The controls on earthquake ground motion in foreland-basin settings: The effects of basin and source geometry

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Summary

Rapid urban growth has led to large population densities in foreland basin regions, and therefore a rapid increase in the number of people exposed to hazard from earthquakes in the adjacent mountain ranges. It is well known that earthquake-induced ground shaking is amplified in sedimentary basins. However, questions remain regarding the main controls on this effect. It is, therefore, crucial to identify the main controls on earthquake shaking in foreland basins as a step towards mitigating the earthquake risk posed to these regions. We model seismic-wave propagation from range-front thrust-faulting earthquakes in a foreland-basin setting. The basin geometry (depth and width) and source characteristics (fault dip and source-to-basin distance) were varied, and the resultant ground motion was calculated. We find that the source depth determines the amount of near-source ground shaking and the basin structure controls the propagation of this energy into the foreland basin. Of particular importance is the relative length scales of the basin depth and dominant seismic wavelength (controlled by the source characteristics), as this controls the amount of dispersion of surface-wave energy, and so the amplitude and duration of ground motion. The maximum ground motions occur when the basin depth matches the dominant wavelength set by the source. Basins that are shallow compared with the dominant wavelength result in low-amplitude and long-duration dispersed waveforms. However, the basin structure has a smaller effect on the ground shaking than the source depth and geometry, highlighting the need for understanding the depth distribution and dip angles of earthquakes when assessing earthquake hazard in foreland-basin settings.

Key Words:

- Earthquake ground motions, Wave propagation, Earthquake hazards, Continental tectonics: compressional, Seismicity and tectonics

1 Introduction

A foreland basin is typically a wedge-shaped, sedimentary basin that forms adjacent to a mountain front in a fold-thrust belt, in response to lithospheric flexure during orogenesis [e.g. DeCelles and Giles, 1996]. This work will primarily focus on foreland basins in continental collisional settings, as earthquakes within the continental interiors have a long record of producing catastrophic damage and loss of life [e.g. England and Jackson, 2011].

Foreland basins pose several hazards to people who reside in cities built upon them. Due to rapid urban growth, $\sim$ 56% of the world's population now lives in urban areas, with cities and even megacities (e.g. Delhi; Population >28 million) existing within some foreland-basin settings [Cox, 2019] (see Fig. 1). These cities are built on thick layers of sediment [Burbank et al., 1996; Campbell et al., 2013; Gavillot et al., 2016; Grützner et al., 2017], located on or near active faults [Tapponnier and Molnar, 1979; Thompson et al., 2002; England and Jackson, 2011; Abdrakhmatov et al., 2016] and have a history of large destructive earthquakes [Bilham, 2004; Lavé et al., 2005; Pathier et al., 2006; England and Jackson, 2011; Lay et al., 2017]. Fig. 1 illustrates the location of known faults, foreland and intermontane basins, and past seismicity within Central Asia. When compared with the inset map showing population density [CIESIN, 2016], it can be seen that areas with high population densities often overlay basins in seismically active regions, or occur along range fronts. It is well established that ground shaking due to earthquakes is amplified in sedimentary basins [Bard and Bouchon, 1985; Sanchez, 1987; Rial et al., 1992; Olsen et al., 2003; LeBrun et al., 2002; Aagaard et al., 2008; Lozano et al., 2009; Taborda and Bielak, 2013; Galetzka et al., 2015; Meza-Fajardo et al., 2016; Bowden and Tsai, 2017; Rupakhety et al., 2017]. However, questions remain as to the relative importance of the factors that control this effect, which is a mixture of source characteristics and the wave-propagation effects. In this paper we investigate the controls on earthquake shaking in foreland basins, to better understand the seismic hazard that these regions pose to their inhabitants.

There are several methods for modelling earthquake ground shaking, each with distinct benefits and limitations. Ground Motion Prediction Equations (GMPEs) (also commonly known as Ground Motion Models (GMMs)) are used to estimate the expected ground shaking for a given area based on earthquake magnitude and mechanism, source-to-site distance and local geological conditions [e.g. Douglas, 2019, and references therein]. GMPEs are empirical fits to specific observations and use regression analysis between recorded ground motion and an intensity measure like damage statistics [Wu et al., 2003] or Modified Mercalli Intensity [Wald et al., 1999]. This technique is particularly useful for real-time applications such as performing earthquake-loss assessments for emergency response and disaster management purposes in the immediate aftermath of an earthquake [Wu et al., 2003]. However, most GMPEs do not account for spatial variations in path and/or site effects, which are known to significantly affect ground motions [Lastrico et al., 1972; Drake, 1980; Bard and Bouchon, 1985; Sanchez, 1987; Kawase and Aki, 1989; Olsen and Schuster, 1995; Joyner, 2000; Day et al., 2008; Frankel et al., 2009; LeBrun et al., 2002; Taborda and Bielak, 2013; Bhattarai et al., 2015; Bowden and Tsai, 2017; Rajaure et al., 2017; Rodgers et al., 2018; Wirth et al., 2019]. Furthermore, only limited regions have the required density of observations to allow GMPEs to be derived, leading to location bias in the resulting equations [Abrahamson and Silva, 1997; Sadigh et al., 1997; Campbell and Bozorgnia, 2008; Boore and Atkinson, 2008; Chiou and Youngs, 2008; Idriss, 2008; Power et al., 2008; Campbell et al., 2009; Gülerce et al., 2013]. Therefore, there is a reason for considering other techniques in parallel, which allow us to investigate the effects of lateral heterogeneity in the crust on the duration and intensity of ground motion.

Modelling of seismic-wave propagation is an alternative method for investigating seismic ground motion, which allows for a variety of geological structures to be incorporated into the models. This method is becoming more frequently used as computational resources have improved and more accurate 3-D velocity models of the Earth's structure have been produced [Rodgers et al., 2018]. Many studies have applied these methods to specific locations, such as in Southern California [Olsen, 2000; Graves and Pitarka, 2004; Olsen et al., 2003; Aagaard et al., 2008; Day et al., 2008; Graves et al., 2008; Harmsen et al., 2008; Aagaard et al., 2010; Bielak et al., 2010; Hartzell et al., 2010; Taborda and Bielak, 2013; Graves and Pitarka, 2016; Rodgers et al., 2018, 2019; Rodgers, 2019], Utah [Olsen and Schuster, 1995], Cascadia [Frankel et al., 2009; Wirth et al., 2019], Grenoble [Chaljub et al., 2005, 2010] and western Japan [Asano et al., 2016]. In this paper, we take a slightly different approach and perform calculations using an idealised geological structure, which can then be applied to a range of locations.

This study aims to identify the main controls of earthquake shaking in foreland basins on a regional scale and over a broad range of frequencies. Rather than model a specific basin, we aim to examine ground motion in a generic foreland-basin geometry. We intend to establish the main source and structural controls on the ground motion and therefore deduce underlying principles that can then be applied to a range of specific locations.

2 Methodology

SW4 (*Seismic Waves, 4th Order*) is a finite difference code [Petersson and Sjögreen, 2017a] that we have used to simulate seismic-wave propagation through a foreland-basin setting from a thrust-faulting earthquake along the range front of a mountain belt. The code solves the elastic and viscoelastic wave equations and is fourth-order accurate in time and space [Petersson and Sjögreen, 2012; Sjögreen and Petersson, 2012; Petersson and Sjögreen, 2014, 2015, 2017b]. SW4's capabilities made it an appropriate tool for use in this study, especially its ability to use a damping layer on all model boundaries (except the surface) to reduce artificial reflections from far-field boundaries [Petersson and Sjögreen, 2012, 2014, 2015, 2017b]. SW4 has already been applied to investigate seismic ground motions and has successfully reproduced ground motions consistent with GMPEs in the areas where they are defined [Imperatori and Gallovič, 2017; Rodgers et al., 2018, 2019]. Therefore we chose to use SW4 to model seismic ground motion in a typical foreland-basin system, to investigate the main controls on the ground motions.

In the paper, we concentrate exclusively on the basin-scale controls on the ground motions. It is well established that the shallow (e.g. top 30 m) velocity structure can have a large effect on the amplification of ground shaking, and can vary dramatically over short horizontal distances (e.g. hundreds of metres) [Anderson et al., 1996; Catchings and Lee, 1996; Boore and Joyner, 1997; Bowden and Tsai, 2017; Rajaure et al., 2017]. These smaller-scale effects from the shallow velocity structure will, however, be superimposed on the larger source and basin-scale geometrical effects (which we focus on in this paper), which control the characteristics of the waves entering the near-surface. We emphasise that the effects of the shallow velocity structure will be superimposed on these larger-scale effects that we study here.

2.1 Model Setup

We use a simple geometrical model for a foreland basin, as illustrated in Fig. 2. The model encompasses crystalline basement underlying a foreland basin, with seismic waves being produced by a planar, thrust-faulting earthquake at the basin margin. Our model is designed to replicate a typical foreland-basin setting. We impose slip on a fault that underlies the mountain range (although the topography itself is not included in our models), and in which rupture does not propagate into the adjacent basin. The rupture terminates in an up-dip location analogous to being beneath the range front of the mountain belt, as seen in Fig. 1 and observational studies from this tectonic setting

[Beaumont, 1981; Baranowski et al., 1984; Allen et al., 1986; Nelson et al., 1987; Abers et al., 1988; Fan and Ni, 1989; Molnar and Lyon-Caen, 1989; Cotton et al., 1996; DeCelles and Giles, 1996; Ghose et al., 1997, 1998; Bilham, 2004; Sloan et al., 2011; Avouac et al., 2015; Galetzka et al., 2015; Gavillot et al., 2016; Ainscoe et al., 2017; Wesnousky et al., 2018]. The majority of earthquakes in this tectonic setting do not rupture to the surface (the exception being rare, large events, such as a subset of those on the Himalayan range front [Wesnousky et al., 2017]). We therefore use a geometry in which the slip remains buried at depth, but note that for a minority of earthquakes in this setting there is sometimes some surface slip. The computational domain was set to be wide enough along-strike (X direction in Fig. 2) to encompass a complete earthquake rupture and deep enough (Z direction in Fig. 2) to accurately capture all wavelengths of interest (described below). The length of the domain (Y direction in Fig. 2) was determined based on the width of the basin being modelled. The density (ρ) of the crystalline basement was set to 2,700 kg m^{-3} and the P- (V_P) and S-wave (V_S) velocities were 6,000 m s^{-1} and 3,500 m s^{-1} respectively, yielding a V_P/V_S ratio of 1.7, following the findings of Hetényi et al. [2006]; Srinagesh et al. [2011] and Mitra et al. [2011]. Foreland-basin geometries vary significantly depending on the topographic load from neighbouring mountain ranges and the elastic thickness of the foreland, which together control their depth and their width [e.g. Allen et al., 1986; DeCelles and Giles, 1996; Naylor and Sinclair, 2008]. Basin depth and width are varied throughout this study to investigate what control basin structure has on earthquake ground motion. These variables are discussed further in Section 2.2. To replicate the characteristic wedge-shape for the basin, Turcotte and Schubert's (2014) end-load model was adopted. The maximum basin depth was set to the value chosen for each model ('d' on Fig. 2). The shape of the basin-basement interface to the right of the deepest point in Fig. 2 (in the area marked 'w') was calculated using a flexural profile, with an elastic thickness selected to match the basin width chosen for the model. The basin depth and width are varied between successive models. The basin boundary at the left-hand edge of Fig. 2 was set to dip at 30° after Ford [2004] and Hetényi et al. [2006]. The material properties of the basin fill (ρ = 2250 kg m^{-3}, V_P = 4375 m s^{-1}, V_S = 2500 m s^{-1}, V_P/V_S ratio = 1.75) were selected based on a compilation of global foreland-basin studies [Knopoff, 1971; DeCelles and Giles, 1996; Day et al., 2003; Olsen et al., 2003; Hauksson and Shearer, 2006; Hetényi et al., 2006; Mitra et al., 2011; DeCelles, 2012; Rodgers et al., 2018; Chen and Wei, 2019; Rodgers et al., 2019], and we describe later the effects of changing the chosen values. One of the main assumptions in the model is that both the basement and basin mediums are homogeneous. This simplified approach removes the local effects of internal layering, compaction and porosity variations, which are beyond the scope of this study. The calculated ground motions are therefore wholly a result of the larger basin-scale structure. In our initial models, we use no anelastic attenuation, to highlight the effects of source characteristics and basin geometry. We will then describe the results of models that include attenuation.

