (1.5–2.5 km/s) are observed in the upper 200–500 m thick shallow layer due to the presence of incoherent, sediments and/or volcanic products. At greater depths (3 km is the maximum investigated depth) *P* velocities abruptly increase to values that range 4–6 km/s, which are probably related to the presence of the shallow discontinuity between alluvial/volcanic sediments and the carbonatic substratum.
2. *P* velocities obtained by 1-D modelling are consistent with the laboratory measurements of mechanical properties of volcanic rocks.
3. Synthetic models have shown the low to high velocity discontinuities by a flexure in the $v(z)$ curves; the depth of the discontinuity fits well the flat area of the flexure.
4. The application of the methods described above enabled us to estimate the depth and velocity contrasts associated with the carbonate top interface beneath Mt. Vesuvius along sections with different directions. This confirms the result obtained in the previous experiment (ZOLLO *et al.*, 1998), and it is a new result because previous seismic studies were only concerned with the area outside the volcano edifice.
5. This study confirms the evidence for a high velocity body located NW of Mt. Somma-Vesuvius as detected from a previous tomographic sounding in the area.
6. This paper provides one 2-D interpolated model to be used for preliminary structural interpretations and serves as an example of a starting model for detailed further 3-D tomographic analyses.

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