3D seismic wave amplification in the Indo-Gangetic basin from spectral element simulations

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ABSTRACT

This study investigates seismic wave amplification effects in the Indo-Gangetic (IG) basin for possible large earthquakes in the region using spectral-element simulations. The Indo-Gangetic basin is a large and deep sedimentary basin that covers the northern part of India, in which several mega-cities are located, including the capital city of New Delhi. The seismicity in the region due to presence of many active tectonic faults is an important matter of concern for engineers. The damage caused in a future large earthquake could affect a huge population and hinder the development of numerous large-scale industrial establishments. Due to local soil conditions and the structural complexity of the sedimentary basin, seismic wave amplification is expected. However, the absence of seismic data for large earthquakes and limited knowledge of the structure of the basin poses challenge in estimating shaking amplifications. Therefore, we model the 3D structure of the basin using Spectral Finite Element method (Specfem3D) including the topography of the Himalayan mountains, and compute synthetic seismograms for a suite of simulated rupture scenarios. First, we use two past earthquakes in the basin to calibrate our 3D model by comparing the simulated ground motions with the recorded data. Later, we consider realizations of potential future large earthquake (M_w 7.1), by generating different kinematic rupture models. We simulate earthquake scenarios for different source parameters to quantify the statistics of expected ground shaking levels. We then infer seismic wave amplification as a function of both frequency and basin depth for complex seismic sources. Our results indicate a maximum amplification of 16 in Peak Ground Velocity (PGV) and 19–35 in Spectral Accelerations (Sa) at long periods. The results presented in this study may be useful for engineers to predict ground motions for future large earthquakes in absence of any available seismicity data.

1. Introduction

Quantifying amplification of ground motion in sedimentary basins is one of the key issues faced by seismologists and earthquake engineers. The “basin effect” is manifested in the surface waves that are generated from the interaction of body waves with the boundaries of the basin. Further propagation of surface waves through the different sediment strata, alters the frequency content of the waves due to the sediment-crust impedance contrast [1]. Depending on their frequency content, the waves are either attenuated or amplified by repeated reflections and refractions within the basin. Thus, the spectral amplitudes largely depend on the material properties of the sediment layers as well as the geometry of the basin. There are several empirical and analytical equations to estimate the fundamental resonant frequency of the basin that are based on basin width, depth and shear wave velocities [1,2]. Amplification occurs when this resonant frequency approaches any of the frequencies of earthquake

ground motion. Observations from many past earthquakes - 1985 Mexico City Earthquake, 1999 Chi-Chi Earthquake, 2010 Darfield Earthquake, 2011 Christchurch Earthquake - have provided insights into seismic wave amplification at basin sites [3–6].

Engineers widely use “Ground Motion Prediction Equations” (GMPEs) to compute ground motion intensity measures, like Peak Ground Accelerations (PGA), Peak Ground Velocity (PGV), Peak Ground Displacement (PGD) or Spectral Acceleration (Sa at different natural periods). Including amplification effects in these intensity measures as a function of basin depth and frequency of seismic waves is not a trivial problem using GMPEs, because the GMPEs consider the complexities of seismic source and the propagating medium only in a simplified manner. The local site conditions are accounted in the GMPEs using Vs30 (Shear wave velocity of top 30m from the surface) and the spectral decay parameter kappa (κ). However, accurate estimation of these parameters requires a large number of strong motion data and shear wave velocity profile measurements [7,8].

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Also, in large and deep sedimentary basins, the site amplification quantified by Vs30 might not be sufficient because the deeper sediment layers and the geometry of the basin plays a major role. We investigate this issue for the Indo-Gangetic (IG) basin, a large and deep seismically active sedimentary basin that covers almost the central part of India.

The IG basin is a sedimentary basin that is formed between the subducting Indian plate and the Eurasian plate, and comprises thick alluvial deposits. This basin is the depressed part of peninsular India (also known as Peninsular Indian shield) with several hidden faults [9,10]. The thickness of the sediment fill ranges between few tens of meters and 6 km, increasing progressively from south to north in the form of an asymmetrical wedge. The seismo-tectonics of the region is shown in Fig. 1. The IG basin is shown as three distinguished regions in Fig. 1- “Indus Plain” in the West, “Ganges Plain” in the center, and “Brahmaputra Plain” in the East. These regions are named after the important rivers that contribute to the alluvial deposition. Although these are shown separately, we consider the entire region in our simulation as one sedimentary basin. A schematic layout of the shape of IG basin is also shown in Fig. 1. The active hidden faults embedded within the basin pose a significant seismic hazard to this region. Furthermore, the estimated rise in population by 41% due to large industrial establishments increases the vulnerability of many urban agglomerations in this region [11,12]. An earthquake of magnitude Mw 8, Mw 7, Mw 6 and Mw 5 near an urban agglomeration can cause fatalities up to 0.3 million, 40,000, 10,000 and 100 people respectively [13].

