Ground-Motion Modeling of Hayward Fault Scenario Earthquakes, Part II: Simulation of Long-Period and Broadband Ground Motions


Abstract We simulate long-period ($T > 1.0-2.0$ s) and broadband ($T > 0.1$ s) ground motions for 39 scenario earthquakes ($M_w 6.7-7.2$) involving the Hayward, Calaveras, and Rodgers Creek faults. For rupture on the Hayward fault, we consider the effects of creep on coseismic slip using two different approaches, both of which reduce the ground motions, compared with neglecting the influence of creep. Nevertheless, the scenario earthquakes generate strong shaking throughout the San Francisco Bay area, with about 50% of the urban area experiencing modified Mercalli intensity VII or greater for the magnitude 7.0 scenario events. Long-period simulations of the 2007 $M_w 4.18$ Oakland earthquake and the 2007 $M_w 5.45$ Alum Rock earthquake show that the U.S. Geological Survey’s Bay Area Velocity Model version 08.3.0 permits simulation of the amplitude and duration of shaking throughout the San Francisco Bay area for Hayward fault earthquakes, with the greatest accuracy in the Santa Clara Valley (San Jose area). The ground motions for the suite of scenarios exhibit a strong sensitivity to the rupture length (or magnitude), hypocenter (or rupture directivity), and slip distribution. The ground motions display a much weaker sensitivity to the rise time and rupture speed. Peak velocities, peak accelerations, and spectral accelerations from the synthetic broadband ground motions are, on average, slightly higher than the Next Generation Attenuation (NGA) ground-motion prediction equations. We attribute much of this difference to the seismic velocity structure in the San Francisco Bay area and how the NGA models account for basin amplification; the NGA relations may underpredict amplification in shallow sedimentary basins. The simulations also suggest that the Spudich and Chiou (2008) directivity corrections to the NGA relations could be improved by increasing the areal extent of rupture directivity with period.

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Online Material: Comparison of ground-motion intensities from Hayward synthetics to NGA results.

Introduction

The most recent large rupture of the Hayward fault, a magnitude 6.8 event (Bakun, 1999), occurred on 21 October 1868 and caused widespread damage throughout the sparsely populated region east of San Francisco Bay and significant damage in the city of San Francisco (Boatwright and Bundock, 2008). Moreover, as a result of the level of damage in San Francisco, the 1868 earthquake was called the “Great San Francisco Earthquake” until the 18 April 1906 magnitude 7.9 earthquake (Stover and Coffman, 1993). The San Francisco Bay area is currently home to approximately seven million people (2000 census, see Data and Resources section) with heavy urbanization along the entire length of the Hayward fault. In this study, we compute ground motions for a wide variety of plausible ruptures on the Hayward–Rodgers Creek fault system, which carries the highest probability of producing a magnitude 6.7 or larger event in the next 30 years (Working Group on California Earthquake Probabilities (WGCEP), 2008). We employ a suite of 39 earthquake scenarios involving rupture of the Hayward fault.
Part I (Aagaard et al., 2010) discusses the construction of the scenarios in detail. Some of the scenarios also involve rupture of a 23 km portion of the Calaveras fault (six scenarios) or rupture of the Rodgers Creek fault (four scenarios).

In previous work, we estimated ground motions for the 1906 earthquake and scenarios rupturing that same 480 km portion of the northern San Andreas fault (Aagaard, Brocher, Dolenc, Dreger, Graves, Harmsen, Hartzell, Larsen, McCandless, et al., 2008) as well as the 1989 magnitude 6.9 Loma Prieta earthquake (Aagaard, Brocher, Dolenc, Dreger, Graves, Harmsen, Hartzell, Larsen, Zoback, et al., 2008). The simulations demonstrated consistent amplification of shaking associated with sedimentary basins, such as the Cupertino Basin west of San Jose, the Cotati and Windsor basins under Santa Rosa, the Livermore basin, and the Great Valley. The pattern of shaking within the San Francisco Bay urban region also displayed a strong sensitivity to the hypocenter with significantly stronger shaking for north-to-south rupture compared to south-to-north rupture. Although we attribute some of this observation to the location of a majority of the urban area lying south of San Francisco and Oakland, the geologic structure also appears to play a significant role. Having characterized the ground motions generated by ruptures along the San Andreas fault on the west side of San Francisco Bay, in this study we focus on ground motions generated by ruptures of the Hayward, Rodgers Creek, and Calaveras faults on the east side of San Francisco Bay.

