

Regional-Scale Controls on Earthquake Ground Motions and Their Implications for Assessing Historical Earthquake Magnitudes

Aisling O’Kane^{*1,2} and Alex Copley¹

ABSTRACT

Historical earthquake magnitudes estimated from building damage distributions are essential for constraining the potential magnitudes of future earthquakes on fault systems and their accumulated slip deficits, which underpin seismic hazard assessments. However, earthquake-induced damage patterns reflect complex trade-offs between earthquake magnitude and location, wave propagation effects, site conditions, and building vulnerability. The relative influence of these factors remains poorly quantified in continental foreland basin settings, where thrust earthquakes generate destructive ground shaking that propagates through laterally varying geological structures. This study addresses this issue using seismic wavefield simulations with varying earthquake magnitudes, thrust fault locations, and basin structures. Combining these simulated ground motions with established peak ground velocity–intensity–magnitude relationships, we quantify how source and basin characteristics can contribute to over- or underestimation of historical earthquake magnitudes that are based upon intensity observations. The simulation results show that proximity of the fault rupture to the foreland basin and earthquake magnitude are the dominant controls on the resultant ground motions. Increasing the fault’s distance from the range front by ~50 km produces ground-motion reductions equivalent to decreasing earthquake magnitude by 1–2 M_w units. Basin structure superimposes an additional control on these ground motions, increasing peak ground accelerations by factors of 1.5–2 near the basin margin but becoming negligible for more distant events. In range-front settings, this study quantifies which locations and magnitudes of earthquakes are likely to be unrecorded in the historical record. In addition, we show that whether or not the presence of foreland basin structure is accounted for when analyzing macroseismic intensity distributions can lead to misestimation of historical earthquake magnitudes by up to 0.5 M_w , with total source–path–site uncertainties reaching ~1 M_w . These systematic biases have significant implications for assessing historical moment release in continental collision zones.

KEY POINTS

- Complex trade-offs obscure the relative roles of the source, site, and path effects on earthquake damage.
- Earthquake location and magnitude dominantly control ground motions and damage.
- Source and basin effects can bias magnitude estimates for historical earthquakes.

INTRODUCTION

Historical earthquake magnitudes estimated from building damage distributions provide critical constraints on slip deficits for major fault systems and the possible magnitudes of future events, informing seismic hazard assessments in

tectonically active regions. However, interpreting historical damage and intensity patterns to infer earthquake source parameters requires understanding complex trade-offs between the earthquake’s magnitude and rupture location, wave propagation through subsurface geology, and site conditions. Empirical ground-motion models (GMMs) account for

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source, path, and site effects (e.g., Joyner and Boore, 1981; Campillo *et al.*, 1989; Anderson *et al.*, 1996; Atkinson and Silva, 1997; Boore *et al.*, 2014; Kotha *et al.*, 2020). Modern GMMs (e.g., Abrahamson *et al.*, 2014; Boore *et al.*, 2014) capture the well-known ground-motion attenuation relationships with distance, arising from geometric spreading and anelastic attenuation (Joyner and Boore, 1981). These relationships have been empirically derived from earthquakes across a wide range of magnitudes and tectonic settings, including the United States ($M_w \sim 2$ to 8; Campbell and Bozorgnia, 2014; Pasyanos, 2015), Japan ($M_w \sim 5$ to 8; Kanno *et al.*, 2006; Zhao *et al.*, 2006), and New Zealand ($M_w \sim 3$ to 7; Bradley, 2013; Atkinson, 2024). However, disentangling the trade-offs between earthquake magnitude and location in controlling ground shaking remains challenging, as even GMMs incorporating site or basin terms cannot fully account for the 3D influence of subsurface structure or be uniquely isolated from observational data (Bakun and Wentworth, 1997; Gasperini *et al.*, 2010). Accurately quantifying the relative influence of earthquake magnitude, rupture location, and subsurface structure on ground-motion intensity is therefore essential for reliably interpreting historical damage patterns, which has implications for assessing the rate of past moment release, the magnitude of past events, and the size of slip deficits along fault systems.

Earthquake magnitudes estimated from structural damage reports often vary from those derived from direct observations (Bindi *et al.*, 2013; Middleton *et al.*, 2016; Feng *et al.*, 2020; Ou *et al.*, 2020). For example, the 1905 Kangra earthquake in India has been assigned magnitudes ranging from M_s 7.5–8.6 based on instrumental records and M_i 7–8 from the Middlemiss (1910) intensity-based damage survey (Ambraseys and Bilham, 2000), with improved analysis and geodetic constraints converging on M_w 7.8 (Ambraseys and Bilham, 2000; Hough *et al.*, 2005). Different observational methods can also yield inconsistent results, providing an additional, geological source of magnitude uncertainty distinct from the wave-propagation effects examined in this study. For example, geomorphological observations (e.g., surface ruptures identified in paleoseismic studies) in continental collision zones often indicate smaller earthquake magnitudes than seismological or building damage studies, due to some of the seismic slip being buried at depth. The 1911 Chon-Kemin earthquake in the northern Tien Shan (Kyrgyzstan and Kazakhstan) demonstrated this effect, with analog seismic records yielding a moment magnitude of 8.0 (Kulikova and Krüger, 2015), whereas Delvaux *et al.* (2001) and Arrowsmith *et al.* (2017) determined smaller magnitudes ($M_w \sim 7.8$) from preserved surface ruptures. Similarly, the 1992 Suusamyr earthquake in Kyrgyzstan, despite having a surface-wave magnitude of 7.3, only produced a series of short, discontinuous surface ruptures totaling 7.6 km in length (Ghose *et al.*, 1997; Ainscoe *et al.*, 2019). Even great earthquakes do not always produce clear surface ruptures, such

as the 1934 M_w 8.4 Bihar–Nepal earthquake, whose surface rupture evidence is debated (Sapkota *et al.*, 2013; Wesnousky *et al.*, 2018). Beyond surface rupture production and preservation, it is important to consider how underlying wave-propagation effects influence magnitude estimates derived from damage distributions, for accurate interpretation of historical earthquakes.

Estimating the location of historical earthquakes introduces similar uncertainties, as damage distributions are influenced by site conditions (e.g., structural vulnerability; Malhotra, 1999, and subsurface geology; Campillo *et al.*, 1989; Anderson *et al.*, 1996; Joyner, 2000), path effects (e.g., directivity; Koketsu *et al.*, 2016, and surface-wave amplification; Campillo *et al.*, 1989), and are inherently weighted toward populated regions with the most records.

We focus on range-bounding thrust faults of continental mountain ranges in this study, for two reasons. First, these faults produce some of the largest-magnitude earthquakes in the continents (Molnar and Ghose, 2000; Bilham, 2019), and often have large populations concentrated in the adjacent foreland basins (Bilham *et al.*, 2001; Tiwari *et al.*, 2018), making them an important source of hazard to understand more fully. Second, the large spatial variations in subsurface structure that occur in the mountain belts and adjacent foreland basins (Hetényi *et al.*, 2006; DeCelles, 2011; Mitra *et al.*, 2011) present an ideal opportunity to investigate the effects of this structure on ground shaking, and the effects on the assessment of historical earthquakes. These range-bounding faults are often the most seismically active structures within collisional settings (Stevens and Avouac, 2015), and have historically caused catastrophic building damage and loss of life (England and Jackson, 2011), as exemplified by the 1905 M_w 7.8 Kangra earthquake discussed earlier, which caused $\sim 20,000$ deaths and $\sim 100,000$ building collapses (Ambraseys and Bilham, 2000). Although these thrust earthquakes rupture within the mountain belt, their most damaging ground motions can be generated in the adjacent foreland basins, where kilometer-thick low-velocity sediments amplify ground shaking (Bard and Bouchon, 1985; Joyner, 2000), and where some of the world's largest and most rapidly growing cities are concentrated (e.g., Delhi, Lahore; Cox, 2025). Although distance–attenuation relationships predict that earthquake ground motions decay systematically with distance, the rate of decay varies with magnitude (Boore *et al.*, 2014) and is strongly influenced by subsurface structure, particularly in basin settings (Wirth *et al.*, 2019; O'Kane and Copley, 2021). Therefore, gleaning information about earthquakes in the interiors of mountain belts from studying the distribution of damage in the adjacent, more populated foreland basins requires systematic quantification of the trade-offs between source characteristics, path effects, and site effects.

In this study, we use physics-based seismic wavefield simulations to systematically quantify the relative contributions of

earthquake magnitude, location, and basin structure to the resultant ground motions in continental range-front settings. Many of the model parameters we discuss subsequently are based upon the Himalaya, due to the wealth of available observational constraints and the large at-risk populations. However, the model geometry and tectonic characteristics are similar to those of continental range-front settings globally (e.g., Allen and Homewood, 1986; DeCelles, 2011), and the findings are therefore likely to apply to a range of foreland settings, including the Tien Shan, Zagros, European Alps, and Caucasus. We then convert these simulated ground motions to macroseismic intensities using established peak ground velocity (PGV)–intensity relationships (e.g., Wald *et al.*, 1999; Worden *et al.*, 2012; Caprio *et al.*, 2015), and use these intensities to estimate the earthquake magnitudes that would be inferred if these events were to be present in the historical record. This process allows us to test the sensitivity of historical earthquake magnitude estimates to whether foreland basin structure has been fully accounted for in the analysis. This approach demonstrates the potential for systematic biases in historical earthquake magnitude estimates derived from intensity distributions, with implications for assessing historical magnitudes and slip deficits in active collision zones.

First, a series of 18 3D seismic-wave-propagation models is conducted, with nine models containing a foreland basin and nine control models without. For each scenario, the seismic wavefield is calculated as a function of distance from the rupture, systematically varying the earthquake magnitude (M_w 5.5, 6.5, and 7.5) and fault location (0, 50, and 100 km horizontal distance between the up-dip edge of the rupture and the basin margin, hereafter referred to as the “distance from basin”). Next, the simulated ground motions from the basin and control models are compared to isolate the relative roles of source location, magnitude, and foreland basin structure on the ground-motion amplitude and frequency content. Finally, by converting modeled PGVs to modified Mercalli intensity (MMI), and analyzing these synthetic MMI distributions using techniques commonly used to estimate the magnitude of historic earthquakes, we quantify how source and basin characteristics can contribute to over- or underestimation of historical magnitudes.