An earthquake with a moment magnitude (M_w) of 6.5 was simulated on a planar fault with a 10 km diameter, within the basement material directly down-dip of the base of the foreland basin. We choose this magnitude because such earthquakes are relatively common (Fig. 1) and because the compact rupture allows us to limit the size of our computational domain so that a large number of numerical experiments can be performed. The principles revealed by our results allow us to generalise to other earthquake magnitudes, as discussed below. The rupture begins at the down-dip edge of the rupture patch, in the along-strike centre of the fault plane, and travels radially outwards across the fault plane with a rupture velocity of 2.5 km s^{-1}, analogous to observations of past continental thrust-faulting earthquakes [Cotton et al., 1996; Copley et al., 2011; Yi-Ying et al., 2012; Denolle and Shearer, 2016; Kumar et al., 2017; Hayes, 2017]. The imposed slip pattern is circular, with a slip distribution set using the expressions for a circular crack given by Bürgmann et al. [1994]. Using Aki's (1967) relationship between seismic moment, stress drop and fault dimensions for a circular fault, our modelled earthquake has a stress drop of 2.6 MPa, similar to that recorded for the M_w 7.8 25th April 2015 Gorkha mainshock ($\Delta\sigma \approx 3.2 - 3.4$ MPa) [Lay et al., 2017; Prakash et al., 2016]. The earthquake is formed of sub-sources that we place at 25 m intervals across the rupture patch. A Gaussian source time function was used for each of the sub-sources. Each sub-source has a set angular frequency of 20 Hz which corresponds to a fundamental frequency (f_0) of 3.18 Hz [Sjögreen and Petersson, 2012; Petersson and Sjögreen, 2017b]. For a Gaussian time function, the maximum frequency (f_{max}) is $2.5 \times f_0$. The minimum resolvable frequency (f_{min}) in our model is set by the size of the model domain. This domain is set to be large compared with the rupture dimension so that all frequencies that are produced with appreciable amplitudes by the source are resolvable. For the models below in which attenuation is included, the minimum resolvable frequency is $f_{max} \times 10^{-2}$, [Sjögreen and Petersson, 2012; Petersson and Sjögreen, 2017b]. Therefore, we model frequencies in the range of 0.08 – 7.95 Hz, covering the frequencies typical for building oscillation [Murty et al., 2012; Parajuli and Kiyono, 2015; Idham, 2018; Bońkowski et al., 2019], which is important when considering the seismic hazard of a region and building resilience to earthquake shaking. We discuss later the effects of using different source characteristics.

To enable the maximum frequencies produced by our source to be modelled, we require at least six grid points per minimum wavelength (W_{min}), which can be calculated by dividing the minimum shear-wave velocity in the model by the maximum frequency, to give a W_{min} of 314 m [Petersson and Sjögreen, 2012, 2015, 2017b]. Consequently, the grid size was set at 50 m so that all frequencies within the range of 0.08 – 7.95 Hz could be accurately resolved.

We extract waveforms for analysis from a line of synthetic stations positioned linearly across the modelled foreland basin, across-strike from the centre of the rupture patch and the hypocentre (Fig. 2). This geometry means that the resulting ground motions are entirely within the plane of the section, with no out-of-plane motions. Therefore, in the subsequent sections, where the horizontal component of the ground motion is discussed, we are referring to that component within the plane of the cross-section.

2.2 Model Parameter Ranges

Firstly, we conducted a series of simulations with a range of source depths but no foreland basin, to act as a control in order to evaluate the basin effects in subsequent models. We then varied both the basin and source geometries with the aim of identifying the main controls on the ground shaking. Table 1 outlines each of the parameters that we varied in these models.

The basin geometry was varied first, starting with the basin depth. The depth was modelled between 1 km and 5 km at the deepest part ('d' on Fig. 2), spanning the majority of observed foreland-basin depths (see Fig. 1 for a selection, and DeCelles and Giles [1996]). The across-strike basin width was then varied, which we define as the distance between the deepest part of the basin and the furthest edge (i.e. the width of the flexural profile, marked as 'w' on Fig. 2). The basin width is controlled by the elastic thickness of the foreland and in this study we consider basin widths in the range 50 – 200 km, as this spans the observed widths of foreland basins (including in northern India, where the foreland basin is ∼200 km wide).

The source geometry was subsequently varied. Because low-angle thrust earthquakes can occur across a range of distances from a foreland basin (as seen across the Himalayan arc), an earthquake rupture can either rupture to the range front [Lavé et al., 2005; Kumar et al., 2006; Malik et al., 2010; Kumahara and Jayangondaperumal, 2013; Sapkota et al., 2013; Bollinger et al., 2014; Gavillot et al., 2016; Wesnousky et al., 2017] or remain buried at depth [Molnar and Lyon-Caen, 1989; Avouac et al., 2015; Galetzka et al., 2015; Wesnousky et al., 2018]. To account for this variety, we varied the distance between the rupture plane and the basin between 0 km and 100 km. Likewise, the fault dip was varied between the angle determined from the down-dip extrapolation of a flexural profile for a given basin width (minimum of 13° in the models below) and a maximum of 45°. This range in fault dip is in line with studies carried out by Hendrix et al. [1992]; Allen et al. [1993, 1994]; Bilham et al. [2003]; Hetényi et al. [2006] and Middleton and Copley [2014] and is illustrated by the fault-plane dips on the focal mechanisms shown in Fig. 1.

Subsequent simulations were conducted to determine the effect of the material properties on the ground motion. Table 1 outlines these parameters and the ranges over which they were varied. The seismic-wave speeds within the basin were varied to account for different basin compositions and degree of compaction. The shear-wave velocities (V_S) were varied between 2500 m s^{-1} and 3500 m s^{-1}, based on the findings of Hetényi et al. [2006] and Mitra et al. [2011], whilst keeping the V_P/V_S ratio constant at 1.75. In the calculations described so far, all simulations were run under purely elastic conditions, however, most materials are not elastic and attenuation plays a role in the ground motion produced by earthquakes [Bowden and Tsai, 2017]. Therefore, we conducted a series of simulations with the addition of attenuation. A Quality factor (Q; the inverse of attenuation) is set to be equal for P-waves and S-waves, and is varied in the range 75 to 300, based on studies by Olsen [2000]; Singh et al. [2004]; Hauksson and Shearer [2006]; Shearer et al. [2006]; Srinagesh et al. [2011] and Sharma et al. [2014].

We analyse the Peak Ground Velocities (PGVs) across the computational domain. This ground-motion parameter was selected following studies from Wald et al. [1999] and Wu et al. [2003], which found that PGV has a closer correlation with intensity measures and damage statistics than Peak Ground Acceleration (PGA). Similarly, SW4 simulation results from Rodgers et al. [2018] showed near-zero bias when compared with GMPEs, indicating that the PGV is consistent with the Abrahamson et al. [2014] (ASK14) GMPE predictions in the places they are defined and that the simulations can reproduce the observed path and site effects. In addition to calculating the PGV, we performed spectral analysis by calculating fast Fourier transforms (FFTs) using the Welch [1967] method to determine the frequency dependence of the ground motions. The source magnitude, dimension, and frequency content are kept constant across all the simulations, and we describe later the effects of changing these source characteristics.

3 Results

Fig. 3 shows a vertical-component ground-motion time series from models with and without a foreland basin. The cross-sectional profiles illustrate the velocities calculated along a plane positioned in-line with the along-strike centre of the fault plane (see Fig. 2). The cross-sections in Fig. 3 illustrate the lateral propagation of low-amplitude body waves, followed by higher-amplitude, lower-frequency Rayleigh waves. The surface waves dominate the PGV in both scenarios. The dominant wavelength of the Rayleigh waves in the case with no basin is ∼6 km, which is set by the rupture dimension, the rupture velocity, and the wave-propagation velocity. Fig. 3(b) shows a much more complex wavefield than Fig. 3(a), which arises from two main effects. The existence of the low-velocity sedimentary basin causes dispersion of the surface waves. Additionally, the interaction of the surface waves with the interface between the basin and the basement (including the basin edge) causes the generation of body waves, resulting in a longer and more complex series of S-waves. There is also the transmission of some energy into the basin from the body waves propagating beneath the interface. We will investigate these effects below from the perspective of the controls on the ground motion within the basin.

Spectral analysis was carried out for each of the basin and no-basin reference models illustrated in Fig. 3. Spectra were calculated following Welch [1967], at regular intervals (10 km) across the model setting, to identify the frequency dependence of the ground motions. Each panel in Fig. 4(b) illustrates how the Power Spectral Density (PSD) amplitudes of the ground motions vary across the range of resolvable frequencies (0.08 – 7.95 Hz). It is apparent that the chosen spectra and dimensions of the source dominate the signal, specifically the dominant frequency of the surface waves that are produced. Although there are minor peaks that correspond to body-wave resonance in the basin (at frequencies above 1 Hz, as described below), the frequency of the peak ground motions is controlled by the source spectra. We discuss below the effects of changing the source characteristics and dominant frequency.

3.1 Effects of basin depth

Fig. 5 illustrates the effect of basin depth on wave propagation, by comparing a basin that is shallow (maximum depth of 1 km) relative to the dominant wavelength of the surface waves (4 km in the basin, which is lower than in the basement due to the difference in the propagation velocity) with a basin that has a depth which is similar to the dominant wavelength (maximum depth of 5 km). For the shallow basin, the surface waves are strongly dispersed, leading to a long wave train and waveforms that clearly show the earlier parts of the surface waves have lower frequencies than the later parts. For the deeper basin, there is minimal dispersion because the surface waves are dominantly contained within the basin, rather than also sampling the faster underlying basement. However, complex waveforms are visible in the near-field and low-amplitude, high-frequency arrivals are visible before the surface waves in the distant part of the basin (Fig. 5). These features are due to the basin being deep enough for S-waves to resonate within the low-velocity sediments, although they are of lower amplitude than the surface waves.

Fig. 6 shows the results of the same simulations as in Fig. 5 (and additionally for a basin depth of 3 km), plotted as peak ground velocity as a function of distance. The PGVs for the no-basin reference models are also included for comparison. For both cases, we plot the vertical and horizontal components, of which, the vertical ground motions are larger for all models of varying basin depths, due to the geometry of the source. Fig. 6(c) illustrates that the PGVs are greatest for shallow basins, which is a result of the source being positioned down-dip of the flexural-base of the basin (dashed lines in Fig. 6a). This geometry means that shallow basins in our models are associated with shallow, thrust-faulting earthquakes, which result in larger PGVs at the surface. Ulloa and Lozos [2020] also discussed the effect that source depth has on ground shaking, with shallower events resulting in higher ground motions. This source-depth effect dominates the signal. By normalising with respect to the peak PGV for each model (Fig. 6b) we can isolate the effects of the basin geometry, which can be seen across two different length scales: short ($\sim$10 – 20 km) and long ($\sim$100 km). Through the centre of the basin (between distances of 40 – 140 km) the surface waves cause pronounced differences in the normalised PGV values, producing variations in the ground motions in a pattern that spans the entire width of the basin. Body-wave resonance within the basin causes short length-scale undulations in the normalised PGVs, which is superimposed upon the surface-wave effects, with a wavelength of kilometres to tens of kilometres.

The differences between model results that occur on the length scale of the entire basin result from the relation between the dominant wavelength of the surface waves and the depth of the basin. For the case of the shallow basin, dispersion of the surface waves results in a rapid decrease in PGV over a short distance, as the energy is spread over a longer-duration but lower-amplitude wave train. There are also differences in the PGV values between models where the basins are of a similar or larger depth than the dominant surface wavelength within the basin (which is $\sim$4 km). This effect arises because where the surface wavelength is similar to the basin depth and the waves interact with the velocity contrast at the base, surface-wave amplification occurs [Bard and Bouchon, 1980; Joyner, 2000; Olsen, 2000; Day et al., 2008; Denolle et al., 2014; Bowden and Tsai, 2017]. Where the basin is too deep for the base to interact with the surface wave, this effect does not occur. The PGV in both of these cases decays rapidly at the distal part of the basin, where it becomes shallow and dispersion occurs. This effect is why the 1 km-deep basin model in Fig. 6(c) has much lower PGVs in the far-field (especially the vertical component), than the corresponding no-basin reference model (in which such dispersion doesn't occur). The effects of the lateral separation between the source and the edge of the basin will be discussed below.

The basin depth also has a small effect on the position of the peak PGV. In Fig. 6, the basin and source depths increase together, as the fault is a down-dip extension of the base of the foreland basin. In Fig. 6(b) the location of the maximum PGV moves basinward for deeper basins. This effect is not seen in the numerical experiments in Fig. 7(b) where the basin depth remains the same but the source depth varies. This finding implies that this effect arises due to the changes in basin depth and not source depth. However, this movement in the peak PGV has a minor control on the ground motions when compared with the other factors considered in this paper, such as the relationship between the basin depth and dominant wavelength of the source.

