INTRODUCTION

The present study is focused on estimating the probabilistic seismic hazard for the capital city of Ecuador, Quito, the population of which currently exceeds 2 million inhabitants at present. Quito is located at 2800 meters above sea level within the Interandean Depression, bounded by the equatorial line to the north, in an earthquake-prone environment (Chatelain et al., 1999; Fig. 1). The city and its suburbs have developed in a piggy-back basin on the hanging wall of a reverse fault system (Fig. 2) that has been recognized as seismically active in historical, geomorphologic, geologic, and geodetic studies (Soulas et al., 1991; Ego and Sebrier, 1996; Hibsch et al., 1997; Egred, 2009; Champenois et al., 2013; Alvarado et al., 2014).

The historical record, spanning five centuries, shows that the city experienced MSK intensities (Medvedev et al., 1963) in the VII–VIII range at least five times, with damages reported in churches and houses (Pino and Yepes, 1990; Egred, 2009). The most destructive earthquake was the 1859 event (I\text{max} VII–VIII). Analyzing its intensity dataset, Beauval et al. (2010) proposed a mean magnitude $M_I$ 7.2 (Fig. 12 of their paper, $M_I$ is an intensity magnitude equivalent to moment magnitude) and a location at an intermediate depth in the slab; a location on a shallow crustal fault cannot, however, be excluded. The 1797 Riobamba earthquake ($M_I$ 7.6), 160 km south of Quito, produced an intensity VII in the city. A similar shaking level was experienced during another large earthquake, the 1868 Ibarra event, probably generated on the Otavalo fault located 50 km north of Quito ($M_I$ 7.2, fig. 13 in Beauval et al., 2010). Two more earthquakes, described by a sparse set of intensities and therefore difficult to characterize, could have been produced by one of the Quito fault segments. These events occurred in 1587 ($I_{\text{max}}$ VIII, $M_I$ 6.3–6.5, fig. 15 in Beauval et al., 2010) and 1755 (observations available only in Quito, corresponding to intensity VII). In the last 150 years, no major earthquake hit Quito. The most recent significant earthquakes on the city fault system were in 1990 ($M_w$ 5.3), and in 2014 (12 August, $M_w$ 5.1); in both cases three inhabitants were killed in the northern suburb of Pomasqui (location indicated in Fig. 2). In an attempt to extend the observation time window, Hibsch et al. (1997) analyzed earthquake-induced deformation phenomena in lacustrine sediments in the northern part of the Quito basin. They studied approximately a 1500-year time span prior to the historical record. One major event was identified (intensity evaluated to X, greater than the maximum intensity in the historical record) between the tenth and the sixteenth centuries, which they believed could have ruptured the entire Quito fault reaching a magnitude 6.5–7.0. This is the unique evidence for such a large earthquake on the fault system. More recently, the seismic potential of the Quito fault was confirmed on the basis of observations covering a much shorter time window. Analyzing Global Positioning System (GPS) measurements at sites with 10–15 years of recordings, east–west horizontal shortening rates were estimated in the 4.3–5.3 mm/yr range across this blind thrust (Alvarado et al., 2014).

Since 2007, a joint collaboration between France and Ecuador has been in place to prepare all the necessary inputs for computing PSH for Ecuador. The research laboratories involved are the Geophysical Institute (IG) in Quito, ISTerre in Grenoble, and Géoazur in Nice. The IG has a governmental mandate to monitor earthquakes and provide national seismic-hazard estimations. PSH maps are the basis for establishing seismic building codes. A PSH assessment requires a seismicity model, a description of the probability of occurrence of future earthquakes, and a ground-motion prediction equation (GMPE), which gives the probability of occurrence of accelerations as a function of magnitude and distance. As the Ecuadorian strong-motion network is very young, GMPEs established elsewhere in the world must be imported.