Larsen et al. (2000) simulated long-period ground motions for 20 \( M_w \) 7.0 scenario events on the Hayward fault using a simple three-dimensional (3D) seismic velocity model. They found high amplitude motions in the sedimentary basins, such as the San Pablo basin underneath San Pablo Bay, the Evergreen basin in the Santa Clara Valley, and the Livermore basin in the Livermore Valley. Harmsen et al. (2008) improved the characterization of ground motions from large earthquakes on the Hayward fault by studying the long-period (\( T > 1.0 \) s) ground motions in the Santa Clara Valley from six scenario events that included variations in the magnitude, hypocenter, rupture speed, and seismic velocity model. These six scenarios were part of a large suite of 20 scenarios with ruptures on other faults in the region. The six Hayward earthquakes involved rupture of 57 km of the southern portion of the Hayward fault in \( M_w \) 6.9 and \( M_w \) 7.0 events. Harmsen et al. (2008) noted a strong dependence of the peak ground velocity on the hypocenter and significant basin amplification, particularly in the Evergreen basin east of San Jose.

Whereas Harmsen et al. (2008) focus on the variability of ground motions in the Santa Clara Valley for a variety of earthquake sources, including six Hayward fault ruptures, in this study we focus on the ground motions throughout the San Francisco Bay area for a larger suite of Hayward fault scenarios, some of which include rupture on additional faults. Harmsen et al. (2008) did not include the effects of large regions with creep (aseismic slip) on the coseismic slip distribution but did prescribe a uniform tapering in slip at depths above 5 km to limit the contribution of shallow slip to the ground motions. As described in Aagaard et al. (2010), we include two different approaches that account for the effects of creep on the coseismic slip distribution. Thus, we build upon the efforts of Harmsen et al. (2008) in our suite of 39 scenarios by including variations in the rupture length, slip distribution, hypocenter, rise time, rupture speed, and how creep affects the coseismic slip distribution.

Wave Propagation Codes

Simulation of ground motions for the 39 scenario events involved five different ground-motion modeling groups, Aagaard, Graves, Larsen, Ma, and Rodgers, each using a different wave propagation code. As we will discuss later, each group computed ground motions for two well-recorded moderate earthquakes and a common subset of the scenario earthquakes. These simulations demonstrate consistency among the modeling groups and permit tying together results from the different modeling groups’ exploration of a subset of the scenarios. The codes employed by Aagaard, Graves, Larsen, and Rodgers (Larsen and Schultz, 1995; Graves, 1996; Aagaard et al., 2001; Nilsson et al., 2007) were used in studies of the 1906 earthquake (Aagaard, Brocher, Dolenc, Dreger, Graves, Harmsen, Hartzell, Larsen, McCandless, et al., 2008) and the 1989 Loma Prieta earthquake (Aagaard, Brocher, Dolenc, Dreger, Graves, Harmsen, Hartzell, Larsen, Zoback, et al., 2008). Since the simulations of the 1906 earthquake, the code used by Rodgers has been improved to include topography using a curvilinear grid approach (Apelö and Petersson, 2008). Ma previously applied his code (Ma and Liu, 2006) to study wave propagation in the 3D heterogeneous structure of southern California (Ma et al., 2008) and to examine the effects of topography (Ma et al., 2007).

Table 1 and Figure 1 summarize the modeling domains and features of the wave propagation codes used in this study. The domains generally span the region covered by the detailed portion of the U.S. Geological Survey’s (USGS) Bay Area Velocity Model version 08.3.0, which we discuss in the next section. Graves and Larsen resolve waves with periods greater than 1.0 s, whereas Aagaard, Ma, and Rodgers resolve periods greater than 2.0 s. Each of the modeling groups imposes a minimum shear-wave speed of 500 m/s to 700 m/s in their simulations. These choices for the minimum period and shear-wave speed reflect the computational resources available to each group and the overall efficiency of the wave propagation implementation. The similar choices for the minimum shear-wave speed contribute to the strong consistency among the modeling groups.

