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Accurate hypocentre locations in the Middle-Durance Fault Zone, South-Eastern France

Research Article

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Abstract: A one-dimensional velocity model and station corrections for the Middle-Durance fault zone (south-eastern France) were computed by inverting P-wave arrival times recorded on a local seismic network of 8 stations. A total of 93 local events with a minimum of 6 P-phases, RMS 0.4 s and a maximum gap of 220° were selected. Comparison with previous earthquake locations shows an improvement for the relocated earthquakes. Tests were carried out to verify the robustness of inversion results in order to corroborate the conclusions drawn from our findings. The obtained minimum 1-D velocity model can be used to improve routine earthquake locations and represents a further step toward more detailed seismotectonic studies in this area of south-eastern France.

Keywords: earthquake location • minimum 1-D model • Middle Durance Fault Zone (MDFZ) • France

1. Introduction

Accurate earthquake locations are of primary importance when studying the seismicity of a given area because they provide important information on the ongoing seismotectonic processes. In standard location techniques, the velocity parameters are kept fixed to a-priori values, which are assumed to be correct, and the observed travel-time residuals are minimized by adjusting the hypocentral parameters. However, the use of an unsuitable velocity model can introduce systematic errors in the hypocentre locations [1, 2]. Precise hypocentre locations and error estimates, therefore, require the simultaneous solution of both velocity and hypocentral parameters.

In this paper, we define a reference P-wave velocity model for the Middle-Durance Fault Zone (hereafter MDFZ; south-eastern France), using the approach by Kissling et al. [3]. We apply information from surface geology, drilling, and refraction and reflection seismic surveys [4]. This procedure also allows us to compute station corrections, which can be used in standard location methods to account for the heterogeneous velocity structures around individual stations. Special attention was paid to test the stability of the inversion results.

The concept of minimum 1-D model, which represents a first step towards more detailed seismic studies, is widely used around the world. One of the first applications of this method was in north-western Italy [5], but afterwards it was used in northern Chile [6], Costa Rica [7], New Zealand [8], and central and southern Italy [9–13].

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The calculated minimum 1-D model must satisfy the following conditions, with regard to an a-priori model derived from independent geological and geophysical data observation e.g. [14]:

a) the P-wave velocity of each layer is the area weighted average of the velocity sampled at that depth interval by the data;

b) the topmost layer and the station corrections reflect the basic features of near-surface structure;

c) equally high precision locations should be found for all well-locatable earthquakes occurring anywhere within the seismic network.

2. Geological and seismotectonic features

The Middle-Durance Fault lies in the western Provence area (south-eastern France). At present, the Provence region is mainly characterized by a NNW to N-trending compression, as evidenced by focal mechanism solutions [15, 16] and geodetic measurements [17–19].
The MDFZ is an 80-km long fault system with a moderate but regular seismicity and some palaeoseismic evidence of larger events [20–23]. It behaves like an oblique ramp with a left-lateral reverse fault slip and has a low strain rate. It is made up of two fault zones: (1) the Middle-Durance fault Zone (MDFZ) to the north and (2) the Aix Fault Zone (AFZ) to the south (Figure 1), which is connected with a right offset. Its historical seismicity is documented within the SISFRANCE French historical database [24]. Intermittent seismic activity has been reported since the 16th century, with epicentral intensities ranging from VII to VIII MSK. The strongest events were located near the town of Manosque (13/12/1509, Io = VIII and 14/08/1708, Io = VIII).

3. Local earthquake data and a reference 1-D velocity model

In the MDFZ, a local seismic network of eight stations (Figure 2) has been installed since 1983 [25]. Each station is equipped with a short-period vertical seismometer (Kinemetric SS-1 Ranger) with a natural frequency of 1 Hz and a damping of 0.7%, an amplifier-modulator-digitizer and an aerial with a sender UHF. Signals are recorded on magnetic tapes and transferred and processed in two steps: i) a multiplexed radio record the signals at the central receiving station at Pic de Bertagne (BERF); ii)

