Investigation of geothermal structures by magnetotellurics (MT): an example from the Mt. Amiata area, Italy

Gianni Volpi\textsuperscript{a,b}, Adele Manzella\textsuperscript{a,}\textsuperscript{*}, Adolfo Fiordelisi\textsuperscript{b}

\textsuperscript{a}CNR—Institute of Geosciences and Earth Resources, Via Moruzzi 1, 56124 Pisa, Italy
\textsuperscript{b}ENEL Greenpower, Via A. Pisano 120, 56100 Pisa, Italy

Received 9 April 2002; accepted 25 February 2003

Abstract

During 1999 a magnetotelluric (MT) survey was carried out on the southern margin of the Mt. Amiata geothermal region (Tuscany, Italy), with the aim of defining the shallow and deep electric structures related to the local geothermal reservoirs and system heat recharge. Local and remote data were collected along a SW–NE profile and processed with two different robust algorithms. After a detailed study of the EM strike, the data were inverted and two-dimensional (2D) models of electrical resistivity and impedance phase were computed. The interpretation revealed a good correlation between the features of the geothermal field and resistivity distribution at depth. In particular, a shallow conductor (0.5–4 km) detected by the MT survey shows a good correlation with the areal extension of the geothermal reservoirs.

© 2003 CNR. Published by Elsevier Science Ltd. All rights reserved.

Keywords: Geophysics; Magnetotellurics; Geothermal exploration; Mt. Amiata; Italy

1. Introduction and geological setting

The aim of this project was to define the shallow and deep electrical structures of the southern side of the Mt. Amiata geothermal field. An MT survey carried out in 1994 on the western margin of the Mt. Amiata area highlighted its complex structure and the relations between conductive anomalies and the geothermal reservoir (Manzella et al., 1999; Fiordelisi et al., 2000).
The most productive areas of the Amiata geothermal fields were explored in an MT survey carried out during the summer of 1999, to investigate the relationship between the resistivity anomalies defined by MT data interpretation and the geothermal features of the Amiata area, in an area whose shallow structure is well known.

Mt. Amiata is an extinct, recent volcano (0.3 Ma) located in southern Tuscany (Italy). The Mt. Amiata geothermal field is a water-dominated system, with temperatures reaching 350 °C at 3 km depth, which is exploited mainly for the production of electric power. The geothermal system is characterised by a shallow reservoir within the carbonate allochthonous units of the Tuscan Nappe at depths of 0.5–1 km, and a deeper reservoir in the local metamorphic basement at depths greater than 2 km. The entire system is fed by meteoric fluid circulation through fractures and faults and heated by the ancient magma chamber of the volcanic system. Granitic dykes, thermometamorphic aureolas and the gas composition of the extracted fluids testify to the presence of a magmatic body (see e.g. Gianelli et al., 1997). The gravimetric and seismic data indicate the presence of a low-density, seismically transparent body at a depth of about 6–7 km below Mt. Amiata (Bertini et al. 1995, Orlando et al., 1994). The body is believed to be broader in width than the volcano and the known geothermal system and to be partially molten; it probably deepens and then disappears beyond the volcanic edifice (ENEL, 1997).

From a geo-structural point of view, southern Tuscany is characterised by a NNW–SSE system of normal faults defining tectonic depressions filled with Neogenic deposits. It is a tectonically active area with a thin crust and regional heat flow values of more than 100 mW/m². The compressional tectonic phase, linked to the Apenninic orogenesis, was followed by a period of extension and ground uplift, so that older units of allochthonous flysch and carbonatic sequences belonging to the Tuscan Nappe outcrop in some places. Excluding the volcanics, which cover parts of the volcano, the geological and corresponding electrical sequences here are similar to the rest of southern Tuscany and can be easily recognised in deep-well resistivity logs recorded in the geothermal fields of southern Tuscany. The shallow units, which cover most of the area not covered by the volcanites, consist of clayey Neogenic deposits that overlie shaley allochthonous flysch units, all of which are electrically conductive (10–20 ohm-m). The carbonate formations of the Tuscan Nappe comprise limestones, sandstones and clayey schists, and are only moderately resistive (usually around 100 ohm-m). Anhydritic formations of the same Nappe, interbedded with the carbonates, can, however, be very resistive, reaching values of 1000–1500 ohm-m. Beneath the surface sedimentary cover are two metamorphic units: (1) an upper unit of quartzites, phyllites (with some small amounts of graphite), and micaschists, and (2) a lower unit, primarily of gneiss. The resistivities of these metamorphic units were measured in a few deep wells in the Larderello area, and are quite low (about 100 ohm-m) compared to the anhydrites. It is possible, however, that these resistivities are representative of the upper part of the metamorphic basement only, which may have been fractured when the overlying allochthonous units slid over them. In places where the units are not fractured and massive, we can expect higher resistivity values on the order of several hundred ohm-m.
A deep (>6 km), partially molten magmatic body is believed to act as the heat source for the geothermal system. Due to the volcanic nature of the zone and the rifting that affects the entire Tyrrhenian area, a large part of the upper crust in the Amiata area is characterised by low resistivity values. These conductive conditions make it difficult to detect 3D conductors.