The Graves and Larsen modeling codes incorporate the effects of anelastic attenuation through the use of memory variables to approximate a frequency independent \( Q \) over a wide frequency band (Day and Minster, 1984; Robertson et al., 1994; Day and Bradley, 2001). The other modeling groups do not include anelastic attenuation. With the exception of the more distant sites in the Great Valley as discussed
Table 1
Wave Propagation Codes and Modeling Domains

<table>
<thead>
<tr>
<th>Modeling Domains and Features of Wave Propagation Codes</th>
<th>Aagaard</th>
<th>Graves</th>
<th>Larsen</th>
<th>Ma</th>
<th>Rodgers</th>
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<td>-123.4900, 38.2100</td>
<td>-123.3592, 38.1664</td>
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<td>USGS version 08.3.0</td>
<td>USGS version 08.3.0</td>
<td>USGS version 08.3.0</td>
<td>USGS version 08.3.0</td>
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<td>$Q_s = 2Q_p$</td>
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<td>USGS version 08.3.0</td>
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<td>point sources</td>
<td>point sources</td>
<td>point sources</td>
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<td>10,500</td>
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<td>3D geologic model</td>
<td>3D geologic model</td>
<td>3D geologic model</td>
<td>3D geologic model</td>
</tr>
</tbody>
</table>

The corners of the bounding boxes of each domain are given in longitude and latitude (WGS84 horizontal datum).
later, we find that our results are quite insensitive to the inclusion of anelastic attenuation, or the specific form of the \( Q \) model.

The seismic velocity model includes topography, so the ground-motion simulations either explicitly include topography or warp the seismic velocity model by squashing the topography into a flat planar surface. In squashing, the uppermost portion of the Earth (usually a few kilometers thick) is deformed in the vertical direction so that the free surface is flat and aligned at a reference elevation, typically the bias. However, several iterations were required to fine-tune the velocities in the sedimentary units and to remove

\[
D(t) = \begin{cases} 
D_f C_n \left( 0.7t - 0.7 \frac{\pi}{\tau_1} \sin \frac{\pi t}{\tau_1} - 1.2 \frac{\pi}{\tau_1} \cos \frac{\pi t}{\tau_1} \right) & 0 \leq t < \tau_1 \\
D_f C_n \left( 1.0t - 0.7 \frac{\pi}{\tau_1} \sin \frac{\pi t}{\tau_1} + 0.3 \frac{\pi}{\tau_1} \sin \frac{\pi (t - \tau_1)}{\tau_1} + 1.2 \frac{\pi}{\tau_1} \tau_1 - 0.3 \tau_1 \right) & \tau_1 \leq t < 2\tau_1, \\
D_f C_n \left( 0.3t + 0.3 \frac{\pi}{\tau_1} \sin \frac{\pi (t - \tau_1)}{\tau_1} + 1.1 \tau_1 \right) & 2\tau_1 \leq t < \tau 
\end{cases}
\]

Here, \( D(t) \) is the slip at time \( t \), \( D_f \) is the final slip, \( \tau_{95} \) is the rise time (as measured by the time it takes for 95% of the slip to occur), and \( \tau, \tau_1, \tau_2, \) and \( C_n \) are constants.

Seismic Velocity Model

For this study, we updated version 05.1.0 of the USGS Bay Area Velocity Model that we used for calculating the ground motions for the 1906 earthquake (Aagaard, Brocher, Dolenc, Dreger, Graves, Harmsen, Hartzell, Larsen, Zoback, et al., 2008). Waveform modeling (30 \( s > T > 4 \) s) of moderate earthquakes (\( M_w 4-5 \)) in the San Francisco Bay region showed that version 05.1.0 of the model predicted surface waves arriving about 5% faster than observed (Rodgers et al., 2008). Analysis of arrival times for small to moderate earthquakes and refraction shots confirmed a bias of about 5% too fast for both dilatational-wave speed (\( V_P \)) and the shear-wave speed (\( V_S \)) in version 05.1.0 of the model (Douglas Dreger, personal comm., 2007). To quantify the discrepancy in wave speeds, we compared the wave speeds for each geologic unit in the seismic velocity model with the corresponding wave speeds from the Thurber et al. (2007) seismic tomographic model. We updated the relations between seismic wave speed and depth to improve the fit to the wave speeds and gradients in the Thurber et al. (2007) tomography model at depths below several kilometers. We continued to use sonic well log data (Brocher, 2005) as the target at shallower depths where it provides tighter constraints.