3.2 Effects of basin width

Fig. 7 illustrates the effect of basin width on PGV. The vertical and horizontal components of the ground motion are plotted, and similarly to Fig. 6, the vertical ground motions are largest. The PGVs for the no-basin reference models are also included for comparison. Fig. 7 demonstrates the short- and long-wavelength characteristics of body-wave resonance and surface-wave propagation respectively, as described above. The PGVs for wide basins (200 km) decrease gradually over large distances. In comparison, narrow basins (50 km) see a rapid decrease in PGV over short distances (Fig. 7c) as the shallow, distal part of the basin is encountered and a dispersive wave train is produced, causing the duration of shaking to increase and the amplitude to decrease. This effect is a result of when the basin becomes shallow enough that significant surface-wave dispersion begins to occur. Therefore, it is the basin width that defines when appreciable dispersion begins.

Unlike basin depth, Fig. 7(c) illustrates only small differences (<7%) in the PGV at the edge of the basin closest to the source. This effect is due to the underlying effect that source depths vary slightly with changes in the dip of the basin floor (and it's down-dip continuation which hosts the earthquake), which are in turn caused by varying the basin width (Fig. 7a).

3.3 Effects of the source-to-basin distance

Fig. 8 illustrates the effect of the source-to-basin distance on wave propagation. This numerical experiment is based on the observation that earthquake ruptures sometimes reach the range front, but in other cases remain buried at depth [Bilham, 2004; Lavé et al., 2005; Kumar et al., 2006; Malik et al., 2010; Kumahara and Jayangondaperumal, 2013; Sapkota et al., 2013; Bollinger et al., 2014; Avouac et al., 2015; Gavillot et al., 2016; Wesnousky et al., 2017, 2018]. In this plot, Fig. 8(b) shows the vertical PGV whilst Fig. 8(c) illustrates the horizontal PGV, both of which have the no-basin reference models included for comparison. The vertical ground motions are larger than the horizontal ground motions, consistent with the results in Sections 3.1 and 3.2. Similarly, within the basins, the no-basin reference models have lower PGVs compared to the basin models, irrespective of their source-to-basin distance. Beyond the basins, the PGV values for models including basins are lower than those for models without, due to the surface-wave dispersion that occurs as the waves propagate through the shallow parts of the basins. These numerical simulations also show the short- and long-wavelength characteristics of body-wave resonance and surface-wave propagation, as discussed above.

The near-source PGVs differ depending on the source depth, and the values in the proximal part of the basin depend on the position of the source relative to the basin. The ruptures that are more distant from the basin result in lower increases in PGV as the waves enter the basin (Figs 8b and 8c). The increase happens because of the change in material properties between the basement and the basin. This effect is lower (in percentage terms, compared to the near-source PGV) for sources distant from the basin because as the source moves away from the basin, a smaller proportion of the total energy is directed into the basin (due to it occupying a smaller proportion of the cross-sectional area into which waves are radiated by the source). For sources positioned 25 – 50 km from the basin, the amplification in the basin is such that the PGV there roughly equals that in the near-source region, producing a double-peak in the ground motion pattern (Figs 8b and 8c).

3.4 Effects of fault dip

Fig. 9 illustrates the effect of the fault dip on PGV. The vertical and horizontal components of the ground motion are plotted, as well as the no-basin reference models for comparison. This numerical experiment was inspired by the observation that some mountain range fronts are characterised by low-angle thrusting along planes down-dip of the foreland basin (e.g. the Himalayas), whilst others are characterised by higher-angle faulting on planes dipping at $\sim 45°$ beneath the mountains (e.g. the northern Tien Shan), as seen in Fig. 1. Shallow-dipping faults, like the one that is in line with the base of the basin (13.5°) on Fig. 9(a), produced higher PGVs than the steeper-dipping faults (25° & 45°). This is a result of two controls: source depth and the angle of the incident wave.

For a given depth to the top of the earthquake rupture, steepening the dip moves the centroid (the slip-weighted average depth of slip) to deeper depths, resulting in lower PGV values (as described above). A second important effect revealed by Fig. 9 relates to the resulting wave propagation. The amount of energy that is either reflected or refracted off the basin-basement interface is controlled by the angle of the incident wave. Shallow fault dips result in waves propagating into the basin from shallow angles, and therefore being trapped by internal reflection within the basin (using Snell's law, we determined that the basin has a critical angle of 46°). For rupture on steeply-dipping faults, a greater proportion of the energy is incident on the surface and basin floor at higher angles. Therefore, more of this energy is transmitted into the interior of the Earth, rather than trapped in the basin. Also plotted on Fig. 8 is the PGV in the horizontal component. As in the models described above, this component is lower-amplitude than the vertical. The two components become more equal as the fault dip increases, because more equal amplitudes of vertical and horizontal motion in the resulting seismic waves are produced by steeper-dipping faults. However, both components decrease as the fault dip increases (due to the change in source depth), which is a larger control on the ground motion.

3.5 Effect of material properties

The material properties of the basin fill also affect the amount of ground shaking that is produced. Fig. 10 illustrates the effect of basin seismic velocity on wave propagation. These results are expressed as a function of the S-wave velocity, but the P-wave velocity has also been varied to maintain a V_P/V_S ratio of 1.75. The vertical and horizontal components of the ground motion are plotted on Fig. 10, as well as the no-basin reference models for comparison. As seen previously, the vertical PGVs are higher than the horizontal values. This effect results from the source geometry. Both components decrease as the basin S-wave velocity increases. All the resultant ground motions from the basin models exceed the no-basin reference PGVs in the locations of the basins, but are lower in the far-field due to the surface-wave dispersion that occurs as the waves propagate through the shallow parts of the basins.

There are three main effects shown on Fig. 10. Firstly, the absolute value of the PGV varies, because the velocity controls the degree of amplification caused when the waves enter the basin. Secondly, the wavelengths of the body-wave resonance and the broad signal caused by surface-wave amplification and dispersion are changed slightly. This is due to the variable basin velocities and different reflection coefficients at the basin-basement interface, resulting in different dominant wavelengths and resonant frequencies in the basin interior. Thirdly, the far-field PGV is slightly higher for higher basin velocities, because the lower velocity contrast with the basement means that less energy is lost due to body-wave excitation by the surface waves along the interface between the basin and basement. However, the differences between these models are small compared with the effects of the source and basin geometry described above.

Fig. 11 illustrates the effect of attenuation on the distribution of PGV in the basin. The quality factor is set to be equal for P-waves and S-waves and is varied in the range 75 – 300, based on the results of Olsen [2000]; Singh et al. [2004]; Hauksson and Shearer [2006]; Shearer et al. [2006]; Srinagesh et al. [2011] and Sharma et al. [2014]. In this plot, Fig. 11(b) shows the vertical PGV whilst Fig. 11(c) illustrates the horizontal PGV, both of which have the no-basin reference models included for comparison. In agreement with the previous sections, the vertical ground motions are larger than the horizontal ground motions, and the no-basin reference models have lower PGVs than the basin models. Similar to previous numerical simulations, Fig. 11 shows the short- and long-wavelength characteristics of body-wave resonance and surface-wave propagation. As expected, as attenuation increases the PGV decreases, and this effect is most pronounced in the distal part of the basin, where the waves have propagated furthest. A quality factor of $\sim$100 is likely to be relevant to the bulk of the basin fill (i.e. not the near-surface sediments), with quality factors $\geq$400-500 for the deeper crustal material [Schlotterbeck and Abers, 2001; Hauksson and Shearer, 2006; Shearer et al., 2006]. In comparison to the model results described above, attenuation has a similar-sized effect on the magnitude of PGV within the foreland as the basin geometry, but a smaller effect on the magnitude of PGV than the source geometry (i.e. depth and dip).

4 Discussion

The modelling results described above show that the source characteristics have a larger effect on PGV than the basin geometry. Of particular importance are the source depth, location relative to the basin margin, and fault dip, all of which can vary significantly between mountain ranges and sometimes along-strike within a given range (e.g. Fig. 1; Maggi et al. [2000]; Sibson and Xie [1998]; Bilham et al. [2003]; Bilham [2004]; Jackson et al. [2008]; Middleton and Copley [2014]; Bai et al. [2019]). However, these quantities can often be difficult to estimate in advance of earthquakes (e.g. it was not widely expected that some large earthquakes on the Himalayan megathrust in Nepal would fail to rupture to the surface, as was the case with the 2015 Gorkha earthquake [Avouac et al., 2015; Galetzka et al., 2015]). The results presented above suggest that these source attributes have a more important impact on basin-scale ground shaking than the basin geometry itself.

The basin geometry does, however, also play a role in controlling ground shaking. The relative length scale of the basin depth and the dominant wavelength of the surface waves controls whether appreciable surface-wave dispersion occurs, resulting in longer-duration but lower-amplitude ground motions. Basin depth and width both contribute to controlling the locations where appreciable dispersion occurs, for a given earthquake. This concept allows us to extend our analysis to a wider range of source magnitudes and spatial sizes than that considered above. Increasing the magnitude of the source also involves increasing the spatial size of the rupture, due to the observed relationship between magnitude and fault dimension [Scholz, 1982; Scholz et al., 1986; Cowie and Scholz, 1992; Scholz, 1997]. Increasing the earthquake magnitude will increase the resulting PGV (because of the amount of energy release), in addition to increasing the dominant wavelength of the surface waves (due to the increasing fault size), and therefore change the range of basin geometries over which surface-wave dispersion becomes important. These effects are conceptually displayed in Fig. 12. The red curve represents the case for a given magnitude, such as the M_w 6.5 considered here. Deep basins and the associated deep earthquakes produce low-amplitude ground shaking. Shallow basins and the associated shallow earthquakes result in high-amplitude ground motions that are rapidly dispersed during propagation through the basin. There is a middle ground in which the basin and source are shallow enough that high-amplitude surface waves are produced, but that the basin is deep enough to produce little dispersion across most of the basin width. If the magnitude of the earthquake is increased, the basin needs to be deeper to prevent dispersion, but the PGV is increased for all basin depths due to the magnitude increase. The effect is, therefore, to move the curve up and to the right on the graph shown in Fig. 12. A corollary of this effect is that lateral differences in basin depth (e.g. as shown across the Indo-Gangetic and Tarim basins in Fig. 1; Chatterjee [1971]; Lee [1985]; Graham et al. [1990]; Nishidai and Berry [1990]; Cobbold et al. [1993]; Royden [1993]; Huafu et al. [1994]; Yang and Liu [2002]; Bilham et al. [2003]; Hetényi et al. [2006]; Mitra et al. [2011]; Srinagesh et al. [2011]; Li et al. [2013]; Wei et al. [2013]; Morin et al. [2019]) can have an important consequence in terms of the magnitudes of earthquakes for which PGV values will be similar across large areas of the basins, or decay rapidly with distance.

In addition to the spatial size of the fault rupture, other effects can also play a role in controlling the dominant wavelength of the resulting surface waves. For example, the rupture velocity of the earthquake (which controls the importance of directionality effects) and the intrinsic frequency content of the source (e.g. relating to the length-scale of individual asperities within the rupture patch) can control the wavelength of the resulting waves. The stress drop of the earthquake, which controls the spatial size of the rupture for a given moment release, can have a similar effect. Likewise, the seismic velocity of the material in which the source is embedded. It is beyond the scope of this manuscript to consider each of these effects. However, our results presented above provide a means to infer their role in the resulting ground motions using our finding that, following source depth, the next most dominant control on the ground motions is the relative length scales of the basin depth and dominant surface-wave wavelength. Therefore, all effects that involve increasing the dominant wavelength of the surface waves (reducing the intrinsic frequency of the source, reducing the stress drop, reducing the rupture velocity and increasing the ambient seismic velocity) will have an effect that is equivalent to the subset of the consequences of increasing the seismic moment that are based on the effects on the dominant wavelength, as described above. Such changes will therefore result in surface-wave dispersion effects being more important for a given basin depth, or less important if these parameters are changed in the opposite direction. Based on the geological setting and mode of formation of foreland basins, we have concentrated on thrust-faulting earthquake ruptures. However, we note that the effects of the relative sizes of the dominant wavelength of the waves and the basin depth will also be true for other types of events (i.e. strike-slip earthquakes within the mountains bounding a basin), but the specific ground motions (e.g. the relative importance of vertical and horizontal motions) will depend on the details of the source geometry.

Having identified the main controls on earthquake shaking in foreland basins from range-front thrust earthquakes, we considered the controls on the amount of ground shaking produced by normal-faulting events, often observed in the flexing, underlying crystalline basement in foreland-basin systems [DeCelles and Giles, 1996]. Assessing the seismic hazard resulting from such normal faults is difficult as they are often too deep to observe any expression of the extension at the surface, but it is worthwhile comparing their likely effects with those of the range-front thrust-faulting events.