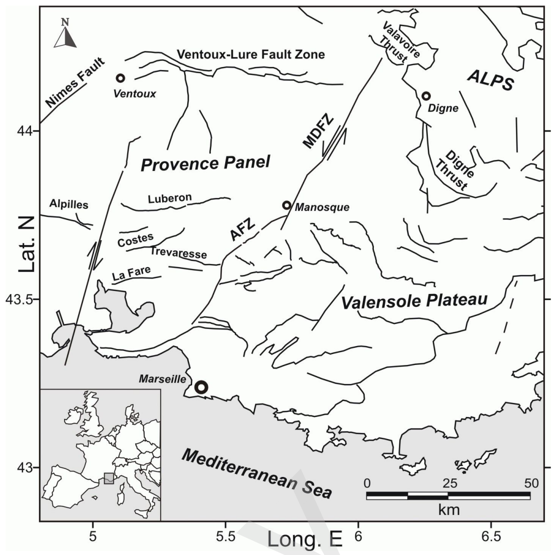

Figure 1. Simplified tectonic map of south-eastern France (modified from Cushing et al., 2007): MDFZ and AFZ represent, the Middle-Durance Fault Zone and the Aix Fault Zone respectively.

data are acquired and transferred to a secondary computer, where the seismic signals are routinely processed in the seismologic centre in Saint Jéróme, Marseille. Data are sampled at 75 Hz; the corner frequency of the anti-alias filter is centred at 30 Hz. The arrival times of the P- and S-waves, when detectable, are picked with an accuracy that is generally within a few tens of milliseconds.

At the Saint Jéróme seismologic centre, earthquakes are routinely located using the software Hypoinverse [26] and the a-priori velocity model proposed by Fourno et al. [25]. In the inversion process we use additional information to select the a-priori model, as suggested by Klimes [27] and Kissling [28] to ensure that we are minimizing arrival-time residuals, rather than minimizing residuals resulting from the kinematic hypocentre determination. In this case, the initial layering of the a-priori velocity model was chosen considering the local geological and geophysical data [4]. We used two layers to approximate the crust, and a half-space for the mantle below the Moho. The thickness of the first layer (7 km) accounts for the Meso-Cenozoic uppermost crust [29]. The discontinuity at 7 km corresponds to the top of the pre-Triassic basement. The Moho is placed at 30 km depth, based on seismic refraction and reflection studies [30].

The resulting velocity model (Figure 3) has reduced average station residuals when calculating earthquake locations compared to other models.

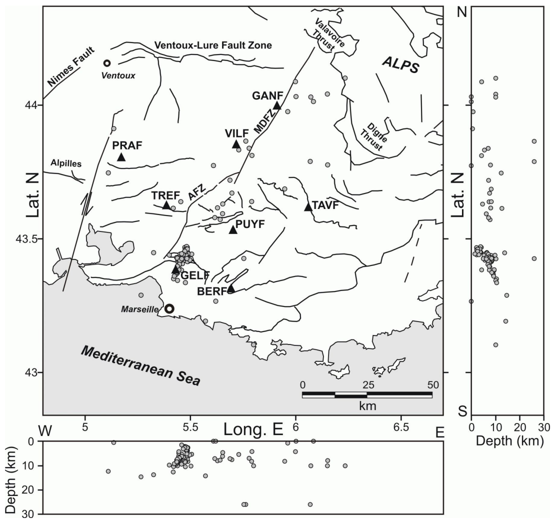

Figure 2. Seismotectonic Map, E -W and N-S vertical sections of the earthquake hypocentres (grey circles) recorded from 1988 to 2007 that were selected for the inversion. Black lines represent main thrust and faults; triangles represent seismic stations.

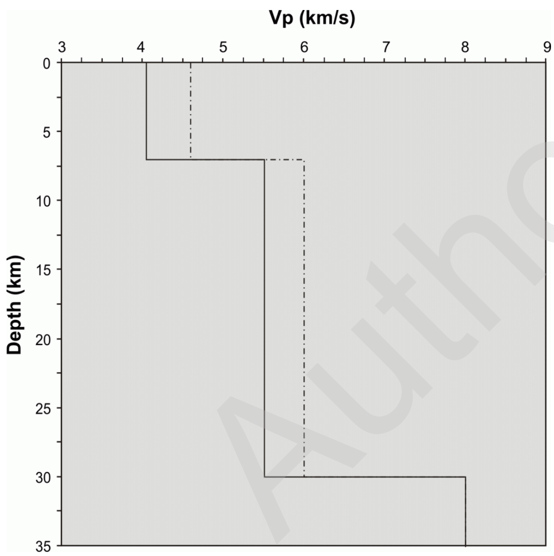

Figure 3. Starting P- wave 1-D velocity model (dotted line) from Fourno et al. (1993) and our computed Minimum 1-D velocity model (solid line).