2. Methodology and data acquisition

The sensitivity of electrical conductivity to the presence of small quantities of interconnected fluids makes the electric (DC) and electromagnetic (EM) methods particularly applicable to geothermal areas, especially where a water-dominated reservoir is heated by a hot, partially molten magmatic body. Both the water-reservoirs and the magmatic body produce conductive anomalies that can be defined by EM methods.

Of all the EM methods, magnetotellurics seems to be the most appropriate since the investigation depth of MT can easily reach several km, being correlated to the different frequencies of the Earth's natural, time-variable electromagnetic field. The ability of MT to detect conductors embedded in resistive media is related to the dimensions of the conductor, to the resistivity contrast between the conductor and the host rocks and to the specific electrostratigraphic distribution of the study area (Newman et al., 1985).

The high conductivity of the shallow subsurface prevents our using direct current (DC) resistivity methods since the penetration depth of the latter is limited to the top of the first resistive unit. Wide-band (100–0.001 Hz) MT data can also explore the earth structures in conductive environments. However, the data from the natural, weak-source MT method applied in an intensely exploited geothermal zone can be affected by the presence of EM noise. An MT survey carried out in 1994 on the western margin of the Mt. Amiata field (Manzella et al., 1999; Fiordelisi et al., 2000; B–B' in Fig. 1a) and another MT survey in the Larderello geothermal area (Fiordelisi et al., 1995) provided interesting results for defining the characteristics of the EM noise affecting the geothermal areas of southern Tuscany.

EM noise sources were recognised in the high and low frequency bands. At high frequencies, geothermal wells and electric power plants caused the greatest problems. Most of this noise, however, can be effectively removed by notch filtering and synchronous acquisitions, using the remote-reference acquisition technique (Gamble et al., 1979). A reference station at a distance of 1–2 km is quite capable of dealing with high frequency noise, including the noise from power lines.

A more difficult task is the elimination of low frequency EM noise, which mainly derives from a particularly strong source of low frequency-correlated signal related to the local DC electrified railways (Fiordelisi et al., 1995; Larsen et al., 1996). This noise is produced by the electric discharge to the ground as a train passes, when an EM wave is generated by the dipoles created between the train and the discharging points. This EM signal extends for a number of kilometres, depending on ground resistivity, and can mask any natural EM signal. Due to its dipolar nature, the E and
Fig. 1. (a) Geological sketch map of the Mt. Amiata area, showing the known surface evidence of the main faults. Line A–A′: the acquired MT profile. Line B–B′: the MT profile acquired in a previous survey (Manzella et al., 1999). (b) Detailed layout of magnetic and telluric sensors for synchronous recording. Inside the circle: a main site (see text).
H fields are closely correlated and in-phase. This so-called “train effect” appears as a linear increase in the log–log diagram of apparent resistivity versus period, with a slope of one or even higher and a corresponding E/H phase shift near to zero degrees.

Synchronous recordings on properly spaced sites and specifically designed techniques for computing MT transfer functions can, however, significantly improve data quality even in the presence of this noise (Larsen et al., 1996). An island was chosen for the remote site since the study area is close to the coast and there is a large expanse of conductive seawater that effectively screens the correlated noise. The amplitude of the correlated noise signals diminishes to trivial levels at distances greater than $1.5 (\rho T)^{1/2} \text{km}$, where $\rho$ is the resistivity and $T$ is the period (Strangway et al., 1973). If the maximum period of interest is taken to be $1000 \text{s}$ and sea-water resistivity is $0.25 \text{ohm-m}$, this means that the distance must be greater than $24 \text{km}$. The island of Capraia, lying $50 \text{km}$ from the coast and $80 \text{km}$ from the target site (Fig. 1), was chosen as the remote MT recording site. The island effect, which smoothly distorts the MT transfer function, does not degrade the coherence between local and remote time series.