EARTHQUAKE GROUND-MOTION MODELING

Methodology

The finite-difference code SW4 (seismic waves fourth order; Petersson and Sjögreen, 2017a) was used to perform 3D seismic-wave-propagation modeling of thrust-faulting earthquake scenarios along the range front of a mountain belt, with and without an adjacent foreland basin. This type of methodology has grown in popularity over recent years as an alternative to empirical GMMs, due to the advancement of computational resources (McCallen *et al.*, 2021), more accurate Earth velocity models (e.g., Aagaard *et al.*, 2019), a better understanding of

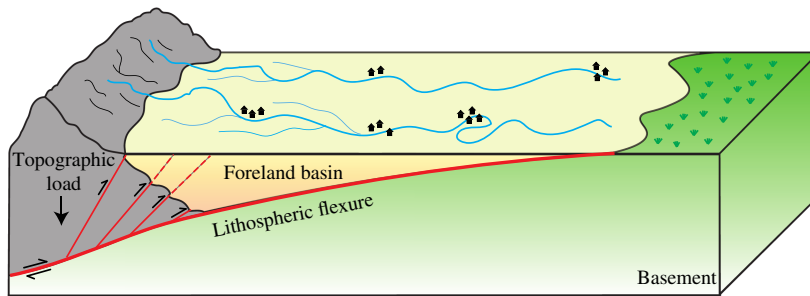
earthquake rupture processes (Hayes, 2017), and recognition that GMMs may not fully account for seismic-wave effects, such as surface-wave amplification (Bowden and Tsai, 2017). The physics-based method is particularly beneficial in seismically active areas that lack sufficient strong-motion records (e.g., the Himalayas; Dixit *et al.*, 2015), or that have not yet experienced enough recorded earthquakes, to allow robust empirical GMMs to be derived. This approach has been shown to accurately reproduce observations in well-instrumented settings such as California (Imperator and Gallovič, 2017; Rodgers *et al.*, 2018), and to estimate ground-motion amplitudes comparable to those derived from empirical GMMs where those models are well constrained by regional data (Rodgers *et al.*, 2019). Hybrid broadband simulations of Himalayan megathrust scenarios within the Indo-Gangetic foreland basin have shown that ground-motion intensities near the source at deep-basin sites can exceed those predicted by empirical GMMs (Jayalakshmi *et al.*, 2021), showing the utility of using multiple approaches.

Our modeling focuses on investigating the effects of earthquake magnitude and location (defined here as the distance perpendicular to strike from a foreland basin margin), and the presence or absence of the low-wave-speed foreland basin fill, on the ground motions produced. Therefore, SW4’s capability to account for source (e.g., Hall *et al.*, 1995; Bowden and Tsai, 2017), site (e.g., Sanchez-Sesma, 1987; Lebrun *et al.*, 2001; Lozano *et al.*, 2009), and path (e.g., Hollingsworth *et al.*, 2007; Chen *et al.*, 2018) effects, which control the earthquake-induced ground motions, make it an appropriate tool for this study. This approach differs from the empirical GMMs commonly used in probabilistic seismic hazard analysis (e.g., Abrahamson *et al.*, 2014; Boore *et al.*, 2014; Campbell and Bozorgnia, 2014; Chiou and Youngs, 2014), which cannot fully account for the same range of effects. SW4’s fourth-order accuracy in time and space, and ability to simulate ground motions at a regional scale (Sjögreen and Petersson, 2012; Petersson and Sjögreen, 2015), make it an appropriate methodology to accurately resolve frequencies relevant for city infrastructure (McCallen *et al.*, 2021), as detailed subsequently. Previous studies have shown that earthquake simulations of the type used in this study can be used to test the seismic response of buildings (Safak, 1999), perform seismic hazard analyses (Cui *et al.*, 2013) and disaster risk calculations (Lu and Guan, 2017).

Model setup and simulations

As illustrated in Figure 1, a simple geometrical model was used to represent a continental mountain range front, comprising a planar thrust fault bounding a sedimentary foreland basin, with an underlying crystalline basement. We use a 3D model to calculate the ground motions resulting from a range of earthquake sources. This three-dimensionality allows us to use fault planes with geometries that approximate those seen

(a) Cartoon schematic of a continental range front with adjacent foreland basin and basal thrust fault



(b) Computational seismic wavefield model of an idealised foreland basin setting

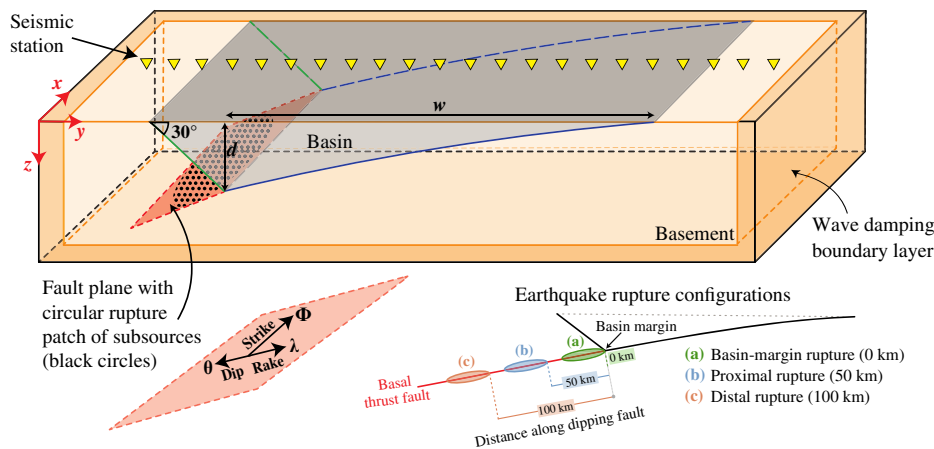


Figure 1. Conceptual and computational setup of the model used to simulate the seismic wavefield of range-front thrust faults in this study. (a) Schematic cross-section of a continental range front (gray) with basal thrust fault (red) dividing the adjacent foreland basin (yellow) and underlying crystalline basement (green). (b) 3D computational model domain comprising a low-wavespeed foreland basin (gray) overlying higher wavespeed crystalline basement (light orange). Yellow-inverted triangles represent a selection of the modeled seismic stations where ground motions are calculated, aligned with the along-strike center of the fault plane at 1 km intervals across the computational domain (in the y direction). A wave-damping boundary layer (dark orange) surrounds the model domain to prevent artificial reflections. The thrust fault (red) contains a circular rupture patch discretized into sub-sources (black circles). The up-dip location and depth of the earthquake rupture vary between models. The cross section on the lower-right of (b) illustrates the three earthquake rupture configurations considered in this study: (a) basin-margin rupture, in which the up-dip termination of the rupture patch is located at the basin margin (0 km); (b) proximal rupture, located 50 km along the fault plane from the basin margin; and (c) distal rupture, 100 km from the basin margin. The color version of this figure is available only in the electronic edition.

in continental collision zones, and to capture the rupture directivity effects on the ground motions that result from the earthquake slip propagating over those planes. Because flexure results in significantly larger strike-perpendicular than along-strike variations in geometry, we only vary the properties of the model in the vertical and strike-perpendicular directions. This feature means that the strike-perpendicular basin edges and depth variability play a role in controlling the modeled ground motions, but there are no effects from along-strike basin edges. We use a constant stress drop in our models, based on the scaling we use between earthquake magnitude and fault

dimension, and we subsequently discuss the effects of this choice.

The model domain, basin structure, earthquake rupture characteristics, and material properties follow the setup established in O’Kane and Copley (2021), which was developed based on observations of global mountain belt range-fronts (Allen and Homewood, 1986; Ford, 2004; Hauksson and Shearer, 2006; Hetényi *et al.*, 2006; Naylor and Sinclair, 2008; DeCelles, 2011; Mitra *et al.*, 2011; Srinagesh *et al.*, 2011; Chen and Wei, 2019), and the individual model parameters used in this study are outlined in Table 1. One of the main assumptions is that both the foreland basin and crystalline basement are homogeneous. Material properties were assigned to the basin sediments ($V_P = 4.4 \text{ km s}^{-1}$, $V_S = 2.5 \text{ km s}^{-1}$, $\rho = 2250 \text{ kg m}^{-3}$) and crystalline basement ($V_P = 6.0 \text{ km s}^{-1}$, $V_S = 3.5 \text{ km s}^{-1}$, $\rho = 2700 \text{ kg m}^{-3}$) based on observations from Himalayan foreland settings (Hetényi *et al.*, 2006; Tiwari *et al.*, 2006; Mitra *et al.*, 2011; Srinagesh *et al.*, 2011). To account for anelastic attenuation, quality factors were assigned as $Q_P = 120$ and $Q_S = 40$ for the basin sediments, and $Q_P = 160$ and $Q_S = 70$ for the crystalline

basement. These values reflect higher attenuation (i.e., lower Q) in softer sedimentary materials compared to hard crystalline rocks, consistent with observations from sedimentary basin settings (Olsen *et al.*, 2003; Hauksson and Shearer, 2006). Although it is well-known that lateral variations in the shallow velocity structure (e.g., the top 30 m) can have a significant effect on the amplification of earthquake ground motions in the near field (Anderson *et al.*, 1996), the focus of this study is to investigate the ground-motion distributions on a basin-wide scale, as opposed to individual building footprints. V_{S30} -based amplification factors (Borcherdt, 1994; Stewart and Seyhan, 2013) were

TABLE 1

SW4 Model Constants Used in the Seismic-Wavefield Simulations

Model Constant	Notation	Value
(a) Model Domain		
Along-strike length	x	50 km
Strike-perpendicular length	y	360 km
Depth	z	8 km
Grid spacing	h	0.05 km
(b) Basin Structure		
Maximum depth	d	3 km
Flexural width	w	200 km
Dip of fault bounding the basin margin	θ	30°
(c) Earthquake Rupture Characteristics		
Rupture velocity	V_{rup}	2.5 km/s ⁻¹
Angular frequency	ω	20 Hz
Fundamental frequency	f_0	3.18 Hz
Minimum frequency	f_{min}	0.08 Hz
Maximum frequency	f_{max}	7.95 Hz
(d) Material Properties		
Basin P -wave velocity	V_P	4.4 km/s ⁻¹
Basin S -wave velocity	V_S	2.5 km/s ⁻¹
Basin P -wave attenuation quality factor	Q_P	120
Basin S -wave attenuation quality factor	Q_S	40
Basin density	ρ	2250 kg m ⁻³
Basement P -wave velocity	V_P	6.0 km/s ⁻¹
Basement S -wave velocity	V_S	3.5 km/s ⁻¹
Basement P -wave attenuation quality factor	Q_P	160
Basement S -wave attenuation quality factor	Q_S	70
Basement density	ρ	2700 kg m ⁻³

SW4, seismic waves fourth order.

