Istituto Nazionale di Geofisica e Vulcanologia
Centro Nazionale Terremoti

First INGV BBOBS campaign in the Ionian sea: crustal velocity model inferred from seismic data recorded

G. D'Anna, G. Mangano, A. Amato, A. D'Alessandro, N. Piana Agostinetti, G. Selvaggi

Introduction

The Ionian Sea represents one of the keys needed to better understand the evolution of the geodynamics of the Mediterranean (*Catalano et al., 2001, Finetti and Del Ben, 2005*). It is one of the most seismically active areas in the Mediterranean region, having experienced numerous destructive earthquakes many of which were followed by forceful tsunamis (*Vannucci et al., 2004; Tinti et al. 2004*). The main seismic activities take place along the Hellenic, Aegean and Calabrian arcs, in Eastern Sicily and on the Hyblean-Malta escarpment. However, most of the seismicity occurring in the Ionian basin is unknown, due to the lack of seismic stations on the seafloor.

In May 2007, the Gibilmanna OBS Lab of the INGV deployed three Broad Band Ocean Bottom Seismometers (BB-OBS) in the Southern Ionian Sea at a depth of 3500-4000 meters, so as to better characterize the seismicity of the Ionian region and collect a lot of seismological data. This was effected with the aim of constructing a reliable Vp and Vs velocity model for the Ionian crust, and was done within the monitoring activities carried out in cooperation with the Italian *National Civil Protection Department* (DPC) and within the European **project NERIES** (*activity NA6*).

Fig. 1: *The three OBS's before the deployment in the Ionian sea*

Setting

The Ionian region and surroundings are seismically very active, and are currently undergoing rather rapid deformation (*Vannucci et al., 2004*).

From a review of the geophysical data of the Ionian Basin, Catalano e al. (2001) proposed that the Ionian Sea is an abyssal plain of oceanic nature, bounded by two conjugate passive continental margins (Apulia and Malta escarpments). Low heat flow values (<40 mW/m2), high positive Bouger gravity anomalies (>200 mGal) and a thick lithospheric mantle (70-90 km) suggest an old age for this oceanic embayment (*Scrocca et al., 2003*).

Many authors (*De Voogd et al., 1992; Catalano et al., 2001, Finetti and Del Ben, 2005*) have proposed that the Ionian Sea was floored by 8-11 km of oceanic crust, with 6-8 km of sedimentary cover of Mesozoic and Tertiary age and that it opened as a consequence of the detachment and anticlockwise rotation of the Adria plate. Nevertheless, some authors (Farrugia and Panza, 1981; Ismail-Zadeh et al., 1998) have proposed that the Ionian Basin crust is of a continental-intermediate type (thinned and streched).

So far, no passive body-wave tomography has been made in the Ionian Sea because it requires the deployment of BB-OBS, needed for the appropriate resolving power, through the measurement of travel times of almost vertical rays. Furthermore, the surface wave tomography suitable in oceanic and continental-ocean transition zones suffers from low resolving power due to the lack of an adequate marine seismic network coverage. However, since 1969 a large numbers of active seismic experiments have been conducted.

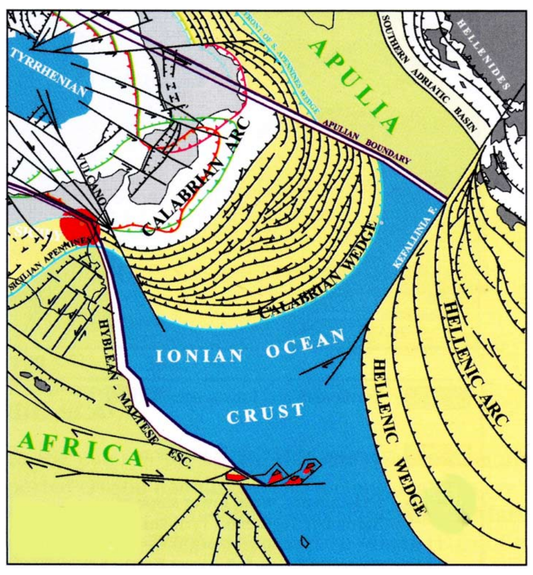

Fig. 2: *Schematic tectonic map of the Ionian area from Finetti and Del Ben, 2005.*

The experiment

To better understand and monitor the earthquake processes taking place in the European-Mediterranean region, in the last decades over 100 seismic monitoring systems and observatories have been created. Nevertheless, **seismic monitoring of the European-Mediterranean region by only land networks is very difficult due to the locations of the earthquakes' epicenters that are often in open sea.** The resulting gaps in coverage produce a biased and incomplete image of Mediterranean seismicity.