By analyzing seismicity distribution, active faults and plate margins, as well as geodetic measurements, Alvarado (2012, Fig. 3) subdivided the region into seismic sources. This seismotectonic zoning is the one used in our 2011 PSH calculations that we provided to the Ecuadorian committee in charge of establishing the new Ecuadorian building code (MIDUVI-CCQ, 2011). Beauval et al. (2010, 2013) created a unified and homogeneous earthquake catalog for Ecuador and borders covering five centuries and integrating instrumental and historical events (©) see Tables S1 and S2, available in the...
The present study proposes to estimate PSH in Quito based on the Alvarado (2012) zoning coupled with the newly published earthquake catalog. The OpenQuake engine is used for all PSH calculations (Global Earthquake Model, www.globalquakemodel.org/ (last accessed October 2014); Crowley et al., 2013; Pagani et al., 2014). We focus on hazard estimate at 475 years return period, corresponding to a 10% probability of exceedance over 50 years. Our approach is first to identify the controlling parameters and then to evaluate the uncertainties on these parameters and the corresponding impact on the hazard levels.

**PROBABILISTIC SEISMIC HAZARD IN QUITO: THE HOST ZONE IS CONTROLLING THE HAZARD**

The territory of Ecuador has been subdivided into seismic sources, producing a seismotectonic model with 26 source zones. The criteria used for deriving this model are described in Alvarado (2012). Crustal sources enclose fault systems or zones of diffuse seismicity. Recurrence curves are modeled based on the observed seismicity rates using the earthquake catalog (see Tables S1 and S2 and the time windows of completeness published in Beauval et al. (2013) and reported in Table 1. Different complete time periods have been determined for shallow earthquakes occurring in the Cordillera and for subduction earthquakes. For example, all earthquakes with magnitudes \( M_w \geq 4.5 \) are considered complete since 1963, whereas earthquakes with magnitudes \( M_w \geq 6.0 \) are complete since 1860 in the Cordillera and since 1900 for the plate margins (interface events) and at depth (inslab events). Recurrence rates are estimated for magnitudes 4.5 and above, together with the \( b \)-value (applying Weichert, 1980; see, e.g., Beauval and Scotti, 2003). All crustal sources are modeled as a real zones, with seismicity rates distributed over depth according to the observed depth distribution of instrumental earthquakes in each zone (up to 35 km in the Cordillera). Based on selected earthquake focal mechanisms and tectonic analysis (Alvarado, 2012), predominant focal mechanisms are identified and further taken into account in the prediction of ground motions. The subduction interfaces are modeled as dipping fault planes, the segmentation along the trench relies on the rupture zone estimated from past megaeartquakes (Esmeraldas zone, ...
corresponding to the rupture plane of the 1906 $M_w$ 8.8 event) or on the seismicity pattern observed along the interface zone. The subduction inslab zones are modeled as volumes, with seismicity distributed between 35 and 200 km. Maximum magnitudes in each source zones are based on the maximum length of fault segments identified as active using the magnitude–length equations published in Leonard (2010) and Strasser et al. (2010). We checked that these magnitudes inferred from active tectonics are always higher or equal to the maximum historical magnitudes recorded in the zones.

The first hazard calculations identify the contributions to the hazard in Quito of crustal sources, inslab volumes, and interface planes. Calculations are performed for a rock site located in Quito (coordinates $-78.51$ in longitude and $-0.2$ in latitude, Figs. 2 and 3), applying the GMPE for a $V_{S30}$ of 760 m/s (shear-wave velocity in the top 30 m). To begin with, the GMPE Zhao et al. (2006) is used for the three tectonic regimes to predict the ground motions produced by earthquakes. Zhao et al. (2006) is based on the rupture distance (closest distance to the fault plane, $R_{rup}$). The minimum magnitude considered in the probabilistic calculation is $M_w$ 5.0; Gaussian distributions predicted by GMPEs are truncated at $\pm 3\sigma$. Hazard curves are calculated taking into account all source zones and then considering separately the crustal source zones, the inslab sources, and the interface sources (Fig. 4a, for peak ground acceleration [PGA]). Using the GMPE Zhao et al. (2006) for the three tectonic regimes, an acceleration of 0.41 g is obtained at 475 years, when either all sources or only the crustal host zone is considered. These results show that, for a site on rock ($V_{S30}$ 760 m/s), there is only one source zone—the host crustal zone—that contributes significantly to the seismic hazard in Quito (PGA) for 475 years return period. The same calculations, using other GMPEs (for crustal sources: Boore and Atkinson, 2008; Akkar and Bommer, 2010, both based on Joyner and Boore distance $R_{JB}$; for subduction sources: Youngs et al., 1997), indicate that the host source always controls the hazard in Quito at 475 years. This result is also valid for spectral frequencies over the 0.1–1 s range. The uniform hazard spectra displayed in Figure 4b show that the spectral accelerations at 475 years based on the full seismicity