Several investigators attempted to understand basin effects for the IG basin from many past earthquakes outside the basin. PGA and PGV at hard sites were computed for large earthquakes (1897 Mw 8.1 Assam, 1905 Mw 7.8 Kangra and 1934 Mw 8.2 Bihar-Nepal) and basin amplifications up to a factor of 3 were inferred near the river banks and flood plains [14]. The damaging 1934 Mw 8.2 Bihar-Nepal earthquake, the 2001 Mw 7.6 Bhuj earthquake, and the most recent 2015 Mw 7.9 Nepal earthquake have illuminated the ground motion amplification in the IG Basin [15–17]. To deepen the understanding of the structure of IG basin, there have been efforts to improve broadband instrumentation in IG basin. An array of ten broadband seismographs in north-south direction near Lucknow city (marked in Fig. 1) provided a few data points to constrain the sedimentary thickness and shear wave velocities and the first estimates of amplifications (of the order of 6–12 times for PGV) were reported from observations and stochastic source and attenuation models [18,19]. Later, network expanded and is currently known as the Central Indo-Gangetic Network (CIGN), consisting of 26 strong motion velocity seismographs. This network recorded PGA and PGV at basin sites 3–4 times higher than the adjacent hard rock sites during the 2014 Mw 6.1 Bay of Bengal Earthquake and the 2015 Mw 7.9 Nepal Earthquake [16]. However, the urban agglomerations in our study region still lack dense data to sufficiently constrain the basin structure and to capture seismic wave amplifications for possible large earthquakes in future.

In absence of broadband data, simulation-based approaches using numerical models of the basin and seismic sources, are more insightful in quantifying ground motions. One recent study models IG basin

Fig. 1. Top: The tectonic setting of Indo-Gangetic basin and surrounding region. Seismicity (2005–2017) is indicated as circles. [Note: The study region is the one marked in red box]. Few active faults are marked as MBT (Main Boundary Thrust), MDF (Mahendragarh-Dehradun Fault), ADF (Aravalli-Delhi Fault), LF (Lucknow Fault). Bottom: Schematic diagram (not in scale) of a section in North-south direction through the basin. (For interpretation of the references to color in this figure legend, the reader is referred to the Web version of this article.)

for seismic sources within the basin [20] using the available shear wave structure of the basin. They reported statistics of expected shaking levels of PGD and PGV on bed rock sites for many large scenario earthquakes. However, their study was not focussed to study basin effects, but to identify the source parameters that caused most severe ground motions in the region. Later, amplifications were estimated in ground motions at basin sites due to Himalayan seismic sources, using a 2D finite element model for IG basin without including the topography of Himalaya [21]. However, seismic amplification can be captured more accurately by including the 3D structure of the IG basin.

In the present study, we develop a 3D finite element model of the IG basin (between Latitude 24.5–32.5°N and Longitude 74.5–82.5°E, Fig. 1). The geometry of the 3D basin structure is available from the “Seismo-tectonic atlas of India” assembled by Geological Survey of India [22]. We extrapolate the information on material properties of the sedimentary layers in the basin (P-wave velocity V_p , shear-wave velocity V_s , density ρ and thickness H) [19,23]. Most of the studies discuss basin effects due to active Himalayan seismic sources that are located outside the basin. The seismic activity in the basin is moderate compared to the Himalaya, and most of the earthquakes are associated with oblique strike-slip fault mechanism on the “Mahendragarh-Dehradun-Fault” (MDF). Seismic sources that are located far away from the basin (Himalaya and Peninsular Indian shield) also produce seismic wave amplification locally at various basin sites. However, a large earthquake on the MDF fault is capable of exciting the entire basin significantly and hence, it is important to quantify the possible amplifications at various basin sites. Therefore, we use the Spectral Finite Element method [24] to model basin effects in ground motions due to large scenario earthquakes (Mw 7.1) within the IG basin. We present amplifications of the simulated ground motions over a grid of stations between Latitude 25–31°N and Longitude 75–81°E at both surface and bedrock levels.