For this reason, in 2002 the INGV (Istituto Nazionale di Geofisica e Vulcanologia, Italy) extended its seismic monitoring network to offshore areas under the *GEOSTAR project* (*Favali et al. 2002*). In fact, it created the first Submarine Network-1 seafloor observatory (SN-1) that was deployed about 25 km off-shore Eastern Sicily.

One of the goals of the NERIES project, activity number NA6, is the extension of the capabilities of the existing land-based infrastructure so as to improve seismic risk assessment and management. Under this project, in May 2007, the Gibilmanna OBS Lab deployed three BB-OBS in one of three "key-sites" proposed by ESONET (European Sea Floor Observatory Network): the Southern Ionian Sea (*Fig. 3, D'Anna et al, 2008*).

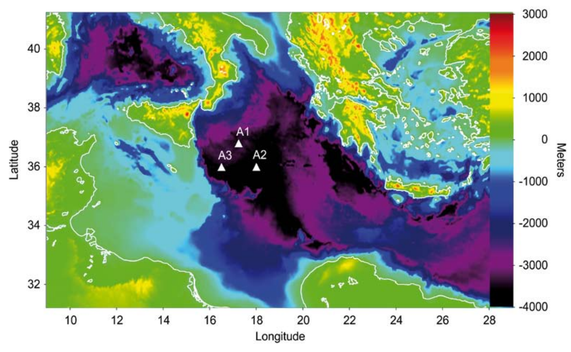

Fig. 3: *The BB-OBS of INGV and its deployment location.*

BB-OBS A1 and A3 were successfully recovered on 2nd February 2008 while A2 was recoverd on 15th March 2008; it was then replaced by another BB-OBS so as to achieve the continuous long-term seismic monitoring task (until May 2010) as planned by the NERIES project.

These instruments, entirely developed by the Gibilmanna OBS Lab as part of a pool of eight BB-OBS ready to be deployed, were equipped with a **Nanometrics Trillium 120P** seismometer, installed in Nautilus gimbals for levelling, and a Cox-Webb 500s-2Hz **Differential Pressure Gauge** (DPG).

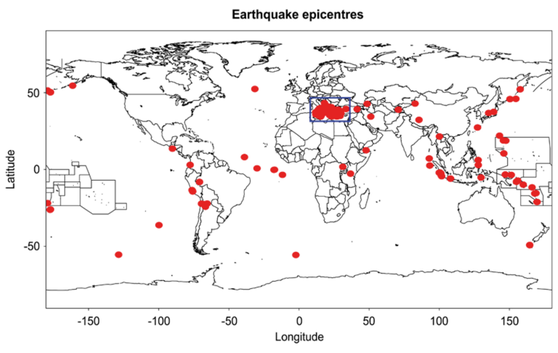

Fig. 4: *Earthquakes with magnitude larger than 5*

During the experiment, the OBS's array recorded more than 450 events; **92 teleseismic events**, **252 regional events** - also recorded by the seismic networks onshore - *and* **more than 115 events that were not recorded by any seismic network on land**.

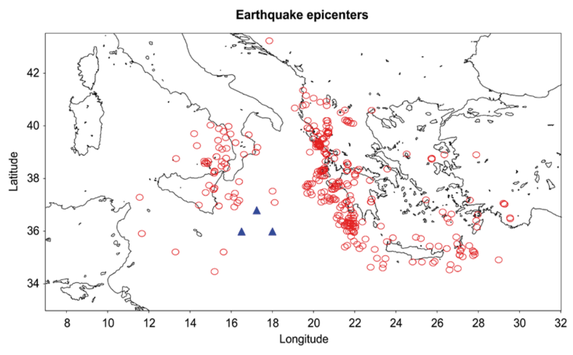

Fig. 5: *Distribution of epicentres of the regional events*

Fig. 4 and 5 show the epicentre distribution of the recorded events. The events were located by INGV, EMSC, USGS and the Greek National Seismological Network and were reported by them in their respective seismic bulletins. Most of the regional events recorded were located along the Calabrian and Aegean arcs and are related to the Ionian subduction.