The Mt. Amiata MT survey discussed in this paper was carried out in June 1999. The four horizontal electric and magnetic MT field components in the frequency band $0.000556–384 \text{Hz}$ were collected at 10 sites at an average distance of $1 \text{km}$ along a SW–NE oriented profile (Fig. 1a, profile A–A'). These sites have been termed the “main sites”. Thirty-seven additional recordings were performed among the main sites at the same frequency band, acquiring only the electric (telluric) components. These dipoles, spaced $200 \text{m}$ apart, have been termed the “satellite sites”. The telluric dipoles of the main sites and the satellite sites were $200 \text{m}$ long and were aligned parallel and perpendicular to the profile direction, wherever topography and vegetation permitted. The satellite sites data, combined with the synchronous magnetic data at the central main sites, provided a continuous MT profile that has proved effective in defining the details of the shallower structures (Manzella et al., 1999; Fiordelisi et al., 2000). Data were acquired during the night, when test soundings indicated that noise tended to be at a minimum. The duration of the acquisition was $15 \text{h}$.

In the MT profile layout (Fig. 1b), each main site is located at the centre of a group of five sites (with two satellite sites on either side of the main site). The main site of each group is identified by the suffix “.3” which follows the number of the group (e.g. 7.3 is the main site of group 7 while 7.1, 7.2, 7.4 and 7.5 are the satellites).

Two types of remote site were used to tackle the problem of EM noise. A far-remote MT recording site was set up $80 \text{km}$ away on the island of Capraia (in Fig. 1b), where correlated train signals were absent. Data from this site were used to separate the Amiata measurements into MT impedances and correlated noise signals. In order to improve high frequency data, time series data were acquired simultaneously on two main sites (local-remote referencing). Considering the low resistivity values of the shallower formations, a spacing of $1 \text{km}$ was felt to be long enough to reference sites in the high frequency band. At a distance of $1 \text{km}$ and with an average resistivity of $10 \text{ohm-m}$, the amplitude of the correlated noise can be
neglected at frequencies higher than $\rho(1.5/1)^2 = 22.5$ Hz. Synchronous recordings were then performed nightly on two main sites, on four–six satellite sites and on the Capraia reference site.

3. Data analysis

Time series at all sites were processed using two different robust algorithms, the first of which was a standard robust algorithm provided by Phoenix Geophysics and based on Jones et al. (1989). In the second phase, all time series data were re-processed using the new version of the robust code specifically designed to extract MT transfer functions from time series highly contaminated by correlated EM noise signals (Larsen et al., 1996). Larsen’s code separates the noise from the main signal; first it discards the uncorrelated noise, and then, by comparing the local H fields with the remote H field, it attempts to separate the correlated noise. Based on the assumption that resistivity is mainly determined by the local E field, and taking into account a large number of synthetic standard impedance curves, the algorithm produces the best smooth robust noise-free estimate of the transfer functions (see Larsen et al., 1996 for details).

Data from each site were processed in single-site mode (SS), local-remote reference mode (LR) and far-remote reference mode (FR) in order to assess data quality in general. The best data were then chosen and combined, for both high and low frequencies.

Data obtained by the Jones and Larsen algorithms provided interesting indications on the impedance strike direction and the impedance skew. Good estimates of these parameters are of fundamental importance for inferring the geometry and the dimensionality of the target structure.

The strike values were similar at all frequencies and showed values ranging between $-20^\circ$ and $+20^\circ$, with the exception of groups 6 and 7 that showed strike values up to $+45^\circ$. Indeed, central sites (groups 6, 7 and 8) were particularly affected by wide-band EM noise because of their vicinity to geothermal wells and power stations. The average strike direction of $-10^\circ$ was in good agreement with the regional geological strike, defined by the NNW–SSE Apennine direction. Skew values below 0.4 also indicated that the study area could be approximated to a 2D geometry. A 2D interpretation, however, can be considered only partially appropriate since the area is very complex. Ledo et al. (2002) have demonstrated the importance of removing near-surface galvanic distortion in any 2D interpretation of 3D MT data, in order to reduce the error sources in the 2D interpretation as well as in the 3D interpretation.

