ELSEVIER

DEVELOPMENTS IN VOLCANOLOGY 7

THE SOUTH AEGEAN ACTIVE VOLCANIC ARC

PRESENT KNOWLEDGE AND FUTURE PERSPECTIVES

EDITED BY

M. FYTIKAS AND G.E. VOUGIOUKALAKIS

THE SOUTH AEGEAN ACTIVE VOLCANIC ARC

Edited by
Fytikas

Included in series
Developments in Volcanology,

Contents

1. Tertiary and Queternary tectonics in the Aegean area (D. Mountrakis). 2. A model of the Aegean geodynamic zone (B. Ranguelov). 3. Upper mantle structure of the Aegean (Greece) derived from phase velocities of fundamental mode Rayleigh waves (I. Kassaras, K. Makropoulos). 4. Deep structure and active tectonics of the Southern Aegean volcanic arc (B.C. Papazachos *et al.*). 5. A west-east traverse along the magmatism of the South Aegean volcanic arc in the light of volcanological, chemical and isotope data (L. Francalanci *et al.*). 6. The South Aegean active volcanic arc: relationships between magmatism and tectonics (G. Pe-Piper, D.J.W. Piper). 7. Our bubbling earth (R.D. Schuiling). 8. Magmatic evolution processes as recorded in plagioclase phenocrysts of Nea Kameni rocks (Santorini volcano, Greece) (A.P. Santo). 9. Volcanic hazards in the Aegean area, relative risk evaluation and monitoring state of the active volcanic centers (G.E. Vougioukalakis, M. Fytikas). 10. Recent seismic activity (1994-2002) of the Santorini volcano using data from local seismological network (I.M. Dimitriadis *et al.*). 11. Geodetic evidence for slow, small scale inflation of the Santorini caldera (S.C. Stiros, A. Chasapis, V. Kontogianni). 12. Long-term prediction of the next eruption in Thera volcano from conditional probability estimates (G.A. Papadopoulos, K. Orfanogiannaki). 13. Late-Holocene coastal uplift in the Nisyros volcano (SE Aegean Sea): Evidence for a probable new phase of slow shallow intrusive activity (S.C. Stiros*et al.*). 14. Investigating the formation of a superficial fracture on Nisyros Island, Greece with the DC resistivity method (D. Galanopoulos, G. Kolettis). 15. Gravity monitoring of Nisyros volcano activity: 2001-2003 preliminary results (M. Di Filippo, B. Toro). 16. Real time monitoring of gas-geochemical parameters in Nisyros fumaroles (M. Teschner *et al.*). 17. The Vani manganese deposit, Milos island, Greece: A fossil stratabound Mn-Ba-Pb-Zn-As-Sb-W-rich hydrothermal deposit (G.P. Glasby *et al.*). 18. An FT-Raman, Raman and FTIR study of hydrothermally altered volcanic rocks from Kos Island (Southeastern Aegean, Greece) (D. Papoulis *et al.*). 19. Rhyolitic dikes of Paros Island, Cyclades (A. Hannappel, T. Reischmann). 20. Investigation of non-pristine volcanic structures acting as probable hosts to epithermal gold mineralization in the back arc region of the active Aegean Arc, used combined satellite imagery and field data: examples from Lesvos volcanic terrain (C. Vamvoukakis*et al.*). 21. Tertiary high-Mg volcanic rocks from Western Anatolia and their geodynamic significance for the evolution of the Aegean area (S. Agostini *et al.*). 22. A mathematical model for the morphological evolution of a volcano on an island (G.A. Skianis, D. Vaiopoulos, V. Tsarbos).

Bibliographic & ordering Information
Hardbound, ISBN: , publication date: 2004
Imprint: ELSEVIER

Recent Seismic Activity (1994-2002) of the Santorini Volcano Using Data from Local Seismological Network

I.M. Dimitriadis[1,*], D.G. Panagiotopoulos[1], C.B. Papazachos[1], P.M. Hatzidimitriou[1], E.E. Karagianni[1] and I. Kane[1]

[1] *Geophysical Laboratory, School of Geology, Aristotle University of Thessaloniki, P.O. Box 352-1, GR-54124, Thessaloniki, Greece*

Abstract

The South Aegean Active Volcanic Arc consists of a chain of five volcanic centers, the most active of which is the Santorini Volcano. A local radio-linked seismological network is installed on the island consisting of five permanent and four temporary stations. The temporary stations have been in operation periodically during the period 1994-1996 and two of them were installed on adjacent islands. All stations are equipped with vertical-component short period seismometers. During the period 1994-2002 a significant number of earthquakes has been recorded, with local (duration) magnitudes, M_D, up to 5.0 and focal depths varying between 0 km and 35 km. Two clusters of epicenters have been located in the broader area of the Santorini Volcano. The first cluster is located in the caldera of the volcano and is associated with the volcanic process in the Kameni Island. The second (larger) cluster is located near the northern edge of the Santorini Island at the Columbo Reef and is connected with the volcanic process at this reef. These clusters can be appropriately associated with the two main tectonic features (faults) in the area under study. The first one (N60^{o}E direction) corresponds to the continuation of the Amorgos fault in the area, while the secondary tectonic line (EW direction) is probably related with the southern edge of a submarine graben, which is located between the islands Amorgos and Santorini. Using the data set of the best-located earthquakes, recorded during the period 1994-2002, an attempt has been made to derive an appropriate equivalent 1-D earth model for the area under study, in order to improve the accuracy of the determined hypocenters, as well as to obtain a preliminary knowledge of the volcano structure.