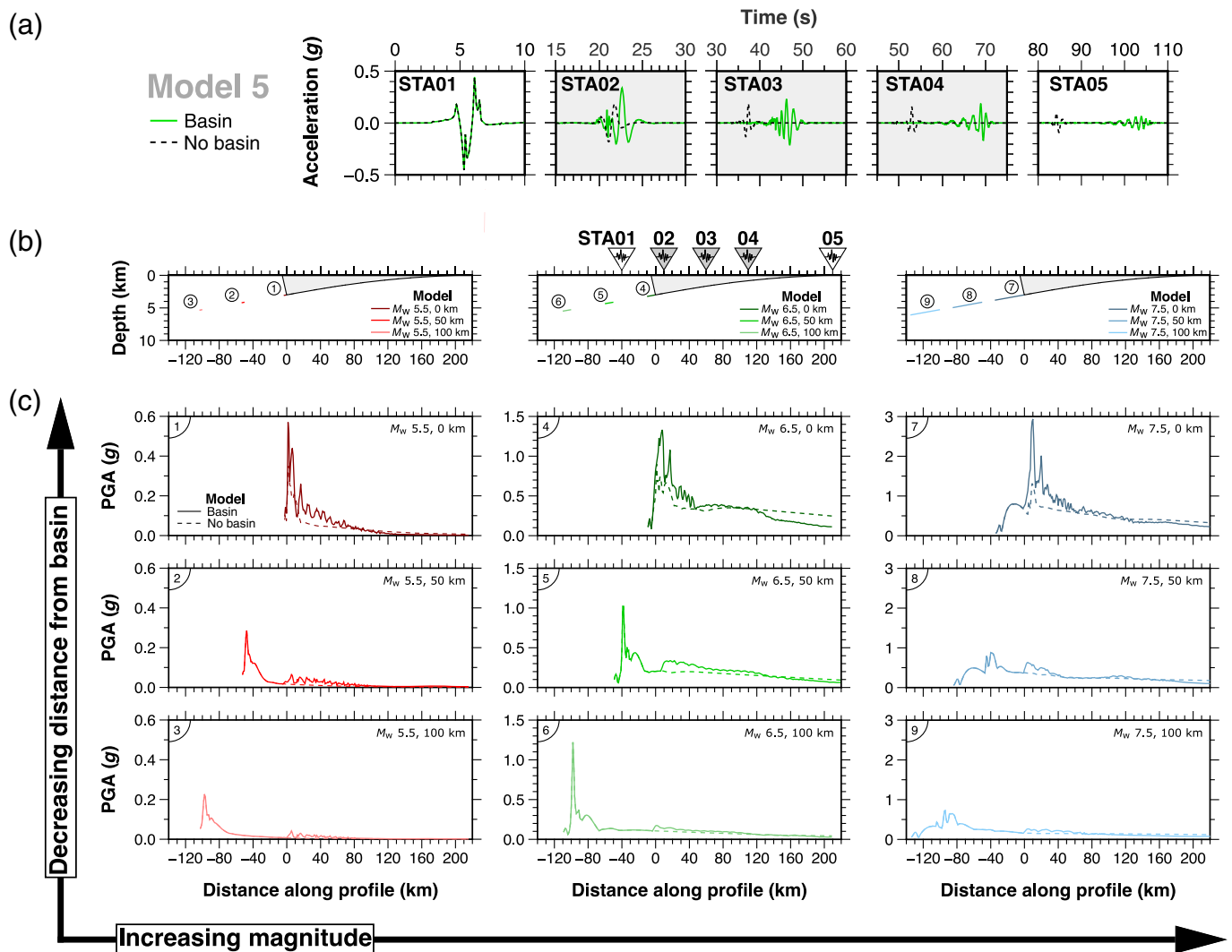
applied as a postprocessing sensitivity test to verify that shallow site effects do not significantly alter the basin-scale effects that we discuss here.

In this study, 18 earthquake scenarios were simulated. All model parameters outlined in Table 1 were kept constant, with the following parameters varied: (1) the earthquake magnitude, (2) the location of the earthquake relative to the range front, and (3) whether a low-seismic-velocity foreland basin is present. The earthquake magnitude also controls the fault dimensions, as the fault radii are scaled according to the Aki (1967) relationship with seismic moment for a circular crack, assuming a constant stress drop across all magnitude scenarios, and resulting in circular rupture diameters of ~ 3.5 , ~ 10 , and ~ 35 km for the M_w 5.5, 6.5, and 7.5 scenarios, respectively. Keeping the model domain, basin structure, material properties, and all other rupture characteristics constant between each simulation allows the ground-motion variations between models to be wholly the result of the changes in these three parameters. Although we note that our study therefore does not consider the effects of stress-drop variations between earthquakes. Foreland-basin geometries vary considerably depending on the elastic thickness of the lithosphere and topographic load exerted by the adjacent mountain ranges

(Allen and Homewood, 1986; DeCelles and Giles, 1996), which control their widths and depths. The basin geometry modeled in this study, with a maximum depth of 3 km (d in Fig. 1b) and strike-perpendicular flexural width of 200 km (w in Fig. 1b) represents a geometry that is consistent with commonly observed foreland basin dimensions (including the Himalayan foreland basin in northern India; O’Kane *et al.*, 2022). To replicate the distinctive wedge shape of a typical foreland basin, the Turcotte and Schubert (2014) end-load flexural model was adopted for the basement-sediment interface in the flexural part of the basin and the interface at the range-front side of the basin was set to dip at 30° (as illustrated in Fig. 1b), in line with observations by Ford (2004) and Hetényi *et al.* (2006). The fault is positioned as a down-dip extension of the flexural base of the foreland basin, resulting in a fault dip of $\sim 13.5^\circ$ for the 200 km wide, 3 km deep basin geometry modeled here, consistent with low-angle thrust-faulting observed along continental collisional zones globally (Ford, 2004; Talebian and Jackson, 2004; Hetényi *et al.*, 2006; O’Kane *et al.*, 2022).

We focus our analysis of the ground motions within the foreland basin region, as the flat topography in these regions means that the population densities are generally much greater than in the adjacent mountains. As we are aiming to test the sensitivity of historical earthquake magnitude estimates to the different earthquake source configurations explored here, we do not use an exposure model, but rather calculate and analyze the ground-motion metrics throughout the model domain.

In each of the model configurations (Fig. 2b), a thrust-faulting earthquake with an imposed moment magnitude of 5.5, 6.5, or 7.5 was simulated on a planar fault. This magnitude range is common on continental range-front thrust faults (e.g., in the Himalayas; Bilham, 2019; Andes; Devlin *et al.*, 2012; Tien Shan; Tapponnier and Molnar, 1979; Nelson *et al.*, 1987; Zagros; Maggi *et al.*, 2000; Talebian and Jackson, 2004, and Caucasus; Priestley *et al.*, 1994; Jackson *et al.*, 2002). In addition to the moment magnitude, an earthquake’s location is a major control on the amplitude and duration of the ground motions generated (O’Kane and Copley, 2021). Thrust-faulting earthquakes in foreland regions can rupture from the base of the seismogenic layer to the range front (e.g., 2008 Wenchuan earthquake; Hao *et al.*, 2009), or partially rupture at depth (e.g., 2015 Gorkha earthquake; Avouac *et al.*, 2015). To capture this variability, the up-dip edge of each rupture was positioned at 0, 50, or 100 km horizontal distance from the basin margin (defined as the place where the basin is deepest; Fig. 1). Both the magnitude and location of the earthquakes were independently varied so that the effects of each parameter can be directly isolated in the resultant ground motions. Simulations were run for each of the earthquake scenarios illustrated in Figure 2b, both with and without a foreland basin present, to isolate the effect that the basin sediments have on the ground motions. In total, this approach results in 18 models, with the variations in fault magnitude and geometry resulting



in 9 different configurations, and each of these being run with and without a foreland basin present in the velocity model.

Each earthquake rupture patch was formed of a network of sub-sources, positioned 25 m apart across a circular rupture area, each with an imposed Gaussian source time function, and with the entire array of sub-sources having an elliptical distribution of slip amplitude (i.e., the variation of slip with distance along the diameter of the rupture patch), after Bürgmann *et al.* (1994). Each subsource has a set angular frequency (ω) of 20 Hz, corresponding to a fundamental frequency ($f_0 = \omega/2\pi$) of ~ 3.18 Hz, and a rise time ($\sigma = 1/\omega$) of 0.05 s (Sjögreen and Petersson, 2012; Petersson and Sjögreen, 2017b). Analogous to past continental thrust-faulting earthquakes, each earthquake initiated at the down-dip edge of the rupture patch and radiated along-strike and up-dip from the along-strike center of the fault plane at 2.5 km s^{-1} (Wen *et al.*, 2012; Avouac *et al.*, 2015; Hayes, 2017), resulting in total rupture durations of ~ 1.4 , ~ 4.0 , and ~ 14.0 s for the M_w 5.5, 6.5, and 7.5 scenarios, respectively. The dense network of sub-sources ensured that each earthquake had a smooth rupture propagation along the fault plane. A computational grid

Figure 2. Ground acceleration as a function of earthquake magnitude and the distance between the rupture and the foreland basin, for modeled range front thrust-faulting earthquake scenarios. (a) Horizontal acceleration time series for a common earthquake scenario, model 5 (M_w 6.5 rupture located 50 km from the basin margin) at five stations across the model domain, three located on foreland basin sediments (gray shading) and two on crystalline basement material. The solid green line represents the results from the model with a foreland basin, whereas the black dashed line illustrates the results from the same model setup but without a foreland basin present. (b) Cross sections of each of the model setups, grouped and colored by magnitude. The left, middle, and right cross sections represent M_w 5.5 (models 1–3), M_w 6.5 (models 4–6), and M_w 7.5 (models 7–9), respectively, and the station locations for the seismograms shown in (a) are shown. (c) Plots illustrating the maximum peak ground acceleration (PGA), calculated at synthetic seismograms aligned with the along-strike center of the fault plane at 1 km intervals across the model domain (in the y direction in Fig. 1). Each plot shows the results from two model simulations, using the same solid- and dashed-line conventions as in (a). Note the different PGA scales for each column. The color version of this figure is available only in the electronic edition.

spacing of 50 m for the wave propagation simulation was selected to allow frequencies of 0.08–7.95 Hz to be accurately resolved, which encapsulates the frequencies at which most urban infrastructure oscillates (Ellis, 1980; Parajuli and Kiyono, 2015).