2. Modelling of IG basin

The 3D model of IG basin is developed using Spectral Element Method (Specfem3D), a numerical code that is based on continuous Galerkin spectral element method that can efficiently model seismic wave propagation in sedimentary basins by using arbitrary unstructured hexahedral meshes. The detailed modelling procedure and the characteristics of the finite element mesh is described in the following sub sections.

2.1. Geometry and material properties

As depicted in Fig. 1, the boundary that separates Himalaya and IG basin lies along the Main Boundary Thrust (MBT) and Main Frontal Thrust (MFT). The IG basin is a huge plain spreading to the south, approaching the Peninsular Indian Shield. The spatial domain used in the present study ranges over a distance of 760km in the east direction, 910km in the north direction and extends to a depth of 80km. The depth of the IG basin increases progressively northwards and reaches a maximum of 6km at the foot of the Himalaya (Fig. 1b), comprising sediment layers with varying material properties. These material properties include the reported values of density (ρ), p wave velocity (V_p), s wave velocity (V_s) and quality factor (Q) for attenuation in the study region [19,23]. V_s of the bed rock beneath the basin is assumed as 3.2km/s. The lowest shear wave velocity at shallowest part of the basin is as low as 475m/s. The thickness and material properties (V_p , V_s , ρ) of each basin layer and the other regions including the Indian Shield and Himalaya are tabulated in Table 1. These values are consistent with the material properties reported in many other studies for the region [36,37,38]. Thus, our model integrates the effect of both shallow sediment layers and deep

Table 1

Material Properties for the different regions of the 3D Model.

Region	Depth (km)	Density (kg/m ³)	V_p (km/s)	V_s (km/s)	Quality factor
Indian Shield	0–80	2680	5.600	3.200	800
IG Basin	0–0.25	1586	0.740	0.475	150
	0.25–1	1870	1.324	0.850	150
	1–2.25	1950	1.558	1.000	150
	2.25–4.75	2425	3.740	2.400	150
	4.75–6	2560	4.675	3.000	150
Himalaya	0–1	2495	4.200	2.400	250
	1–6	2632	5.200	2.970	250
	6–80	2680	5.600	3.200	800

basin structure. We interpolate the material properties onto our simulation grid that covers the spatial domain of $160 \times 160 \times 160$ points in x, y and z directions. From the known boundary of the IG basin, the shape of the basin profile is constructed by using “control points” within the basin. At each control point, the depth of the basin is interpolated between 0 at the southern end (where the basin begins) and 6 km at the northern end (where the foothills of Himalaya begins). For each longitude, there are 12 control points that are used to define the geometric profile of the basin, and the total depth of sedimentary layer progressively increases from south, reaching a depth of 6 km at the northern end. At each point along the depth, the properties are interpolated from Table 1. Uniform material properties are assumed for the shield region and for depths greater than 6 km. The corresponding 3D velocity model is shown in Fig. 2.

2.2. Discretization of the spatial domain

Once the material properties are identified for each region in the model, the next step is to generate the spectral element mesh. To achieve this, the model domain has to be discretized into a number of finite elements in x, y and z directions. In spectral element method, hexahedral elements are used to mesh the spatial domain. The high-resolution (30 m interval) topography model available from the US National Oceanic and Atmospheric Administration (NOAA) is smoothened and implemented in the model as surface elevation. The highest elevation used in the model is approximately 6.8 km. The number of spectral elements in x and y directions is $1024 \times 1024 \times 44$ with dx and dy as 0.7 km $\times$ 0.9 km (Figure showing the spectral element mesh is given in the Electronic Supplement). Table 2 provides the characteristics of the spectral element mesh used in the spatial domain. Each element is further divided into $5 \times 5 \times 5$ grid points. Therefore, the resolution, or distance between grid points on the surface, obtained from this discretization is approximately 0.2 km. In vertical (z) direction, the top 0 to 1 km is divided into 2 elements with $dz = 0.5$ km. Further down, 1 to 6 km is discretized into 5 elements with $dz = 1.0$ km. Beyond 6 km, 37 elements are used with an approximate $dz = 2.0$ km up to the depth of 80 km. In order to optimize the computational cost, we used “mesh doubling” to increase the spatial resolution of the mesh at depths of 1 km (1.4×1.8 km) and 6 km (2.8×3.6 km). The process of doubling also ensures that all elements have an approximate cube shape for stable solution. The size of the mesh is fixed using a trial and error approach, by testing the minimum resolving period (or the highest frequency) for the lowest shear wave velocity. Considering the low shear wave velocity of 475 m/s, we found that our numerical model can accurately resolve periods greater than 2.0 s (or frequencies less than 0.5 Hz) for the assumed grid size (approximate distance between gauss points $0.14 \text{ km} \times 0.18 \text{ km} \times 0.10 \text{ km}$) and a computational time step of 0.0025 s.