We conducted a series of simulations using the same geometrical model setup for a foreland basin as illustrated in Fig. 2, but changed the fault mechanism and location in order to simulate a normal-faulting earthquake in the basement, underlying the basin. The normal fault was positioned with the up-dip termination of the fault at the base of the foreland basin at a depth of 2 km, with a dip of 45°. The remaining source parameters (magnitude, dimension, rupture velocity and frequency content) and material properties (seismic velocities and densities) remained unchanged from our original setup outlined in Section 2.1, to allow for comparisons to be made between the thrust- and normal-faulting earthquake ground motions. Therefore, as the rupture dimension, the rupture velocity and the wave propagation velocities remained the same for both earthquake scenarios, the dominant wavelengths in the basement material ($\sim$6 km) and foreland basin ($\sim$4 km) also remain the same.

Fig. 13(a) illustrates the cross-sectional setup and a snapshot of the resultant wavefield produced by the normal-faulting earthquake rupture. Fig. 13(b) shows the results of the simulation, plotted as PGV as a function of distance. Both vertical and horizontal components of the ground motion are illustrated, with the vertical PGV being higher in both the thrust- and normal-faulting models. Fig. 13(a) demonstrates the lateral propagation of low-amplitude body waves, followed by higher-amplitude lower-frequency Rayleigh waves which dominate the PGV, as was the case with the range-front thrust events modelled above. After the initial up-dip rupture through crystalline basement material producing high PGVs, the surface waves disperse causing a rapid decrease in PGV over a short distance ($\sim$10 km) from the fault (Fig. 13b), as a result of the shallow ($\sim$1.7 – 2.1 km) basin depth. The PGVs for the waves that propagate towards the range front at distances of $\sim$20 – 45 km are higher than the foreland-propagating waves at distances of $\sim$70 – 105 km (Figs 13a and 13b). The laterally-varying basin depth therefore plays a role in counteracting the hanging-wall effect, which tends to increase the ground motions in the hanging wall relative to the footwall. As the waves propagate towards the range front, the basin increases to a maximum depth of 3 km and therefore gets closer to the dominant wavelength of the surface waves. The waves interact with the velocity contrast at the basin-basement interface, causing surface-wave amplification and higher PGVs [Bard and Bouchon, 1980; Joyner, 2000; Olsen, 2000; Day et al., 2008; Denolle et al., 2014; Bowden and Tsai, 2017]. The higher PGVs in the mountainward direction compared to the basinward direction are also partially due to rupture-directivity effects. The waves propagating away from the range front, however, are strongly dispersed, and the basin becomes too shallow for the S-waves to resonate within the low-velocity sediments.

When comparing the normal- and thrust-faulting ground motions, there are two controlling variables: the source depth and the fault dip. As the thrust fault has both a shallower source depth and a more shallowly-dipping fault plane than the normal fault, it produces higher PGVs (Fig. 13b). In terms of length scales, the thrust-faulting earthquake resulted in longer-wavelength, basin-wide effects, whilst the normal-faulting earthquake yielded shorter-wavelength effects which were more localised to the fault region. This effect arises because of the dip effects discussed above, with more of the normal-faulting energy being reflected into the deep Earth rather than propagating through the basin, and due to the waves being generated in a region with a shallower basin depth.

Although we have changed a number of geometrical parameters between the normal-faulting and thrust-faulting earthquakes in this comparison, these changes are based upon observations from foreland-basin settings. Although the details of the comparison depend upon our chosen parameters, some overall concepts can be demonstrated. When comparing the two rupture scenarios (range-front, thrust-faulting vs distal, normal-faulting) in a foreland collisional setting, it is clear that range-front thrust faults yield larger-magnitude ground motions than buried normal faults. Wirth et al. [2019] also showed that shallow, thrust earthquakes produced higher amplification in the Seattle and Tacoma Basins, compared to deep, normal earthquakes which could suggest that the source depth remains the dominant control on ground motion, despite the tectonic setting. However, the results presented above demonstrate that, for a given magnitude, normal faulting in the underlying basin can result in higher PGV for localised regions of the basin than for an equivalent range-front thrust.

5 Conclusion

Large populations are present in cities built on or near foreland basins, and often information about their seismic risk is either unknown or limited. Although body-wave resonance has long been a well-understood phenomenon, surface waves and their path effects are less understood, often resulting in an underestimation of the seismic hazard in some regions. Seismic-wave-propagation modelling in this study has shown that the amount of initial ground motion produced largely depends on the source depth, whilst the basin structure (width and depth) determines how much of this energy gets dispersed. The maximum ground velocities are produced when the basin depth matches the dominant wavelength produced by the source. The basin width, however, determines how rapidly this ground motion decreases with distance, given that the width determines where the basin becomes shallow enough for dispersion to begin.

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7 Data Availability

The SW4 open-source code was developed at Lawrence Livermore National Laboratory and is distributed by the Computational Infrastructure for Geodynamics (`www.geodynamics.org/cig/software/sw4`). In Fig. 1 we used an array of freely-available datasets for topographic, earthquake, fault and population density data. ETOPO2 (2-minute gridded global relief) data was downloaded from the NOAA National Centers for Environmental Information web page (`https://www.ngdc.noaa.gov/mgg/global/etopo2.html`). Earthquake data was sourced from various individual studies referenced in the manuscript and has been compiled by Wimpenny [2020] into a Global Waveform Catalogue, available at `https://comet.nerc.ac.uk/gwfm_catalogue/`. Fault traces were plotted from Taylor and Yin [2009] and Styron et al. [2010]. Gridded population density data (GPWv4) was sourced from the Socioeconomic Data and Applications Center web page (`https://sedac.ciesin.columbia.edu/data/collection/gpw-v4`), whilst all other population-relation statistics were retrieved from Cox [2019]. All figures have been produced using Generic Mapping Tools v6 [Wessel et al., 2019] and Inkscape v0.92.3 [Inkscape Project, 2018].

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Tables

Table 1: Geometrical parameters and material properties that were varied during the seismic-wave-propagation modelling and the ranges over which they were varied.

Variable	**Notation**	**Range**
Geometrical Parameters		
Basin depth	d	0 – 5 km
Basin width	w	50 – 200 km
Source-to-basin distance*	–	0 – 100 km
Fault dip	–	13.5 – 45.0°
Material Properties		
Basin shear-wave velocity	V_S	2.5 – 3.5 km s^{-1}
Attenuation (Quality Factor)	Q	75 – 300

*This refers to the distance between the up-dip termination of the rupture patch and the maximum basin depth.

Figures

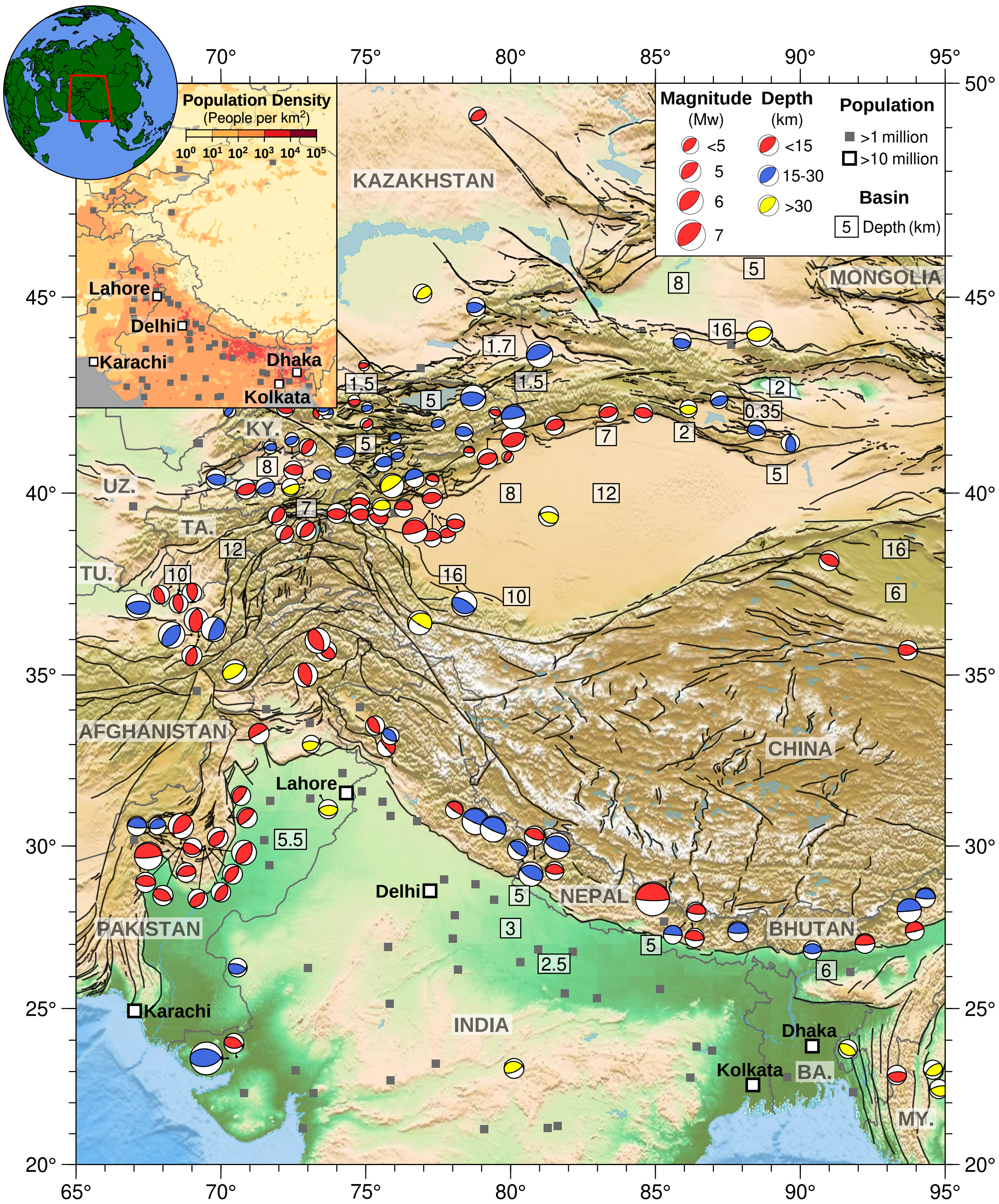

Figure 1: Map of Central Asia illustrating the interplay between topography, seismicity and population. Earthquake focal mechanisms are shown for thrust-faulting earthquakes; scaled in size by their moment magnitudes and coloured according to their centroid depth in kilometres [Molnar and Tapponnier, 1978; Kirsty and Simpson, 1980; Molnar and Chen, 1983; Baranowski et al., 1984; Eyidogan and Jackson, 1985; Nelson et al., 1987; Abers et al., 1988; Chen, 1988; Fan and Ni, 1989; Molnar and Lyon-Caen, 1989; Chen and Molnar, 1990; Holt et al., 1991; Burtman and Molnar, 1993; Fan et al., 1994; Cotton et al., 1996; Ghose et al., 1997, 1998; Berberian et al., 2000; Bernard et al., 2000; Jackson et al., 2002; Chen and Yang, 2004; Bayasgalan et al., 2005; Mitra et al., 2005; Sloan et al., 2011; Craig et al., 2012; Ainscoe et al., 2017]. Active faults, according to known fault databases,

Figure 1 ***(previous page)*:** are represented by black lines [Taylor and Yin, 2009; Styron et al., 2010]. The depths of foreland and intermontane basins are plotted in kilometres [Chatterjee, 1971; Lee, 1985; Khaimov, 1986; Carroll et al., 1990; Graham et al., 1990; Nishidai and Berry, 1990; Allen et al., 1993; Cobbold et al., 1993; Royden, 1993; Allen et al., 1994; Huafu et al., 1994; Hendrix et al., 1992; Coutand et al., 2002; DeBatist et al., 2002; Yang and Liu, 2002; Bilham et al., 2003; Hetényi et al., 2006; Sobel et al., 2006; Zhou et al., 2006; Fang et al., 2007; Xuezhong et al., 2008; Yin et al., 2008; Bian et al., 2010; Goode et al., 2011; Mitra et al., 2011; Srinagesh et al., 2011; Kober et al., 2013; Li et al., 2013; Wei et al., 2013; Zhao et al., 2013; Macaulay et al., 2016; Bande et al., 2017; Bosboom et al., 2017; Brunet et al., 2017; Pei et al., 2017; Voigt et al., 2017; Yu et al., 2017; Kufner et al., 2018; Chapman et al., 2019; Morin et al., 2019]. Major cities have been plotted according to their population size [Cox, 2019]. Solid, grey lines mark out the national borders with countries labelled in capitals. Six were abbreviated as follows: BA = Bangladesh, KY = Kyrgyzstan, MY = Myanmar, TA = Tajikistan, TU = Turkmenistan and UZ = Uzbekistan. The inset map shows the population density of Central Asia [CIESIN, 2016]. An orthographic projection outlines the geographical extents of the main figure.