Each simulation recorded the seismic wavefield at synthetic receiver stations placed at 1 km intervals in the strike-perpendicular direction and aligned with the along-strike center of the model domain (see Fig. 1b). We model the full waveform at each station, and then use this waveform to calculate a range of ground-motion metrics (e.g., peak ground accelerations [PGAs] and displacements, and spectral accelerations [SAs] and displacements) for use in the subsequent analysis. SAs and displacements were calculated with 5% critical damping, using the established approach of Nigam and Jennings (1969).

To verify the simulated ground motions are robust to shallow site effects, V_{S30} -based amplification factors were applied as a sensitivity test. The V_{S30} parameter (average shear-wave velocity in the upper 30 m) was introduced by Borchardt (1994) to classify sites with similar seismic response and remains widely adopted in seismic design standards. The PGA amplification factors from Stewart and Seyhan (2013) were applied, as these are updated from Borchardt (1994) to account for nonlinear soil behavior and extend applicability to strong ground motions exceeding 0.5g, which is the case in several of the simulated earthquake scenarios. Seismic stations within the modeled basin were assigned National Earthquake Hazards Reduction Program site class D ($V_{S30} = 180 - 360$ m/s), representative of soft alluvial sediments in a foreland basin, while basement receiver stations were assigned Site Class B ($V_{S30} = 760 - 1,500$ m/s), consistent with crystalline rocks (Pudi *et al.*, 2022). Short-period amplification factors (F_{PGA}) for site class D decrease from 1.6 at weak shaking levels ($PGA < 0.1g$) to 1.1 at strong shaking levels ($PGA \geq 0.5g$), reflecting nonlinear soil behavior. Site class B stations were assigned a constant $F_{PGA} = 0.9$ at all shaking levels, to account for the slight deamplification at hard-rock sites relative to the reference velocity of $V_{S30} = 760$ m/s used in Stewart and Seyhan (2013).

Seismic-wavefield simulation results

We present the results of the model simulations using a range of different ground-motion metrics. We begin with PGA, a familiar metric that is widely used in the interpretation of historical earthquakes, before presenting SA and spectral displacement (SD), which provide a more complete representation of the ground-motion characteristics at a given location.

Figure 2 shows the results from the 18 model simulations conducted in this study. Figure 2a shows horizontal acceleration–time series at five stations across the model domain (in the y direction in Fig. 1b) for a common earthquake scenario

(model 5; M_w 6.5 rupture 50 km from the basin margin), with station locations shown in Figure 2b. Station STA01, proximal to the rupture, records the largest accelerations ($\sim 0.5g$) of the five stations shown, with ground-motion amplitudes decreasing with distance from STA01 to STA05. Stations on foreland basin sediments (STA02 to STA04) show greater amplitudes and longer duration shaking in the basin model relative to the no-basin control model, due to amplification and surface-wave dispersion within the basin and the generation of additional body waves through interaction with the basin–basement interface, as discussed in O’Kane and Copley (2021). The basin model waveforms also arrive later at these stations, with the separation in arrival times between the basin and no-basin models increasing with distance, as waves propagate more slowly through the low-velocity sediments than through the underlying crystalline basement. In contrast, the distal station STA05, located at >200 km from the basin margin, shows lower accelerations for the basin model than the no-basin equivalent, as amplitudes are decreased by surface-wave dispersion in the shallow distal portion of the basin (O’Kane and Copley, 2021).

Figure 2b illustrates the different model setups and Figure 2c presents the PGAs calculated at synthetic stations across the model domain, aligned with the along-strike center of the fault plane. As expected, the models with faults located at the basin margin (models 1, 4, and 7), and those with larger magnitude earthquakes generated the greatest PGAs. The models with basin-margin sources produce the highest ground motions because the shallow rupture depths generate strong near-source shaking, geometric focusing concentrates seismic energy at the basin edge, and resonance within the basin amplifies the ground motions. In contrast, sources located 50–100 km from the basin margin are deeper, and experience dispersion and attenuation before the waves enter the foreland basin, such that although body-wave amplification occurs within the basin, the resulting PGAs remain much lower than basin-margin equivalents. The mechanisms that control these source–basin interactions are described in detail by O’Kane and Copley (2021).

The model results provide insights into the trade-offs between earthquake location and magnitude in controlling the ground motions in a foreland basin region. For example, the PGAs within the foreland basin generated by model 9 (M_w 7.5 with fault rupture 100 km from the basin margin) are similar to those from model 1, which had a range-front rupture location, but was two moment magnitude units smaller (M_w 5.5). These magnitude–distance trade-offs have critical implications for interpreting historical earthquake magnitude estimates, as explored in the following sections.

The modeled PGA values are broadly consistent with observations of buried thrust-faulting earthquakes, supporting our choice of approach. The 1978 M_w 7.3 Tabas earthquake, which ruptured a shallow, gently dipping (20°) thrust fault, generated horizontal and vertical PGAs of 0.94g and 0.74g, respectively

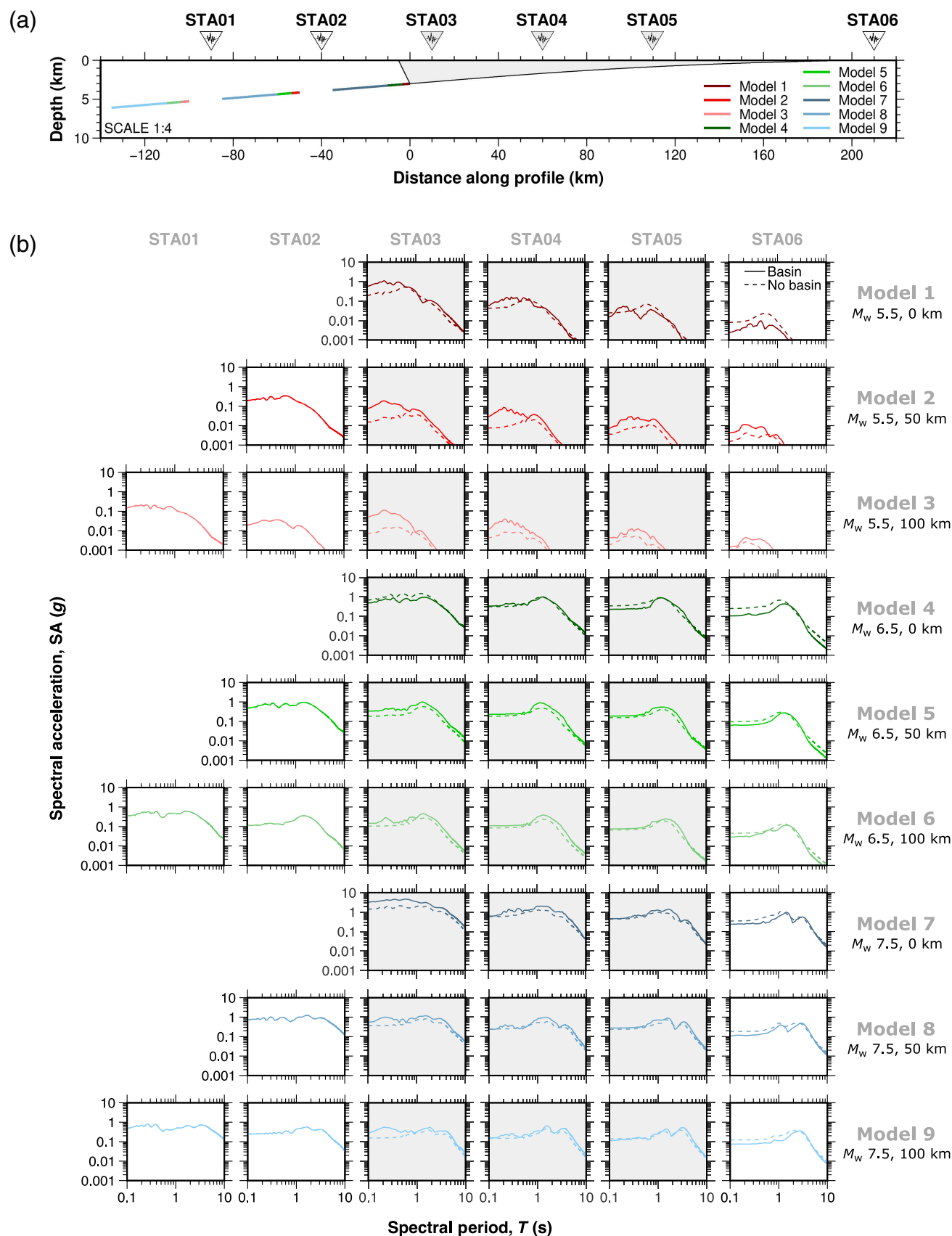


Figure 3. Spectral accelerations (SAs) recorded at select seismograms aligned with the along-strike center of the fault plane. (a) Cross-section illustrating the nine different model setups with faults differentiated by color. (b) Peak

SAs for six stations across the model domain, three located on foreland basin sediments (gray shading) and three on crystalline basement material. The color version of this figure is available only in the electronic edition.

