Perceptible earthquakes in the broad Aegean area

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Abstract

A probabilistic estimate of seismic hazard can be obtained from the spatial distribution of earthquake sources, their frequency–magnitude distribution and the rate of attenuation of strong ground motion with distance. We calculate the earthquake perceptibility, i.e. the annual probability that a particular level of ground shaking will be generated by earthquakes of particular magnitude, by weighting frequency–magnitude data with the predicted felt area for a given level of ground shaking at a particular magnitude. This provides an earthquake selection criterion that can be used in the anti-seismic design of non-critical structures. We calculate the perceptibility, at a particular value of isoseismal intensity, peak ground acceleration and velocity, as a function of source magnitude and frequency for the broad Aegean area using local attenuation laws. We use frequency–magnitude distributions that were previously obtained by combining short-term catalogue data with tectonic moment rate data for 14 tectonic zones in Greece with sufficient earthquake data, and where contemporary strain rates are available from satellite data. Many of the zones show a ‘characteristic earthquake’ distribution with the most perceptible earthquake equal to the maximum magnitude earthquake, but a relatively flat perceptibility between magnitudes 6 and 7. The maximum perceptible magnitude is in the fastest-deforming region in the middle of the Aegean sea, and tends to be systematically low on the west in comparison to the east of the Aegean sea. The tectonic data strongly constrain the long-term recurrence rates and lead to low error estimates (±0.2) in the most perceptible magnitudes.

Keywords: Perceptibility; Ground motion; Seismic hazard; Aegean area

1. Introduction

Different physical, numerical, and statistical models have been applied over the last decade in order to define the seismotectonic environment and the future behaviour of a region (e.g. Ito and Matsuzaki, 1990; Cowie et al., 1993; Lomnitz-Adler, 1993; Rundle and Klein, 1993; Kagan and Jackson, 1994; Main, 1995). These models have various proportions of deterministic and probabilistic elements, and are capable, in principle of validation in particular regions once sufficient data become available. All of these models predict that the frequency–magnitude distribution at low magnitudes takes the form of the Gutenberg–Richter law

\[ \log N = a - bm, \] (1)
where \( a \) and \( b \) are model parameters and \( N \) is the cumulative number of earthquakes with magnitude greater than or equal to \( m \). This relation defines the annual probability of occurrence of a given earthquake magnitude in a given region, and is a strong primary constraint on seismic hazard. However, the distribution at high magnitudes is less certain due to the small number of data points for the rare, large events.

A variety of different approaches have been adopted in an attempt to overcome this problem of uncertainty in recurrence rates for large events in the calculation of seismic hazard (Burton et al., 1984; Makropoulos and Burton, 1985; Goes and Ward, 1994; Lutz and Kiremidjian, 1995; Shimazaki et al., 1999; Papazachos, 1999; Papaioannou and Papazachos, 2000). In particular, geological and geodetic data on deformation rates can be used to place strong constraints on the long-term recurrence. For example, Goes and Ward (1994) applied a seismicity model based on the concept of fault segmentation and the physics of static dislocation, that allows for stress transfer between fault segments. The application of their model allows to obtain recurrence statistics and long-term probability estimates for shocks with \( M \geq 6.0 \) on the San Andreas Fault. In this approach, constraints are provided by geological and seismological observations of segment lengths, characteristic magnitudes and long term recurrence rates. Interaction between faults is also a theme in the stochastic earthquake occurrence model (Lutz and Kiremidjian, 1995). This process can be used to estimate the hazard due to large, spatially and temporally, dependent earthquakes. A similar stochastic method known as “renewal process” is described by Shimazaki et al. (1999) and applied to the recurrence of large earthquakes in Japan.

Estimates of seismic hazard depend not only on the recurrence rates of events of different magnitudes, but also on the attenuation of seismic energy with distance (due to a combination of elastic geometric spreading and anelastic absorption). Burton et al. (1984) combined earthquake magnitudes obtained through Gumbel’s third asymptotic distribution of extreme values with regional intensity attenuation laws to determine the relative ‘perceptibility’ of earthquakes of different magnitude at a given level of ground shaking. The earthquake perceptibility \( P(x|m) \) is defined as the conditional probability that a site perceives ground shaking at least of level \( x \) (where \( x \) may be intensity, peak ground acceleration, velocity, etc.) given the annual probability of occurrence of an earthquake of magnitude \( m \) (Burton, 1978). Hence the most perceptible earthquake is the event that is most likely to occur and be perceived or felt at a given level of ground motion at a site or in an area.