Initial changes to the seismic velocity model indeed appropriately reduced the travel times and significantly reduced the bias. However, several iterations were required to fine-tune the velocities in the sedimentary units and to remove
Testing with Moderate Earthquakes

In order to examine the ability of our long-period ground-motion simulations to reproduce recorded waveforms throughout the San Francisco Bay area for earthquakes on the Hayward fault, we simulated two recent moderate earthquakes: the 31 October 2007 $M_s$ 5.45 Alum Rock event and the 20 July 2007 $M_s$ 4.18 Oakland event. These events occurred on the Hayward fault and were recorded throughout the Bay Area.

4. For granitic rocks above 2 km depth, the velocities honor borehole sonic data (Brocher, 2005), which is considered more reliable, whereas below 2 km we honored the gradient observed in the Thurber et al. (2007) tomographic model.

5. For Great Valley sequence units below 3 km, the velocities honor the Thurber et al. (2007) tomographic model.

6. For Tertiary-Cenozoic sedimentary units at depths above 750 m, we attempted to honor the $V_p$ relation of Hartzell et al. (2006) for the Cupertino basin; otherwise, above 4 km depth, the wave speeds honor the sonic well log data (Brocher, 2005), which are considered more reliable than tomography at these depths; below 4 km the wave speeds honor the Thurber et al. (2007) tomographic model.

7. For sedimentary units in the La Honda Basin, wave speeds were increased 10–20% except at depths greater than 6 km, where it remains unchanged:

$$V_p = \begin{cases} 
2.50 + 0.625d & \text{for } 0 < d \leq 4 \text{ km} \\
5.00 + 0.200(d - 4) & \text{for } 4 \text{ km} < d \leq 6 \text{ km} \\
5.40 & \text{for } d > 6 \text{ km}
\end{cases} \tag{7}$$

where $V_p$ is in km/s and $d$ is depth in km. The sonic log for the Champlin Petroleum borehole indicates a $V_p$ of 3.8 km/s at 1 km, a $V_p$ of 4.2 km/s at 2 km, and a $V_p$ of 4.7 at 3 km (Brocher et al., 1997). This borehole sampled a steeply dipping section of Butano sandstone for its entire length, so this sonic log may not be representative of the basin as a whole. Williams et al. (1999) interpret a seismic refraction line along the La Honda Basin with the upper 3 km of the basin modeled with a $V_p$ of 4 km/s. The relations given by equation (7) yield slightly faster wave speeds than those for the Great Valley sequence units to a depth of 4 km. Between 4 and 6 km, the $V_p$ lies between that for the Great Valley sequence units and the new Tertiary-Cenozoic sedimentary unit relation from the Cupertino Basin.

8. For Cenozoic sedimentary rocks near Half Moon Bay, we applied separate relations compared to other older Tertiary deposits in order to prevent very strong amplification that does not fit observations from the Loma Prieta earthquake.

9. We updated the attenuation quality factors, $Q_p$ and $Q_s$, to the values given in Brocher (2008).

Figure 1. Bounding boxes of the domains used by the five ground-motion modeling groups and the detailed portion of the USGS Bay Area Velocity Model version 08.3.0. (dotted black box). The thick lines show the extent of rupture on the surface traces in our scenario earthquakes on the Hayward, Rodgers Creek, and Calaveras faults. The color version of this figure is available only in the electronic edition.

- a spurious velocity contrast that had been introduced across the San Andreas fault in the La Honda Basin, which was inconsistent with 1989 Loma Prieta travel-time observations. In the following paragraphs we summarize the major changes made to the velocity model to produce the updated 08.3.0 version used in this study. We applied these changes prior to simulating the two moderate earthquakes and scenario events discussed in the following sections. The electronic supplement contains the relations between wave speed and depth in version 08.3.0, which supersede the relations given in Brocher (2008).

1. For the upper mantle, we decreased the wave speeds by 2.5% to better match the Thurber et al. (2007) tomographic model, and we added a small positive gradient based on seismic refraction results.

2. For the mafic lower crust/Great Valley ophiolite, we reduced $V_p$ in the upper 18 km by as much as 1 km/s and increased it below 18 km depth to a maximum of about 7.4 km/s.

3. For Franciscan units, we reduced $V_p$ at depths of 1–3 km by 1–2%, at depths of 4–10 km by 9%, and we increase $V_p$ below 20 km by about 3%.