<table>
<thead>
<tr>
<th>Magnitude interval</th>
<th>4.0–4.5</th>
<th>4.5–5.0</th>
<th>5.0–5.5</th>
<th>5.5–6.0</th>
<th>6.0–6.5</th>
<th>6.5–7.0</th>
<th>7.0–7.5</th>
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<tr>
<td>Time period of completeness: subduction sources</td>
<td>1997</td>
<td>1963</td>
<td>1963</td>
<td>1963</td>
<td>1900</td>
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Considering the interface sources alone, an acceleration of 0.12g is obtained in Quito at 475 years return period (Fig. 4a). The interface contributing is the Esmeraldas zone, modeled as a dipping plane (20°) between 4 and 44 km depth, extending over the rupture area of the 1906 event (from latitude −0.5° to +4°, approximately 500 km long, Fig. 3a), with maximum magnitude 8.8 ($M_w$). The Gutenberg–Richter recurrence curve established from the seismic catalog predicts one earthquake with magnitude $\geq 7.0$ on average every 25 years in this source zone, and one earthquake with magnitude $\geq 8.0$ on average every 110 years. However, this interface is located at 200 km from Quito, and its contribution to the annual exceedance rate for an acceleration of 0.41g is negligible (calculations at rock).

**VARIABILITY OF THE UNIFORM HAZARD SPECTRUM AT 475 YEARS**

**Host Zone: Frequency–Magnitude Distributions**

We now focus on the host zone controlling the seismic hazard in Quito. The characterization of the Quito reverse fault system has improved significantly with the use of Interferometric Synthetic Aperture Radar (InSAR) by Champenois et al. (2013) and the use of geomorphology, tectonics, and GPS by Alvarado et al. (2014). However, the information available does not yet permit to propose a recurrence model for each fault segment. Rather than considering individual segments in the PSH calculations, an areal seismic source has been delineated enclosing the whole reverse fault system (Figs. 3b and 5a, c). The prolongation of the Quito fault system to the south is the reverse Latacunga system (see e.g., fig. 16 in Beauval et al., 2010, named Poalo-Saquisili-Yambo fault), both dipping to the west. Between these systems (between latitudes $−0.4°$ and $−0.7°$), no fault trace has been clearly identified, although historical earthquakes, for example in 1923, have occurred there (Beauval et al., 2013). Following Alvarado (2012), two options are proposed for delineating the source zone enclosing Quito (host zone). The first one (option S1, Fig. 5a) is a large area enclosing both fault systems, extending from latitude $−1.15°$ to $+0.5°$ (Quito–Latacunga zone). This option has been used in the previous calculations (see Probabilistic Seismic Hazard in Quito: The Host Zone is Controlling the Hazard section). As Alvarado et al. (2014) note, the north–south-trending Quito and Latacunga reverse faulting systems are geodynamically similar because they form the western boundary of the “Quito–Latacunga microblock” (∼150 km long). The fault systems located to the north and to the south of this microblock trend northeast–southwest and have right-lateral strike-slip mechanism. The microblock is characterized by discontinuous Quaternary folds and piggy-back basins related to blind thrusts with evidences of tilted young volcanoclastic deposits (Lavenu et al., 1995; Alvarado et al., 2014). Available focal mechanisms for $M_w \geq 5.0$ earthquakes suggest a common present day behavior along strike. The second option isolates each fault system and divides the large zone into three smaller ones (Quito, Machachi, and Latacunga zones, Fig. 5b,c). As the Latacunga source zone contains more earthquakes than the Quito source zone, the hazard evaluated in Quito will be higher for the large host zone option (S1) than for the small Quito host zone option (S2).