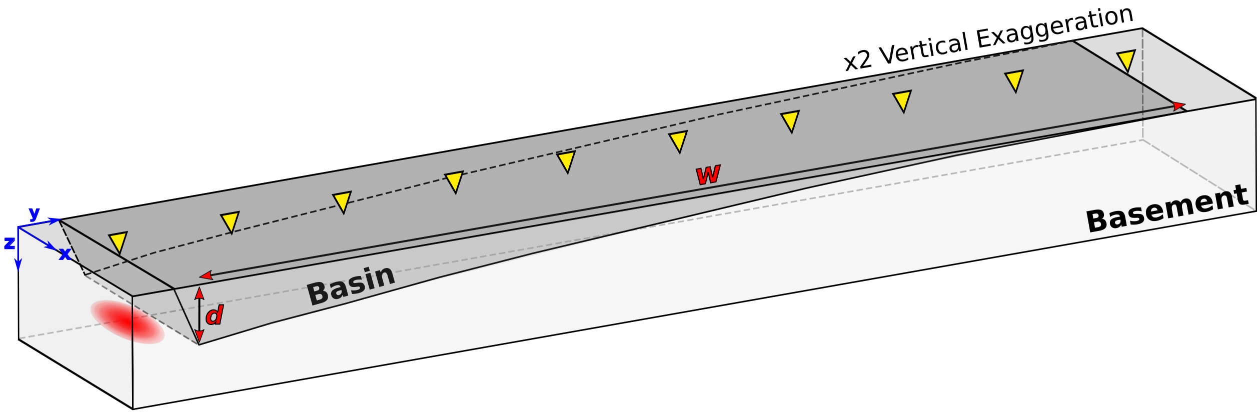

Figure 2: Schematic setup of the model used in this study. The model represents a simplified cross-section of a foreland-basin system orientated perpendicular to the range front. The model comprises two homogeneous mediums representing crystalline-basement rocks and foreland-basin sediments. The material properties for the basement rocks and foreland sediments are outlined in Section 2.1. The red shaded area, down-dip of the foreland basin represents a circular thrust rupture, with a diameter of 10 km and is planar in cross-section. Basin depth (d), basin width (w), fault dip and the distance between the fault source and the basin were varied to determine the effect that each variable had on the ground motion. The yellow triangles represent a selection of the modelled receiver stations that were aligned with the along-strike centre of the fault plane at kilometre intervals across the computational domain (Y direction).

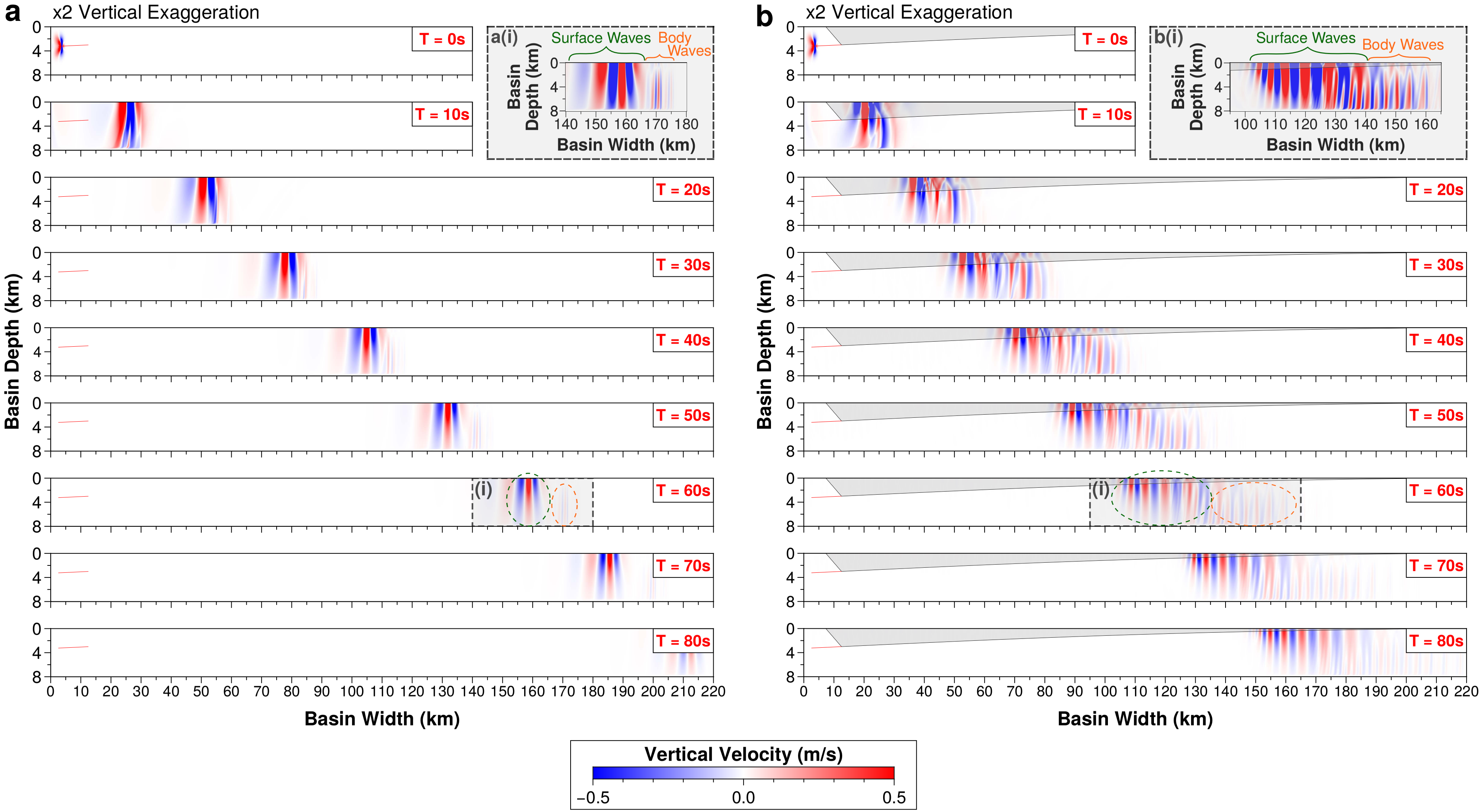

Figure 3: Time-series of wave propagation showing resultant ground motions from a M_w 6.5 thrust-faulting earthquake across two different computational domains. The cross-sections illustrate the vertical velocities calculated on a plane aligned with the along-strike centre of the fault plane (see Fig. 2). (a) illustrates the propagating wavefield through homogeneous crystalline basement. (b) illustrates the propagating wavefield through a foreland basin (shaded grey) and crystalline basement material. The material properties for the basement rocks and foreland-basin sediments are outlined in Section 2.1. The solid red line, orientated down-dip of the foreland basin, represents a circular thrust fault, with a diameter of 10 km and is planar in cross-section. The maximum vertical velocity is $\approx$4 m s^{-1}, however, the scale bar has been saturated to illustrate all wave effects. a(i) and b(i) have been plotted on a further saturated scale to provide a clearer view of the surface waves (indicated in green) and body waves (indicated in orange).

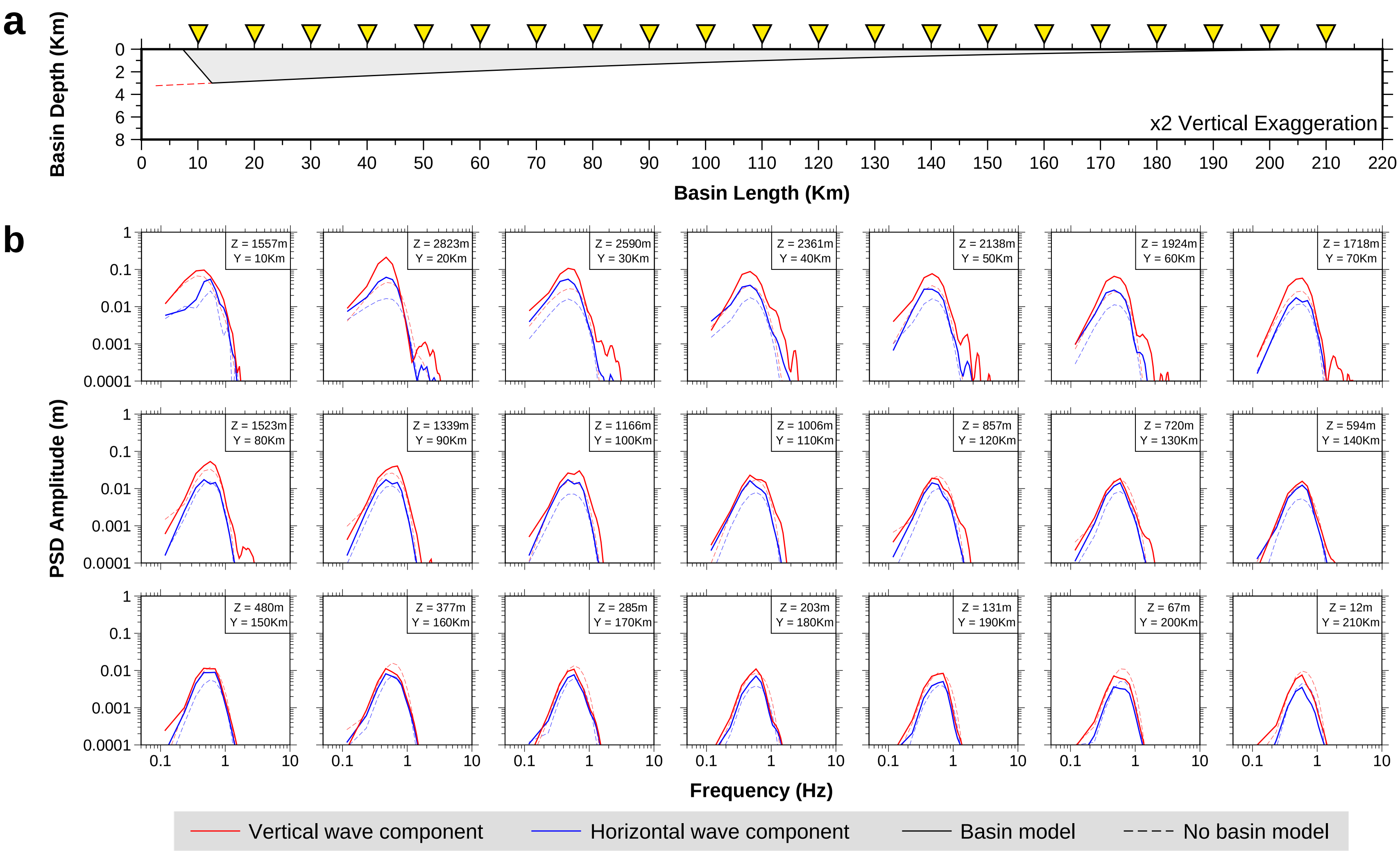

Figure 4: Frequency spectra of the ground motions. (a) illustrates the cross-sectional setup for this model, comprising a 10 km planar earthquake rupture (dashed red line) orientated down-dip of a 3 km deep, 200 km wide foreland basin (shaded grey). The yellow triangles represent a selection of the modelled receiver stations, which are aligned with the along-strike centre of the fault plane, across the computational domain (Y direction in Fig. 2). (b) shows calculated spectra at each of the receiver station locations. Each plot shows how the Power Spectral Density (PSD) amplitudes vary across the range of resolvable frequencies. The red and blue solid lines represent the vertical- and horizontal-wave motion for the basin model respectively. The dashed lines show the equivalent values for the reference model with no foreland basin.