* Corresponding Author. Tel.: +30-2310998535 Fax: +30-2310998528
E-mail address: danis@lemnos.geo.auth.gr (I.M. Dimitriadis)

Keywords: Santorini Volcano; Seismic activity; hypocentral estimation; 1-D velocity model

1. Introduction

1.1 Geographic and Geotectonic Setting of the Santorini Volcano

The Hellenic arc is one of the dominant tectonic features of the southern Aegean area. It separates the Aegean Sea from the Mediterranean Sea and has the main properties of a typical island arc (Papazachos and Comninakis, 1971; McKenzie, 1978; Le Pichon and Angelier, 1979, 1981). It consists of an outer (southern) "sedimentary arc", which is a link between Dinaric Alps and the Turkish Taurids, and of an inner (northern) "volcanic arc" with Quaternary volcanoes.

Between the sedimentary and the volcanic arc is the Cretan Sea, with a maximum water depth of about 2 km. On the convex side of the arc (eastern Mediterranean) a system of troughs and trenches (Ionian, Pliny and Stravo) is found, which is called the Hellenic Trench with a maximum water depth of about 5 km. A well-defined Bennioff zone of intermediate depth earthquakes, dipping from the Mediterranean Sea to the Aegean Sea and associated with this arc, was first identified by Papazachos and Comninakis (1969).

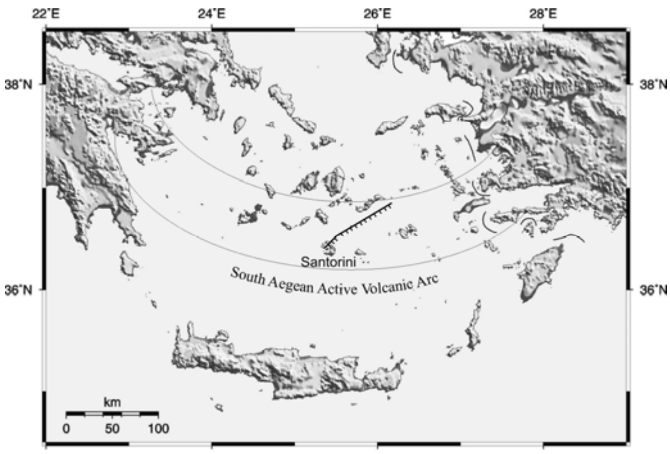

Figure 1. The Amorgos Fault and the South Aegean Active Volcanic Arc (SAAVA).

Santorini is one of the five volcanic centers, which make up the South Aegean Active Volcanic Arc (SAAVA) (Fig.1). The volcanic centers include three Quaternary Volcanoes (Santorini, Nisyros and Methana), solfataras and fumaroles fields (Sousaki, Milos and Kos)

(Georgalas, 1962; Fytikas et al. 1985). The volcanoes, along with the clusters of epicenters of shallow and intermediate depth earthquakes, form the five seismovolcanic clusters (Sousaki, Methana, Milos, Santorini and Nisyros) (Papazachos and Panagiotopoulos, 1993). These are connected with tectonic zones of weakness, which follow a direction N60^{0}E. The most active volcanic center is the Santorini Volcano, which has erupted at least nine times in the last 600 years (1457, 1508, 1573, 1650, 1707, 1866, 1925, 1939 and 1950) (Papazachos, 1989; Fytikas et al. 1990).

The main tectonic feature in the area of Santorini is the Amorgos fault with a NE-SW direction, where the large earthquake in 1956 with M=7.5 occurred (Shirokova, 1972; Papazachos et al. 2001; Pavlides and Valkaniotis, 2003) (Fig.1). Another important tectonic feature in the same area is the graben northern of Santorini, which has an approximately ENE-WSW direction (Perissoratis, 1996).

1.2 Geology of Santorini

Santorini is a group of five islands (Thera, Therasia, Aspronisi, Palaea Kameni and Nea Kameni). Thera, Therasia and Aspronisi enclose a sea-flooded caldera. Apart from a small non-volcanic basement found in the southeastern part of Thera, these islands are composed of volcanic rocks. Volcanism in the area of Santorini started about 2 million years ago with the extrusion of dacitic vents on the Akrotiri peninsula and continued to produce different kinds of lavas and pyroclasts (Friedrich, 1994, 2000). However, the most characteristic type of activity over the last 200.000 years has been the cyclic construction of shield volcanoes interrupted by large explosive and destructive events like the Minoan eruption (Bond and Sparks, 1976; Heiken and McCoy, 1984, 1990; Druitt et al. 1989, 1999). In particular, a thick layer of white pumice, which has been laid down by the Minoan Eruption, covers a great part of Santorini (Fig.2).

Non-volcanic rocks are exposed at the Profitis Ilias Mountain and at other parts of the main island (Kilias et al. 1996; Mountrakis et al. 1996). The oldest volcanic rocks can be found on the Akrotiri peninsula, while the youngest volcanic rocks can be found on the Palaea and Nea Kameni islands, which formed during eruptions in historic times (Druitt and Francaviglia, 1992; Seidenkrantz and Friedrich, 1992).

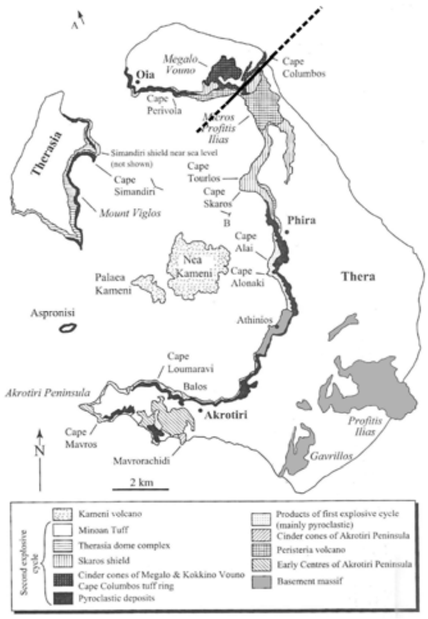

Figure 2. Geological map of the Santorini Islands; the heavy line represents the continuation of Amorgos fault (modified from Druitt et al., 1999).