The recurrence curve corresponding to the Quito zone relies on very few events (Fig. 6b). Seven events with magnitudes
\( M_w \geq 4.5 \) belong to the source zone; the greatest magnitudes are \( M_I 5.8 \) (10 August 1938, Sangolqui event, Beauval et al., 2010) and \( M_I 6.4 \) (31 August 1587, Guayllabamba event, Beauval et al., 2010). Six events fall inside the completeness time windows \( 4.5 \leq M_w \leq 5.8 \). One option is to work with the rate and \( b \)-value inferred from this scarce catalog \( (b = 0.81) \). Around 0.11/year events with \( M_w \geq 4.5 \) occur within this source zone, equivalent to 1 event \( M_w \geq 4.5 \) every 9 years. Another option is

\[ a/b = 0.72 \]

\[ a/b = 0.97 \]

\[ a/b = 0.81 \]

\[ a/b = 0.97 \]

\[ a/b = 0.81 \]

\[ a/b = 0.97 \]

\[ a/b = 0.81 \]
Figure 7. Uniform hazard spectra for a rock site in Quito, at 475 years return period. (a) Testing two geometries for the host source zone, S1 and S2 (see Variability of the Uniform Hazard Spectrum (UHS) section), and three ground-motion prediction equations (GMPEs) for crustal events (AB2010: Akkar and Bommer, 2010; BA2008: Boore and Atkinson, 2008; Zetal2006: Zhao et al., 2006). (b) UHS resulting from the final logic tree including two geometries for the host zone, two optional b-values, and three GMPEs. Black indicates host source zone option S1 and blue indicates host source zone option S2. All source zones are taken into account in the calculations (crustal, interface, inslab), although interface/inslab contributions are negligible at this return period.

Variability of the Uniform Hazard Spectrum (UHS)

To begin with, three equations are selected to explore the epistemic uncertainty of the ground-motion predictions from crustal earthquakes: Zhao et al. (2006), established from Japanese data; Akkar and Bommer (2010), based on European and Middle East data; and Boore and Atkinson (2008), using western United States strong motions. Tested against diverse strong-motion datasets, these equations proved to be robust and stable models over the full frequency range (e.g., Delavaud et al., 2012; Beauval, Tasan, et al., 2012). As inslab and interface sources have a negligible contribution to the hazard at 475 years, only one GMPE is considered for these sources, namely, Zhao et al. (2006), which was identified among best-fitting subduction models for South America in Arango et al. (2012) and Beauval, Cotton, et al. (2012). We confirmed that the present results are similar if Youngs et al. (1997) is used.

We estimated PSH using two definitions for the host zone contour (S1 or S2, seismic rates for M 4.5+ and b-values based on the zone dataset), and the three selected GMPEs. The six combinations of source zones and GMPEs result in six uniform hazard spectra at 475 years (Fig. 7a). As expected, using the zoning S2 provides lower hazard estimates in Quito than when using the zoning S1. For the PGA, results vary between 0.32g and 0.55g, depending on the zonation used and on the GMPE selected. Based on Akkar and Bommer (2010), the PGA increases from 0.46g (S2) to 0.55g (S1); whereas in the case of Boore and Atkinson (2008), the PGA increases from 0.32g to 0.41g. Keeping the zoning fixed (S2), the PGA increases from 0.32g to 0.46g if using Akkar and Bommer (2010) instead of Boore and Atkinson (2008) (0.41g–0.55g in the case of S1). These calculations show that the hazard in Quito at 475 years is controlled both by the definition of the areal host zone (~20%–30% variability at the PGA, ~30%–35% variability at 0.5 s) and by the choice of the GMPE selected for this zone (~35%–45% variability at the PGA, ~30%–35% variability at 0.5 s), for all spectral periods up to 0.5 s (Fig. 7a). At the spec-
tral period of 1 s, the impact of the choice of the host zone is much higher than the impact of the GMPE choice.