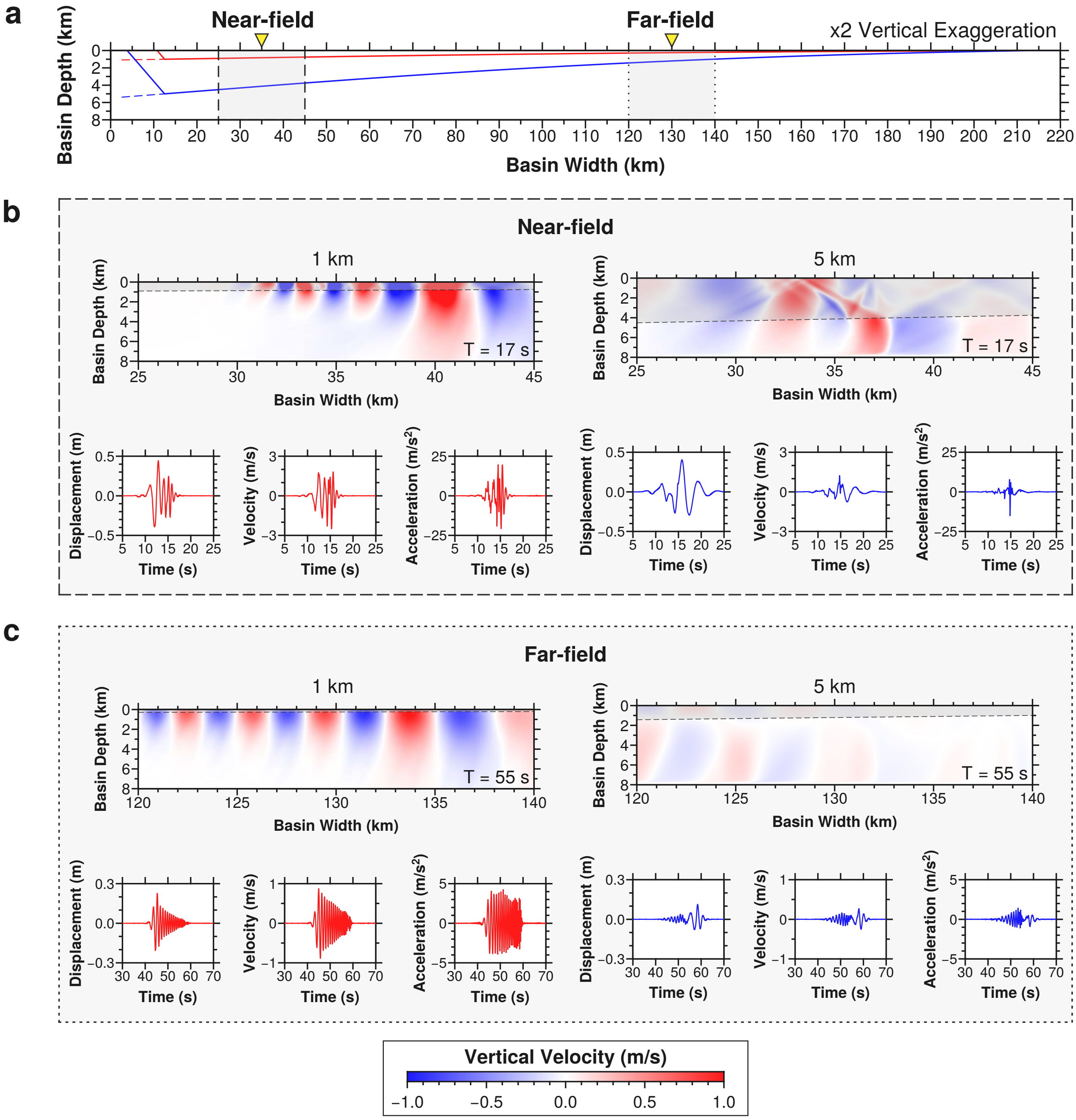

Figure 5: Calculated vertical velocities showing the near- and far-field effects of basin depth. (a) demonstrates the cross-sectional setup for two model simulations of varying basin depths; a 1 km deep basin denoted by a red line and a 5 km deep basin represented by a blue line. A 10 km planar fault is orientated down-dip of each foreland basin and illustrated as a dashed line. The shaded grey boxes outlined by dashes and dots represent the location of the near- (b) and far-field (c) results respectively, with waveforms shown at the yellow triangles, which represent receiver stations. (b) illustrates the resultant wavefield calculated for both the shallow (1 km) and deep (5 km) basins at a particular distance and time (35 km/17 s) in the near-field. (c) illustrates the resultant wavefield calculated for both the shallow (1 km) and deep (5 km) basins at a particular distance and time (130 km/55 s) in the far-field. The maximum vertical velocity for (b) and (c) is ≈3 m s^{-1}, however, the scale bar has been saturated to illustrate all wave effects.

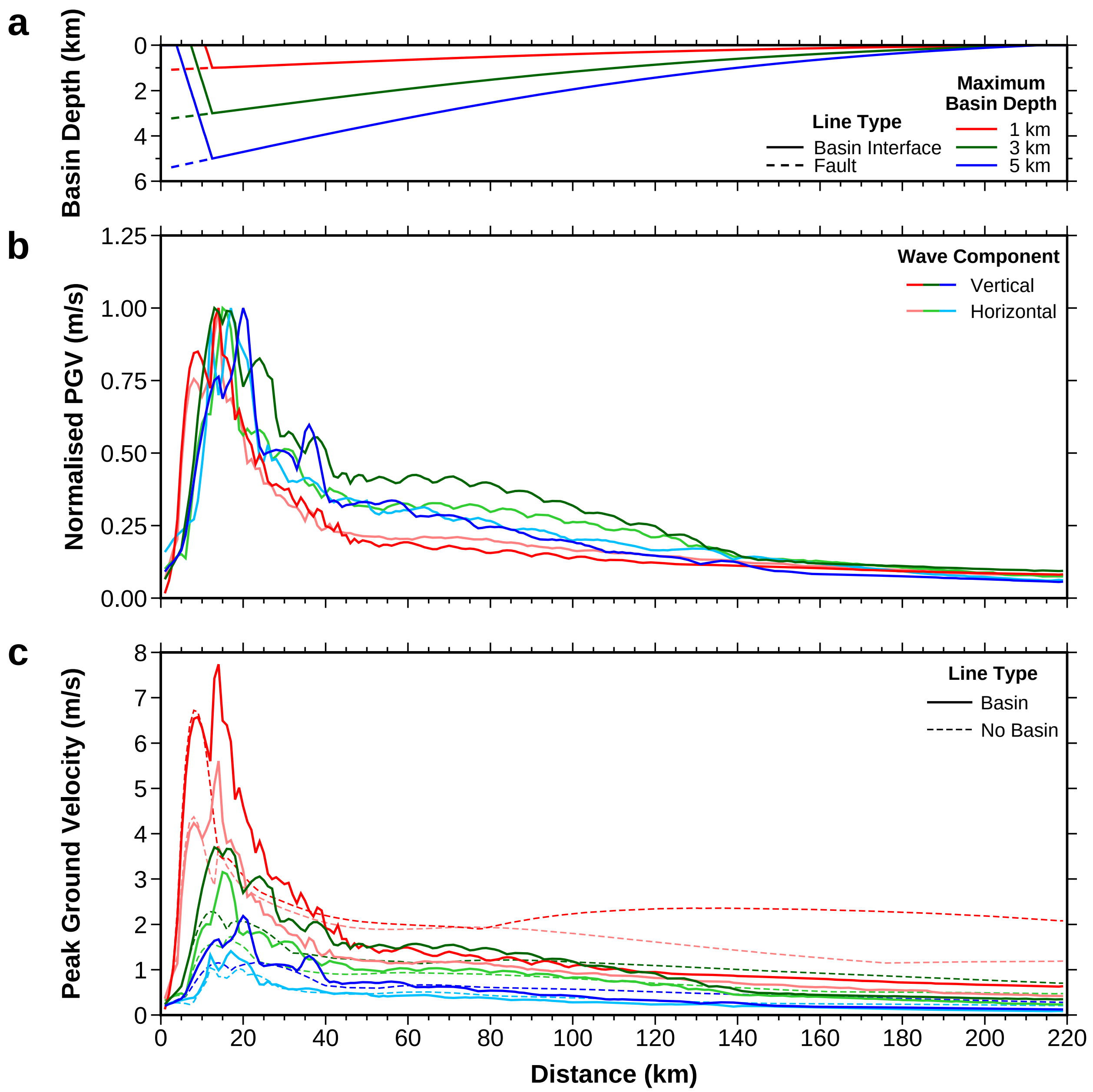

Figure 6: Peak ground velocity plotted as a function of distance across a foreland basin for three different basin depths. The basin-basement interface in (a) and resultant PGV in (b) and (c) for 1 km, 3 km and 5 km deep basins are denoted by red, green and blue solid lines respectively. The vertical and horizontal components for each basin in (b) and (c) are illustrated by dark- and light-coloured lines respectively. (a) demonstrates the cross-sectional setup for model simulations of varying basin depths. A 10 km planar earthquake rupture is orientated down-dip of each foreland basin and is illustrated by a dashed line. (b) shows PGV as a function of distance plotted for different basin depths, normalised by the maximum value of PGV. (c) shows PGV plotted as a function of distance for a range of source depths. The dashed lines show the equivalent values for the reference models with no foreland basin.

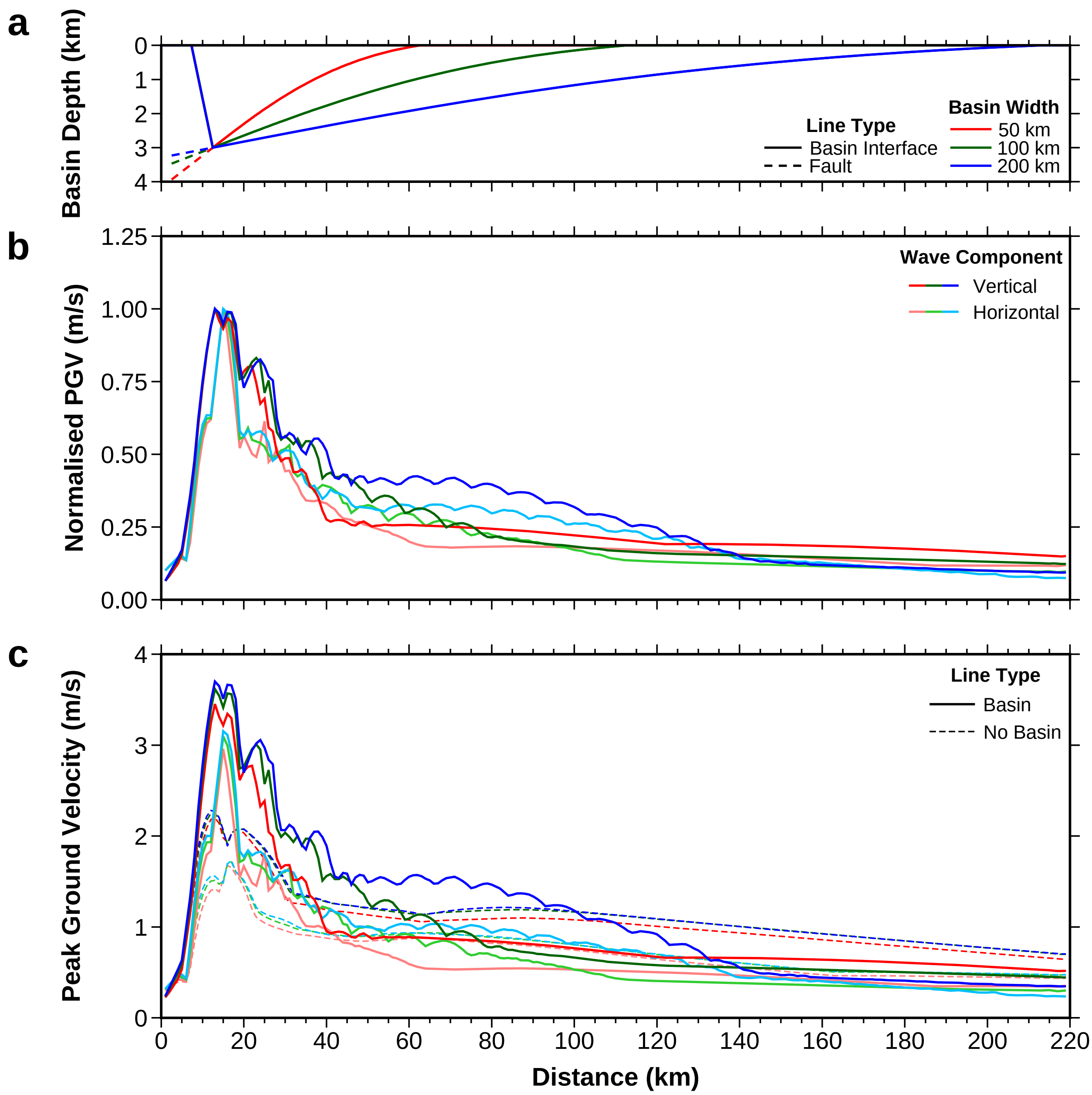

Figure 7: Peak ground velocity plotted as a function of distance across a foreland basin for three different basin widths. The basin-basement interface in (a) and resultant PGV in (b) and (c) for 50 km, 100 km and 200 km wide basins are denoted by red, green and blue solid lines respectively. The vertical and horizontal components of ground motion for each basin in (b) and (c) are illustrated by dark- and light-coloured lines respectively. (a) demonstrates the cross-sectional setup for model simulations of varying basin widths. A 10 km planar earthquake rupture is orientated down-dip of each foreland basin and is illustrated by a dashed line. (b) shows PGV as a function of distance plotted for different basin widths, normalised by the maximum value of PGV. (c) shows PGV plotted as a function of distance for a range of basin widths. The dashed lines show the equivalent values for the reference models with no foreland basin.