(Niazi, 1986), comparable to model 8's horizontal ground motions (0.88g). Model 9 in Figure 2 illustrates an M_w 7.5 earthquake rupture located 100 km from the basin margin, which is roughly analogous to the 2015 Gorkha earthquake. Both the model and strong-motion recordings for this earthquake at far-field basin distances demonstrate lower-than-expected PGA values ($<0.1g$; Singh *et al.*, 2017) for a large earthquake, indicating strongly attenuated ground motions at these distances. More broadly, Suzuki and Iervolino (2017) compiled PGAs of 0.7–2.5g from eight M_w 5.6–7.0 global thrust-faulting earthquakes. These values demonstrate that the ground motions generated from the M_w 5.5–7.5 synthetic wavefield models (with maximum PGAs of 0.2g–3.0g) are sufficiently close to observed values that they can be used to investigate the controls on macroseismic intensity distributions in continental range-front settings.

Figure 3 presents the SAs analyzed at six synthetic seismograms, which are aligned with the along-strike center of the fault plane across the model domain (y direction in Fig. 1b). The spectra with 5% critical damping were computed following the approach of Nigam and Jennings (1969), at three stations within the foreland basin (shaded gray, STA03–STA05) and three on the surrounding crystalline basement (STA01, STA02, and STA06), to assess the level of shaking expected at different periods relevant to buildings. Figure 3b presents results for all stations within each model domain, however, stations STA01 and STA02 fall outside the model domain for some scenarios and are therefore not plotted.

The SA patterns are similar across the basin and basement stations, with peak periods controlled by the earthquake magnitude. The basin's fundamental frequency varies laterally as the basin depth changes, but is >0.2 Hz (<5 s) for basin depths of ≤ 3 km (using a simple 1D approximation of $f_0 = V_s/4H$). The effects of the basin on the frequency content of the ground shaking can be seen in Figure 3 by comparing the solid and dashed lines, which represent models with and without a basin, respectively. The basin amplification is most pronounced for the M_w 5.5 earthquake scenarios because the dominant surface-wave wavelength (~ 4 km) is comparable to the basin depth (3 km), meaning the surface waves are largely confined within the low-velocity sediments, resulting in minimal dispersion and sustained amplification (O'Kane and Copley, 2021). The larger M_w 6.5 models (models 4–6) demonstrate one clear peak in the SAs at a period of 1.5 s (~ 0.6 Hz), which is consistent with the corner frequency set by the M_w 6.5 source. Similarly, for the M_w 7.5 models, there is significant energy at periods of ~ 3 s, reflecting the high-amplitude, lower-frequency surface waves which dominate in the larger magnitude earthquake simulations. Similar SA patterns have been observed in past earthquakes, including the 1994 M_w 6.7 Northridge blind-thrust rupture, which produced spectral peaks at 0.6–2.0 s attributed to source characteristics (Malhotra, 1999), and are comparable to equivalent M_w 6.5

synthetic wavefield models in this study. At the most distal station (STA06 in Fig. 3), the no-basin models (dashed lines) produce higher SAs than their equivalent basin models (solid lines) for the M_w 6.5 and 7.5 scenarios. This effect occurs because the dominant wavelengths for these magnitudes (10 and 35 km, respectively) are large relative to the basin depth (3 km), meaning that surface waves undergo dispersion during propagation through the basin, reducing the far-field amplitudes (O'Kane and Copley, 2021). Therefore, the control models without a foreland basin maintain higher SAs at far-field distances.

Figure 4 illustrates the maximum and mean PGA calculated from each of the 18 simulated earthquake ruptures. To investigate the effect of earthquake magnitude and location on the PGA, mean values were calculated across various sections of the foreland basin (i.e., the proximal ≤ 50 km and distal ≥ 150 km, as shown in Fig. 4). Figure 4a–d shows the direct trade-off between magnitude, location, and the effect of the foreland basin on the resultant earthquake ground motions. Figure 4e–h presents the same metrics after applying V_{S30} -based amplification factors (as described earlier) to test the sensitivity of the ground motions to shallow site effects. The maximum PGA of $\sim 2.94g$ (as shown in Fig. 4a) and mean PGA of $\sim 0.64g$ (as shown in Fig. 4b) was generated by model 7, as expected, since it has the largest magnitude (M_w 7.5) and the fault rupture is located at the basin margin.

As expected, the PGA values are lower for the models with smaller magnitude and more distant earthquakes, and these plots allow the relative effects of earthquake magnitude and distance to be quantified. For example, in the part of the basin proximal to the range front (i.e., Fig. 4c), for every 50 km increase in the source distance between the earthquake and the edge of the basin, the PGA is reduced by an amount equivalent to a 1–2 source magnitude-unit decrease. The effect is most pronounced for the maximum PGA (Fig. 4a), which lacks the spatial averaging inherent in the mean values. In the distal parts of the basin (Fig. 4d), the same principles apply, but with lower absolute PGA values and reduced sensitivity to earthquake location due to the attenuation and dispersion of waves as they propagate through the basin. Importantly, in the proximal parts of the basin there can be a difference of a factor of ~ 1.5 to 2.0 in the PGA values between models with and without a basin present.

Figure 4e–h presents the same PGA metrics after applying V_{S30} -based amplification factors from Stewart and Seyhan (2013) to test the sensitivity of the ground motions to shallow site effects, not resolved in the seismic-wavefield simulations. The corrections amplify basin PGAs by 10%–60% depending on shaking amplitudes ($F_{PGA} = 1.1 - 1.6$ for site class D sediments), whereas no-basin PGAs are reduced by 10% ($F_{PGA} = 0.9$ for site class B crystalline rock). As a result, V_{S30} -corrected maximum PGAs (Fig. 4e) increase by up to 14% in the basin models (e.g., model 7 increased from 2.94g

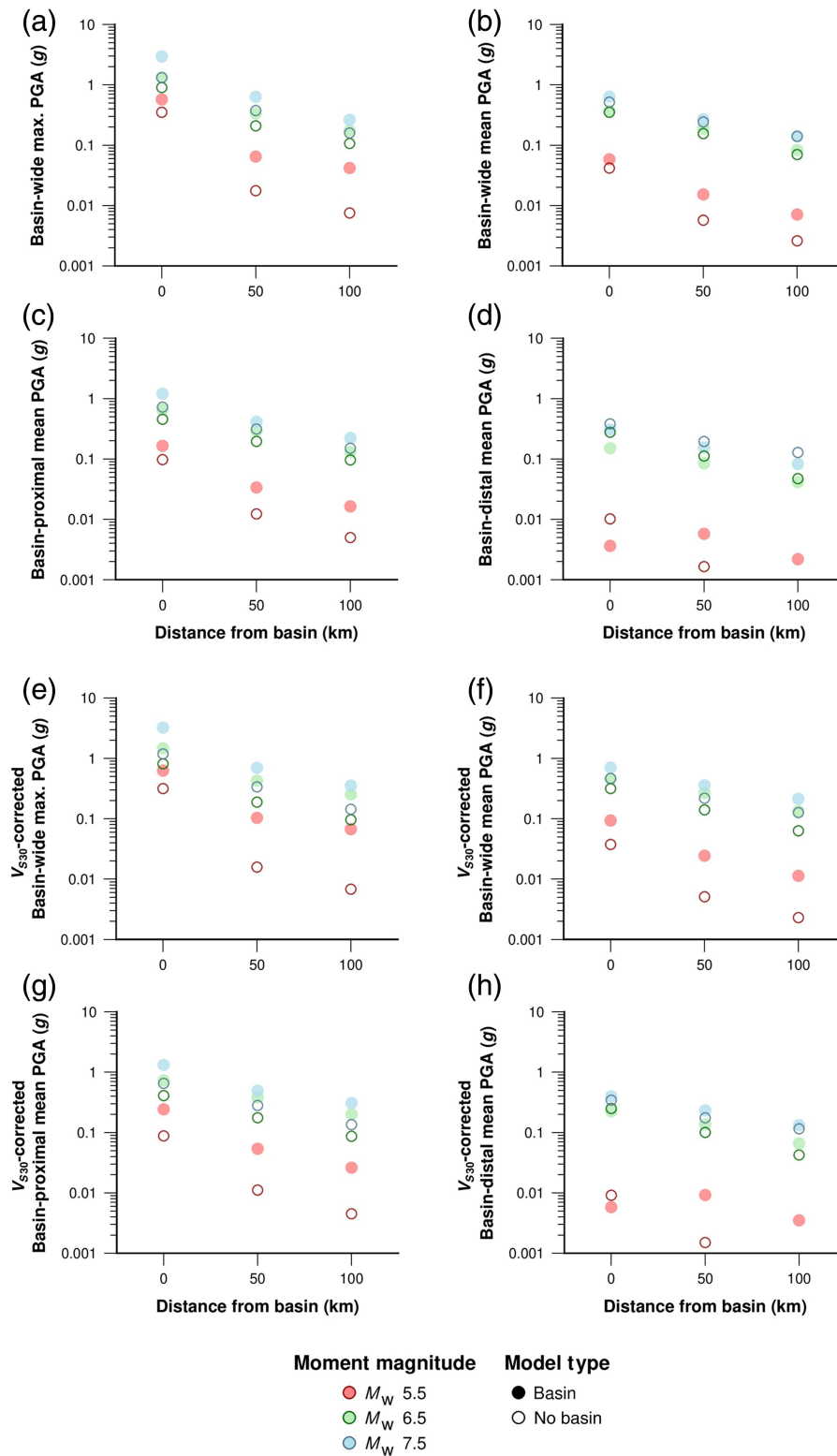


Figure 4. Maximum and mean PGAs calculated as a function of earthquake distance and magnitude for the 18 wavefield models conducted in this study. (a) Maximum PGAs recorded across the entire basin. (b) Mean PGAs recorded across the entire basin. (c,d) Mean PGAs recorded at proximal (0–50 km) and distal (150–200 km) distance intervals from the basin margin. (e–h) V_{S30} -corrected versions of (a–d), with amplitude-dependent site amplification factors from [Stewart and Seyhan \(2013\)](#) applied to basin models (site class D, $V_{S30} \sim 180\text{--}360$ m/s, $F_{PGA} = 1.1\text{--}1.6$) and bedrock models (site class B, $V_{S30} \sim 760\text{--}1500$ m/s, $F_{PGA} = 0.9$). Note that the M_w 5.5 no-basin scenarios in (d) and (h) fall below the minimum-plotted PGA value, as these earthquake scenarios generate negligible ground motions at distal distances. The color version of this figure is available only in the electronic edition.

to 3.23g; +10%), whereas mean PGAs increase by 28% (Fig. 4f). Amplification varies spatially, with smaller increases in proximal regions (<20% at distances of 0–50 km, Fig. 4g) and larger increases at greater source-to-basin distances (<45% at distances of 150–200 km, Fig. 4h), reflecting the amplitude-dependent nature of the V_{S30} corrections. Despite changes in absolute PGA values, the magnitude–distance trade-offs identified in Figure 4a–d are preserved (e.g., 50 km distance offset being equivalent to 1–2 magnitude units in regions proximal to the fault, Fig. 4g), as the earthquake magnitude and source-to-basin distance continue to be the primary controls on the ground-motion distributions. As this study focuses on the relative basin-scale controls rather than absolute ground-motion amplitudes, these V_{S30} adjustments serve only as a sensitivity test, and all subsequent analyses use the simulated wavefield values without these corrections applied.

