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Path-specific, dispersion-based velocity models and moment tensors of moderate events recorded at few distant stations: Examples from Brazil and Greece

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ABSTRACT

Centroid moment tensor (CMT) determination in intraplate regions like Brazil can be very difficult, because earthquakes are often recorded just at few and distant stations. This paper introduces a methodology for datasets like that. The methodology is based on waveform inversion in which each source-station path has its own velocity model. The 1-D path-specific velocity models are derived from the Rayleigh- and Love-wave dispersion curves. The waveform inversion is accompanied by posterior check of numerous P-wave first-motion polarities. An important innovation is the use of so-called frequency range test. The test basically consists in calculating CMT's for many different frequency ranges to assess the stability and uncertainty of the solution. The method is validated on two Brazilian earthquakes and a well-known Greek event. An offshore event (mb 5.2) in SE Brazil is inverted with four stations, at epicentral distances 300–400 km. The other Brazilian earthquake (mb 4.8 in Central Brazil) is even more challenging – only two broadband stations at 800–1300 km are at disposal for waveform inversion. The paper unambiguously demonstrates that the path-specific velocity models significantly increase the reliability of the CMT's. While standard models (e.g. IASP91) typically allow waveform modeling up to epicentral distances of the order of a few (~10) minimum shear wavelengths (MSW), using the path-specific velocity models we successfully inverted waveforms up to > 20 MSW. Single-station waveform inversions are thoroughly tested, but multi-station joint inversions are shown to be preferable. The new methodology of this paper, providing a reasonable estimate of focal mechanisms and their uncertainties in case of highly limited waveform data, may find broad applicability in Brazil and elsewhere.
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1. Introduction

The moment tensor solution is an important tool to understand earthquakes. It provides a simple point-source rupture model, characterized by centroid position and time, nodal planes (strike, dip, and rake) and moment magnitude (Mw). These data are essential for other research fields such as seismic hazard assessment (Convertito and Herrero, 2004; Morrato et al., 2007) and seismotectonic studies (Presti et al., 2013; Herman et al., 2014).

In areas with few earthquakes and sparse station distribution, many events are recorded only at large epicentral distances, which complicates determination of the moment tensor. In particular, in the intraplate areas, where seismic attenuation is low (Hwang et al., 2011; Barros et al., 2011), earthquakes of moderate magnitudes can be well recorded up to distances ~ 1000 km. However, the determination of moment tensor at large regional distances can suffer from inherent inaccuracy of the velocity model (e.g., Nayak and Dreger, 2014). A particular challenge is the moment tensor determination from a single station (Fan and Wallace, 1991; Dreger and Helmberger, 1993; and Kim and Kraeva, 1999).

In this work, we discuss the importance of 1-D velocity models specifically derived for each source-station path. We show, similarly to Assumpção et al. (2011) and Herrmann et al. (2011), that using surface-wave group-velocity dispersion inverted into 1-D velocity models *for each source-station path* can significantly improve the reliability of the moment tensor (MT) determination. We also present a new tool to check the stability of the MT solution: the frequency range test. It consists of performing the waveform

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inversion for many different frequency ranges and systematically investigating variation of the MT solution and its quality. The solution quality is characterized by waveform fit (quantified by variance reduction). The P-wave first-motion polarities are used as additional information to constrain the solution. As such, not only the best-fitting MT solution is calculated, but also a group of solutions well-fitting the data is identified, providing information about the focal-mechanism uncertainty.

We illustrate the methods on the focal mechanism determination of two recent moderate-size earthquakes (mb 5.2 and 4.8) in Brazil that were recorded only at few (2–4) regional stations. Furthermore, we test our methodology also on one earthquake with well-determined focal mechanism in the western Corinth Gulf in Greece. As a special case, single-station inversions are also discussed.

2. Methods

2.1. Surface wave analyses

We build a 1D velocity model using the Love- and Rayleigh-wave group velocity dispersion along each source-station path to calculate the Green's function for the waveform inversion following the steps below:

1. Rotation of horizontal components into transversal component.
2. Measure of the Rayleigh and Love surface waves group velocity dispersion in the vertical and in the transversal components, respectively.
3. Creation of more than 3000 initial velocity models with different criteria to execute the inversion.
4. Inversion of the surface wave velocity dispersion into finals shear velocity model.
5. Selection of the best models according to their misfit and construction of a weighted mean to be used for the waveform inversion.