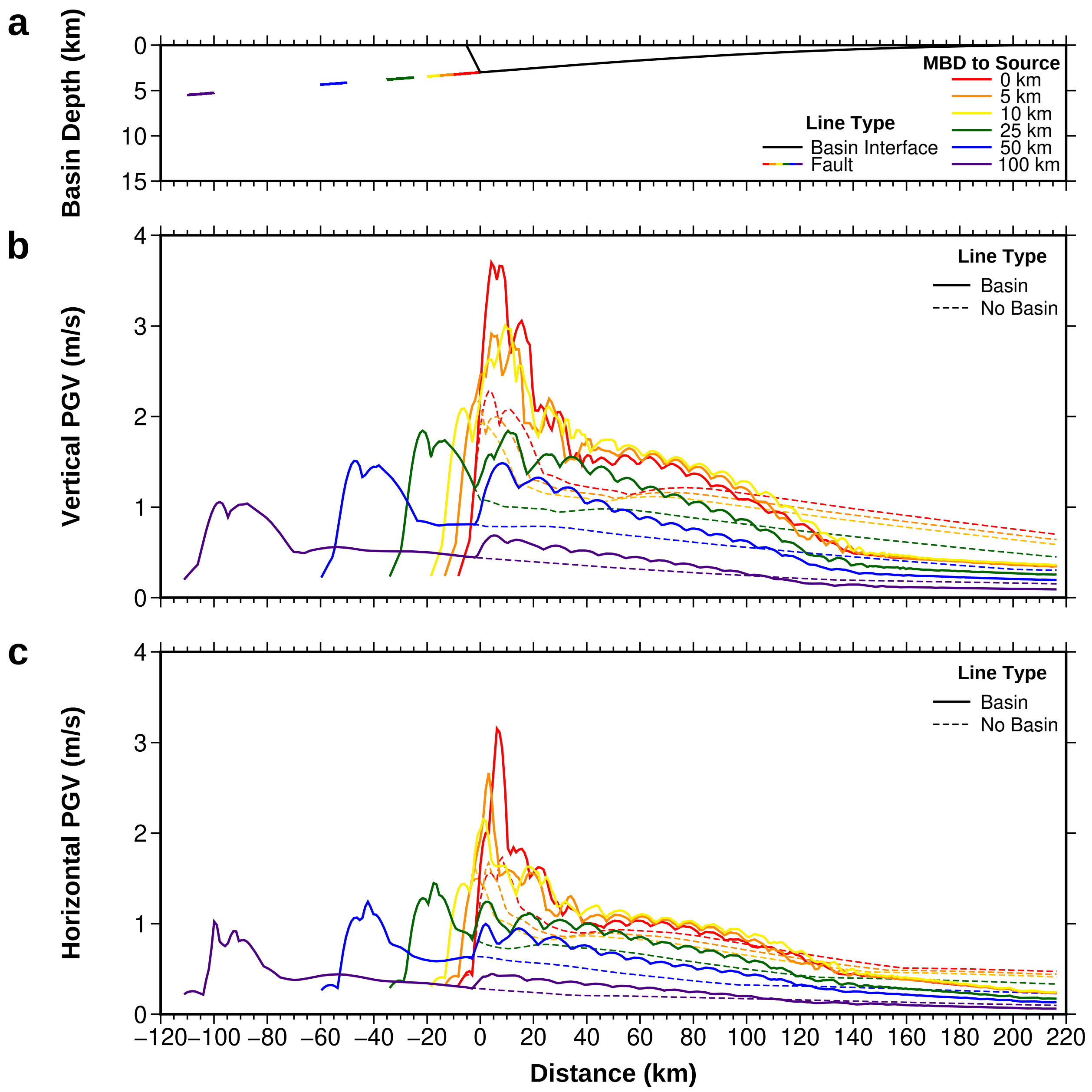

Figure 8: Peak ground velocity plotted as a function of distance across a foreland basin for six faults with different basin-to-source distances. Each fault is illustrated in (a), (b) and (c) by red, orange, yellow, green, blue and indigo coloured lines for basin-to-source distances of 0 km, 5 km, 10 km, 25 km, 50 km and 100 km respectively. The dashed lines in (b) and (c) show the equivalent values for the reference models with no foreland basin. (a) demonstrates the cross-sectional setup for model simulations of varying distances between the Maximum Basin Depth (MBD) and sources. A 10 km planar earthquake rupture is orientated down-dip and positioned at various distances from a 3 km deep, 200 km wide foreland basin which is outlined in black. (b) shows vertical PGV plotted as a function of distance for a range of faults with varying basin-to-source distances. (c) shows horizontal PGV plotted as a function of distance for a range of faults with varying basin-to-source distances.

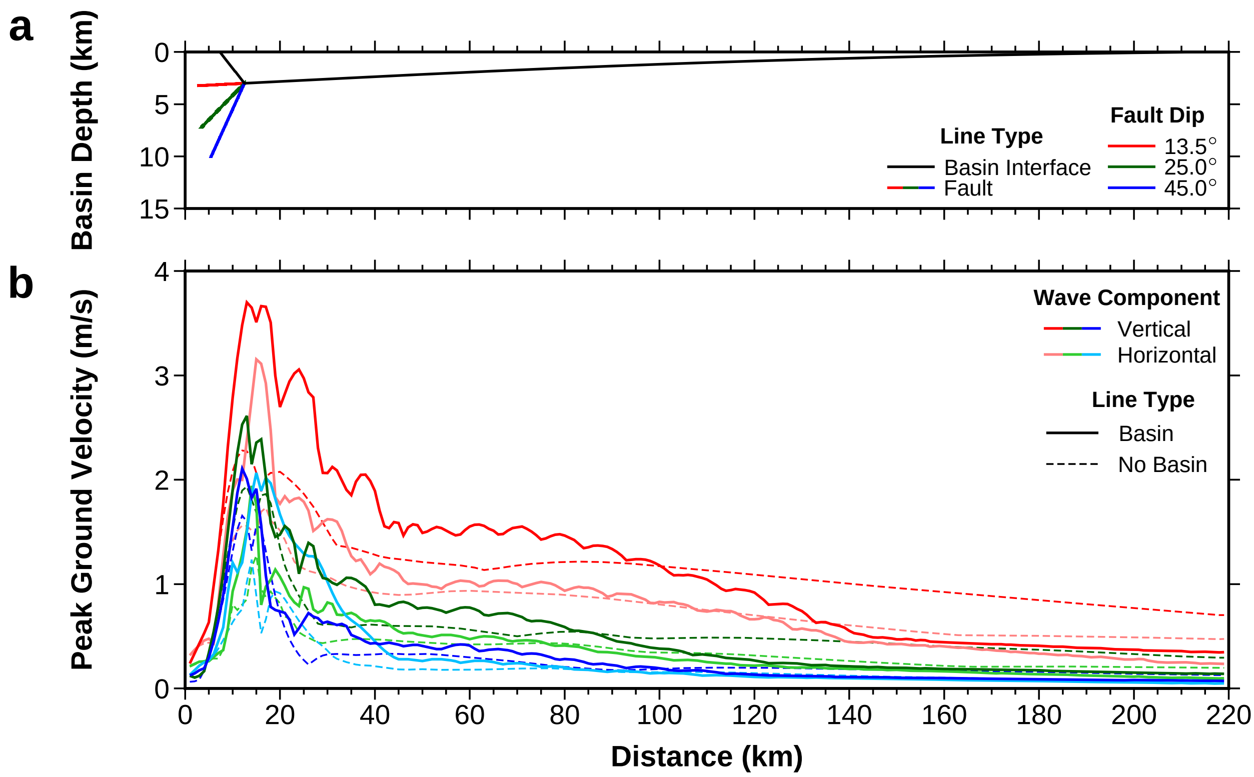

Figure 9: Peak ground velocity plotted as a function of distance across a foreland basin for three faults with different dips. Each fault is illustrated in (a) and (b) by red, green and blue lines for dips of 13.5°, 25.0° and 45.0° respectively. (a) demonstrates the cross-sectional setup for model simulations of varying fault dips. A 10 km planar earthquake rupture is orientated at different angles, down-dip of a 3 km deep, 200 km wide foreland basin which is outlined in black. (b) shows PGV plotted as a function of distance for a range of source dips. The vertical and horizontal components are illustrated by dark- and light-coloured lines respectively. The dashed lines show the equivalent values for the reference models with no foreland basin.

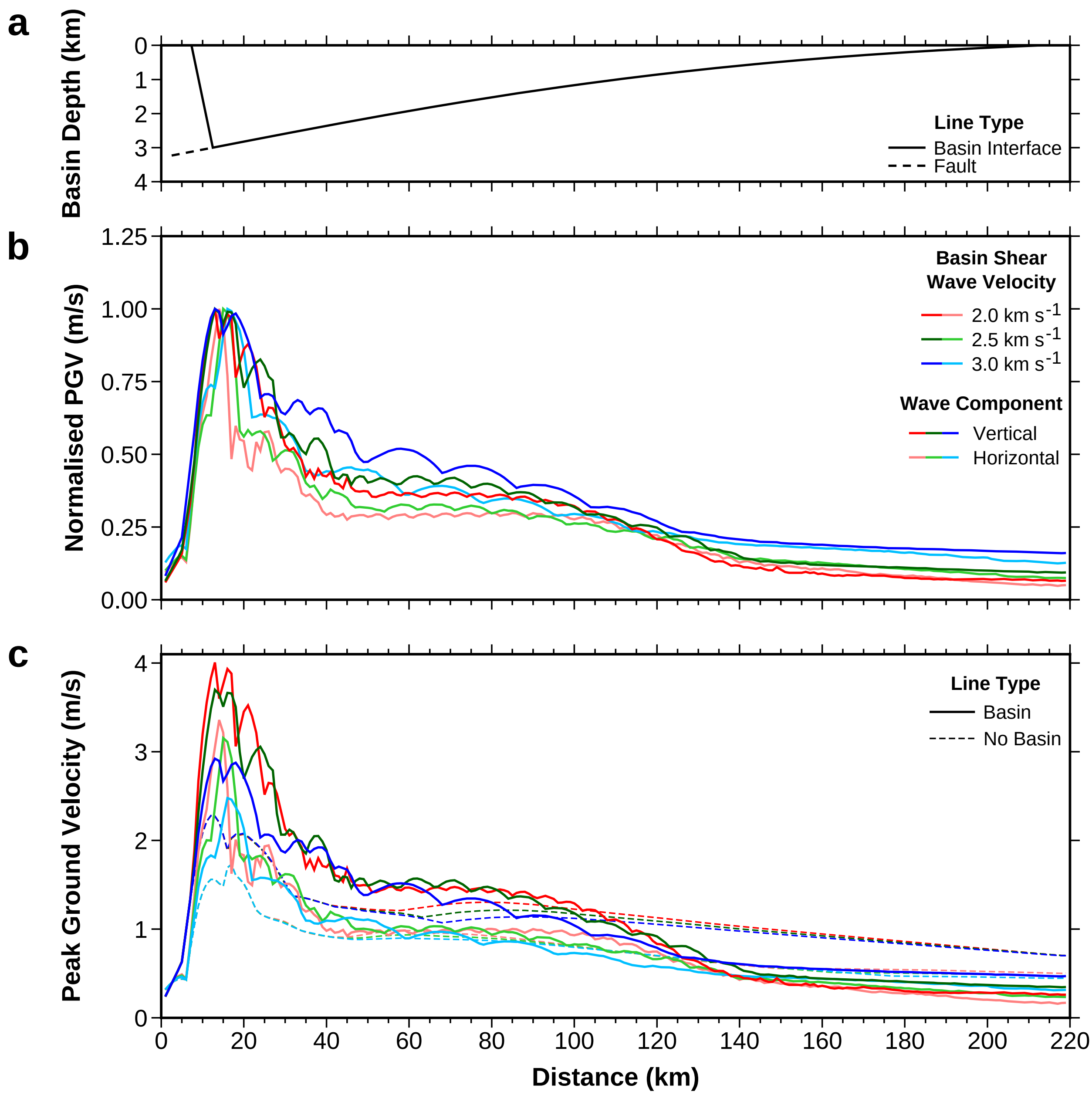

Figure 10: Peak ground velocity plotted as a function of distance across a foreland basin for three different basin shear-wave velocities (with the P-wave velocity also varied to keep a constant V_P/V_S ratio of 1.75). The resultant PGVs in (b) and (c) for shear-wave speeds of 2.0 km s^{-1}, 2.5 km s^{-1} and 3.0 km s^{-1} are denoted by red, green and blue lines respectively. The vertical and horizontal components in (b) and (c) are illustrated by dark- and light-coloured lines respectively. (a) demonstrates the cross-sectional setup for the model simulations, comprising a 10 km planar earthquake rupture (dashed black line) orientated down-dip of a 3 km deep, 200 km wide foreland basin (solid black line). (b) shows PGV as a function of distance plotted for different basin S-wave velocities, normalised by the maximum value of PGV. (c) shows PGV plotted as a function of distance for a range of basin S-wave velocities. The dashed lines show the equivalent values for the reference models with no foreland basin.