The recurrence model in the source zone S2 is not well constrained. The rate for events with \( M_g \geq 4.5 \) and the \( b \)-value, used for extrapolating this rate to higher magnitudes, rely on very few events (see Host Zone: Frequency–Magnitude Distributions section). To take into account the uncertainty on the recurrence model, two more recurrence models are included in the analysis. Gutenberg–Richter curves based on the rates of \( M \geq 4.5 \) coupled with the well-constrained 0.97 \( b \)-value of the Sierra region (Fig. 6) are now considered. This \( b \)-value is larger than the \( b \)-values estimated from the zone dataset, producing lower seismic rates over the magnitude range [4.5–7.0]. The uncertainty in the predicted recurrence of earthquakes with magnitudes 6 and larger is significant. The mean recurrence time in the Quito–Latacunga zone S1 varies between 33 and 90 years, depending on the \( b \)-value selected. In the case of the smaller Quito zone S2, the recurrence time for an earthquake \( M \geq 6.0 \) varies between 166 and 285 years. Results for this simple logic tree, obtained from the combination of four seismicity models for the host zone and three crustal GMPEs, are displayed in Figure 7b. The PGA in Quito at 475 years varies between 0.28g and 0.55g, with the mean value around 0.39g. At 0.2 s (5 Hz), accelerations vary between 0.63g and 1.28g, with a mean around 0.86g, whereas at 0.5 s (2 Hz), accelerations vary between 0.32g and 0.7g, with a mean value around 0.45g.

Frequency–Magnitude Distributions Based on the Slip Rate and Corresponding Hazard
Recent development of geodetic (GPS) networks in Ecuador provided the first present-day estimates of crustal deformation and slip rates on major faults (Nocquet et al., 2014). In the Quito area, GPS results spanning a period of ~15 years show a horizontal shortening rate at ~4 mm/yr between sites located west and east of the Quito fault system (Alvarado et al., 2014). Alvarado et al. (2014) further show that GPS data are well modeled by a single fault with an associated slip rate ranging from 4 to 5.3 mm/year. Moreover, a sharp velocity gradient observed across the Quito fault system indicates that only a fraction of the fault plane is presently accumulating elastic stress, available for future earthquakes. More precisely, Alvarado et al. (2014) found that the depth over which elastic stress is presently accumulated is in the range of 3–7 km.

The average slip-rate estimates can be used to propose alternative earthquake occurrence relations independent of the earthquake catalog (see e.g., Anderson and Luco, 1983; Youngs and Coppersmith, 1985; Bungum, 2007). Under the assumption that deformation remains steady in time, geodetically derived fault slip rates can be used to propose earthquake frequency–magnitude distributions consistent with the annual rate of moment deficit accumulation. The Quito zone is an area of approximately 70 km × 40 km, which encloses the Quito fault. Most of the seismicity in the zone is related to the Quito fault system. This fault system is composed of several blind thrust segments, the exact geometries and extensions in depth of which are yet to be defined (Alvarado et al., 2014). For the purpose of the calculation, a simplified geometry is considered. The fault is modeled as a single segment of 50 km (subsurface length, latitudes 0.0275° to −0.4238° and longitudes −78.385° to −78.50°, Fig. 2).

Although the fault slip rate constrains the seismic moment rate to be released on the fault, a model is required to distribute it through earthquakes of various magnitudes. Anderson and Luco (1983) propose a frequency–magnitude distribution constrained by the slip rate, the \( b \)-value, and the maximum magnitude \( M_{\text{max}} \) on the fault. They have reviewed several forms of recurrence relationships that have been developed using slip-rate constraints. Following the work achieved in the SHARE project (Woessner et al., 2012), the model number 2 providing the activity rate \( N_2 \) is selected (table 3 in Anderson and Luco, 1983; equation 7 in Bungum, 2007). The cumulative number of earthquakes with magnitude greater than \( M \) is modeled by an exponential function truncated at \( M_{\text{max}} \). The density function decreases continuously to zero as \( M \) approaches \( M_{\text{max}} \). The number of earthquakes \( N \) above magnitude 5.0 \( (M_{\text{min}} \) in the PSH calculation) is calculated as follows:

\[
N(m \geq 5.0) = \frac{\beta}{\ln(10)} \frac{d \ln(10)}{ \ln(10)} \ln(b \ln(10) \times (M_{\text{max}} - M) - 1) \times e^{-\beta \ln(10) \times (M_{\text{max}} - M)}
\]

in which

\[
\beta = \sqrt{\frac{\alpha M_0(0)}{\mu W^2}}; \quad \alpha = \frac{D}{L};
\]

\[
\log M_0 = \epsilon + dM(\epsilon = 160.5, d = 1.5)
\]