The concept of the “most perceptible earthquake” essentially involves a trade-off between small earthquakes, which are common but not widely felt, and large earthquakes, which are uncommon but felt widely. For any site, there will be a magnitude that represents a maximum, determined by the variables governing frequency and impact.

The concept is analogous to that of “design earthquake” in hazard analysis. In a very recent study Tsapanos (2003) applied the “design earthquake” concept and estimated the “design PGA” values for a seismic hazard scenario in the main cities of the island of Crete. A given level of ground motion hazard is associated with a specific probability (the typical output of a hazard study for design purposes), so it is reasonable to ask which earthquake parameters are most likely to produce this ground motion. Is the hazard mostly from small local earthquakes or large distant ones? The answer will be different for different sites. This problem of disaggregation of hazard can be addressed analytically (McGuire, 1995) or through simulation (Musson, 1999).

Makropoulos and Burton (1985) used the notion of perceptibility to assess seismic hazard in Greece using attenuation laws for ground acceleration. Most recently Papaioannou and Papazachos (2000) estimated time-dependent hazard based on the whole earthquake catalogue (not just the extreme values). They calculated the conditional probability of occurrence for macroseismic intensity for 144 Greek cities, based on a method that assumes fluctuations in \( b \) to be much lower than those in \( a \) in Eq. (1) above (Papazachos, 1999). This is based on the hypothesis that \( b \) values depend on material properties and the seismotectonic setting and therefore should vary relatively smoothly in space.

There have been many studies of seismic hazard in Greece, due both to the availability of different types
of data and practical concerns for the population. On a global scale, Greece is ranked sixth in terms of seismic hazard (Tsapanos and Burton, 1991). The country has often experienced strong and catastrophic earthquakes, with significant casualties and damage to buildings infrastructure, e.g. the recent Athens earthquake ($m = 5.9$, 7 September 1999). The geographical distribution of seismic hazard, based on such seismic sources, has been determined by Papazachos et al. (1991), and more recently quantified as a function of various design parameters. These include: the maximum expected macroseismic intensity (Shebalin et al., 1976; Papaioannou, 1984), the peak ground acceleration or velocity (Algermissen et al., 1976; Makropoulos and Burton, 1985), and the duration of strong ground motion (Margaris et al., 1990; Papazachos et al., 1992).

In this paper we calculate the perceptibility of ground motion (intensity, peak ground acceleration and velocity) for the Aegean area. We use frequency–magnitude relations from a recently-published data set of historical and instrumentally recorded earthquakes, constrained by recent satellite data on deformation rates in Greece and the surrounding areas (Holt et al., 2000; Koravos et al., 2003). We allow the distribution to deviate from Eq. (1) at high magnitudes, and calculate the annual probability of ground shaking at a particular level as a function of earthquake magnitude, based on appropriate ground attenuation laws for intensity acceleration and velocity. The results follow from a previous paper (Koravos et al., 2003) where we used a new catalogue for earthquake recurrence in Greece and the surrounding area to determine the best-fitting frequency–magnitude relations some of the 16 tectonic zones defined by a geodetic study of crustal deformation rates (Holt et al., 2000). These authors considered two zones to have insufficient data for the comparison of seismic and tectonic information rate (zones 15 and 16), and Koravos et al. (2003) found a further two (zones 1 and 2) insufficient for analysis of maximum magnitude. The deformation rates were used to constrain extrapolations from the frequency–magnitude relation, and to estimate the maximum credible magnitude, a crucial parameter of interest to earthquake design engineers. Here we combine these frequency–magnitude distributions with local attenuation laws to predict the distribution of perceptibility of ground motion as a function of magnitude for the same tectonic zones.