4. For granitic rocks above 2 km depth, the velocities honor borehole sonic data (Brocher, 2005), which is considered more reliable, whereas below 2 km we honored the gradient observed in the Thurber et al. (2007) tomographic model.

5. For Great Valley sequence units below 3 km, the velocities honor the Thurber et al. (2007) tomographic model.

6. For Tertiary-Cenozoic sedimentary units at depths above 750 m, we attempted to honor the $V_p$ relation of Hartzell et al. (2006) for the Cupertino basin; otherwise, above 4 km depth, the wave speeds honor the sonic well log data (Brocher, 2005), which are considered more reliable than tomography at these depths; below 4 km the wave speeds honor the Thurber et al. (2007) tomographic model.

7. For sedimentary units in the La Honda Basin, wave speeds were increased 10–20% except at depths greater than 6 km, where it remains unchanged:

$$V_p = \begin{cases} 
2.50 + 0.625d & \text{for } 0 < d \leq 4 \text{ km} \\
5.00 + 0.200(d - 4) & \text{for } 4 \text{ km} < d \leq 6 \text{ km} \\
5.40 & \text{for } d > 6 \text{ km}
\end{cases} \tag{7}$$

where $V_p$ is in km/s and $d$ is depth in km. The sonic log for the Champlin Petroleum borehole indicates a $V_p$ of 3.8 km/s at 1 km, a $V_p$ of 4.2 km/s at 2 km, and a $V_p$ of 4.7 at 3 km (Brocher et al., 1997). This borehole sampled a steeply dipping section of Butano sandstone for its entire length, so this sonic log may not be representative of the basin as a whole. Williams et al. (1999) interpret a seismic refraction line along the La Honda Basin with the upper 3 km of the basin modeled with a $V_p$ of 4 km/s. The relations given by equation (7) yield slightly faster wave speeds than those for the Great Valley sequence units to a depth of 4 km. Between 4 and 6 km, the $V_p$ lies between that for the Great Valley sequence units and the new Tertiary-Cenozoic sedimentary unit relation from the Cupertino Basin.

8. For Cenozoic sedimentary rocks near Half Moon Bay, we applied separate relations compared to other older Tertiary deposits in order to prevent very strong amplification that does not fit observations from the Loma Prieta earthquake.

9. We updated the attenuation quality factors, $Q_p$ and $Q_s$, to the values given in Brocher (2008).
the area of interest for this study (Fig. 2). The Alum Rock event samples the crust around the southern end of the Hayward Fault, while the Oakland event samples the crust around the northern end of the fault. The ruptures we consider in this study span the region between these events.

These moderate earthquakes are much smaller than our scenario events so that the event rupture processes are much simpler, which allows us to evaluate the seismic velocity model and path propagation effects with relatively little bias due to source processes (e.g., spatial and temporal evolution of slip, rupture speed, and rise time). These events are large enough that some finite-source effects are evident, so we do use simple finite-source models. Furthermore, these events were recorded with a high signal-to-noise ratio and occur more frequently than large events. These simulations complement those done previously by Aagaard, Brocher, Dolenc, Dreger, Graves, Harmsen, Hartzell, Larsen, Zoback, et al. (2008) for the 1989 M\textsubscript{w} 6.9 Loma Prieta earthquake, which occurred along the San Andreas fault on the west side of San Francisco Bay and included 3D source and basin effects. In simulating these two moderate earthquakes, we are most interested in evaluating our ability to match arrival times, amplitudes, and dominant features of the waveforms. Consequently, we use only deterministic long-period (T > 1–2 s) simulations and compare against recorded waveforms; we do not compare the synthetics against the Next Generation Attenuation (NGA) ground-motion prediction equations, which are limited to peak amplitudes.