4. Accurate hypocentre location

The seismic wave travel-time is a non-linear function of the hypocentral parameters and the seismic velocities sampled along the ray path between the hypocentre and the station. The dependence on hypocentral parameters and seismic velocity is called the coupled hypocentre-velocity model problem [1, 31, 32]. It can be linearized and written in matrix notation as [3]:

$$t = Hh + Mm + e = Ad + e, \tag{1}$$

where t is the vector of the travel-time residuals, H is the matrix of the travel-time partial derivatives with respect to hypocentral parameters, h is the vector of the hypocentral parameter adjustments, M is the matrix of the travel-time partial derivatives with respect to the model parameters, m is the vector of the velocity parameter adjustments, e is the vector of the travel-time errors, which includes contributions from errors in measuring the observed travel-times, errors in t_{calc} due to errors in station coordinates, use of the wrong velocity model and hypocentral coordinates, and errors caused by the linear approximation, A is the matrix of all partial derivatives and d is the vector of hypocentral and model parameter adjustments. In a standard location procedure, the velocity parameters are maintained fixed to a-priori values and the observed travel-time residuals are minimized by perturbing the four hypocentral parameters (origin time, epicentre coordinates, and focal depth). Neglecting the coupling between hypocentral and velocity parameters during the location process, however, can introduce systematic errors. Precise hypocentre locations and error estimates, therefore, demand the simultaneous solution of both velocity and hypocentral parameters. Kissling et al. [3] concur with Thurber [1] that the correct hypocentral coordinates are most reliably achieved by solving the coupled hypocentre-velocity model problem, rather than alternating independent hypocentre and velocity adjustment steps. The obtained minimum 1-D model represents a velocity model that reflects the a-priori information and leads to a minimum average of rms values for the best-selected earthquakes used in the inversion. Each velocity value in a given layer of the Minimum 1-D model is the weighted average over all rays within that depth interval. To account for lateral variations in the subsurface, station corrections are included in the 1-D inversion process. The applicability of the Minimum 1-D model, even in areas characterized by dipping structures and significant Moho topography, and its performance for high-precision earthquake locations have been documented by numerous tests [32]. Moreover, using the minimum 1-D model as the initial reference model, no significant and systematic shift in hypocentre locations is observed when inverting parameters for the identification of a 3D velocity structure. Therefore the minimum 1-D model is the most appropriate for uniform high-precision earthquake locations in the MDFZ, outperforming any velocity model based only on a-priori information. An outline of the inversion procedure for the computation of the minimum 1-D velocity model is presented in Figure 4.

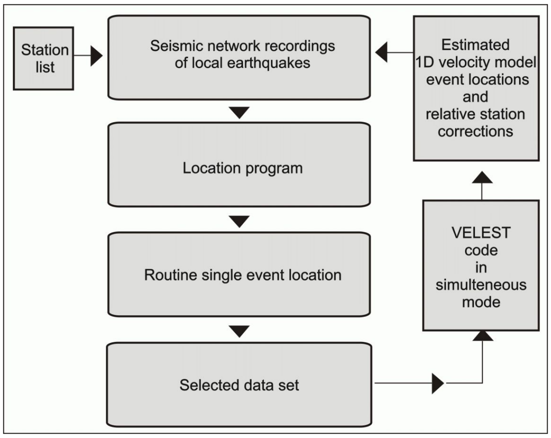

Figure 4. Block diagram of the inversion procedure used to compute the Minimum 1-D model.

5. Computation of a minimum 1-D model for the Middle Durance Fault Zone

As a reference 1-D velocity model must approximate a weighted average of the data but must also reflect the gross features of the structure, the computation of a reference model starts with the definition of three elements: 1) the zone of study, 2) the geometry of the initial 1-D velocity model, 3) the selection of a high-quality set of local earthquake data.

For the identification of an optimum 1-D P-wave velocity model we used the widely known software VELEST [5]. The program, for local earthquakes, comes with two options: in 'simultaneous mode', it solves the coupled hypocentre-velocity model problem; in 'single-event-mode' it computes only the earthquake locations, keeping the velocity parameters fixed. In both approaches the forward problem is solved by ray-tracing from source to receiver, computing the direct, refracted, and the reflected rays passing through the 1-D model. The inverse problem is solved by inversion of the damped least square matrix. Because the problem is non-linear, the solution is obtained iteratively, where one iteration consists of solving both the complete forward problem and the complete inverse problem once.