2. Data used and method applied

Papazachos et al. (2000) compiled a comprehensive catalogue of instrumental and historical earthquakes in Greece and the surrounding area for the time span 550 BC–1999 AD. The magnitudes are quantified in terms of the seismic moment magnitude scale which we should denote $m$. The completeness of the catalogue and the corresponding errors in magnitude, depth and epicenter estimates are given by Papazachos et al. (2000). The published catalogue is available at (http://www.jjlahr.com/iaspei/europe/greece/the). In a later study (Koravos et al., 2003), it was shown that some of the large events included in Papazachos et al. (2000) catalogue have systematically higher values. Because they were calibrated using the events at the beginning of the 20th century (1900–1912 or so) which were not corrected for the effect of using undamped narrow-band seismic instruments (Abe and Noguchi, 1983). For example, we corrected the event of 1904 (41.80N 23.00E) that occurred in zone 14 from $m_s = 7.7$ to $m_s = 7.1$, as well as the event of 1905 (040.26N 24.33E) generated in zone 10 from $m_s = 7.5$ to $m_s = 6.8$, based on the instrumental determination of Abe and Noguchi (1983) and Ambraseys (2001). Since these events are amongst the largest recorded events in the catalogue, they also have a disproportionate effect on the calibration of the historical catalogue for the largest large events. This systematic effect is on average of the order of $-0.4$ (Koravos et al., 2003) and this was then applied to the historical data. We also added one additional year of data (for the year 2000) from the preliminary annual bulletins of the seismological station of the Aristotle University of Thessaloniki. The data were divided geographically into the 14 tectonic zones considered to have adequate data by Holt et al. (2000), and used to estimate the range of complete reporting for different magnitude thresholds, the frequency–magnitude relation and maximum magnitudes, constrained by the tectonic deformation rates (Koravos et al., 2003). Fig. 1 depicts the examined zones and shows the spatial distribution of the epicenters in the study area for the time period 550 BC–2000 AD. For consistency the
numbers of the tectonic zones correspond to those in Holt et al. (2000).

Here we calculate the perceptibility based on the felt area at a given level of seismic intensity $I$ (modified Mercalli scale), ground motion acceleration $a$ in cm/s$^2$ and velocity $v$ in cm/s. The attenuation formula for $I$ as a function of hypocentral distance $R$ for the broad Aegean area is:

$$I = 1.063 + 1.5222M_s - 1.1021\ln R - 0.0043R \quad (2)$$

where $M_s$ is the surface magnitude (Musson, 2000). The corresponding attenuation law of Eq. (2) in terms of epicentral distance $\Delta$ in km (for $\Delta > h$) is:

$$I = 1.063 + 1.5222M_s - 1.1021\ln \left( \sqrt{\Delta^2 + h^2} \right)$$

$$- 0.0043 \left( \sqrt{\Delta^2 + h^2} \right) \quad (3)$$

where $h$ is the depth in km.

The attenuation relationship for horizontal peak ground acceleration, $a_g$ (in cm/s$^2$) and velocity, $v_g$ (in cm/s), in the broad Aegean is:

$$\ln a_g = 4.37 + 1.02M_s - 1.65\ln(\Delta + 15) + 0.31S + 0.66P^* \quad (4)$$

$$\ln v_g = -0.18 + 1.29M_s - 1.62\ln(\Delta + 10) + 0.22S + 0.73P^* \quad (5)$$

(Theodulidis and Papazachos, 1992), where $M_s$ is the surface magnitude, $\Delta$ is the epicentral distance in km, the site effect $S$ takes the value of zero at “alluvium” sites and one at “rock” sites. The parameter $P^*$ accounts for the scatter in the data about the best fitting line. $P^* = 0$ for 50 percentile values and = 1 for 84 percentile values. The attenuation law of Eqs. (3), (4) and (5), is shown in Fig. 2A, B and C, respectively for $M_s = 6.0$, $h = 10$ km, $S = 0.5$ and $P^* = 0$. 