The singular value decomposition of the impedances was then carried out, obtaining, by eigenstate analysis, the maximum and minimum impedance amplitude and phase (La Torraca et al., 1986). After decomposition most of the MT sounding curves show a strong 2D effect with a clear separation between the curve with the electric field parallel to the strike (TE mode) and the curve related to current circulation normal to the strike (TM mode). This 2D effect confirmed the indications derived from the impedance skew.
In order to eliminate the problem of static shift, the data were compared to the geoelectrical data available in the area. The static shift problem arises from local resistivity perturbations that mainly affect the electric fields, causing a frequency-independent shift of the apparent resistivity curves; phase data, on the other hand, are not affected. The static shift of MT data is usually corrected by an independent geophysical method close to the MT site in order to determine an undistorted 1-D shallow model. The MT apparent resistivity curves can then be corrected to the undistorted values calculated from this model. In this work we have used Schlumberger VES data to correct the MT data. Although in a purely galvanic technique such as the VES there are also boundary charges affecting the measurements, the mobility of the source and receiver electrode arrays during the VES data acquisition could contribute to reduce the effects, so they are not as harmful as in MT.

The analysis of the apparent resistivity curves revealed some common features in adjacent sites along the acquisition profile. It was possible to distinguish the following main categories of sites (Fig. 2).

Fig. 2. Common features of adjacent sites along the acquisition profile. Three categories of sites can be recognised: external sites (left), central geothermal sites (centre) and peripheral geothermal sites (right). The fig. shows the apparent resistivity curves of sites 2.3, 6.3 and 4.3, respectively, obtained by LR processing and La Torraca decomposition.
1. **External sites**, on the south-western margin of the profile and 5 km from the most intensely exploited geothermal zones (groups 1, 2 and 3). These are characterised by the highest shallow resistivity values (40 ohm-m) and by TM over TE mode data. A typical resistive–conductive–resistive sequence is visible;

2. **Central geothermal sites** (groups 6 and 7). Located in the most intensely exploited part of the geothermal field, very close to wells and electric power stations, these sites have the lowest resistivity values (<10 ohm-m) in the high frequency band (384–1 Hz). The TE over TM mode data testify to the presence of a deep conductor;

3. **Peripheral geothermal sites** (groups 4, 5, 8, 9 and 10). They border the central part of the field. The TM over TE mode data highlight their peripheral position with respect to the most conductive parts of the deep geothermal structures.

Data quality varies between the high and low frequency results. The **high frequency** data were of good quality and there was not much difference between SS and LR or FR reference processing in this band, both for the Jones and Larsen algorithms.

For the **low frequency** band, local SS as well as LR processing produced very small error estimates, but the low frequency response was corrupted by the correlated noise signals. In the FR reference results the error bars at low frequencies are larger than those obtained from SS or LR processing, since a part of the data was eliminated by the referencing algorithm. The use of Larsen’s code in the FR processing provided slightly better results in the low frequency band for the sites of groups 4 and 5 only, whereas the other sites did not improve with respect to the results of the

---

**Fig. 3.** Apparent resistivity plots of site 4.3 using the island of Capraia site as remote: Jones’ standard robust remote-reference processing on the left; Larsen’s robust remote-reference processing on the right. Larsen’s code was slightly more effective for extracting good MT transfer functions.
Jones’ processing (Fig. 3). Larsen’s code was expected to provide the best estimate in the case of this particular noise, but it proved to be extremely dependent on the data quality of the remote site, which was particularly good only during the recording of the groups of sites 4 and 5. In order to obtain the best estimate from Larsen’s code, three conditions must be fulfilled: (1) the remote site must not be affected by noise; (2) the local signal must not be overwhelmed by the noise; (3) the acquisition must be long enough to provide many data segments even after the reduction operated by the processing technique. The latter conditions imply that the higher the noise, the longer the acquisition must be. In our case, the acquisition was not very long and the remote site was not really quiet enough, except when recording sites 4 and 5. In general, for both Jones’ and Larsen’s processing results, the FR data provide the correct shape of the apparent resistivity and phase curves but the data appear to be scattered. By comparing LR and FR data, however, we observe that the data are affected by correlated noise only in a restricted frequency band, between 1 and 0.1 Hz at the most. The lowest frequency parts of the apparent resistivity and phase curves (>0.1 Hz) maintain the correct shape but are shifted for the effect of noise. This type of behaviour is illustrated in Fig. 4 for the main site 8.3; the same effect can be observed in all the site data. Since the LR data were of very good quality and appeared to be affected by noise only in a limited frequency band, we used the decomposed LR data, masked the data in the noisy frequency band and, comparing them with FR data, shifted the apparent resistivity and phase values to their correct position. We then proceeded to the modelling.