Fig. 2. 3D model of IG basin showing the variation of shear wave velocity in the study region. The model is developed in SPECFEM3D. Bottom Right: A zoomed view of the cross-section of IG basin, showing the progressively increasing basin depth.

Table 2
Characteristics of Spectral element mesh.

Characteristic	Value
Dimension (km × km)	760 × 910
Depth (km)	80
Type of element	Hexahedral
No. of spectral element on each side	1024 × 1024 (top-surface)
Total no. of elements	7.8×10^6
Min size of element (km × km × km)	0.70 × 0.90 × 0.50
Max size of element (km × km × km)	2.80 × 3.60 × 2.00
No. of GLL points in the entire mesh	9.4 Million
Total Degree of freedom	28.3 Million
Time Step used for simulation	0.0025 s

The simulations are performed at the Kaust Supercomputing Laboratory (KSL). Each simulation is performed on 32 compute nodes (1024 processor cores). The runtime of a complete simulation on each node is about 6 h to produce a 250 s long seismogram at all the specified station locations.

3. Comparison with strong motion data

To calibrate our basin model, we selected two past earthquakes in the IG basin to compare recorded ground motions with simulated results. The Indian strong motion network maintained by PESMOS (Program for Excellence in Strong Motion Studies) comprises about 300 digital accelerographs in the highly seismic zones including northern and north-eastern India. Among these data sets, we selected strong motion records from two earthquakes in the IG basin - 1) Mw 4.9 event (5th March 2012) 2) Mw 4.3 event (7th September 2011), both on the MDF fault segment. The source location, focal mechanisms, along with the locations of selected recording stations are shown in Fig. 3 for the respective events. The strong motion data for the events are available at 13 and 10 sites, most of them in the basin. The source is represented as a point double couple, with a Gaussian source time function. The rise time of slip is obtained from

the relation given in Equation (1) [25].

$$T_r = 2.303 \times 10^{-9} M_0^{(1/3)} \quad (1)$$

where M_0 is the seismic moment (dyne-cm). The rise times computed using Equation (1) are 0.26 s and 0.11 s for the two events. The event depths are known from PESMOS database and are located at 14 km and 8 km with strike, dip and rake angles 347° , 47° and 131° respectively. Assuming a material rigidity of $4 \times 10^{10} \text{ N/m}^2$, the mean slips used to compute the components of moment tensor are obtained as 10 cm and 5 cm respectively for both the events using the relations of [26,27]. Both recorded and simulated waveforms are bandpass filtered in the frequency range of 0.01–0.5 Hz using a second order Butterworth filter.

For the first event, the comparison of observed and simulated waveforms and Fourier amplitude spectra for three sites (DL1-02, DL1-05 and DL1-12) is shown in Fig. 4. For the second event, the comparisons are shown for sites DL2-02, DL2-07 and DL2-10 in Fig. 5. The model bias is calculated as the ratio of logarithm (base 10) of PGV (recorded) to PGV (simulated). The bias in both horizontal (geometric mean of EW and NS components) and vertical component of PGV for all the sites is depicted in Fig. 6. The simulated waveforms closely agree with the recorded data at almost all sites in the basin. The peak amplitudes (PGV) in the simulated velocity seismograms match with the recorded ground motion at most sites, including the phase-arrival times. The generated surface waves in the later part of the seismograms are visible in both recorded and simulated data, which means that our 3D model captures the basin effects in the simulated waveforms. The above observations indicate that our 3D basin model is sufficiently accurate for ground motion simulations, allowing us to study seismic wave amplification for possible earthquake scenarios.

4. Scenario earthquakes

We now quantify seismic wave amplifications due to possible large earthquakes in the region. Although there have not been any large earthquakes in this region recently, it is important to consider

Fig. 3. Focal mechanism of past events used in the study. Top Panel shows the entire regional model used in the study, highlighting two earthquakes in the past (a) the Mw4.9 event on March 5, 2012 (Event 1: DL1) and (b) the Mw4.2 event on September 7, 2011 (Event 2: DL2). Bottom Panel shows the locations of the recording stations for the individual events.

these events in probabilistic seismic hazard analysis. In absence of strong motion records for large earthquakes, numerical simulations provide more insights into basin effects on ground motions. Because we consider large earthquakes in our scenarios, the earthquake source plays a major role, especially in the near field region. Therefore, we do not use the point source approximation anymore and include more source parameters which are detailed below.