The recorded motions for the M\textsubscript{w} 5.45 Alum Rock earthquake show evidence of a finite-source process, so we construct a simple finite-source model for this event using the method of Dreger and Kaverina (2000). We compute Green’s functions for the three-component displacement waveforms recorded at Berkeley Digital Seismic Network stations, BKS, CMB, PKD, and KCC with the GIL7 velocity model (Dreger and Romanowicz, 1994; Pasyanos et al., 1996). This 1D velocity model is commonly used in earthquake location and moment tensor analyses in the San Francisco Bay area. The Green’s functions are convolved with a 0.3 s triangular slip velocity function. Both the observed waveforms and Green’s functions are band-pass filtered between 0.01 and 0.3 Hz using a two-pole acausal Butterworth filter. Assuming a rupture velocity of 2.8 km/s (approximately 0.80 V\textsubscript{s}) yields a peak slip of 0.17 m, and a 7 km by 4 km (length by width) rupture patch. The slip extends down-dip and to the southeast of the hypocenter (9 km depth) and was located between 9 and 13 km depth. The directivity associated with this rupture model derived from the waveform modeling is consistent with the directivity inferred from the spatial variation in peak horizontal ground accelerations and peak horizontal velocities (Seekins and Boatwright, 2007). This slip patch was projected onto the Hayward–Calaveras fault surface, which results in a close match between the strike and dip of the Berkeley focal mechanism and the strike and dip of the fault surface.

Each modeling group computed the waveforms at 185 stations for this event using the rupture model described earlier and the USGS Bay Area Velocity Model version 08.3.0. In general, the simulations capture the main features of the observed ground motions and are consistent among modeling groups. Figure 2 shows the two earthquake locations and the stations for which we compare ground-motion time histories. Figure 3 illustrates the variation in shaking across the Santa Clara Valley that is accurately captured by the synthetic waveforms at stations CHR, Q32, H30, and CDOB for T > 2.0 s. The electronic supplement includes plots of the synthetic and observed waveforms low-pass filtered with corner frequencies of 0.25 Hz (T > 4.0 s) and 0.5 Hz (T > 2.0 s) for all 185 stations. For station CHR the observed horizontal velocities exceed the synthetic velocities by about 60%. Because this station sits only about 6.7 km from the epicenter, we attribute this discrepancy to our rupture model that simplifies the slip distribution and rupture propagation. For example, the rupture model may underestimate the amount of up-dip directivity.

The waveforms at station Q32, which sits in the Evergreen basin 10.8 km southwest of the epicenter, exhibit greater complexity than those at station CHR as a result of the complex basin response. The horizontal components are dominated by about 10 s of larger amplitude motion (peak velocities exceed 10 mm/s) followed by another

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**Figure 2.** Map of the San Francisco Bay Area showing the double couple focal mechanisms for the two moderate earthquakes we modeled for validation: 31 October 2007 Alum Rock (M\textsubscript{w} 5.45) in the dark color and 20 July 2007 Oakland (M\textsubscript{w} 4.18) in the light color. Also shown are seismic stations (triangles, color coded by event) used for waveform comparisons in Figure 3 and Figure 4. The color version of this figure is available only in the electronic edition.
10 s of more moderate motion (peak velocities exceed 5 mm/s). The synthetic waveforms reproduce the variation in shaking but fail to capture details in the waveforms at periods near 2.0 s. This discrepancy may be associated with surface waves generated at the basin edge, which are sensitive to the contrast in shear modulus between the Evergreen basin and the adjoining Great Valley sequence. The synthetics closely agree with the observed waveforms at periods of 4.0 s and greater (the comparison of ground-motion intensities from Hayward synthetics to NGA results is available in the electronic supplement to this paper). Moving west across the Santa Clara Valley, the seismic waves leave the Evergreen basin and, near the western edge of the valley, enter the Cupertino basin. At station H30, which sits 25.6 km south-west of the epicenter, Love waves (north–south component) dominate the ground motions with over 20 s of significant motion (peak velocities are near 5 mm/s). The synthetics reproduce the onset, amplitude, and duration of the motion; however, the later surface wave arrivals in the synthetics are slightly delayed relative to the observations. The synthetics at periods of 2.0 s and longer match the observations nearly as well as the synthetics at periods of 4.0 s and longer. In addition to the mismatch in later surface wave arrivals, the primary discrepancy is that the synthetics contain one additional cycle of relatively large amplitude motion associated with the surface waves compared to the observations.

The southward directivity causes the seismic waves radiated north of the epicenter to arrive over a longer time window. As a result, the geologic structures tend to dominate the character of the shaking with less influence from the source. The southward directivity also yields lower amplitude ground motions north of the epicenter. In this region the synthetics generally reproduce the amplitude and duration of motion at periods of 4.0 s and longer but struggle to reproduce the waveforms at periods approaching 2.0 s. For example, at station CDOB in Livermore, 33.2 km north of the epicenter, reverberations within the Livermore sedimentary basin increases the duration of the shaking.