4. Modelling and interpretation

As described in the Introduction and emphasised in Section 3, the Amiata area is characterised by a NNW–SSE geological strike (Apennine mountain chain direction) that runs approximately orthogonal to the MT profile and is consistent with the indications derived from the analysis of the electric strike. 2D modelling should therefore be appropriate for reconstructing the main electrostratigraphic features. There are many wells in the vicinity of the chosen profile, and well stratigraphy, together with the seismic reflection data, has allowed us to define a geological structure below this profile (Fig. 5). The three westernmost wells are not productive, although the temperature is high. All the other wells are productive. The shallow wells produce from the shallow reservoir, and the two deep wells from the deep reservoir.

The data were inverted for the 2D conductivity models using a regularised inversion algorithm (Rodi and Mackie, 2001). The inversions were computed using many a priori models, but of two different types: (1) models with a uniform earth resistivity, and (2) models based on geology, as defined in Fig. 5, using the resistivity values described in the Introduction and testing different values of resistivity for the metamorphic basement. Topography was taken into account, as well as the effects of seawater. All model parameters were allowed to vary freely during the inversion, except for the cells associated with seawater, which were locked to a value of 0.25
Fig. 4. Comparison between local remote (LR) (left) and far-remote (FR) data at site 2.3 (center). The LR apparent resistivity curves are affected by correlated noise in the 1–0.1 Hz frequency range (in the rectangle), with the lowest frequency part of the LR apparent resistivity curve (>0.1 Hz) shifted to unrealistic high resistivity values. On the right, the edited LR data used for inversion: the unrealistic data have been masked and the low frequency data have been shifted following the indications given by FR data.
ohm-m. A noise floor of 5% for the apparent resistivity and of 2% for the phase were used. Several inversions were performed of both the TM and TM–TE modes. Since TM mode typically suffers less 3D distortion than TE (Wannamaker et al., 1984), some inversions took into account only the TM mode data. In order to verify the effect of finite strike on our data, we also inverted the TM mode data starting from an a priori model obtained by joint inversion of TM and TE mode data.

The resulting models are sensitive to structures to depths of approximately 5–6 km, whereas the data show no detectable variations of resistivity at greater depth. Comparison of the many models obtained with 2D inversion allowed us to define the robust features, i.e. the characteristics that do not depend on the a priori model but are required by data.

The most conspicuous feature is the presence of a thick conductive anomaly below the central area of our profile, defined by sites 4–8. The lateral dimensions of this conductive anomaly are robust, whereas its resistivity and thickness depend on the mode used for inversion: the anomaly appears thinner and more conductive when both TE and TM mode data are inverted. Another robust feature is the presence of a resistive anomaly below sites 2 and 3 at depths of more than 1 km b.s.l., whereas the rest of the resistivity values at the same depths depend greatly on the resistivity used in the a priori model. These values range from a few hundred to thousands of ohm-m, becoming uniform at more than 10 km b.s.l.

The best inversion result, i.e. the best match with well data, was obtained with the a priori geological model, whose metamorphic basement was considered very resistive. The joint inversion of TE and TM mode data, as well as the inversion of TM only data, are shown in Fig. 6. The normalized root mean square (rms) misfit achieved was 8.5 for TE–TM inversion, and 4.5 for TM inversion. The difference between the two models is evident, and is a clear effect of the attempt to 2D model 3D data. Ledo et al. (2002) have shown that the effects of finite strike are not
significant when the 3D conductive structure is located below the profile and the structure has a strike extent greater than about one-half of a skin depth. In our case, the length scale of the 3D structure is not very large and the TM only inversion provides the best results. In Fig. 7 we compare the TM only inversion of Fig. 6 with