4.1. Finite fault source models

Here we consider thirty scenarios of Mw7.1 strike-slip earthquake in the IG basin. All events are assumed to be located near the north-south trending MDF fault across the IG basin. Based on regional focal-mechanism solutions and the seismicity depths, we define the strike (55°), the dip (50°), the rake (10°), and the depth of top edge of fault (5 km) for these events. The rupture plane dimensions (L and W) are estimated using the empirical scaling laws of [26,27]. For the scenarios, we use a rupture length $L = 45$ km and width

$W = 25$ km. The rupture models showing the slip distribution, hypocenter location and rupture times are shown in Fig. 7. Each rupture model is divided into subfaults of size $1 \text{ km} \times 1 \text{ km}$. For each scenario, the slip distribution and the hypocenter location is varied on the fault plane using the approaches of [28,29]. The source time function is a Gaussian pulse, with the rise time that scales as cube root of slip (Equation (1)) and we assume constant rupture velocity (0.8 times shear wave velocity). The rise time of the slip in each subfault is 2.2 s. The slip velocity function assumed for a sample scenario event is also indicated in Fig. 7. Note that due to assumption of constant rupture velocity, the rupture expands radially isotropic with distance from hypocenter. Because our numerical simulations are more accurate for low frequencies, we include source complexities that are static in nature (variations in slip distribution and hypocenter locations). Due to our low resolving frequencies ($f_{\max} = 0.5 \text{ Hz}$), we do not consider variations in rupture velocities and onset time of slip, which are the sources of high frequency radiation.

Fig. 4. Left panel: Comparison of velocity time histories at few selected stations for the Mw4.9 event on March 5, 2012 (Event 1: DL1). Black line indicates instrument data, and red line indicates simulated time histories. Right Panel: Comparison of fourier amplitude spectra at the corresponding stations. Both recorded and simulated waveforms bandpass filtered in the frequency range 0–0.5Hz. (For interpretation of the references to color in this figure legend, the reader is referred to the Web version of this article.)

4.2. Receiver configuration

To analyze the simulated wavefield, we used four different arrays of receivers along various azimuth angles ($\theta = 55^\circ, 145^\circ, 190^\circ$) considering the fault trace (Fig. 8). Each array consists of 10 receivers (at interval of 33km) including the one on the top center of the fault. To perform detailed statistical analysis, we also defined a grid of stations (101×101 in both directions, at interval of approximately

6.5km). These locations are identical for all rupture models used in the study. We computed Joyner-Boore distances (R_{JB}), the shortest distance of the receiver from the surface trace of the fault plane, and used these measures to compare our results with GMPE-based predictions. Here, R_{JB} varies between 0km and 500km. For computing amplifications, all receivers are assumed to be located both at top surface level and bed rock level (below 6km). Therefore, in all the simulations, we use a total of 20476 station locations to obtain three component synthetic wave-forms.

Fig. 5. Left Panel: Comparison of velocity time histories at few selected stations for the Mw4.2 event on September 7, 2011 (Event: DL2). Black line indicates instrument data, and red line indicates simulated time histories. Right Panel: Comparison of fourier amplitude spectra at the corresponding stations. Both recorded and simulated waveforms are bandpass filtered in the frequency range 0–0.5 Hz. (For interpretation of the references to color in this figure legend, the reader is referred to the Web version of this article.)

5. Results and discussion

Fig. 9 shows the sample velocity synthetics for event 1 (source model in Fig. 7) at bed rock and top surface sites along the different azimuth arrays A, B, C and D (Fig. 8). We observe several distinct features on the wave-forms at the top surface level of these selected sites. At sites located closer to the source, the synthetics exhibit strong near-fault pulses ($R < 33$ km, along array A) in EW and NS components. The dominant strike-slip faulting mechanism contributes to these directivity pulses that are not seen in the vertical

ground motion (All three components of ground motions are given in the Electronic Supplement). Moving farther away from the source, the synthetics show long period surface waves with extended duration. Along the array line A, the amplitudes of the waves at the top surface level are larger than those at the bed rock level. The amplifications along array line A are due to the progressively increasing basin depth (from $D_{basin} = 2.9$ km to $D_{basin} = 3.5$ km). Near to the source, the rupture process strongly influences the ground velocities (ground motions for stations in all the arrays are shown in E-Supplement). The amplifications observed along the array line B are almost