Fig. 1. Map of the area of study, divided into 16 numbered regions after Holt et al. (2000), showing the spatial distribution of the epicentres of shallow seismicity ($h \leq 50$ km) in the Aegean and the surroundings for 550 BC–2000 AD. The data are from Papazachos et al. (2000).
The surface-wave magnitude $M_s$ is related to the moment magnitude $m$ by
\[
m = 0.56M_s + 2.66 : \quad 4.2 \leq M_s \leq 6.0
\]
\[
m = M_s : \quad 6.0 \leq M_s \leq 8.0
\]
(Papazachos et al., 1997). The felt area of an event of magnitude $m$ producing an intensity of at least $I$ at distance $r$ from the epicentre is:
\[
A(I) = \pi [r(I)]^2
\]
(7)
The annual perceptibility of ground motion at a level $I$ ($a$ or $v$) is then
\[
P(I/m) = F(m)A(I)/A_{\text{max}}
\]
(8)
where $F$ is the incremental annual frequency of occurrence of a magnitude in the range $m \pm \delta m/2$, and $A_{\text{max}}$ is the total area of study.

3. Results

The perceptibility for the 14 zones is shown in Fig. 3 for $I=\text{VIII}$ and a typical focal depth $h=10$ km. The solid lines represent the functional form for Eq. (8), given the frequency distributions of Koravos et al. (2003) and the attenuation laws described above, plotted against the data. The finite focal depth implies that there is a minimum threshold magnitude at the surface that can produce ground shaking at a given intensity (Main, 1995). In Fig. 3, this threshold magnitude for perceptibility of $I=\text{VIII}$ is above magnitude 6.4. For the first two zones (zones 1 and 2) of the examined area, the absence of data above the threshold magnitude of 6.4 for Greece (Koravos et al., 2003) made it impossible to estimate the maximum credible magnitude. These regions are not considered further. Above the threshold magnitude, most of the remaining areas show a steady increase in perceptibility to a relatively flat local maximum or inflexion point, and then a peak at the maximum magnitude. For these plots, the most perceptible magnitude is equal to the maximum magnitude. This is due to the fact that the best fitting frequency–magnitude distribution for these zones has a characteristic peak at the largest event sizes (Koravos et al., 2003), consistent
Fig. 3. The perceptibility $P = F(m)A(I)/A_{\text{max}}$ for the different tectonic zones identified by Holt et al. (2000), predicted from the frequency–magnitude data and curve fits of Koravos et al. (2003), using $I = \text{VIII}$ and $h = 10$ km.
Fig. 4. Perceptibility of ground motion at 0.2 g peak horizontal ground acceleration, assuming $S = 0.5$ for the site conditions and $P^* = 0$ in Eq. (4).
Fig. 5. Perceptibility of ground motion at 10 cm/s peak horizontal velocity, assuming $S=0.5$ for the site conditions and $P^*=0$ in Eq. (5).
with a ‘supercritical’ distribution (Main, 1995). Zone 9 in contrast shows a distinct maximum at \( m = 6.9 \) and zone 10 at \( m = 7.4 \), although zone 10 has a wide range of magnitudes (\( m = 7.0 – 7.7 \)) where the perceptibility remains constant. Main (1995) showed that the perceptibility remains relatively constant for all magnitudes above the threshold for the ‘critical’ (Gutenberg–Richter) distribution. This is consistent with the results for zone 10.

The perceptibility for ground-motion acceleration at a level of 0.2 g, assuming an average site condition \( S = 0.5 \) and \( P^* = 0 \) is shown in Fig. 4. In this case the threshold magnitude of perceptibility is \( m = 5.6 \). Most of the curves show features that are qualitatively similar to the corresponding curves for intensity VIII in Fig. 3. However, the peak probability occurs systematically at lower magnitudes for acceleration 0.2 g in Fig. 4. For example, in zone 9 the most perceptible magnitude \( m_p \) is in the range 6.1(\( \pm 0.1 \)) for acceleration cf. 6.9(\( \pm 0.1 \)) for intensity. These low error values in \( m_p \) reflect the strong constraint of deformation rates on the results. In zone 10 the broad maximum in perceptibility has a lower and narrow range for acceleration (\( m = 6.1 – 6.4 \)), when compared to the range of intensity (\( m = 7.0 – 7.7 \)).

The perceptibility for ground-motion velocity at a level of 10 cm/s for \( S = 0.5 \) and \( P^* = 0 \) is shown in Fig. 5. The threshold magnitude in this case is \( m = 5.5 \) and the general pattern in this figure is similar to the corresponding curves for acceleration in Fig. 4.