\( S \) is the slip rate. The \( b \)-value is assumed equal to the \( b \)-value from the Quito zone dataset (0.81, Fig. 6b). The well-known log–linear relation between the seismic moment \( M_0 \) and moment magnitude is used (Kanamori and Anderson, 1975). The rigidity modulus is fixed to \( 3 \times 10^{11} \) dyn/cm². The parameter \( \alpha \) is the ratio of the average displacement \( D \) in the largest earthquake rupturing the total width to the fault length \( L \). This parameter bears large uncertainties. Considering an average displacement from 1 to 2\( m \) (earthquake with magnitude ~7.0, Wells and Coppersmith, 1994), and assuming a length equal to 50 km, yields the range \( 2 \times 10^{-5} \) to \( 4 \times 10^{-5} \) for \( \alpha \). These values are in accordance with \( \alpha = 2 \times 10^{-5} \) recommended for thrust faults in Anderson and Luco (1983). The equation is applied using both \( \alpha \) values.

Here, we present two sets of calculations. The first set directly uses the geodetically derived slip rate for the Quito fault, and four calculations are then performed based on the minimum (4.3 mm/yr) and maximum (5.3 mm/yr) slip-rate bounds, combined with two \( \alpha \) values. This calculation implicitly assumes that the fault is locked over the entire seismogenic thickness. The second set accounts for aseismic slip on the Quito fault. Assuming that \( \varepsilon \% \) of the fault surface is creeping, the annual seismic moment rate deficit is
in which $A$ is the surface of the fault. It is therefore equivalent to divide $A$ or $S$ to obtain the same annual moment rate deficit. Most faults on continents are locked over the whole seismogenic upper crust that is $\sim 15$ km. The locking depth of 7 km indicated by the GPS results therefore suggests that $\varepsilon$ is close to $\sim 50\%$. We therefore perform the calculation using 50% of the slip rate ($S/2$) to derive a frequency–magnitude distribution for the case of the partially locked Quito fault.

The first set of calculations provides frequency–magnitude distributions with many more earthquakes than has been observed in the past (Fig. 8a). As a consequence, the resulting recurrence models are predicting higher rates than those inferred from the earthquake catalogs. This trend, that is, modeled rates based on fault slip rates higher than rates based on past seismicity, has been observed in other seismic-hazard studies using GPS strain rates, for example, Mazzotti et al. (2011) or in the SHARE project (Woessner et al., 2012). If one divides the measured slip rates into an aseismic and a seismic release, the predicted rates are lower and in agreement with the recurrence inferred from past seismicity (Fig. 8c).

The UHS at 475 years is calculated considering four seismicity models (Quito zone S2, minimum and maximum slip rates, two optional $\alpha$ values) and the three selected crustal GMPEs (Akkar and Bommer, 2010; Zhao et al., 2006; Boore and Atkinson, 2008). (c) and (d) Same as (a) and (b) with a slip rate reduced by 50% (consistent with a partially locked fault).
are significantly higher than those relying on the catalog-based recurrence curves (Fig. 7). Accelerations at the PGA vary between 0.43g and 0.73g, at 475 years, with a mean close to 0.55g. Assuming $c$ equal to 50% (Fig. 8d), the accelerations are close to those relying on the catalog-based recurrence curves. Accelerations at the PGA vary between 0.32g and 0.58g, with a mean around 0.42g.

**Restricting Large Earthquakes to the Quito Fault and Hanging-Wall Effect**

**PGA at 475 Years: Profile Perpendicular to the Fault**

When modeling the seismicity with areal sources in the probabilistic calculations, the hanging-wall effect cannot be taken into account properly. The city is lying over the hanging wall of the Quito reverse fault system, and in case of a large earthquake below the city, ground motions are expected to be much higher in Quito than in the suburbs at the foot of the hills (footwall, see Fig. 2). There are several past examples of this effect, for example, during the Chi-Chi $M_w$ 7.1 earthquake in Taiwan \cite{Chang et al. 2004}. As an attempt to include hanging-wall amplifications, two more GMPEs for crustal earthquakes are included, the Abrahamson and Silva (2008) and Chiou and Youngs (2008) models.