2. Seismic Monitoring at the Santorini Volcano

2.1 Local Seismological Network

Within the framework project of the seismic monitoring of the Santorini Volcano, the Geodynamic Institute of the National Observatory of Athens in cooperation with the Geophysical Laboratories of the University of Athens and Thessaloniki installed a local radio-linked seismological network consisting of eight stations on the islands of Santorini and Ios in 1994 (Panagiotopoulos et al. 1996). In particular, the stations of COLUMBO, AKROTIRI, CENTRAL and OIA have been installed on the Thera island, the stations of RIVA and KERA on the Therasia island, the station of KAMENI on the Nea Kameni island and the station of IOS on the Ios island (Fig.3). The stations of IOS, COLUMBO and RIVA have been in operation during the period 1994-1996. The other five stations (AKROTIRI, CENTRAL, KERA, KAMENI, OIA) are in operation until today.

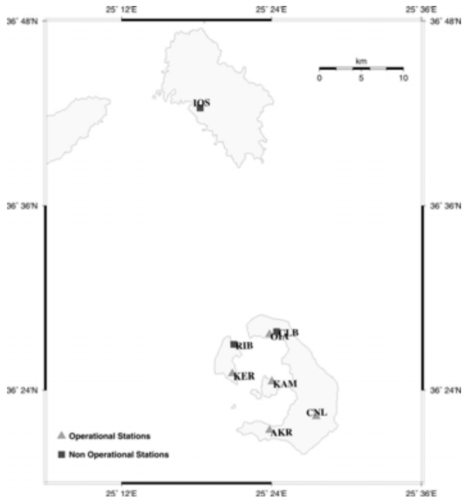

Figure 3. Map of the local seismological network installed on the Santorini and adjacent islands.

All stations are equipped with vertical-component short period seismometers and the analogue seismic signals are transmitted via antennas to the central station, which is situated on the highest point of Thera (Profitis Ilias Mountain, 500m altitude). The main station (observatory) is a relatively small construction that insures the antenna and the computerised recording unit and analysing of the seismological data. The recording and analysing system of the signals consists of three units: the conversion system of analogue to digital signals, the server and the terminal (workstation).

2.2 Estimation of the epicenters and hypocenters

The main objective of this paper is the precise hypocenter estimation of local earthquakes that occurred in the broader area of the Santorini Volcano during the period of 1994-2002. The estimation of the epicenters was performed using the computer program HYPOELLIPSE (Lahr, 1989, 1999) with an initial local velocity model derived from a large-scale 3-D model (Papazachos and Nolet, 1997) for the area under study (Table 1).

Depth (km)	V_P (km/s)	V_S (km/s)
0,0	4,00	2,25
1,0	6,00	3,37
24,0	6,60	3,71
32,0	7,70	4,33
40,0	8,10	4,55

Table 1. The initial local velocity model used for the estimation of the events.

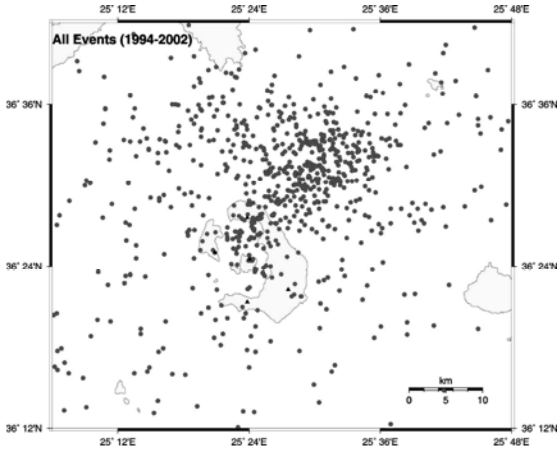

Figure 4. Epicenters of local earthquakes of the period 1994-2002 in the area under study.

During the period 1994-2002, 1076 events have been recorded, with local (duration) magnitudes, M_D, up to 5.0 and focal depths up to 35 km. Despite the poor spatial coverage of the recording stations and the simplified earth model used, two clusters of epicenters have been located in the broader area of the Santorini Volcano. As is shown in figure (4), a large cluster of epicenters is located near the northeastern edge of the island at Columbo Reef, while a small cluster of epicenters is located in the caldera of the volcanic center.

In order to separate the best-located events due to the uncertainty of the determined hypocenters and the lack of P and S phases, earthquakes with: ERH (minimum horizontal error) < 20km, ERZ (minimum depth error) < 25km, RMS (minimum residual error) < 0.5, phases number per event > 6, Azimuthal Gap < 340^0, the minimum focal depth used < 35km, and, finally, a minimum epicentral distance < 30km were selected (Fig.5). The final data set satisfying these conditions consists of 157 events. Figure (6) shows the distribution of the epicenters of the best-located earthquakes, where two very clear tectonic lines are identified. The main one has a NE-SW direction and corresponds to the Amorgos Fault, while the secondary tectonic line has an approximately E-W direction and is related with a graben northern of Santorini.