Figure 3. Observed (top trace) and simulated velocity waveforms at four stations in the San Francisco Bay area for the 31 October 2007 Alum Rock earthquake. The velocity waveforms have been low-pass filtered to a common bandwidth of $T > 2.0$ s using two passes of a two-pole Butterworth filter with a corner frequency of 0.5 Hz. The color version of this figure is available only in the electronic edition.
The synthetics exhibit longer period motions than the observations even at periods of 4.0 s and longer, suggesting that the seismic velocity model does not adequately capture the geometry of the basin and/or variation of physical properties in and around the basin. Thus, the waveforms are not as accurate in this region as they are in the Santa Clara Valley.

In simulating the 20 July 2007 $M_w$ 4.18 Oakland event, we assess the ability of our simulations to reproduce the ground motions throughout the San Francisco Bay area to a source on the northern portion of the Hayward fault. In order to construct a finite-source model for this event, we employ empirical Green’s function deconvolution of the records of a nearby $M_w$ 2.7 event at two borehole stations, CMSB and SM2B. The observed moment-rate function at CMSB has much narrower pulse widths than SM2B, 0.14 s compared to 0.93 s. Based on the moment tensor of the event from the Berkeley Seismological Laboratory, the azimuth to CMSB and SM2B are 322° and 55°, respectively; these stations sit nearly perfectly in the along-strike direction and the fault-perpendicular directions. Assuming a rupture speed of 2.8 km/s, we determine the rupture length, $l$, from the fault-perpendicular directivity relationship,

$$l = \tau_p 2.8 \text{ km/s} = 2.6 \text{ km}, \quad (8)$$

where $\tau_p$ is the pulse width. Similar to the Alum Rock event, the directivity inferred from waveform modeling is consistent with that inferred from the spatial variation in peak horizontal accelerations and peak horizontal velocities (Seekins and Boatwright, 2007). We choose a rupture width of 0.8 km based on a stress drop of 1.0 MPa and the vertical strike-slip fault relationship between stress drop, scalar moment, and source dimension. These rupture dimensions yield a slip of 0.09 m with a rake of 168° from the Berkeley Seismological Laboratory moment tensor solution. As in the case of the Alum Rock event, we project this slip patch onto the nonplanar geometry of the Hayward fault surface.

Figure 4 displays the observed and synthetic velocity waveforms at four sites (2190, BRK, BRIB, and CTA) for this event. The ground motions at these stations illustrate the agreement between the synthetics and observed motions at locations with different source and wave propagation path characteristics. The waveforms have been low-pass filtered with a corner frequency of 0.5 Hz ($T > 2.0$ s). The electronic supplement contains plots of the observed and synthetic velocity waveforms for 115 stations low-pass filtered with corner frequencies of 0.5 Hz ($T > 2.0$ s) and 0.25 Hz ($T > 4.0$ s). Station BRK lies 9.7 km northwest of the epicenter and less than a kilometer west of the Hayward fault and station CMSB (which we used to construct the source model). At station BRK, the synthetics reproduce the simple velocity pulse present in the observed waveforms. Although this site displays the best match between the synthetics and observed waveforms, the synthetics at several other sites north of the epicenter also provide a good fit to the observed amplitude and duration of shaking, especially at periods of 4.0 s and longer.

Station 2190 sits 5.8 km south of the epicenter and less than 2 km from the edge of San Francisco Bay. The synthetics capture the relative amplitude with the greatest motion associated with the shear-wave arrival but underpredict the amplitude and complexity, especially in the east–west component. For periods greater than 2.0 s, the observed peak velocity for the north–south component exceeds 1 mm/s, whereas the peak velocity for the synthetics are all about 0.5 mm/s. At longer periods ($T > 4.0$ s), the discrepancy in amplitude between the observed and synthetic waveforms becomes smaller. We attribute this discrepancy to insufficient amplification in the simulations resulting from a combination of the minimum shear-wave speed imposed in the simulations (which artificially stiffens the soft, near-surface sediments) and complexity in the geologic structure not included in the seismic velocity model. For example, this region south of the epicenter between the San Francisco Bay and the Hayward fault may include locally softer or deeper alluvial sediments than regions north of the epicenter, where the synthetics closely follow the observed motions. The consistency among the modeling groups, however, remains excellent considering the variations in amplitude associated with different minimum shear-wave speeds.