Within the MDFZ about 290 earthquakes were recorded between 1988 and 2007. Since large uncertainties in the hypocentre location would introduce instabilities in the inversion process, we located the events by using the a-priori velocity model and the program VELEST in single event mode before including the earthquakes in the joint inversion of velocity and hypocentral parameters. The database was then filtered matching minimum requests with respect to location quality criteria. Earthquakes were selected using the criteria of at least 6 detectable P-phase arrivals, $rms < 0.4$ s and a maximum GAP of 220°. The maximum GAP is an important parameter that ensures that events can be well located within the local network. We chose to consider also a few epicenters with a gap larger than 180° which are outside the network, because their raypaths help to constrain the study volume.

The resulting dataset includes 93 earthquakes, with a total of 599 P-wave observations. Figure 2 shows the location of the selected events. The depth distribution shows that most of the selected hypocentres are shallower than 15 km.

These events were then inverted using the program VELEST in simultaneous mode to calculate hypocentre locations as well as the parameters of the velocity structure and station corrections. The model damping parameters were chosen following the default values proposed by Kissling (see VELEST user's guide - Kissling, [5]). As the layer depths are kept fixed according to the recommendations of Kissling et al. [3], we began with a large number of thin layers (3 km thick) and then combined layers for which velocities converged to similar values during the inversion process. The inversion process stopped when earthquake locations, station delays and layer velocities did not vary significantly in subsequent iterations.

6. Results

After 6 iterations we obtained a velocity model that is compared in Figure 3 with the initial model. This final model satisfies the following requirements: 1) earthquake locations, station delays and velocity values do not vary significantly in subsequent iterations; 2) the total *rms* value of all events is significantly reduced with respect to the first routine earthquake locations. We obtained a variance improvement of about 61 % and a final *rms* of 0.27 s. The average deviations, after the first iteration, in origin time, x, y and z were 0.5 s, 0.42 km, 1.2 km and 1.2 km, respectively. A map of relocated events is shown in Figure 5. Standard deviations of the velocity values of the proposed model are 0.30 km/s or less. The P-wave velocity in the upper crust decreased with respect to the starting model, to 4.05 km/s, while the P-wave velocity in the second layer of the crust is 5.52 km/s (Figure 3). Below 30 km the starting velocity of 8.0 km/s remains unchanged, due to the few illuminating ray-paths available.

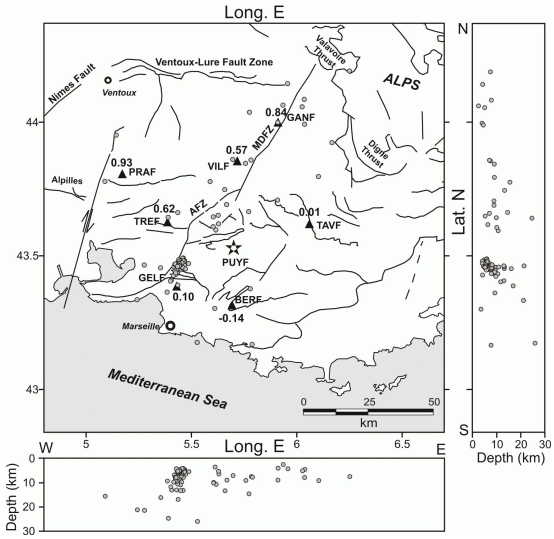

Figure 5. Seismotectonic Map, E -W and N-S vertical sections of the selected events relocated after 1-D inversion using the VELEST code. The station correction values are also shown. The star indicates the reference station. Negative corrections correspond to true velocities that are faster than the model.

6.1. Station corrections

Station corrections are an integral part of the minimum 1-D velocity model since they should partly account for the three-dimensionality of the velocity field that cannot be adequately represented by a 1-D model [28]. Thus, part of the travel-time residuals not explained by the 1-D structure are included in the station correction. Station corrections are strongly coupled with the velocity and structure directly below the station. A change in the velocity structure of the uppermost layers beneath the station is reflected in a more or less constant time-shift of the calculated travel times, which can be compensated for by adjusting the station correction. Typically, they are correlated to a 'reference station', which is preferably chosen close to the geometric centre of the network, and is among the stations with the highest number of readings. The reference station is assigned a correction value of 0. Negative corrections are encountered when the true velocities are higher than the calculated ones, positive corrections occur for lower velocities than predicted by the model. We may exclude biases on the station corrections due to topographic effects because VELEST uses station elevations for the joint inversion of hypocentral and velocity parameters. Consequently, rays are traced exactly to the true station position [6].