Fig. 6. Bias of simulated peak ground velocity (PGV) with respect to the recorded PGV. Horizontal components of PGVs are obtained as the geometric mean of East-West and North-South components of PGV. The maximum bias and the standard deviation of bias are within an acceptable range of $[-0.5, 0.5]$.

constant because all stations in this array have almost equal depth of basin sediments (average $D_{\text{basin}} = 2.6$ km). For the stations in array line C and D, we observe that ground motion amplitudes are stronger at distances nearer to the source with larger basin depths (approximately 1 km).

Next, we analyze the statistics of Fourier Amplitude Spectra by computing the mean and standard deviation for all 30 scenario events (Fig. 10). The average horizontal component that we used in our simulation is *rotd50* [30], which is commonly adopted in GMPE-based approaches. We observe visible amplifications in the spectra up to a corner frequency of 0.5 Hz beyond which the spectra decays. The spectra peaks at specific frequencies for each array line at various basin depths. The frequencies at which the peaks are observed largely depend on the resonant frequencies of the basin for that particular depth ([11]). Along array A, difference between ground motion intensities at top surface level and bed rock level increases progressively with increase in basin depth. Array B stations lie along a line of approximately constant basin depth, therefore, the differences between ground motion intensities at top surface level and bed rock level are observed to be of same order. Interestingly, amplifications observed from arrays C and D that pass through Aravalli ridges (refer to Fig. 8), are due to a combined effect of basin depth and multiple scattering in the ridge-valley boundaries. For example, differences between ground motion intensities at top surface level and bed rock level near the ridges at shallower basin depths (0.3 km) are larger than those at deeper basin sites (1.7 km). Along array C, these differences are not visible at some sites because they are located on the ridge, where basin depth is zero. The peaks that are visible in the decay part of the spectra (beyond 1 Hz) are numerical artifacts that are not important in the range of frequencies resolved by the model.

However, to provide more insights into the frequency dependent aspects of difference between ground motion intensities at top surface level and bed rock level, we calculate site amplification (S.A) in Spectral Acceleration S_a as the ratio of S_a at top surface to S_a at bed rock. This ratio is plotted as a function of natural period for various

basin depths along the individual azimuth arrays A, B, C and D (Fig. 11). The amplifications are visible at specific natural periods (2 s, 4 s, 8 s) in all scenarios. We also observe that the maximum amplification, of the mean from all scenarios, shifts to longer natural periods with increasing basin depths (for example, along array C, maximum amplification is at 2 s for basin depth of 0.4 km, 6 s for basin depth of 0.9 km, and 7 s for basin depths larger than 1 km). Along array line A, the maximum amplification is at $T = 8$ s, and there are more variabilities at lower natural periods (2 s, 4 s) for larger basin depths. Waves propagated from variable sources undergo multiple scattering depending on variation in material property. Along array A, the basin depth progressively increases contributing to larger variation in material property. The stations along array B have almost equal basin depths (average 2.6 km), however, the basin effects periodically occur at natural periods 2 s, 3.5 s and 8 s. Very near to the source, the periodicity is not visible and the variabilities are more spread across all natural periods. The array C is in the strike-normal direction, and as indicated previously, we observe no significant amplification at sites that are located on the Aravalli ridge (at $R = 100$ km, 167 km) as well as those at the start of the basin (at $R = 233$ km, $D_{\text{basin}} = 0$ km). However, sites that are located at ridge-valley interfaces exhibit amplifications due to scattering. Along array D, variability observed at shallower basin depths (0.3 km) is possibly due to multiple scattering at ridge-valley interfaces. Therefore, in our analysis, we observe that basin amplification is a complicated function of basin depth, frequency and multiple scattering at ridge-valley interfaces.

To further analyze our simulation results and to compare them with common engineering approaches, we used standard GMPEs, which are applicable for shallow crustal earthquakes [30–35], to compare against our simulated *Rotd50* PGV. In Fig. 12, we show simulated PGVs at the top surface and bed rock for all scenario events. We applied the top surface shear wave velocity (475 m/s) in all GMPEs and predicted lower PGVs that mark the -1σ bound of simulated PGVs at top surface level. Both simulated and GMPE predicted PGVs capture the distance decay beyond 10 km. The simulated

Fig. 7. Rupture models used in the simulations. The models are generated following the method of [28] for finite fault dimensions 45×25 km. Hypocenter locations are varied on the fault plane and a constant rupture velocity (0.8 times V_s) is assumed. Bottom right: Slip velocity functions at randomly selected sub-faults of the first rupture model. The rise time is 2.2 s (using Equation (1)). All the rupture models used in our simulation are given in Electronic Supplement.