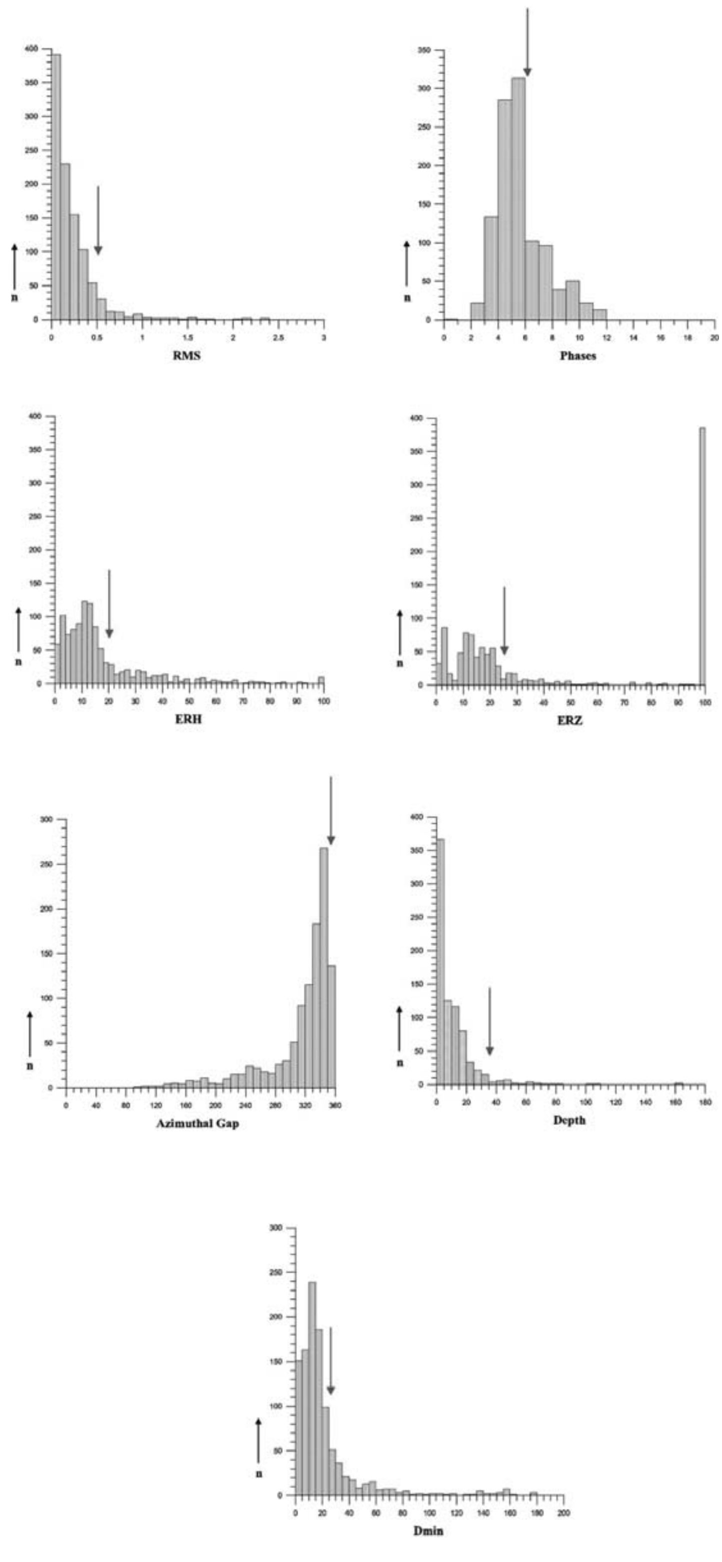

Figure 5. Histograms of the seismic parameters along with the confined parameters (black arrows) used for the selection of the best-located events.

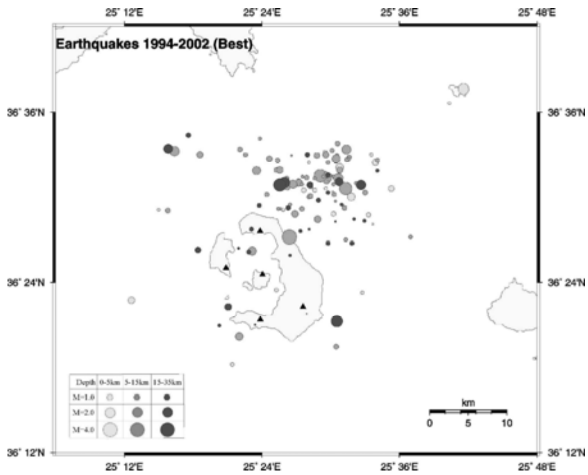

Figure 6. Epicenters of the best-located earthquakes of the period 1994-2002 in the area under study.

3. Magnitude Estimation of the Local Earthquakes

Durations of the recorded waveforms were used in order to estimate the local magnitude of all earthquakes. The equation, which was used for this estimation, is the following one:

$$M_D = c_1 \log T + c_2 \log D + c_3 \qquad (1)$$

where M_D is the duration magnitude, T is the duration of the waveform of (arrival of P waves up to the time where the recorded pp amplitude is less than 2 mm) and D is the epicentral distance.

For the constants c_1 and c_2 we adopted the values, 1.97 and 0.0012 respectively, determined for the broader Greek area (Kiratzi, 1984). Constant, c_3, is calculated for every station used for the magnitude estimation. In order to estimate the value c_3, the recordings of 28 strong earthquakes of the broader area of Santorini from the Greek National Networks and the National Networks of adjacent countries were used. The presence of an active extensional tectonic regime in the area under study is confirmed by the distribution of strong earthquakes ($M_w > 4.2$) along the Amorgos Fault, as is shown in figure (7).