In three zones 4, 6 and 7 the perceptible magnitude shows a difference of the order of \( \pm 0.1 \) in correlation with perceptible magnitudes estimated from intensity and acceleration.

Fig. 6 shows the spatial distribution of the studied zones and the perceptible magnitudes for intensity,

Fig. 6. Spatial distribution of the examined zones and the perceptible magnitudes for intensity, acceleration and velocity. The standard deviation is also given in parenthesis for both magnitudes.
acceleration and velocity with their uncertainties. Inspection of this figure shows that perceptible magnitudes are over 7.0 in almost all the regions with sufficient data, indicating that the seismic hazard is really dominated by the rare, large events in the Aegean area.

4. Discussion

Papazachos and Kiratzi (1996) examined the seismic deformation velocities for the area of Greece and its surroundings. They found high seismic deformation velocities for the middle part of the Aegean sea having in general north–south direction, high to intermediate values for the eastern part and low values for the western part of the Aegean area where the direction changes to east–west. Crustal deformation strain rates were determined by Kahle et al. (1998, 2000) based on the use of GPS measurements in the eastern Mediterranean region. They found high values for the area of the north Anatolian fault (170 nstrain year$^{-1}$) while lower (90–120 nstrain year$^{-1}$) and lowest (less than 40 nstrain year$^{-1}$) values were determined for the southeastern Aegean and central and southwestern Aegean, respectively. Our results are consistent with the more recent data of Holt et al. (2000) deduced by the combination of SLR and earthquake mechanisms. The obtained results ($P(x|m), m_p$) confirm that local deformation rate exercises a first-order control on the seismic hazard. The highest value for $m_p$ is in the fastest-deforming part in the middle of the Aegean sea and tends to be systematically low on the west in contrast to the east of the Aegean sea (Fig. 6). For example in zone 11 where the deformation velocity is 24 mm/year according to Papazachos and Kiratzi (1996) and 28 mm/year according to Holt et al. (2000), $m_p$ magnitudes (deduced from intensity, acceleration and velocity) are equal to 7.6. The other one is zone 12 where the seismic deformation velocity is 16 mm/year (Papazachos and Kiratzi, 1996) and 17 mm/year (Holt et al., 2000), $m_p$ magnitudes are equal to 7.4. The Aegean Sea coincides from north to south with zones 14, 10, 11, 6, 7 and 3 of Holt et al. (2000). We observed that generally the easternmost zones (12, 8 and 4, corresponding to the western part of Turkey), have larger $m_p$ values than the westernmost ones (13, 9, 5 and 1, i.e. mainland Greece). These results confirm a very good correlation that exists between the seismic deformation velocities, deformation strain rates and $m_p$.

5. Conclusion

The perceptibility of ground motion has been estimated for seismic zones corresponding to a grid used to determine tectonic deformation rates, for intensity, acceleration and velocity. Such results can be used to constrain seismic hazard based on combining spatially variable frequency–magnitude and satellite strain rate data with attenuation laws. Most of the deformation zones examined here exhibit a characteristic or supercritical frequency–magnitude distribution, where the moment release and the seismic hazard is dominated by largest magnitudes, and the most perceptible earthquake is equal to the maximum magnitude. The tectonic data strongly constrain the long-term recurrence, and lead to rather low estimates of the error in perceptible magnitudes, on the order of $\pm 0.1$ or $\pm 0.2$ magnitude units. Only two zones, zones 9 and 10 showed critical behaviour with a most perceptible magnitude in the middle of the perceptible range. The location of this peak is lower for acceleration than intensity in both of these zones. Three zones had insufficient data for the analysis presented here, while in the rest of the zones we observed a good correlation between the seismic deformation velocities and the perceptible magnitudes. Zones of high seismic deformation rates are correlated with high values of perceptible magnitudes ($m_p$). The middle of the Aegean has the highest deformation rate and the highest value for the $m_p$. Intermediate to high deformation rates and $m_p$ values are found on the eastern side of the Aegean area, while low deformation rates are associated with relatively $m_p$ parameters on mainland Greece, in the western part of the examined area.

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