We analyze the group velocities with multiple filtering techniques using the codes of Herrmann (2013). The Rayleigh and Love wave dispersion is measured on the vertical and transverse component, respectively. We use records providing clear and unambiguous dispersion curves in a period range at least 10 s long, in both components. Moreover, we also require that at least one component must provide the dispersion curve for periods higher than 30 s.

The inversion of the dispersion curve into shear velocity models is made using methodology of Julià et al. (2000). The code uses damped least-square method; it allows for weighting the initial velocity of each layer and controlling the smoothness of the velocity variation between layers. Initial models of four different patterns are created in the present paper:

a) Models with constant-velocity layers of equal thickness;
b) Models with constant-velocity layers whose thickness increases with depth;
c) Models as (a), but with Moho depth prescribed with a large weight (almost fixed);
d) Models as (b), but with Moho depth prescribed with a large weight (almost fixed).

In the creation of the initial models following the patterns c) and d) we utilized information about Moho depth given by Assumpção et al. (2013). The authors compiled data on crustal thickness studies in South America from receiver function, surface wave analysis and deep seismic refraction. In our initial models, the layer corresponding to Moho has a weight of 10 while other layers have weight 1. Additionally, several different smoothness constraints are applied. All together, we create more than 3000 initial models for each path.

The misfit between the observed (d) and synthetic (s) dispersion curves is defined as follows:

$$M = \frac{W_R \frac{\sum (d_R - s_R)^2}{d_R^2} + W_L \frac{\sum (d_L - s_L)^2}{d_L^2}}{W_R + W_L} \quad (1)$$

where the subscript R and L denotes the Rayleigh and Love waves and W_R and W_L are their weights. The inversion of dispersion curves is performed for each initial model and the fit is measured according Equation (1) discarding the final inverted models with two or more velocity inversions (by velocity inversion we mean a local decrease of velocity with depth). To define velocity models for the waveform inversion, we calculate the weighted mean of the final models whose misfit is between the obtained minimum and some setup maximum. Here we choose the maximum equal 3*minimum. The weights are given by $1/M$.

The inversion of dispersion curves is made for the S-wave velocity, Vs, while the Vp velocities are derived using an assumed Vp/Vs ratio, constant with depth. An example of the dispersion-curve inversion is shown in Fig. 1. The spectrogram used to derive the dispersion curve, i.e., the period versus group velocity diagram, in shown in Fig. S1 (electronic supplement).

2.2. Moment tensor solution

We use *ISOLA* (Sokos and Zahradník, 2008, 2013) to invert full 3-component waveforms into the centroid moment tensor. The code calculates Green's functions by the discrete wavenumber method (Bouchon, 1981; Coutant, 1989). The deviatoric moment tensor is calculated by least-squares fitting of the observed and synthetic seismograms, while the centroid time and depth are grid-searched. The resolvability of the inversion is quantified by condition number (CN): low values, about 2–5, imply that the moment-tensor is relatively well resolved, while large values indicate an ill-posed problem whose solution may have no physical meaning. In our tests, we perform the waveform inversion for stations weighted equal to their epicentral distance, accompanied by a posterior check of first-motion polarity agreement. The waveform fit is quantified by the weighted variance reduction VR ($<= 1$):

$$VR = 1 - \frac{\sum W^2 (d - s)^2}{\sum W^2 d^2} \quad (2)$$

where d and s are the observed and synthetic seismograms, respectively, and W is the weight (equal to the station epicentral distance).

2.3. Frequency range test

The moment tensor solution needs a suitable frequency range. The low-frequency cutoff limit is given by the signal-to-noise ratio, while the high-frequency limit depends on the quality of the velocity model (e.g. Fojtíkovía and Zahradník, 2014; Zahradník et al., 2015). Standard velocity models, for example those used to locate earthquakes, typically enable waveform modeling only at wavelengths greater than 1/10 of the epicentral distance (Zahradník et al., 2015). Specific path-dependent models may increase the standard high-frequency cutoff, and we investigate such a possibility. That is why in this paper we repeat the waveform inversion in several frequency ranges, trying to define the (possibly multiple)

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