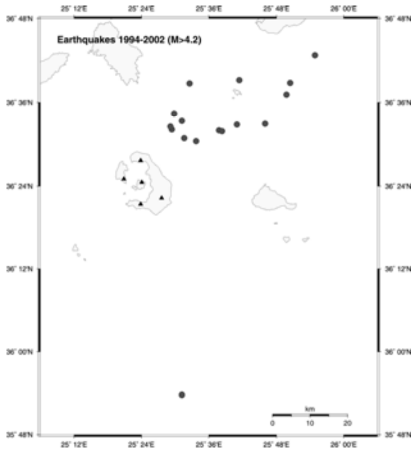

Figure 7. Epicenters of strong earthquakes (M>4.2) of the period 1994-2002 occurred in the broader area of the Santorini Volcano.

The final values, c_3, as well as the number of recordings used for each station are shown in table (2). Equation *(1)* was used for the estimation of magnitudes for all the events and the magnitudes were up to 5.0.

Station	C_3	Obs
AKR	0,685	13
KAM	0,686	20
KER	0,619	11
OIA	0,699	18
RIB	0,545	11

Table 2. Seismological stations and the corresponding values of constant, c_3, along with the number of recordings used for each station.

4. Estimation of 1-D Earth Model

The hypocenters of the earthquakes have been estimated using HYPOELLIPSE (Lahr, 1989, 1999) and scattered over a great depth range due to the simplified earth model used. Hence, it was necessary to determine a new velocity model for the area under study in order to improve the accuracy of the determined hypocenters, as well as to obtain a preliminary knowledge of the volcano velocity structure. For this purpose a 1-D inversion of travel times has been used. The principles of the inversion of travel times from local earthquakes are essentially those described in the original work of Aki and Lee (1976).

The travel time residual is a function of the perturbations of the event's hypocentral parameters and the velocity model. If this function is linearized using an initial approximate solution, a system of equations can be defined:

$$\mathbf{A}\,\mathbf{x} + \mathbf{B}\,\mathbf{v} = \mathbf{r} \qquad (2)$$

where, x and v, are the velocity and hypocentral perturbation vectors, r, is the residual vector and, A and B, are appropriate Jacobian matrices.

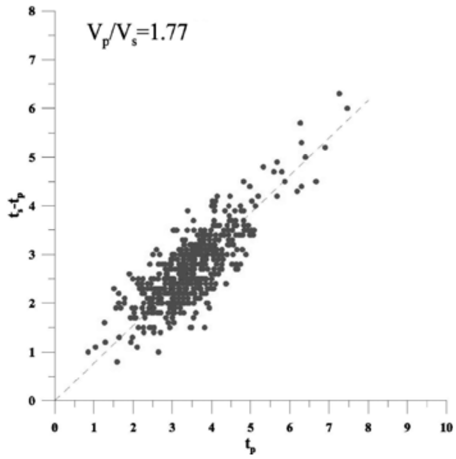

Figure 8. Wadati diagram of the area under study.

The 1-D velocity model for the area under study was derived from the computer program VELEST (Kissling et al. 1994, 1995). Using the final data set of the best-located earthquakes, a Wadati diagram has been made in order to estimate the **Vp/Vs** ratio (Fig.8). The calculated ratio, which has been used for the inversion, has been found equal to 1.77.

To probe the dependence of the solution on the initial model we used three different initial velocity models, with varying model geometry (layer thickness). In particular, we used one model similar to the local velocity model used for the hypocenter estimation (Papazachos and Nolet, 1997), one with extremely high crustal velocities and one model where the velocities follow a gradient. Then we produce for each initial model random models similar to the specified characteristics. Finally, we concluded to a group of approximately 60 initial models (Fig.9a). Using VELEST we calculated final velocity models for every initial model. Then we calculated the three final average 1-D velocity models and we observed that within the well-resolved depth range these were almost identical

(Fig.9b). The final 1-D velocity model with the minimum RMS region was selected as the suggested velocity model for the area under study (the model with the black solid line in Fig.9b).

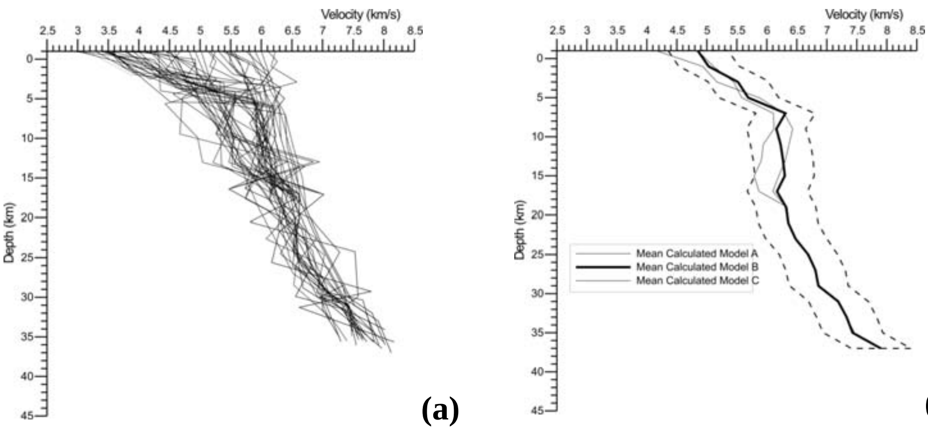

Figure 9(a). Group of the initial velocity models used. (b) Final 1-D calculated velocity models.

Due to the lack of data below 20 km, two velocity models for the broader region of the Aegean Sea have been considered to represent a typical model below this depth. The first velocity model considered has been proposed by Papazachos and Nolet (1997) for P-wave velocities using travel times; while the second velocity model considered has been proposed by Karagianni et al. (2002) for velocities from surface wave inversion. For the final 1-D velocity model we adopted Papazachos and Nolet model, since our data are mainly P-wave arrivals, similar to their study. The model is shown in figure (10).