Figure 5 illustrates the expected level of shaking at periods relevant to low-, mid-, and high-rise buildings. Average periods of 0.15, 0.50, and 1.50 s were selected to represent low-, mid-, and high-rise structures (given the resonant frequencies of buildings with height ranges defined by the HAZUS methodology; [Federal Emergency Management Agency \[FEMA\], 2024](#)), and for each of the 18 seismic-wavefield models, the SDs were plotted. The results in Figure 5a show that the absolute SD values are

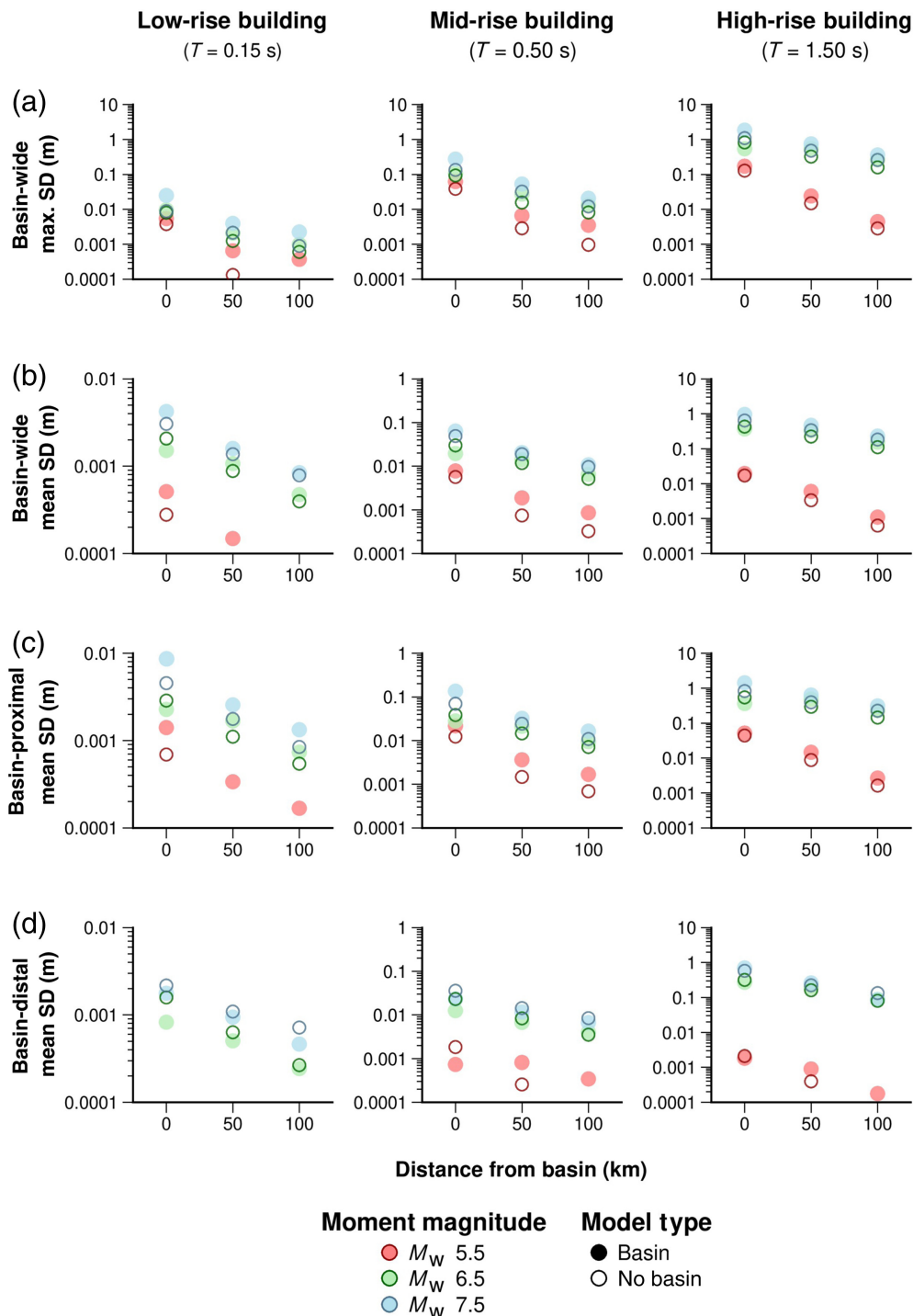


Figure 5. Maximum and mean peak spectral displacements (SDs) at periods relevant for low-rise, mid-rise, and high-rise structures. (a) Maximum SDs recorded across the entire basin. (b) Mean SDs recorded across the entire basin. (c, d) Mean SDs recorded at proximal (0–50 km) and distal (150–200 km) distance intervals from the basin margin. Note that some M_w 5.5 scatterpoints, particularly for low-rise building periods, are not shown as these scenarios generate SDs that are too small to produce appreciable structural response. The color version of this figure is available only in the electronic edition.

significantly higher (up to ~ 2 m of SD) for a period of 1.50 s, sensitive to high-rise structures, compared to the low- or mid-rise building periods, which generated $<15\%$ of the high-rise SDs.

moderate-magnitude earthquakes (M_w 5.5–6.5) at the basin margin can generate PGAs comparable to larger events (M_w 7.5) located 50–100 km away (Fig. 4). These relationships

This pattern of elevated SDs at lower frequencies is a well-known feature that arises due to the frequency content of the source and the influence of basin amplification. This phenomenon was observed in Mexico City, where long-period ground motions during the 1985 Michoacán earthquake were amplified due to double resonance between the infrastructure and shallow lake sediments, causing the most damage to mid- and high-rise buildings across the city (Campillo *et al.*, 1989).

Implications for seismic hazard

The seismic-wavefield simulations presented in Figures 2–5 reveal systematic controls on ground-motion distributions in foreland basin settings, with important implications for seismic hazard assessment. The results show that the ground-motion amplitudes are governed by the earthquake magnitude, rupture location, and basin structure (through mechanisms described in more detail by O’Kane and Copley, 2021). The quantified trade-off, whereby a 50 km increase in source-to-basin distance produces an equivalent reduction in PGA to a 1–2 moment magnitude unit decrease, reflects a compound effect of increasing the horizontal distance between the rupture and the basin, and increasing the source depth as the dipping range-bounding fault moves further from the basin margin, further reducing ground motions relative to basin-margin equivalents. Consequently,

indicate the importance of accounting for both seismicity distribution and basin geometry in seismic hazard assessments, as simplified distance-based attenuation relationships, commonly used in empirical GMMs, may underestimate hazard from basin-proximal sources or overestimate that from distant-source events. Although this study focuses on the earthquake magnitude and location, the fault dip can also exert a significant control on the ground motions, due to the effects of radiation patterns, incidence angles at the basin-basement interface, and the depth range over which different fault dip angles are commonly observed in continental mountain belts. These effects are discussed in detail in our previous article, [O'Kane and Copley \(2021\)](#). The low-angle fault modeled here ($\sim 13.5^\circ$) represents a higher-hazard end-member relative to steeper range-bounding faults observed in other collisional settings (e.g., $\sim 45^\circ$ in the northern Tien Shan; [Ghose et al., 1998](#); [Kober et al., 2013](#)).

Basin structure exerts a first-order control on the ground motions, which is most pronounced in basin-margin ruptures, which generate substantially higher ground motions than equivalent scenarios without a foreland basin (Figs. 2–5). Although this study models a single representative basin geometry, basin depth and width both influence the spatial distribution of ground motions ([O'Kane and Copley, 2021](#)). The impedance contrast between the basin sediments and underlying crystalline basement further controls the pattern of ground shaking, although for generic foreland basin velocity and density structures the influence of material properties on the resultant ground motions is secondary to that of source and basin geometry ([O'Kane and Copley, 2021](#)). The lack of sensitivity of our results to the V_{S30} -based amplification factors (Fig. 4e–h), demonstrates that the large-scale distribution of seismic hazard across foreland basin settings is primarily controlled by deeper structural features including the basin depth, geometry, and fault characteristics, rather than near-surface conditions (although these can be significant factors if they vary on local scales). This finding has important implications for regional seismic hazard assessment, indicating that the 3D heterogeneous velocity structure and geometric effects (e.g., basin-edge focusing and body-wave resonance) should be considered when undertaking hazard assessments. Therefore, physics-based simulations should be considered as part of seismic hazard assessments in these settings.

The frequency-dependent nature of the ground-motion amplification (Fig. 5) has critical implications for building vulnerability. The substantially larger SDs at longer periods, combined with the slower decay with distance of long-period ground motions, mean that tall buildings face disproportionate hazard, not only from proximal earthquakes but also from distant, large-magnitude events (consistent with observations and physics-based simulations of basin amplification elsewhere; [Cipta et al., 2018](#); [Wirth et al., 2018](#); [Kakoty et al., 2021](#); [Ornthammarath et al., 2025](#)). These building-type-specific

responses necessitate structure-dependent hazard assessments in addition to the PGA-based evaluations here.