Within the Quito areal zone, all large earthquakes (e.g., $M \geq 6.0$) are expected to occur on the identified segments of the reverse fault system. Rather than distributing all seismicity rates ($M = 5–7$) homogeneously over the Quito areal zone, an alternative is to distribute the seismicity rates of magnitudes 5–6 over the zone (as background seismicity) and assign the rates of magnitudes 6–7 on the Quito fault plane. To perform this exercise, the simplified fault geometry described in the Frequency–Magnitude Distributions Based on the Slip Rate and Corresponding Hazard section is used. The reverse fault is dipping to the west with an angle of 50°, extending from 3 to 18 km depth (Fig. 9, corresponding to a width around 7.0 km). Wells and Coppersmith, 1994). There is a significant uncertainty in this geometry, however, the dip and extension in depth are compatible with the distribution of microseismicity and with geomorphological characteristics (relocations of tectonic events by Lamarque \cite{2011} and Font et al. \cite{2013}, analysis of local microseismicity by Alvarado et al. \cite{2014}). Each segment of the fault system has a main compressional and secondary dextral strike-slip component, confirmed by the available focal mechanisms \cite{Segovia and Alvarado, 2009; Alvarado et al., 2014}.

At first, calculations are performed applying the frequency–magnitude distribution based on the Quito zone dataset ($b = 0.81$ and rates of $M \geq 4.5$ equal to 0.11, Fig. 6b). Four GMPEs are applied: Akkar and Bommer (2010), Boore and Atkinson (2008), Abrahamson and Silva (2008) and Chiou and Youngs (2008) (distance measures $R_{JB}$, $R_{JB}$, $R_{JB}$, and $R_{JB}$, respectively). The PGA obtained at 475 years, for sites located on a profile perpendicular to the fault, is displayed in Figure 10a (profile in Fig. 2). For comparison, accelerations obtained distributing all the seismicity over the areal source zone case are also superimposed \cite{Variability of the Uniform Hazard Spectrum (UHS) section}. Concentrating the occurrence of magnitudes 6–7 on the fault plane produces an increase of acceleration levels at sites located above the fault plane ($R_{JB} = 0$). Applying the models Akkar and Bommer (2010) and Boore and Atkinson (2008) (no hanging-wall coefficient) or Chiou and Youngs (2008) (with hanging-wall coefficient) at the sites located above the fault plane results in an increase in accelerations up to 20%–30%. Maximum accelerations of 0.53g at 475 years are obtained applying the Akkar and Bommer (2010) ground-motion equation. Applying the equation by Abrahamson and Silva (2008) produces a greater acceleration increase (35% at maximum) with respect to the areal zone case. If using the larger host zone option (zone Quito–Latacunga), the hazard obtained in Quito is higher than when using the smaller Quito host zone. Restricting magnitudes 6–7 to the fault plane (rates proportionated to the Quito zone area) increases accelerations at 475 years by 20%–40% (Fig. 10b), reaching a maximum of 0.68g (Akkar and Bommer, 2010 or Abrahamson and Silva, 2008 models).

**Scenarios and Hazard Values at 475 Years**

The acceleration at 475 years is a threshold, corresponding to the acceleration at the site with a 10% probability of being exceeded at least once over 50 years. All earthquakes included in the model, with low-to-high magnitudes, close or far from the site, with a nonzero probability of producing an acceleration greater than this threshold, contribute to this hazard calculation. It can be interesting, although requires care, to compare these accelerations at 475 years with acceleration levels corresponding to different earthquake scenarios. We believe that such exercises can be helpful to better grasp the meaning of the probabilistic output. The median acceleration to expect in case of a given earthquake on the Quito fault is superimposed on the previous profiles, applying the Abrahamson and Silva...
could be repeated anywhere on the fault. Two larger events are the GMPE, the acceleration levels at 475 years are greater than show that, whatever the decision on the seismicity model or 1997). Results displayed in Figure 10a,b consider that of an earthquake that occurred in 1990 on a northern segment of the fault (Pomasqui, 11 August 1990, Fig. 2), which could be repeated anywhere on the fault. Two larger events are considered: an earthquake with $M_w 6.0$, which probably occurred at least once in the last five centuries (Beauval et al., 2010, 2013), and an earthquake with $M_w 7.0$ corresponding to the complete rupture of the fault (for which there is evidence of only one, from paleoseismology, 500–1000 years ago, Hibsch et al., 1997). Results displayed in Figure 10a,b show that, whatever the decision on the seismicity model or the GMPE, the acceleration levels at 475 years are greater than the acceleration corresponding to the $M_w 5.3$ scenario. Considering the worst scenario in our seismicity model, $M_w 7.0$, the maximum acceleration to expect (if considering only the median) exceeds the maximum acceleration at 475 years, whatever the combination of models chosen.