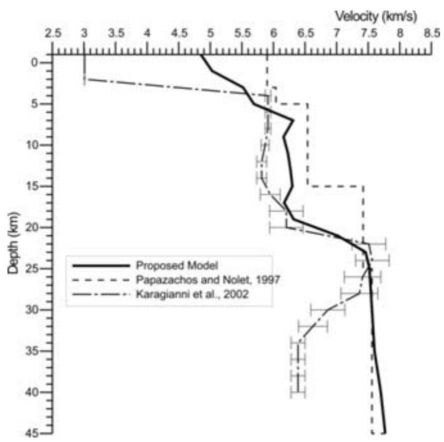

Figure 10. The final model (black solid line) suggested for the area under study.

5. Relocation of the earthquakes

The initial data set of 1076 earthquakes has been recalculated by using the computer program HYPOELLIPSE (Lahr, 1989, 1999) with the new suggested velocity model for the area under study (Table 3).

Depth (km)	V_P (km/s)	V_S (km/s)
0,0	4,85	2,74
1,0	5,03	2,84
3,0	5,52	3,12
5,0	5,69	3,21
7,0	6,31	3,56
9,0	6,16	3,48
11,0	6,23	3,52
13,0	6,27	3,54
15,0	6,30	3,56
17,0	6,17	3,48
19,0	6,32	3,57
21,0	7,02	3,96
23,0	7,46	4,21
25,0	7,52	4,25
30,0	7,56	4,27
35,0	7,60	4,29
40,0	7,70	4,35
45,0	7,77	4,39

Table 3. The final 1-D velocity model suggested for the area under study.

In figure (11) the final epicenter distribution of the best-located earthquakes of the period 1994-2002 is shown. As is shown, there are minor differences in comparison with the epicenters estimated using the initial velocity model, while there are some notable differences in the hypocentral depths. In particular, the new recalculated earthquake foci can be found in shallower depths. In general, the epicenters estimated with the initial velocity model are quite reliable.

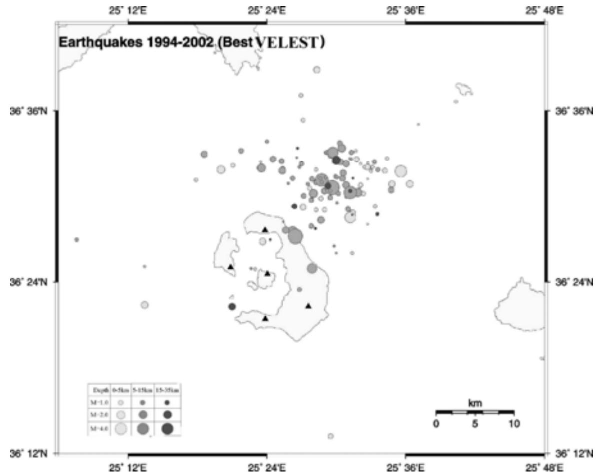

Figure 11. Epicenters of the best-located earthquakes of the period 1994-2002 recalculated using the suggested 1-D velocity model for the area under study.

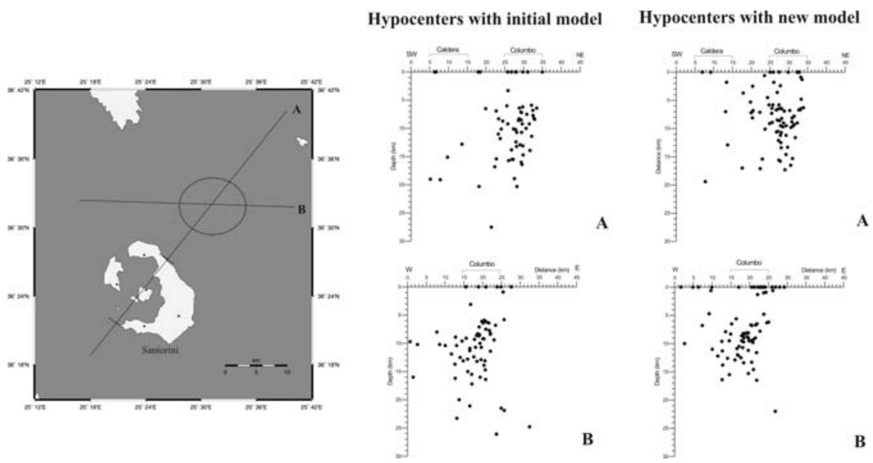

Figure 12. Cross-sections of the best-located earthquake foci along lines A and B determined using the initial 1-D velocity model and the suggested velocity model for the area under study.

Figures (12) and (13) represent cross-sections of the best-located earthquake foci estimated using both the initial and the suggested velocity model. The distribution of the hypocenters has improved by using the new velocity model, as is shown in both figures. Particularly, the new hypocentral depths are scattered over a smaller range than in the initial model. Moreover, a better coverage for the hypocentral depths varying between 0 km and 5 km is observed.

Therefore, it can be concluded that the main seismic activity, which is associated with the volcanic processes as well as with the tectonic regime of the broader area of Santorini Island, takes place, mainly, at the Columbo Reef (Fig.12A and Fig.12B). The seismic activity under the caldera of the volcanic center is quite deep, at a depth range between 10 and 25 km (Fig.12A and Fig.13C). From all cross-sections we can hypothesize the presence of a volcanic submarine edifice in relatively shallow depths at Columbo Reef. In figure (13D) it is clear that the magmatic chamber and the main magmatic vent are located at depths between 5 and 20 km. These results would appear to suggest that the observed earthquakes originate in domains with complex characteristics and the discrimination of the volcanic earthquakes is quite difficult.