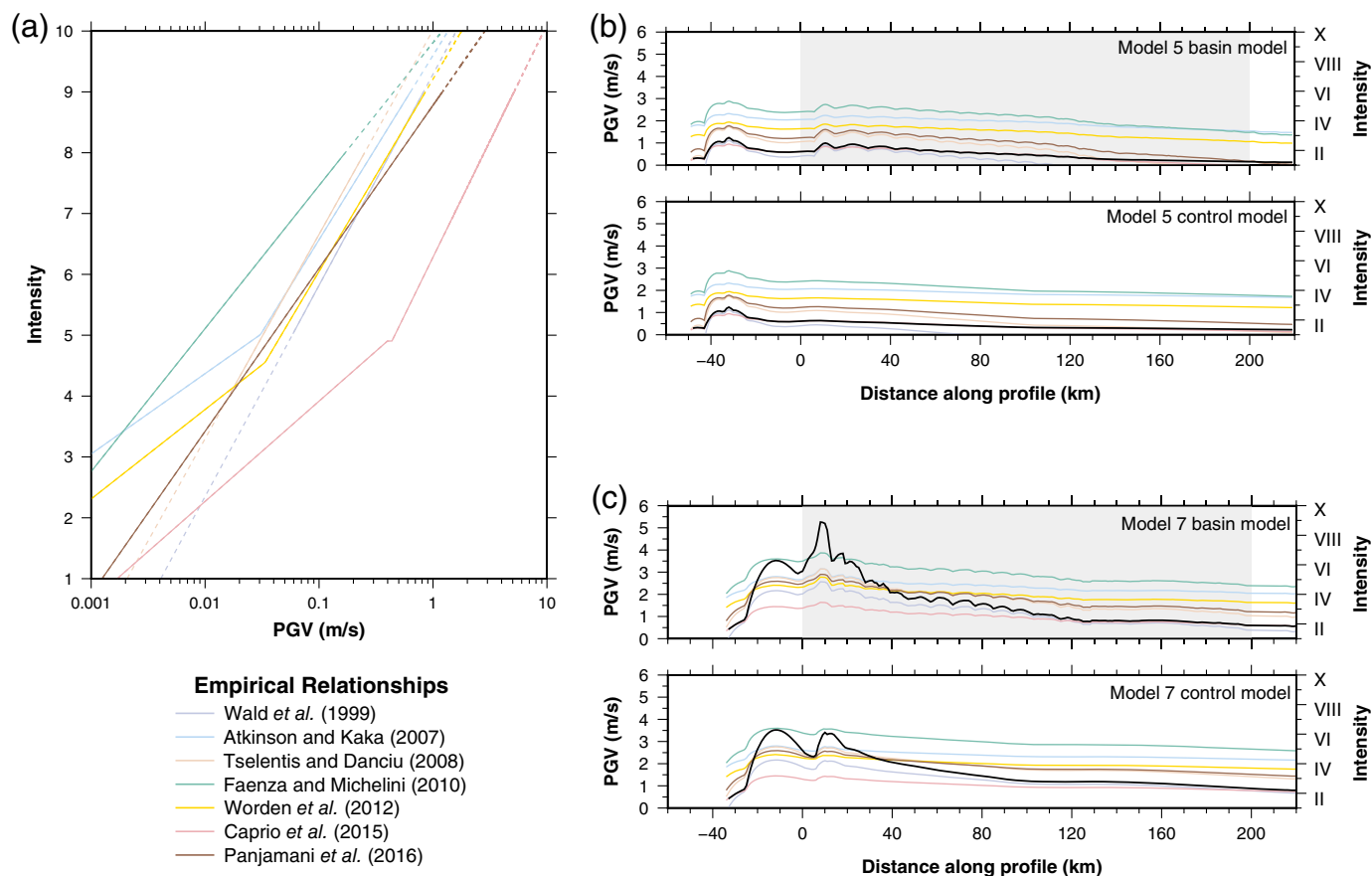
IMPLICATIONS FOR THE INTERPRETATION OF HISTORICAL EARTHQUAKE RECORDS

The previous sections examined the controls on ground motions generated by earthquakes in continental range-front settings. To use these findings to aid interpretations of the historical record of earthquake damage requires converting these ground motions into a form comparable to historical observations, which are recorded as Mercalli-style damage intensities rather than instrumental ground-motion values. Therefore, we convert peak ground motions to MMI, and then use established intensity–magnitude relationships to assess how the source and basin characteristics (e.g., earthquake magnitude, location, and basin sediments) can affect inferred earthquake magnitudes.

Initial conclusions can already be drawn from the ground-motion patterns presented earlier. As expected, M_w 5.5 earthquakes are likely unrecorded in the historical record, except for very shallow events close to population centers, consistent with global evidence that earthquakes of this magnitude range frequently cause no reported casualties or damage ([Nievas et al., 2020](#)). For the tectonic setting modeled here, where the events occur on the down-dip extension of the base of the foreland basin, the ruptures are deep enough that M_w 5.5 events generate negligible ground shaking. Perhaps more surprising is that events up to M_w 7.5 can produce relatively minor ground motions if they occur beneath the mountain belt (e.g., 2015 Gorkha and 1905 Kangra earthquakes), rather than if they rupture the shallower, up-dip portion of the locked fault immediately adjacent to the foreland basin. These two events are well documented because of observations from within the Himalayan foothills. However, similar events where such observations are not available due to the date of occurrence or geographical setting would not be recorded in the historical record if the only observations come from cities within the foreland basins. There is therefore the potential for a large component of the past earthquake moment release in such settings to be unrecognized.

PGV versus intensity relationships

Macroseismic intensity values can be derived from earthquake ground motions through a variety of different methods. In the 1950s, [Gutenberg and Richter \(1956\)](#) and [Hershberger \(1956\)](#) attempted to retrieve MMIs from PGAs. However, more recently, PGVs have proven to be a better ground-motion metric for intensity, due to their alignment with observed intensity measures ([Wald et al., 1999](#)), damage statistics ([Wu et al., 2003](#)), and ShakeMap applications ([Kaka and Atkinson, 2004](#); [Worden et al., 2010](#)). Therefore, for this analysis, the PGV outputs from the GMMs are converted to the MMI scale ([Wood and Neumann, 1931](#)). Unlike the simulated ground motions



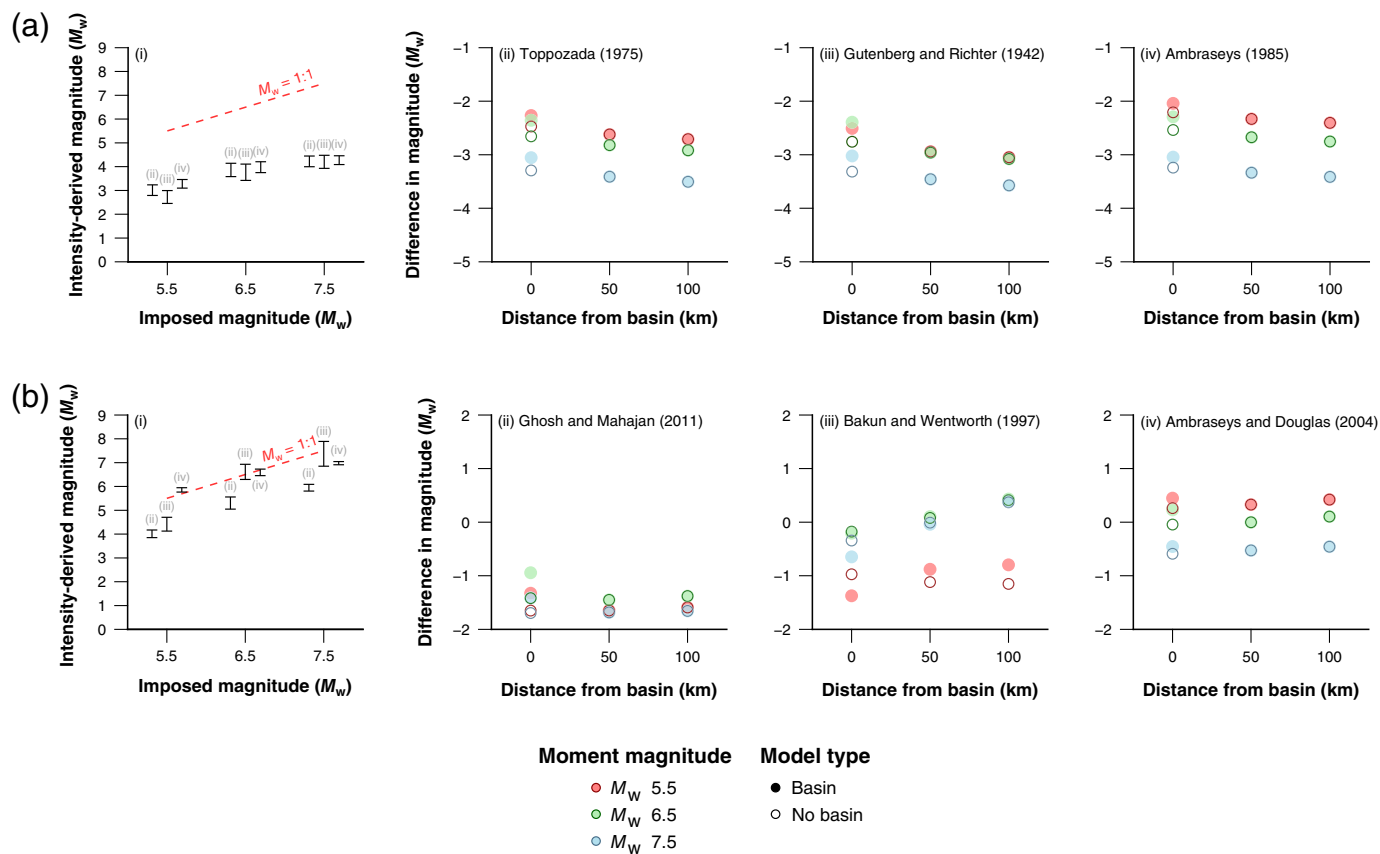
discussed earlier, MMI is ranked on observed effects. It is one of the most widely adopted intensity scales, and correlates well with peak ground-motion metrics (Musson *et al.*, 2010; Worden *et al.*, 2010).