Fig. 8. Location of receivers in the study region. Four arrays A, B, C and D along different azimuth angles θ , each having 10 receivers are indicated by $\blacktriangle$. The simulations are performed on these array of receivers (at every 33 km interval) and also on a grid of stations (at every 6.5 km interval in North and East direction). Receivers are placed both at bed rock (depth 6 km) and top surface level. Important towns and cities are marked as squares.

PGVs for the bed rock sites are below the GMPEs and the simulated PGVs for the top surface level stations are almost above the GMPEs. The most striking feature is that the amplification in simulated PGVs increase not only with basin depth, but also due to additional scattering effects at ridge-valley interfaces as observed for arrays C and D in Fig. 11. Maximum amplification in PGV is observed at the deepest part of the basin, which is possibly due to the basin-edge effect. Additionally, we also noted slight amplification beyond 300 km for zero basin depths, which correspond to sites at the foothills of Himalaya. These amplifications could be possibly due to topographic effects that are also modelled in our study region (refer to Fig. 2).

Furthermore, we investigate the spatial variability of amplification in PGV (RotD50) and Sa at long periods ($T = 4$ s and 8 s). We compute the mean and standard deviation of amplification in PGV, both EW and NS components separately (the statistics are given in Electronic Supplement Figure 5), and for PGV (RotD50) as shown in Fig. 13. Here we observe maximum amplification (max S.A = 16) due to basin effects in the NS component, focused towards north of the seismic source near the edge of basin. The observed pattern of amplification corresponds to the radiated seismic energy field of North-South component of PGV, thus indicating that the seismic wave amplifications are predominantly influenced by both source mechanism (strike-slip, rake = 10°) and basin geometry. In order to

quantify seismic wave amplification as a function of basin depth, we identify 3 different sections AA', BB' and CC' (see Fig. 8). Among the three sections considered, Sections BB' and CC' predominantly pass through ridge-valley interface. Fig. 13 (Bottom panel) shows amplifications in PGV for basin depths along these three sections individually. Along AA', we observe that amplifications at basin depths shallower than 2 km is possibly due to multiple scattering at ridge-valley interface and those beyond 2 km are due to larger basin depths. Along BB' and CC', we note that variability is much more at larger basin depths due to source effects combined with multiple scattering. Furthermore, we compute amplifications in Spectral Accelerations at periods $T=4$ sec and $T=8$ sec and the results are illustrated in Figs. 14 and 15). We observe that at low periods (4 s), amplification is large in strike-normal direction near Aravalli-ridge. Sa near the ridge-valley boundary is almost twice as high as that at larger basin depths. For longer periods (8 s), the effects from multiple scattering are not distinguishable from effects of source and basin depths. The amplifications also exhibit different mode-shapes for different natural periods (or resonant frequencies), thus indicating seismic wave amplification as a function of basin depth and period. We have also indicated the possible amplifications for important towns and cities in the basin in Table 3. The results show that maximum amplifications in both PGV and Sa are observed in the city of

Fig. 9. Synthetic waveforms at surface and bedrock sites for the first rupture scenario [Note: A B C and D corresponds to the azimuth-stations shown in Fig. 8, R is the distance of the station from the top center of the fault and D_{basin} is the basin depth at corresponding receiver locations].

Chandigarh (PGV 12.38, $Sa_{4 \text{ sec}}$ 19.65 and $Sa_{8 \text{ sec}}$ 14.14) for the set of scenarios considered in this study. The spatial maps for amplification in PGV and Sa for individual scenarios are provided in Electronic Supplement.