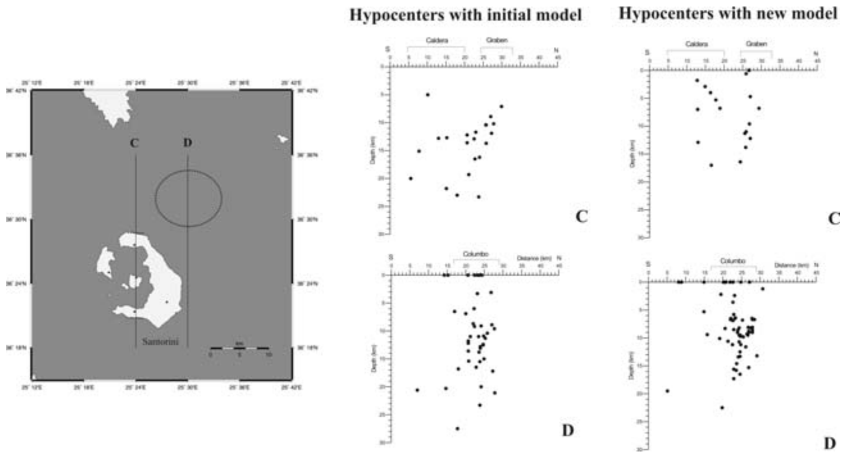

Figure 13. Cross-sections of the best-located earthquake foci along lines C and D determined using the initial 1-D velocity model and the suggested velocity model for the area under study.

6. Conclusions

Low seismic activity is observed in the broader area of the Santorini Island (approximately 1000 earthquakes during a time period of 8 years). In particular, two clusters of epicenters have been identified in the broader area of the Santorini Volcano. The first cluster is located in the caldera of the volcanic center and is associated with the volcanic process in the Nea Kameni Island. The second (larger one) cluster is located near the northeastern edge of the Santorini Island at Columbo Reef and is connected with the volcanic process at this reef. These clusters can be appropriately associated with the two main tectonic features in the area under study. The main one, which has an N60^{o}E direction, corresponds to the continuation of the Amorgos fault in the area, while the secondary tectonic line, which has an approximately EW direction, is probably related with the southern edge of a submarine graben, northern of Santorini.

It is obvious that, these two tectonic lines (faults) and their intersection are related to the volcanic activity of the Columbo Reef but the available data and their accuracy are not enough for a safe explanation. A systematic study of the region, with more and better-distributed seismographs should, in the future, determine the true dimensions of the volcanic edifice at the Columbo Reef with higher precision.

Despite the poor spatial coverage of the recording stations and the lack of S phases, an appropriate equivalent 1-D earth model has been defined, in order to improve the accuracy of the determined hypocenters, as well as to have a preliminary knowledge of the shallow velocity structure of the Santorini Volcano area.

ACKNOWLEDGEMENTS

This work has been funded by the E-RUPTION E.C. project (contract EVR1-2001-00024). We are grateful to the Institute for the Study and Monitoring of the Santorini Volcano (ISMOSAV), Santorini, Greece, for providing the data from the local seismological network. Thanks are also due to Wessel and Smith (1995) for their generous distribution of the GMT software used to generate most of the maps of this study.

REFERENCES

Aki, K. and Lee, W.H.K., 1976. Determination of three-dimensional velocity anomalies under a seismic array using first P arrival times from local earthquakes: A homogeneous initial model. *Journal of Geophysical Research*, 81: 4381-4399.

Bond, A. and Sparks, R.S.J., 1976. The Minoan eruption of Santorini, Greece. *Journal of the Geological Society of London*, 132: 1-16.

Druitt, T.H., Mellors, R.A., Pyle, D.M. and Sparks, R.S.J., 1989. Explosive volcanism on Santorini, Greece. *Geophysical Magazine*, 126: 95-126.

Druitt, T.H. and Francaviglia, V., 1992. Caldera formation on Santorini and the physiography of the islands in the late Bronze Age. *Bulletin of Volcanology 54*, No. 6: 484-493.

Druitt T.H., Davies, M.S., Edwards, L., Sparks, R.S.J., Mellors, R.M., Pyle, D.M., Lanphere, M. and Barreirio, B., 1999. The Santorini

Volcano. *Geological Society Special Memoir, 19*, Geological Society Pub House, London.

Friedrich, W.L., 1994. Feuer im Meer. Spektrum Akademischer Verlag, Heidelberg, Berlin.

Friedrich, W.L., 2000. Fire in the Sea: Volcanism and the Natural History of Santorini. Cambridge University Pr., Cambridge.

Fytikas, M., Innocenti, F., Manetti, P., Mazzuoli, R., Peccerillo, A. and Villari, L., 1985. Tertiary to Quaternary evolution of volcanism in the Aegean region. In: *The Geological Evolution of the Eastern Mediterranean, Special Publ. Geol. Soc.*, 17: 687-699.

Fytikas, M., Kolios, N. and Vougioukalakis, G., 1990. Post-Minoan Volcanic Activity of the Santorini Volcano. Volcanic hazard and risk, forecasting possibilities. In: Hardy, D.A., (Editor), *Thera and the Aegean World III, 2*. The Thera Foundation, London, pp. 183-198.

Georgalas, G.C., 1962. Catalogue of the Active Volcanoes of the World including solfataras fields, Greece.

Heiken, G. and McCoy, F., 1984. Caldera Development During the Minoan Eruption, Thera, Cyclades, Greece. *Journal of Geophysical Research*, 89: 8441-8462.

Heiken, G. and McCoy, F., 1990. Precursory activity to the Minoan Eruption, Thera. In: Hardy, D.A., (Editor), *Thera and the Aegean World III, 2*. The Thera Foundation, London, pp. 79-88.