Established empirical PGV–intensity relationships were used to calculate the maximum intensities produced by every earthquake scenario at each station along the model domain (in the y direction in Fig. 1b). Several commonly used PGV–intensity laws (Wald *et al.*, 1999; Atkinson and Kaka, 2007; Tselentis and Danciu, 2008; Faenza and Michelini, 2010; Worden *et al.*, 2012; Caprio *et al.*, 2015; Panjamani *et al.*, 2016) were selected (Fig. 6a), and MMI values corresponding to the maximum PGV recorded at each station were calculated for each of the 18 model scenarios. The results for a common earthquake scenario (model 5), analogous to the 2015 M_w 6.3–6.5 Pishan earthquake, which occurred in the western Kunlun Mountains near the Tibetan Plateau–Tarim basin range front (Lu *et al.*, 2016; Ainscoe *et al.*, 2017) are illustrated in Figure 6b. Figure 6c, presents a more extreme and infrequent earthquake scenario (model 7), analogous to the destructive 1905 Kangra earthquake (Ambraseys and Bilham, 2000).

For the moderate-magnitude earthquake scenario (M_w 6.5 rupture 50 km from the basin margin; Fig. 6b), intensities range from I in the far field to V in the near field. Shaking is rarely felt at intensity I, which occurs only in the distant regions (beyond 180 km from the fault) for MMIs derived by Wald *et al.* (1999), Tselentis and Danciu (2008), Caprio

Figure 6. Peak ground velocity (PGV)–intensity empirical relationships used to derive modified Mercalli intensities (MMIs) from the seismic-wavefield model ground motions. (a) Seven empirical laws that are used in this study as a function of PGV and MMI. The solid lines indicate the calibrated magnitude range of each relationship, whereas dashed lines represent extrapolated values. (b,c) PGV (black line) and MMI (colored lines) as a function of distance along profile for (b) a common earthquake scenario (model 5), and (c) a more extreme earthquake scenario (model 7). Gray shading represents the location of the foreland basin within the model domain. The color version of this figure is available only in the electronic edition.

et al. (2015), and Panjamani *et al.* (2016). In near- and mid-field regions, however, where Atkinson and Kaka (2007) and Faenza and Michelini (2010) derive MMIs exceeding IV, light shaking is felt by nearly everyone with potential damage to objects such as windows, doors, and crockery. In comparison, the more extreme model scenario (M_w 7.5 rupture at the basin margin) generates substantially higher intensities of up to VII (Fig. 6c), corresponding to very strong shaking felt by all, with slight damage to well-built structures and considerable damage to poorly designed or constructed buildings. The basin models show higher intensities than the no-basin equivalents across both scenarios, especially in the near- to mid-field regions (0–120 km, Fig. 6b–c), where small-wavelength undulations in the intensity profiles reflect amplification from body-wave resonance within the basin sediments. These modeled



intensity distributions agree with observations from historical thrust-faulting earthquakes (Ambraseys, 1985; Wald *et al.*, 1999; Ambraseys and Bilham, 2000; Tselentis and Danciu, 2008; Ghosh and Mahajan, 2011), as expected, given that the empirical PGV–intensity relationships are based on such events and the modeled ground motions fall within the observed ranges, as described previously.

Intensity versus magnitude relationships

To test the sensitivity of historical earthquake magnitude estimates to geological structure and source characteristics, synthetic earthquake magnitudes were back-calculated from the modeled intensities using established empirical relationships. This process mirrors how historical earthquake magnitudes are derived from intensity observations, allowing for the evaluation of systematic biases introduced by foreland basin settings and buried thrust faulting.

The area over which an earthquake was felt, and the degree of damage experienced, can be used to estimate historical earthquake magnitudes through three main approaches: (1) point-value methods using only epicentral intensity (Gutenberg and Richter, 1942; Topopozada, 1975; Ambraseys, 1985), (2) a linear approach using observed intensity at a given epicentral distance (Bakun and Wentworth, 1997; Ambraseys and Douglas, 2004; Ghosh and Mahajan, 2011), or (3) an area-based method using isoseismal area for specific intensity levels (Musson, 1996; Bakun and Wentworth, 1997). Each approach

Figure 7. Comparison of imposed earthquake magnitudes with intensity-derived magnitudes, calculated using published intensity–magnitude relationships for each of the 18 seismic-wavefield models. (a) Approaches of Topopozada (1975), Gutenberg and Richter (1942), and Ambraseys (1985), who determine the earthquake magnitude based on a point-source value of the epicentral intensity alone. (b) Linear regression analyses of Ghosh and Mahajan (2011), Bakun and Wentworth (1997), and Ambraseys and Douglas (2004), which is based on both epicentral intensity and epicentral distance. The color version of this figure is available only in the electronic edition.

relies on empirical relationships calibrated from different earthquake catalogs with varying tectonic settings and faulting styles. Point-value methods are considered less reliable (Ambraseys, 1985), with Topopozada (1975) reporting underestimations averaging 1.3 magnitude units, whereas distance-scaled and area-based methods provide more robust estimates with uncertainties typically quoted to be within 0.5 magnitude units (Topopozada, 1975; Musson, 1996; Bakun and Wentworth, 1997). This study tests the first two approaches using modeled intensities for each of the seismic wavefield scenarios (Fig. 7). The area-based method was not applied because accurate isoseismal mapping would require a significantly larger along-strike model domain, making simulations computationally prohibitive.

Figure 7 presents magnitude estimates for the 18 seismic wavefield models using epicentral intensity (Fig. 7a) and

distance-scaled intensity (Fig. 7b) methods. The point-source relationships (Gutenberg and Richter, 1942; Toppozada, 1975; Ambraseys, 1985) consistently underestimate the magnitudes by 2–4 units (Fig. 7a), with differences of up to 0.5 magnitude units between the basin and no-basin control models. This substantial underestimation reflects the buried fault geometry of foreland thrust systems, which produces lower maximum epicentral intensities than the predominantly surface-rupturing earthquakes used to calibrate these relationships, such as the California events used by Toppozada (1975). Consequently, applying point-source methods to pre-instrumental foreland thrust earthquakes would likely result in substantial unrecognized moment release due to characteristically low epicentral intensities despite potentially large magnitudes.

Distance-scaled intensity measurements (Fig. 7b) derived from the empirical approaches of Bakun and Wentworth (1997), Ambraseys and Douglas (2004), and Ghosh and Mahajan (2011) perform significantly better than epicentral intensity alone, reducing magnitude estimation errors to within 1 magnitude unit for most scenarios. However, the magnitude estimates still vary by up to ~ 0.5 units between models with and without a foreland basin. Given the logarithmic relationship between seismic moment and moment magnitude, this 0.5 unit uncertainty could translate to factors of ~ 3 in seismic moment estimates. In other words, whether a foreland basin is present in a region, and whether or not the empirical magnitude-estimation method that is used is built upon data with or without such a basin present, can introduce significant uncertainties into historical moment-release estimates. These basin-related biases decrease in importance with increasing distance from the range front, becoming negligible for earthquakes 50 km or more from the basin margin, while remaining pronounced for proximal range-front sources where basin amplification is strongest.

It is important to recognize that the biases quantified here represent only part of the overall uncertainty in intensity-based magnitude estimation. Source parameters, such as stress drop (which is kept constant across all magnitude scenarios in this study through the Aki (1967) fault-scaling relationship with seismic moment), have been shown to cause discrepancies between intensity-based magnitudes and instrumental moment magnitudes on the order of one M_w unit, with differences of up to 1.3–2.2 M_w units for individual earthquakes (Hough, 2014; Lucas *et al.*, 2023). That finding reflects the fact that macroseismic intensities are controlled by ground motions at frequencies relevant to the built environment (~ 2 to 8 Hz), which depend strongly on earthquake stress drop (Hanks and Johnston, 1992). The biases quantified in this study are therefore independent of, and additional to, these source-related uncertainties. Unlike the structural path and site biases explored here, which can potentially be reduced by accounting for fault location and basin geometry, the uncertainties arising from stress-drop variability are largely

irreducible from intensity data alone (Hough, 2014; Lucas *et al.*, 2023), and additional evidence such as geological or geotectonic constraints on the rupture dimensions is needed.

CONCLUSIONS

Thrust-faulting earthquakes have caused widespread destruction to cities along the margins of continental mountain ranges throughout history. However, interpreting historical earthquake magnitudes from damage distributions requires understanding the relative contributions of earthquake magnitude, rupture location, and the underlying geological structure, which remains poorly quantified in foreland basin settings. This study quantifies these trade-offs through seismic wave-field modeling and empirical PGV–intensity–magnitude relationships. The results reveal that increasing the source-to-basin distance by 50 km produces effects equivalent to reducing the earthquake magnitude by 1–2 units, with basin structure amplifying ground motions by factors of 1.5–2.0. Consequently, moderate-magnitude earthquakes at the basin margin can generate ground motions comparable to those generated by larger events located 50–100 km away. This study also identifies moderate-magnitude or distal large-magnitude earthquake scenarios that produce negligible ground shaking within the foreland basin, which have likely occurred historically but gone unrecorded, and could represent unobserved historical moment release along continental range fronts. Application of empirical PGV–intensity–magnitude relationships reveals that source and propagation effects can cause up to one moment magnitude unit over- or underestimation of historical earthquake magnitudes, with a ~ 0.5 unit difference between models with and without a foreland basin. These findings have direct implications for seismic hazard assessments in tectonically active foreland regions, including the Himalayas, Tien Shan, and Caucasus.

DATA AND RESOURCES

The SW4 open-source seismic-wave-propagation modeling code (Petersson and Sjögreen, 2017a) was developed at Lawrence Livermore National Laboratory (LLNL) and is available at <https://github.com/geodynamics/sw4>. All figures were produced using Generic Mapping Tools (GMT) version 6 (Wessel *et al.*, 2019) and Inkscape v1.2.1 (<https://inkscape.org/>). All websites were last accessed in November 2025.

DECLARATION OF COMPETING INTERESTS

The authors acknowledge that there are no conflicts of interest recorded.

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