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Explaining anisotropic macroseismic fields in terms of fault zone attenuation—A simple model

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ABSTRACT

In this work, we present a simple model of anisotropic macroseismic field based on the assumption that local and regional geological structures change the shape of the isotropic macroseismic field (as expected in 1D media). Local geological structures, like water saturated stratified media, may increase intensity level by multiple reflections, constructive interference and resonant effects, but inelastic attenuation, significantly stronger in water-saturated soils, as well as destructive interference, may decrease intensities. On the other hand, large geological structures like seismotectonically active fault zones decrease intensities due to energy redistribution and inelastic attenuation. This model has been developed for the Karst region of the Outer Dinarides where site effects may be neglected because of specific building construction. Neglecting of site effects simplifies the model, so we just need a map of seismically active faults acting as modulator of macroseismic field. In order to demonstrate how the model works, we have calculated the standard error for 10 earthquakes and the macroseismic fields for three of them with epicenters in the Outer Dinarides and compared the model to empiric isoseismals.

1. Introduction

Intensity is the oldest but still a very useful parameter in studying the seismic risk, estimating the effects of future earthquakes and calibrating the hazard models against historical data (Musson, 2005). Quickly calculated spatial distribution of intensities (macroseismic field) is crucial information for civil defense in case of a strong earthquake. So, in spite of its limits, macroseismic intensity still has a significant role in modern seismology.

Intensity is not a physical parameter although authors relate it to physical parameters such as particle velocity (e.g., Gupta and Nuttli, 1976), acceleration (e.g., Gutenberg and Richter, 1942), energy density (von Koeveslígethy, 1907; Blake, 1941), power of energy (e.g., Howell and Schultz, 1975), Fourier acceleration spectra (Chernov and Sokolov, 1999; Sokolov, 2002), etc. Velocity, acceleration, and energy are continuous values, but intensity is an integer, so we can conceive macroseismic scale as an analog to digital converter (ADC). In case of strong shaking with damage on buildings, intensity can be assessed in an objective way, but for small nondestructive events (intensity smaller than VI degree of MSK or EMS scale), the declared effects of earthquakes are, more or less, judged by subjective observations. This is why some seismologists prefer instrumental data instead of macroseismic ones. We should keep in mind that the intensity is not affected by acceleration only, but also by some other parameters such as velocity, duration of sufficiently strong seismic signal, spectral characteristics of signal, resonant frequencies of buildings and furniture, human sensibility etc. (Chernov and Sokolov, 1999). Because of these parameters, intensity is poorly correlated to peak ground acceleration (PGA) (Wald et al., 1999; Musson, 2005).

In this paper, we present a method for calculating the macroseismic field based on a map of faults and instrumental data (coordinates of hypocenter and magnitude). The instrumental data are available immediately after the earthquake occurs, so the method may be useful for fast response of civil defense forces, as well as for earthquake risk mitigation.

The method has been tested on earthquakes in the Outer Dinarides, a European mountain chain spreading from the Alps in north-west to the Helenides in south-east along the eastern Adriatic coast. Mountain chains in the Dinarides are separated by quasi-parallel fault zones. The carbonate layer, 7–13 km thick, is one of the most important properties of this region. The area of our study lies in the Outer Dinarides between Slovenia and Montenegro (Fig. 1). Epicenters of the earthquakes of the learning set were situated between the towns of Zadar and Dubrovnik, while the earthquakes from the testing set had occurred between the cities of Split and Ilirska Bistrica in Slovenia. This area has been chosen for several reasons. First, this is seismically the most active part of Croatia and a lot of data are available. Furthermore, the elongation of isoseismals in the direction of mountain chains (regardless of source mechanism) suggests a connection between isoseismal shape and geological structures (Bottari et al., 1984). The third reason is the tradition of building in this karst area, with the houses foundations on rock