6. Conclusion

The present study models seismic wave propagation in the IG basin to quantify amplification of ground motions due to possible large earthquakes within the basin. The amplifications estimated from the spectral element simulations are larger than the GMPE-based predictions for the scenario earthquakes. Our results are important for probabilistic seismic hazard assessment of many cities in the basin, especially for long period ground motions. The seismic wave amplifications are quantified as a function of frequency and basin depth with respect to distance from the source. The observed periodic peaks in amplification at various basin depths are due to the resonance interaction between radiated seismic energy from the source, the basin sediments at certain frequencies and multiple scattering occurring at ridge-valley interface. The variabilities in PGV increase at farther sites ($R_{jb} > 100\text{km}$) due to increase in depth of basin sediments, multiple scattering and presence of topography. A slight influence of topography is observed in $PGVs$ at farther sites which are located at the foothills of Himalaya. Our analysis suggests that the variabilities are controlled by heterogeneous slip and multiple scattering from different sedimentary layers. We also show that

the spatial variability in PGV and Sa amplification is dominated by the radiated seismic energy of shear-vertical wavefield from the source and scattering at ridge-valley interface. Although we have presented amplifications for Mw 7.1 earthquakes, our 3D regional velocity model can be utilized to study many other scenarios that need to be considered in probabilistic seismic hazard assessment.

We consider our study a proof-of-concept and modeling-approach exercise, rather than an exhaustive investigation of shaking amplifications in the IG basin. More range of scenarios covering a range of magnitudes and source parameters would provide a better understanding of spatial amplification map for detailed hazard assessment.

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Fig. 10. Left Panel: Fourier amplitude spectra of velocity time-histories simulated at surface and bedrock from 30 simulations [Note: A B C and D corresponds to the azimuth-stations shown previously, R is the distance of the station from the top center of the fault and D_{basin} is the basin depth at corresponding locations]. The spectra decays at a corner frequency of 0.5 Hz. Right Panel: 1D Slices beneath each array A, B, C and D showing the basin depth variation with distance from the top center of fault.

Fig. 11. Left Panel: Amplification of S_a (RotD50) from 30 simulations [Note: A B C and D corresponds to the azimuth-stations shown previously, R is the distance of the station from the top center of the fault and D_{basin} is the basin depth at corresponding locations]. Amplifications show periodic peaks at certain natural periods for different basin depths. Source-related variabilities also increase at these time periods. Right Panel Panel: 1D Slices beneath each array A, B, C and D showing the basin depth variation with distance from the top center of fault.

Fig. 12. Comparison of simulated PGV (bed rock as pink and top surface as color-coded for different basin depths) from 30 simulations with the existing GMPEs Vs30 of 475 km/s is assumed in computing all the GMPEs. Note the increase in amplification with basin depth. At farther distances, there are few sites at the beginning of Himalaya and the PGV amplifications at these sites are due to the topography effects (see the blue dots at larger R_{jb}). (For interpretation of the references to color in this figure legend, the reader is referred to the Web version of this article.)

Appendix A. Supplementary data

Supplementary data to this article can be found online at <https://doi.org/10.1016/j.soildyn.2019.105923>.

Fig. 13. Top Panel: Mean and Standard deviation of amplification in PGV (RotD50) for the study region from all 30 scenarios. Bottom Panel: Amplification (both mean and standard deviation) versus basin depth.

Fig. 14. Top Panel: Mean and Standard deviation of amplification in S_a (RotD50) at period 4s for the study region from all 30 scenarios. Bottom Panel: Amplification (both mean and standard deviation) versus basin depth.

Fig. 15. Top Panel: Mean and Standard deviation of amplification in S_a (RotD50) at period 8s for the study region from all 30 scenarios. Bottom Panel: Amplification (both mean and standard deviation) versus basin depth.

Table 3

Statistics of amplification in PGV (RotD50) and S_a ($T=4$ s and $T=8$ sec) at important cities in the study region. The results indicate maximum amplification near the city of Chandigarh, that lies north of the source with maximum basin depth.

City	Latitude	Longitude	PGV		S_a ($T = 4$ s)		S_a ($T = 8$ s)	
	(°N)	(°E)	Mean	Std. Dev	Mean	Std. Dev	Mean	Std. Dev
Agra	27.18	78.01	2.47	0.49	4.22	0.68	4.01	0.59
Meerut	28.99	77.70	4.78	1.01	5.78	1.35	9.04	2.16
Panipat	29.39	76.97	4.22	0.94	5.02	0.48	9.33	1.23
Roorkee	29.86	77.89	4.48	0.95	9.11	1.06	5.99	0.86
Delhi	28.64	77.23	5.38	0.94	9.33	3.08	9.61	0.82
Jaipur	26.91	75.78	5.36	0.96	16.13	3.01	4.01	0.67
Kanpur	26.45	80.33	1.97	0.18	2.68	0.29	3.53	0.55
Chandigarh	30.73	76.78	12.38	1.87	19.65	0.65	14.14	2.15

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