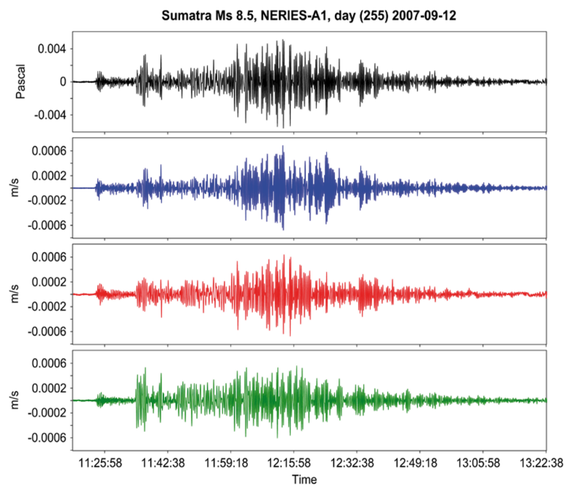

Fig. 6: *Sumatra event of 12th September 2007 (Ms = 8.5).*

Fig. 6 shows the Ms = 8.5 Sumatra event of 12th September 2007, recorded by both DPG and seismometer. Several body and surface phases in both pressure and velocity signals are clearly recognizable on the seismogram.

Optimum P-wave velocity model

So as to obtain an accurate location of the local earthquakes it was necessary to have an optimum P and S waves velocity model for the area analysed. **To define a reference 1D P-velocity model for the Ionian area, we inverted the first P-wave arrival times picked from the regional earthquakes recorded by the OBS's.** We chose only the events with Root Mean Square (RMS) residual lower 0.3 s and standard location errors lower 3.0 km from the entire dataset. On the basis of the a priori information obtained regarding the Ionian Basin and surroundings areas we also discarded the events with shallow hypocentres in intensely deformed areas; for these areas we only selected the events having a hypocentre deeper than 20 km. At the end of this selection **the dataset was composed of 67 earthquakes with a total of 175 P-phase readings; this dataset was used to compute a minimum 1D P-velocity model for the Ionian sea.**

The forward problem of the ray-tracing from source to receiver was solved analytically for critical refracted rays and by the shooting method for direct waves. In the forward problem, station depth was also take into account and consequently the rays were traced exactly to the true station position.

Generally, in the identification of an optimum 1D velocity model both hypocentre locations and parameters of the velocity structure are inverted using the global misfit (sum over all travel-time residuals) as a measure for the goodness of the fit (*Langer et al., 2007*). Nevertheless, since all the earthquakes were located using stations on land, often very remote from the coast line, we only inverted the parameters of the velocity structure.

A critical factor for the inverse problem is the selection of an initial velocity model. We used the velocity model proposed by Finetti and Del Ben (2005) as the start model.

They interpreted the seismo-stratigraphy of the western Ionian crust from wide-scale seismic sequence analyses, specific seismic stratigraphy studies and careful seismic correlations. For these authors, the Ionian Sea is floored by a thin ocean crust that opened during Permo-Triassic times, was covered by a continuous deep-water condensed sedimentation from the Trias to Oligocene, and was overlain by Lower-Middle Miocene, thick Messinian salt and thin Plio-Quaternary. This six layers crustal model was used as a starting model for our data inversion.

In the inversion procedure we only fixed the number of layers and inverted them for both velocity and layer thickness. Because the inverse problem is non-linear the solution is obtained iteratively by random perturbation of the model's parameters until the RMS becomes <0.3s. Fig. 5 shows the P-waves velocity models obtained from P-wave arrival times inversion: the red line is the initial velocity model (Finetti and Del Ben; 2005), the grey line the models obtained from inversion, and the red line the mean model.