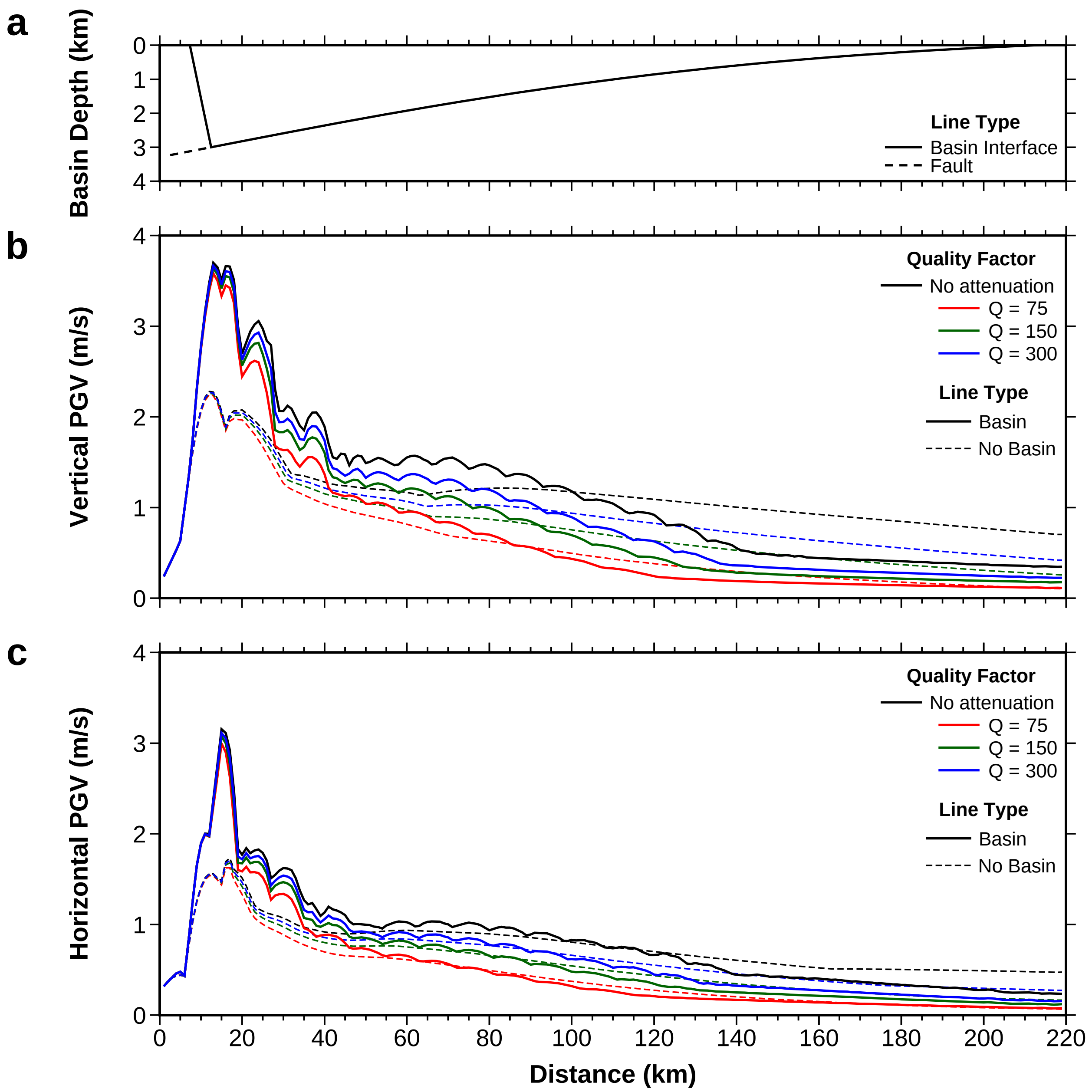

Figure 11: Peak ground velocity plotted as a function of distance across a foreland basin for four simulations with different levels of attenuation. The resultant velocities in (b) and (c) for quality factors of 75, 150 and 300, in addition to the model simulation that was run under elastic conditions (labelled 'No attenuation') are denoted by red, green, blue and black lines respectively. The dashed lines in (b) and (c) show the equivalent values for the reference models with no foreland basin. (a) demonstrates the cross-sectional setup for the model simulations, comprising a 10 km planar earthquake rupture (black dashed line) orientated down-dip of a 3 km deep, 200 km wide foreland basin (black solid line). (b) shows vertical PGV plotted as a function of distance for a range of attenuation quality factors. (c) shows horizontal PGV plotted as a function of distance for a range of attenuation quality factors.

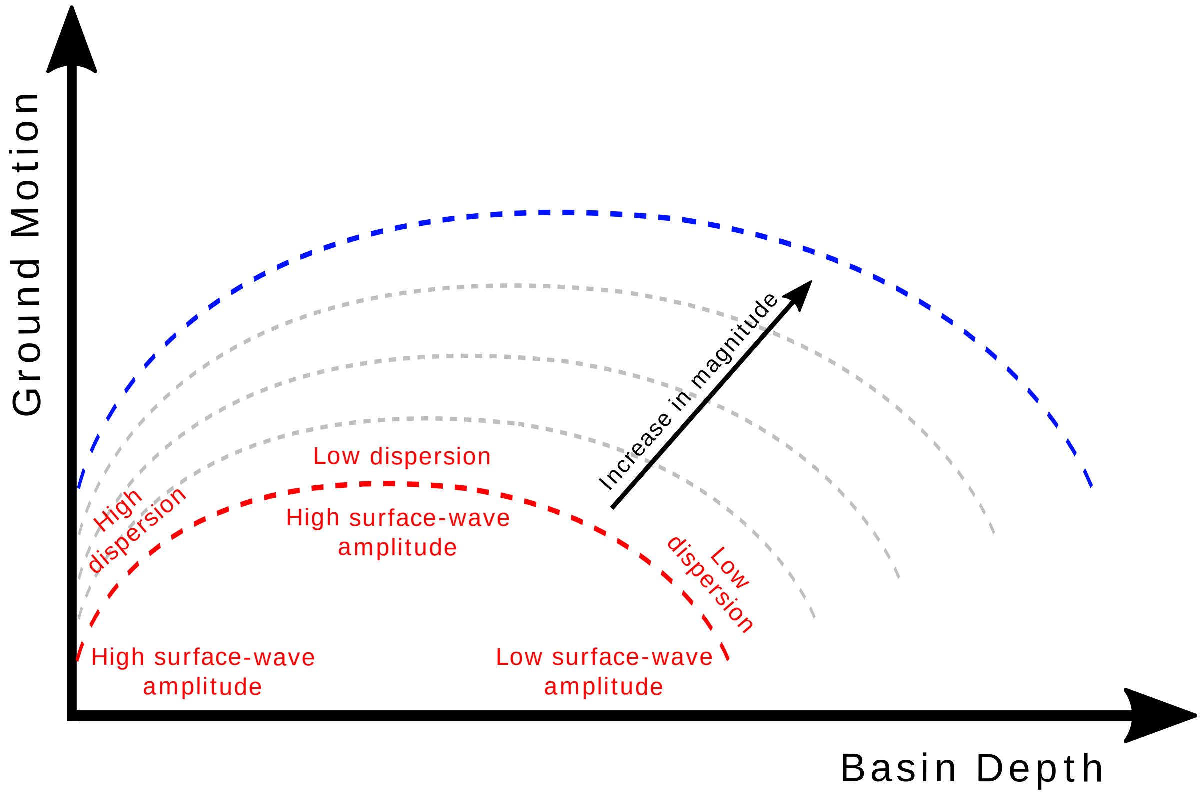

Figure 12: Schematic diagram illustrating the effect of basin depth on ground motion for a location near the centre of the basin. The red dashed line denotes the ground motion trend produced by our modelled M_w 6.5 thrust-faulting earthquake, whilst the blue and grey dashed lines are the expected trends if the source magnitude was increased.

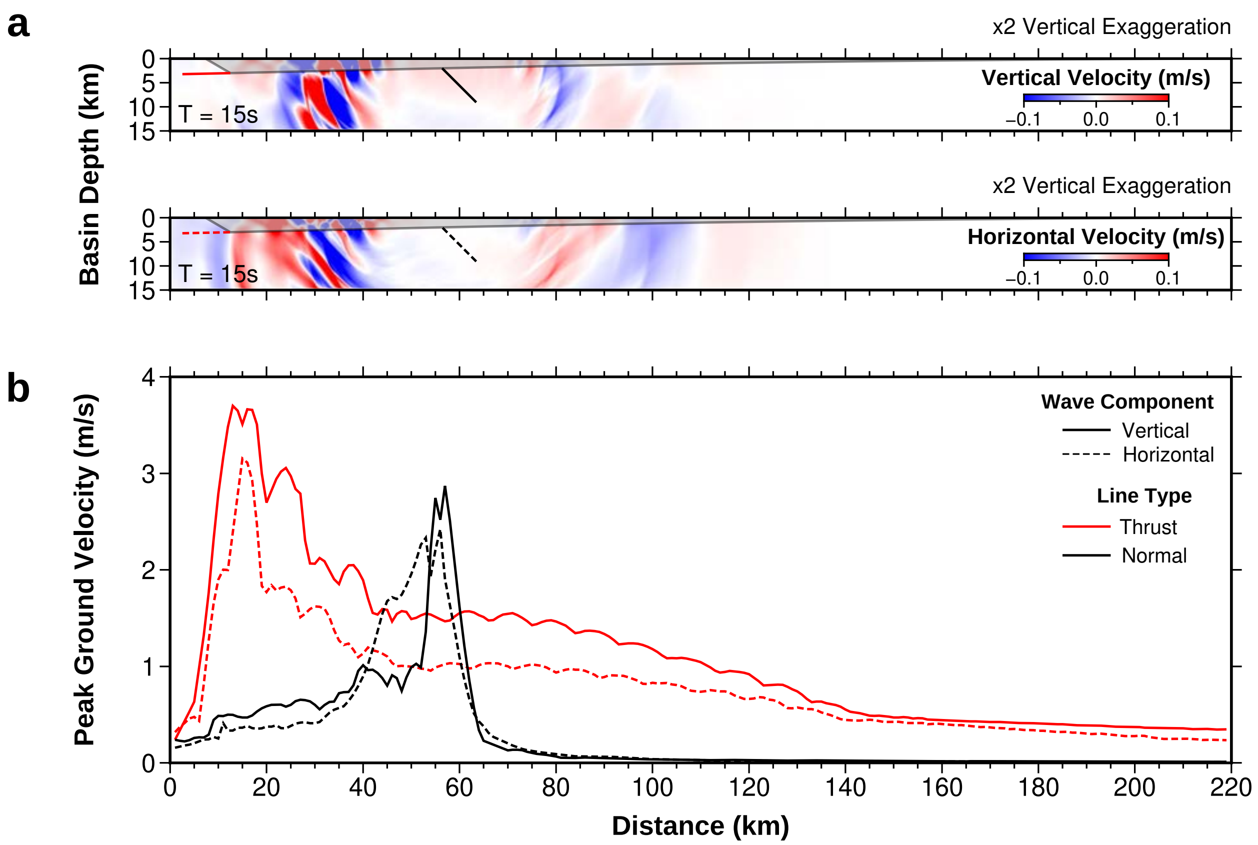

Figure 13: Calculated velocities plotted as a function of distance across a foreland basin for two M_w 6.5 ruptures with different earthquake mechanisms. (a) demonstrates the cross-sectional setup for a normal-faulting earthquake in the underlying basement, overlain with the resultant vertical and horizontal wavefield produced at 15 seconds from the onset of the earthquake rupture. The maximum velocity is ≈3 m s^{-1}, however, the scale bar has been saturated to illustrate all wave effects. (b) shows PGV plotted as a function of distance for a normal-faulting earthquake simulation (black line). For comparison, we have plotted the PGV for a range-front thrust-faulting earthquake (red line) using the same basin geometry. The vertical and horizontal components are illustrated by solid and dashed lines respectively.