Fig. 6. The resistivity models resulting from the inversion of: (a) TM only mode (root mean square = 4.5); (b) both TM and TE modes (rms = 8.5).
the geological and well data of Fig. 5. There is a very good agreement between the conductive anomalies and the most productive wells. The two deep wells at the centre of our profile correspond to the deep anomaly defined by MT data, while all the shallow productive wells fall within the generally conductive shallow subsurface area between sites 4 and 10. The shallow structure reflects the lithological response of shallow formations, which are quite conductive. It is not easy to distinguish between the effects of the shallow, outcropping conductive formations and those of the shallow reservoir in the carbonate and anhydritic rocks. However, where geological information is available and the geometry of the formations is known, the two different effects can be separated, as shown in Fig. 7.

The apparent lack of a resistor between the shallow and deep reservoirs (there are not two conductors but only an extended conductive zone) is probably related to the constraints of 2D modelling of 3D MT data, especially when the target is resistive. It is impossible, with our data only, to define with any precision the separation between the two reservoirs, even by inverting the TM mode only, starting with a model that fits both TE and TM, to resolve the conductor first. This aspect requires further study, and numerical simulation to define the most appropriate distribution of sites to recover this detailed information, and the parameters that are most affected by the presence of a thin resistive layer.

A different picture emerges on the western side of our profile. The three westernmost wells, which are not productive, were drilled in an area defined by the MT data as only slightly anomalous. However, the decrease in resistivity below the sites of group 1 is required by the inversion of TM only data, whereas the joint inversion of both modes suggests that the conductive structure is located laterally off the profile.

Fig. 7. Combination of the model obtained by TM only mode inversion, as in Fig. 6, and the geological section of Fig. 5. The productive wells fall in the areas defined as conductive by MT data. The western part of the AA profile, where the non-productive wells are located, appears resistive. However, a reduction of resistivity is suggested by our data, and is probably located laterally off the profile (see text for details).
The TM only 2D inversion, therefore, probably images a phantom conductive anomaly, indicating that this promising area is not far from our profile.

5. Conclusions

We have demonstrated that MT data can be obtained in intensely industrialised and noisy areas, with due care taken during the acquisition and interpretation phases. We have shown that the acquisition of remote-referenced data is a requisite for retrieving high-quality MT data, but that other conditions must also be fulfilled, such as using a really quiet remote site, and recording very long data segments. In our case the locally-referenced data were not overwhelmed by noise and the remote-reference data, although not good enough to be used on their own, nevertheless provided useful information to separate the noise effect from locally-referenced data. The assumptions we have made in this study will be verified in the future by acquiring very long MT data sets and carefully selecting a remote site.

The most significant result of this study is the presence of electrical discontinuities in the resistive underground below 2 km. We attribute these low resistivities to a single cause, i.e. the deep geothermal conditions of the subsurface, as revealed by drilling data to depths of 4.5 km b.g.l. in the Amiata area. It is evident that geothermal overburden conditions do not produce strong anomalies in density and wave velocities, so we are unable to identify a geothermal reservoir from seismic and gravimetric results alone, although we are able to recognise the geological units and their distribution. The only other physical parameter that is strongly affected by geothermal conditions is heat flow, which shows very high values in all the areas with an anomalously low resistivity, thus corroborating our interpretation. A joint interpretation of seismic and MT data, therefore, provides the most comprehensive picture of the geothermal setting.

We found no indication of a deep conductor in the study area, such as that observed in the Larderello area and in other parts of Mt. Amiata (Fiordelisi et al., 1995; Manzella et al., 1999). We could thus conclude that the heat source of the geothermal system, represented by a cooling, but still hot, magmatic body, does not lie below this profile and is possibly not as hot and extensive as in the Larderello area. We plan to combine the data of this A–A’ profile with data of the B–B’ profile, in order to account for the effects of the 3D setting on MT data, both for shallow and deep structures.

Acknowledgements

The authors are grateful to Phoenix Geophysics Ltd for providing the 10 MTU-24 bit MT systems used in the survey and to Gerry Graham for his invaluable support during the fieldwork. Sandra Trifirò, Annalisa Zaja, Nicola Praticelli and
Giampaolo Girardi are also thanked for their assistance in the survey, and, in particular, Claudio Corsi, who sadly died before the research was completed. Dean Livelybrooks and Gary McNeice are also thanked for their helpful comments and advice, which greatly improved the original manuscript.

References