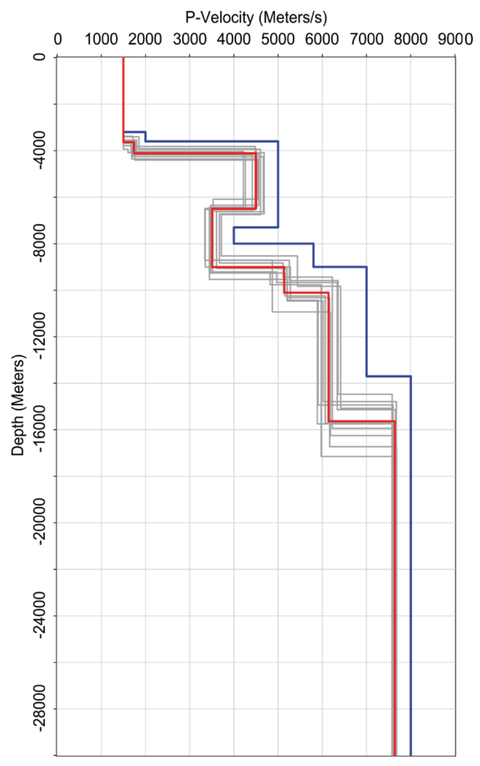

Fig. 7: *P-waves velocity models obtained from P-wave arrival times inversion: the blue line is the initial velocity model (Finetti and Del Ben; 2005), the grey line the models obtained from inversion, and the red line the mean model.*

Teleseismic receiver functions

We analyzed a sample of 4 teleseisms in order to retrieve a preliminary 1D S-velocity profile under the OBS A1.

We first rotated the original waveforms to minimize energy on the transverse components, obtaining the correct seismometer orientation (N152E).

Due to the large water depth, the first multiple phases from the ocean surface are expected after about 4.5 seconds, so we have inverted RF in the first 5 seconds. Also, no water-column multiples were recognized in the vertical component, probably because the geometrical spreading of the wave front and the soft sea-floor attenuated remarkably their amplitudes. For this reason, signals were not corrected for the presence of water-column multiples.

Receiver functions (RF) were computed using a frequency-domain deconvolution technique (*Di Bona, 1998*). RF were stacked to increase the signal-to-noise ratio (*Fig. 8*). The stacked RF has been used as input data for inverting the 1D velocity model (*Fig. 9a*), using the Neighbourhood algorithm (*Sambridge, 1999*). The parameter space boundaries were determined from published results in the Ionian region (*Ferrucci et al., 1991; De Voogt et al., 1992*).