**CONCLUSIONS**

In the present study, PSH estimates at 475 years return period for Quito, capital city of Ecuador, show that the crustal host zone is the only source zone that determines the city’s hazard levels for such return period. Therefore, the emphasis is put on identifying the uncertainties characterizing the host zone, that is, uncertainties in the recurrence of earthquakes expected in the zone and uncertainties on the ground motions that these earthquakes may produce. As the number of local strong ground motions is still scant, GMPEs are imported from other regions. Rather than sampling a complex logic tree, several plausible models are considered and associated with the corresponding uniform hazard spectra.

Exploring recurrence models for the host zone based on different observations and assumptions, and including three GMPE candidates (Zhao et al., 2006; Boore and Atkinson, 2008; Akkar and Bommer, 2010), we obtain a significant variability on the estimated acceleration at 475 years (site coordinates: $\pm 78.51^\circ$ in longitude and $\pm 0.2^\circ$ in latitude, $V_{S30} 760$ m/s):

- Considering historical earthquake catalogs, and relying on frequency–magnitude distributions where rates for magnitudes $6.7$ are extrapolated from statistics of magnitudes $4.5–6.0$ mostly in the twentieth century, the acceleration at the PGA varies between $0.28g$ and $0.55g$ with a mean value around $0.4g$. The results show that both the uncertainties in the GMPE choice and in the seismicity model are responsible for this variability.
- Considering slip rates inferred from geodetic measurements across the Quito fault system, and assuming that most of the deformation occurs seismically (conservative hypothesis), leads to a much greater range of accelerations, $0.43g–0.73g$ for the PGA (with a mean of $0.55g$).
- Considering slip rates inferred from geodetic measurements, and assuming that 50% only of the deformation is released in earthquakes (partly locked fault, model based on 15 years of GPS data), leads to a range of accelerations $0.32g–0.58g$ for the PGA, with a mean of $0.42g$. 

\[
\begin{array}{c}
\text{PGA at 475 years} \\
\text{(a)} \\
\text{Longitud} \\
\end{array}
\]

\[
\begin{array}{c}
\text{PGA at 475 years} \\
\text{(b)} \\
\text{Longitud} \\
\end{array}
\]
These accelerations are in agreement with the catalog-based hazard estimates.

- Restricting the occurrence of magnitudes 6–7 to the Quito fault (a simplified geometry), applying the three initial GMPEs (Akkar and Bommer, 2010; Zhao et al., 2006; Boore and Atkinson, 2008) or GMPEs including a hanging-wall coefficient (Abrahamson and Silva, 2008; Chiou and Youngs, 2008), increases the hazard by 20%–40% at sites located above the fault plane (range 0.42g–0.68g at the considered site). Strong hypotheses are required to define a simple fault plane and to define the recurrence of earthquakes on this fault plane; therefore, these results must be taken with great caution. However, they demonstrate that taking into account faults in hazard calculations can have a major impact.

Modeling the recurrence based on the past earthquake catalog, and relying on an areal source zone model, gives a mean value around 0.4g for the PGA at 475 years in Quito. This mean value is for a site on rock, and site effects need to be further taken into account. These results are in accordance with the acceleration level recently adopted by the new Ecuadorian Building Code (MIDUVI-CCQ, 2011). The seismic provisions of this Building Code are based on our 2011 national PSHA map at rock, based on a single best-estimate model (no logic tree and no exploration of uncertainties). Nonetheless, based on various exercises, we show that if taking into account the fault itself in the hazard calculations, much higher values can be obtained for sites located above the fault. Interdisciplinary studies must be pursued to better understand paleoseismicity and fault kinematics around Quito. Soon there will be enough recordings available from the recently installed strong-motion stations, so that imported GMPEs can be tested to refine the selection. Future research should also focus on understanding better how to include source–site geometry effects such as the hanging wall, as these effects will certainly have direct consequences on the damage distribution in case of a large destructive earthquake.

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