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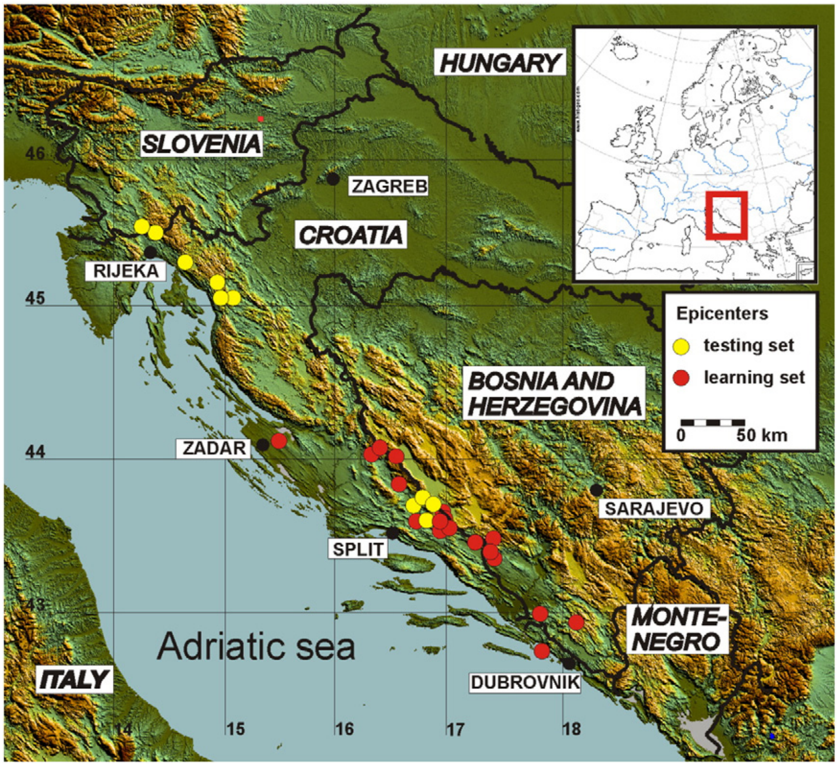

Fig. 1. Epicenters of the earthquakes from learning (red circles) and testing (yellow circles) sets. Almost all earthquakes from the learning set had occurred in Croatia-Herzegovina border region.

because all available soil had to be saved for agriculture. This simplifies our calculations because the soil effect can be neglected.

1.1. Geological setting

The map of the faults used in our model is a part of the Map of Active Faults made in 2006 for monitoring of active tectonic processes (Prelogović et al., 2006) (Fig. 2). Most of the externally positioned faults are thrust ones, while some of the internal faults behave as strike-slip ones. Dips of active faults in the Outer Dinarides vary from 30 degrees in the south-east to 70 degrees in the north-west (Kastelic and Carafa, 2012). The average dip is 50 degrees, but we can neglect its influence because the angle of incidence changes the length of the path through the fault zone by only 16% in the worst case.

The map contains only major faults. We assume that all the faults shown in the map have the same or very similar attenuation properties. The values of the average fault width is valid only for this map. The map does not contain faults in Bosnia and Herzegovina, so the model gives a circular field in that region.

2. Interaction between faults and macroseismic intensity

When the waves pass the fault damage zone amplitude of seismic waves, and consequently macroseismic intensity, can be reduced by attenuation and energy redistribution.

Attenuation reduces wave energy due to several mechanisms like friction, granulation, and viscous dissipation (Alonso-Marroquin et al., 2006). Fault zones consist of the narrow cataclasite containing core and the zone of fractured rock called damage zone (Sibson, 2003), a granular aggregate of incohesive breccia filled with fluid and clay. For the first 4–5 km in depth, breccia fragments are in non-welded contact (Sibson, 1977). Non-welded contact allows continuous stress across the fault zone, but displacement is discontinuous and most of the slip is localized to a fault core. It is known that displacement is discontinuous in the case of fault creeping and coseismic motion (Li and Pyrak-Nolte, 2010). We assume that displacement is discontinuous even in the case of movement forced by incoming seismic waves from a distant source. Since each displacement discontinuity necessarily dissipates energy due to friction, fault zones attenuate seismic waves. Wave energy dissipated in this way heats the volume of the fault zones and the lost energy is significant because fault damage zones could be up to 1.3 km wide (Faulkner et al., 2003). Seismotectonically active zones which consist of several narrow fault zones can be much wider. In the Outer Dinarides, they can be up to 8 km wide (Kuk et al., 2000).

Our assumption is that the main mechanism of energy redistribution at faults is reflection. The faults are fractal structures (Turcotte, 1989; Sammis and Biegel, 1989; Billi et al., 2003; Kim et al., 2004). It means that a fault plane is not a single perfectly flat reflector but it consists of many smaller reflectors like a broken mirror. Energy redistribution by reflections conserves wave energy, but from a macroseismic point of view, reflected energy is "lost." Amplitudes of the reflected waves are small, approximately 10% of the incoming amplitudes. Since the faults are fractal structures, they scatter reflected waves. The interference of the reflected waves or their focusing is not very likely, so they usually do not increase intensity at the incoming side of the fault. Reflected waves have small amplitudes (under threshold level), so they are invisible for ADC of the macroseismic scale.

It seems that damage zones correspond to the low velocity zones (LVZs) since a small amount of clay in pores between rock fragments significantly decreases the shear modulus and, consequently, the wave velocity. The LVZs are well documented (Li and Vidale, 1996; Ben-Zion et al., 2003; Lewis et al., 2005) so we have used the parameters which are usually attributed to LVZ (Q factor, shear wave velocity, width and depth) as the model parameters or as the initial values for calculations.

2.1. Other parameters

Since the main assumption is that attenuation of seismotectonically active fault zones shapes the macroseismic field, we briefly discuss the other factors which may affect intensities.