In Figure 5, the station corrections are given as relative values with respect to the reference station PUYF. They support the validity of the obtained Minimum 1-D model, as it can be related fairly well to the general near-surface conditions inferred from geological evidence. They show zero or negative corrections at stations BERF and TAVF, deployed on compact calcareous or dolomite rocks. Positive corrections are found at the other stations, where outcrops reveal rocks with supposedly low P-wave velocities, such as VILF (sandy clay), GANF (sandy limestone), GELF (sandy limestone), TREF (soft lacustrine limestone) and PRAF (lacustrine limestone and marls) [33–36].

6.2. Earthquake relocation and stability tests

In order to estimate the improvement introduced by using the computed minimum 1-D velocity model and the station corrections, the 93 selected events were relocated using VELEST in 'single event mode' [5] and the errors have been compared those associated with the initial locations. In Figure 6, which shows the difference in *rms* between the two locations, we can note a consistent decrease of *rms* values for the relocated earthquakes.

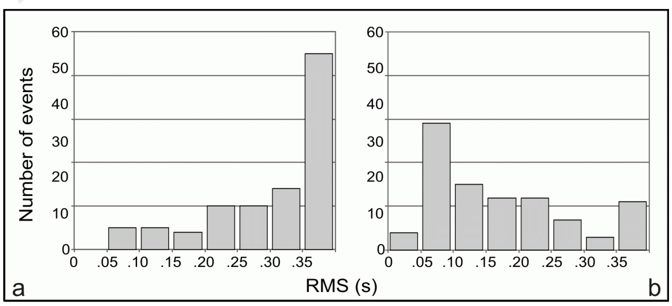

Figure 6. Difference in rms of residuals between (a) the first location with the a-priori model proposed by Fourno et al. (1993) and (b) relocation with the new computed 1-D model and station corrections for the 93 selected events. Note the consistent decrease in the rms value for the relocated earthquakes.

To further explore the robustness of the results we performed two tests which give a direct indication of the inversion stability and/or the model sensitivity with different initial models.

First, we tested the location stability – using the VELEST code but keeping the velocity parameters fixed – by shifting the trial hypocenters randomly in space before including them in the inversion process. This provides a way to check the bias in the hypocentral locations and the solution stability of the coupled problem [6]. If the proposed minimum 1-D velocity model and the available travel-times for each event denote a robust minimum in the solution space, there should be no significant changes in the final hypocentral locations. We have compared the locations with non-perturbed starting solutions, and with starting solutions to which a random perturbation of up to ±5 km was added [6]. The test was repeated five times and the results with the maximum difference between the solutions were considered. This resulted in a conservative estimate of the stability of the hypocentre locations. All tests revealed fairly stable hypocentre determinations for the majority of the events (Figure 7). The difference between the results with non-perturbed starting locations and randomly perturbed ones was fairly low (only 1.5 km or less for 90% of the events).

A second stability test was carried out, as suggested by Haslinger [37], keeping the final hypocentre coordinates of the 93 inverted events fixed, and repeating the inversion process with the same parameters but using different initial velocity models (i.e. with higher and lower velocities with respect to our minimum 1-D model; Figure 8). The convergence of the final inverted models to the minimum 1-D model indicates that this is an adequate 1-D approximation of the upper 30 km of depth.