Karagianni, E.E., Panagiotopoulos, D.G., Panza, G.F., Suhadolc, P., Papazachos, C.B., Papazachos, B.C., Kiratzi, A., Hatzfeld, D., Makropoulos, K., Priestley, K. and Vuan, A., 2002. Rayleigh wave group velocity tomography in the Aegean area. *Tectonophysics*, 358: 187-209.

Kilias, A.A., Mountrakis, D.M., Tranos, M.D. and Pavlides, S.B., 1996. The prevolcanic metamorphic rocks of Santorini island: structural evolution and kinematics during the Tertiary (South Aegean, Greece). Proceedings of 2nd Workshop on European Laboratory Volcanoes, May 2-4 1996, Santorini, Greece, pp. 23-36.

Kiratzi, A., 1984. Magnitude scales for earthquakes in the broader Aegean area. Ph.D. Thesis (in Greek), University of Thessaloniki, Greece.

Kissling, E., Ellsworth, W.L., Eberhart-Phillips, D. and Kradolfer, U., 1994. Initial reference models in local earthquake tomography. *Journal of Geophysical Research*, 99: 19’635-19’646.

Kissling, E., Kradolfer, U., and Maurer, H., 1995. VELEST User’s Guide (Version 3.1). Institute of Geophysics, ETH Zurich.

Lahr, J.C., 1989. HYPOELLIPSE/Version 2.0: A computer program for determining local earthquakes hypocentral parameters, magnitude, and first-motion pattern. U.S. Geological Survey Open-File Report 89-116, p. 92.

Lahr, J.C., 1999. HYPOELLIPSE/Version 1.0: A computer program for determining local earthquake hypocentral parameters, magnitude, and first-motion pattern (Y2K Compliant Version). U.S. Geological Survey Open-File Report 99-23.

Le Pichon, X. and Angelier, J., 1979. The Hellenic arc and trench system: a key to the Neotectonic evolution of the eastern Mediterranean area, *Tectonophysics*, 60: 1-42.

Le Pichon, X. and Angelier, J., 1981. The Aegean Sea, *Phil. Trans. R. Soc., London*, 300: 357-372.

McKenzie, D.P, 1978. Active tectonics of the Alpine-Himalayan belt: the Aegean Sea and surrounding regions, *Geophysical. J. R. astr. Soc.*, 55: 217-254.

Mountrakis, D.M., Pavlides, S.B., Chatzipetros, A., Meletlidis, S., Tranos, M.D., Vougioukalakis, G. and Kilias, A.A., 1996. Active deformation of Santorini. Proceedings of 2nd Workshop on European Laboratory Volcanoes, May 2-4 1996, Santorini, Greece, pp. 13-22.

Panagiotopoulos, D.G., Stavrakakis, G., Makropoulos, K., Papanastasiou, D., Papazachos, C.B., Savvaidis, A.S. and Karagianni, E.E., 1996. Seismic monitoring at the Santorini volcano. Proceedings of 2nd Workshop on European Laboratory Volcanoes, May 2-4 1996, Santorini, Greece, pp. 311-324.

Papazachos, B.C. and Comninakis, P.E., 1969. Geophysical features of the Greek island arc and eastern Mediterranean ridge. Proceedings of the C.R. des Séances de la Conference Reunite a Madrid, 16, pp. 74-75.

Papazachos, B.C. and Comninakis, P.E., 1971. Geophysical and tectonic features of the Aegean arc. *Journal of Geophysical Research*, 76: 8517-8533.

Papazachos, B.C., 1989. Long and short term prediction of the volcanic eruptions in Santorini. In: Proceedings of 3rd International Congress "Thera and the Aegean World III", *Volume 2*, September 4-8 1989, Santorini, Greece, pp. 125-129.

Papazachos, B.C. and Panagiotopoulos, D.G., 1993. Normal faults associated with volcanic activity and deep rupture zones in the southern Aegean volcanic arc. *Tectonophysics*, 220: 301-308.

Papazachos, B.C., Mountrakis, D.M., Papazachos C.B., Tranos, M.D., Karakaisis, G.F. and Savvaidis, A.S., 2001. The fault which have caused the known major earthquakes in Greece and surrounding region between the 5th century BC and today. Proceedings of 2nd National Conference Anti-Seismic Engineering and Technical Seismology, November 28-30 2001, Thessaloniki, Greece, pp. 17-26.

Papazachos, C.B. and Nolet, G., 1997. P and S deep structure of the Hellenic area obtained by robust nonlinear inversion of travel times. *Journal of Geophysical Research*, 102: 8349-8367.

Pavlides, S.B. and Valkaniotis, S., 2003. Tectonic regime of Santorini-Amorgos area, Proceedings of International Conference "The South Aegean Active Volcanic Arc: Present Knowledge and Future Perspectives", September 17-20 2003, Milos Island, Greece. *Book of abstracts*, p. 76.

Perissoratis, C., 1996. The Santorini volcanic complex and its relation to the stratigraphy and structure of the Aegean arc, Greece. *Marine Geology, 128*, pp. 37-58.

Seidenkrantz, M.S. and Friedrich, W.L., 1992. Santorini, Part of The Hellenic Arc: Age Relationship of Its Earliest Volcanism. In: Seidenkrantz, M.S. (Editor), *Foraminiferal Analyses of Shelf Areas. Stratigraphy, Ecology and Taxonomy.* Ph.D. Dissertation, University of Aarhus, pp. 41-65.

Shirokova, E.I., 1972. Stress pattern and probable motion in the earthquake foci of the Asia-Mediterranean seismic belt. In: L.M. Balakina et al. (Editors), *Elastic Strain Field of the Earth and Mechanisms of Earthquake Sources.* Nauka, Moscow, p. 8.

Wessel, P. and Smith, W., 1995. New version of the Generic Mapping Tools. Eos. Trans. Am. Geophys. Union 76, p. 329.