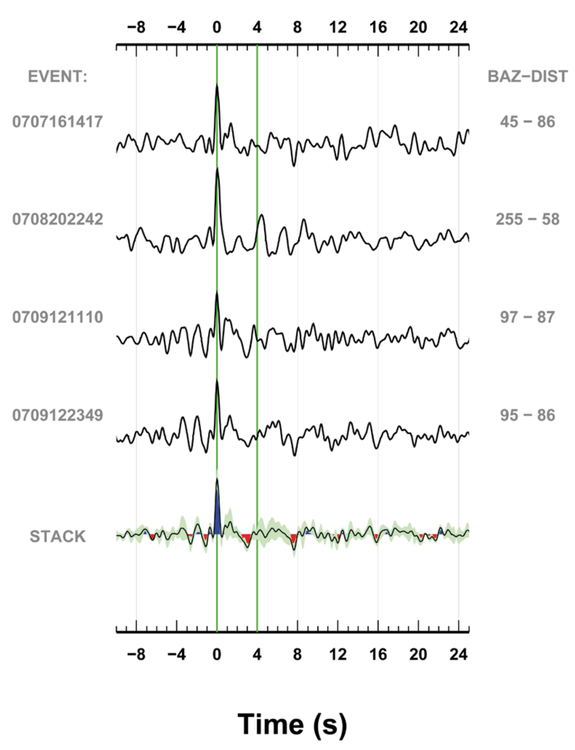

Fig. 8: *Stacked receiver functions.*

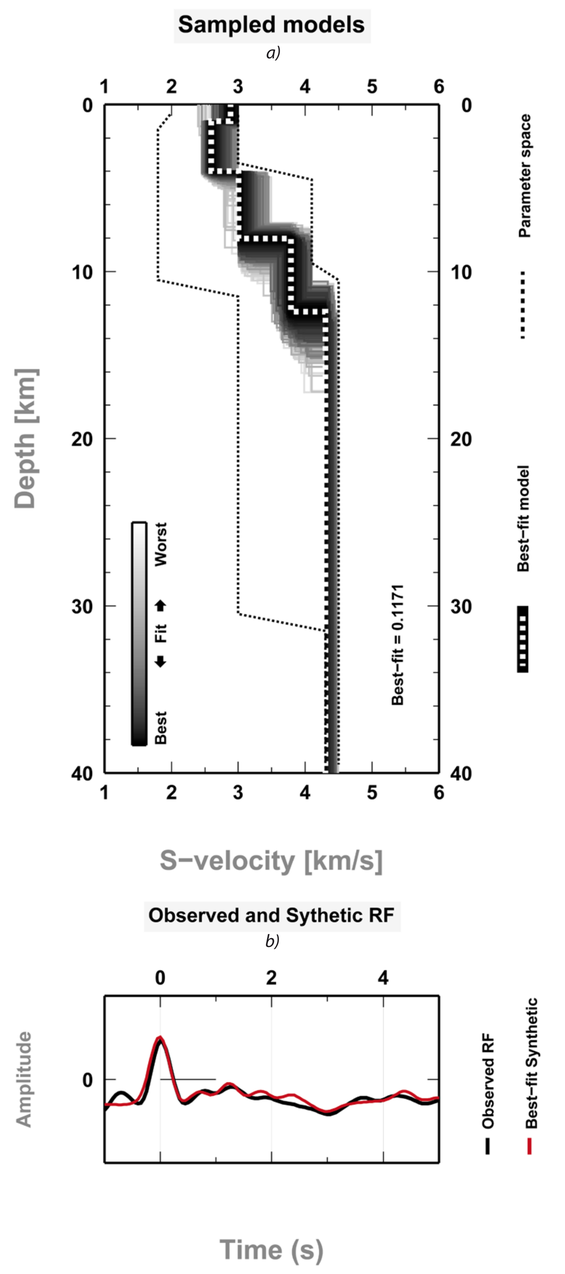

Fig. 9: *a) 1D S-velocity model obtained by inversion of teleseismic receiver functions; b) Comparison between observed stacked (black) and synthetic (red) receiver functions.*

Preliminary results show (*Fig. 9b*) **two well defined Ps converted phases at about 0.8s and 1.2s and a broad negative pulse at about 3.0s.** However, we notice a dependence of the amplitudes of the conversions with backazimuth. More data are needed to analyse this feature in detail.

The RF inversion results (*Fig. 9a*) show an upper, ~1 km thick, high velocity layer, which can be interpreted as the Messinian deposits (*De Voogd et al., 1992*). Below, we find a ~3 km thick lower velocity layer, possibly related to the Meso-Cenozoic terrigenous and carbonate units. Underneath, from ~4 to ~12 km of depth, two layers with Vs between 3 and 4 km/s represent the oceanic Ionian crust, down to about 16 km below the sea level.

Conclusions and future works

The comparison between the preliminary P and S velocity models obtained from the two independent techniques (*Fig. 7 - 9a*) show a good general agreement as for the depth of the main discontinuities. In particular, **both methods agree in constraining the Moho at depth of ~12 km below the ocean bottom, confirming the oceanic nature of the Ionian sea**. Also, a velocity inversion is found for both P and S velocities, although with some difference both in the depth of the low velocity layer (shallower for the RF S velocity profile), and in its thickness and amplitude (more pronounced for the P velocity). This velocity inversion, already found by active seismic experiments (*De Voogd et al., 1992*), is probably associated to the terrigenous Mesozoic-Tertiary sediments underlying the Messinian evaporitic sequence. The oceanic crust appears to be layered, but with the preliminary set of data analysed to date we cannot define precisely its internal structure yet. More RF data will be used in the future to better constrain the S velocity profile. In the next step we will also apply the FTAN method (*Dziewonski et al., 1969*) to the Rayleigh waves recorded by DPG or vertical components to extract its group and phase velocity dispersion curves. The inversion of the dispersion curves will allow us to further constrain the 1D S wave velocity model retrieved by the teleseismic RF with an independent data set. Furthermore, we will relocate the local earthquakes using the 1D P and S velocity models determined for the Ionian area.

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OBS Lab Gibilmanna
obslab@ingv.it