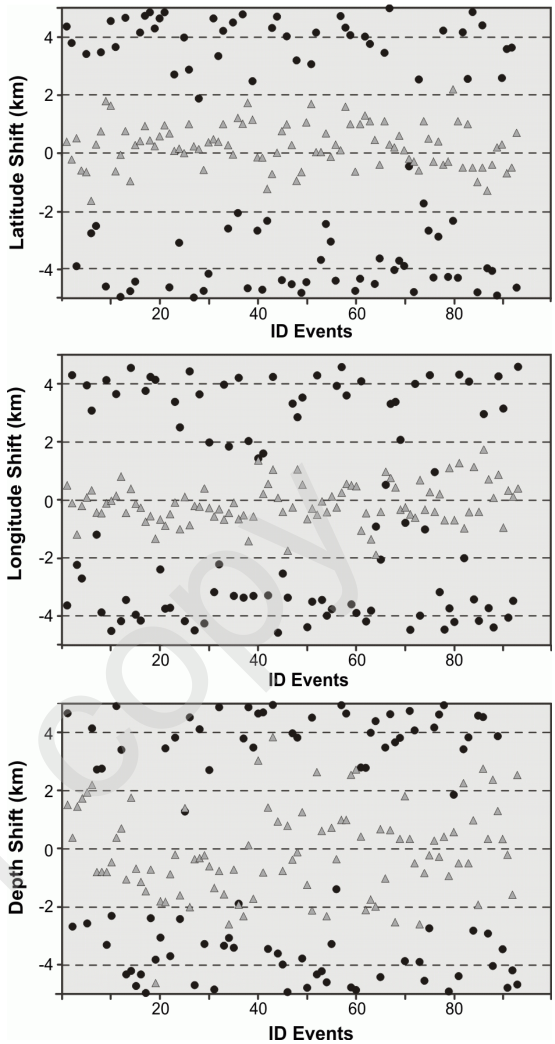

Figure 7. Hypocentre stability test. Black circles represent maximum differences between coordinates of the perturbed trial locations and the original non-perturbed locations encountered during the five random experiments. Grey triangles: maximum differences of final locations (see text for more details).

7. Concluding remarks

We have derived a reference 1-D model and station corrections for the Middle Durance Fault Zone, in southeastern France, by minimizing P-wave residuals for high-quality hypocentre locations according to the procedure of Kissling et al. [3]. We first established the starting a-priori model considering the available geological and geophysical data of the region [4, 29, 30]. This a-priori model is routinely used by scientists to locate earthquakes within the local seismic network in Provence. 93 events were inverted using VELEST [5] in order to calculate adjustments to the P-wave velocity model and to the station corrections. The whole set of local earthquakes was then relocated with VELEST in 'single event mode', using the model obtained from the inversion procedure. As indicated by the resulting lower mean *rms* values and data variance, our minimum 1-D model shows a better fit to the data, which in turn results in more precise and consistent hypocentre locations.

In general, the location of the epicentres suggests a relationship between the seismicity and the main tectonic NE-SW trending faults. Most of the earthquakes are confined to the upper 10 km of the crust; however, as noted by Cushing et al. [14], a few seismic events are located deeper than the sedimentary cover. These deep earthquakes could be linked to potentially seismogenic blind thrusts within the pre-Triassic basement.

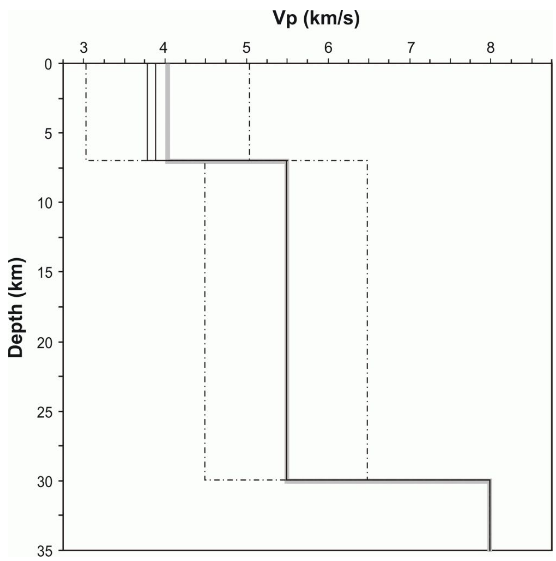

Figure 8. Test of the stability of the minimum 1-D velocity model. The solid grey line shows the minimum 1-D velocity model. The dashed black lines show the input models for the tests with higher and with lower initial velocities. The solid black lines correspond to the resulting models after the inversion (see text for further explanations).

Two tests were performed to determine the robustness of the hypocentre locations and the minimum 1-D model. The inversion process was repeated keeping either the obtained velocity model or the final earthquake locations fixed. Firstly some initial hypocentre locations were perturbed, and these and non-perturbed hypocenters were included in the problem. All events were relocated back to approximately their original positions, indicating robust hypocentre locations. In the second test a range of starting velocity models converge to the same minimum 1-D model, showing that it is an adequate approximation of the crust above 30 km of depth.

In this study the computed minimum 1-D velocity model and station corrections represent a major improvement over other 1-D velocity models for routine earthquake location in the MDFZ. In particular, lateral velocity heterogeneities can be partly accounted for by using the computed station corrections. A better knowledge of the local seismic velocity structure reduces earthquake location errors, to allow us to find a relationship between the seismicity and local tectonic features. These results represent an improvement to the characterization and estimates of the seismogenic potential of this area.

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