Shifting our focus to locations east of the Hayward fault, we find greater complexity in the observed waveforms, especially at shorter periods ($T > 2.0$ s). The simulations have difficulty reproducing this greater complexity; for example, at station BRIB (12.4 km northeast of the epicenter), the observed waveforms include a sharp initial arrival on the east–west component followed by an additional 15 s of shaking. The synthetics also include a sharp arrival on the east–west component, but the arrival is more than a second later than the observed arrival. The synthetics display a similar overall duration of motion but do not replicate the details in the observed waveforms. Furthermore, we find slightly less consistency among the modeling groups, but this is mainly confined to the vertical component. The greater complexity in the observed waveforms and delayed arrival of the shear wave in the synthetics suggests that the average shear-wave speed east of the Hayward fault may be too slow in the seismic velocity model and the geologic structure may be significantly more complex than what is described by the seismic velocity model.

Further northeast at station CTA, 28.2 km from the epicenter, surface waves dominate the waveforms. The synthetics are able to replicate the amplitude and approximate duration of the surface waves but do a poor job of matching the details of the waveforms. In contrast to station BRIB, the arrival time of the shear wave in the synthetics for station CTA match the observed arrival time.

Accuracy of the Seismic Velocity Model

Examination of the waveforms across the region for these two moderate earthquakes suggests that the seismic
velocity model has the greatest accuracy in the Santa Clara Valley; in this region, the synthetics match detailed features of the observed waveforms, such as phase arrival times, duration, and waveform shape. In many other areas, the synthetics are only able to closely match the initial phase arrival and the overall envelope of the waveform. Between the San Francisco Bay and the Hayward fault near Oakland, the simulations fail to capture amplification at periods close to 2.0 s. Examination of the near-surface physical properties in the seismic velocity model in this area indicates that the minimum shear-wave speed imposed in the simulations does not explain adequately this discrepancy, but instead the thickness of the unconsolidated Quaternary deposits in the seismic velocity model may need adjustment. Frankel and Carver (2009) noted similar deficiencies in the seismic velocity model in the region further south between the San Francisco Bay and the Hayward fault using 3D simulations of the ground motions for the 6 September 2008 $M_w$ 4.0 Alamo earthquake. Thus, ground-motion simulations may underpredict the amplitude of the shaking in this region.

East of the Hayward fault, our simulations are only able to reproduce the amplitude and duration of the shaking; the seismic velocity model lacks sufficient detail to reproduce detailed features of the waveforms. The geologic structure east of the Hayward fault has yielded a more complex volume of Cenozoic and Mesozoic rocks compared to the volume dominated by Mesozoic rocks to the west of the fault (Graymer, 2000), so it is not surprising that the seismic velocity model needs to incorporate greater detail in this region. Hence, simulated ground motions from this region east of the Hayward fault are less accurate and are limited to an estimate of the amplitude and duration of shaking. This means they are less suitable for use as input in analyses where the details of the waveforms may be important. Ongoing efforts to refine the seismic velocity model should focus on this region east of the Hayward fault, and validation

Figure 4. Observed (top trace) and simulated velocity waveforms at four stations in the San Francisco Bay area for the 20 July 2007 Oakland earthquake. The velocity waveforms have been low-pass filtered to a common bandwidth of $T > 2.0$ s using two passes of a two-pole Butterworth filter with a corner frequency of 0.5 Hz. The color version of this figure is available only in the electronic edition.
should continue using other moderate earthquakes and future large earthquakes throughout the San Francisco Bay region.

In summary, these comparisons show that the USGS Bay Area Velocity Model version 08.3.0 reproduces important 3D wave propagation features of the observed ground motions throughout the San Francisco Bay region at periods down to 2–4 s for two events along the Hayward fault. Rodgers et al. (2008) drew similar conclusions using version 05.1.0 of the model but noted a persistent bias in the wave speeds as discussed earlier. Kim et al. (2010) demonstrated that version 08.3.0 reduces the average bias in arrival times compared with version 05.1.0 while maintaining a good fit to the peak horizontal velocities over five orders of magnitude for moderate earthquakes. Furthermore, the consistency in ground motions among the five modeling groups using different numerical methods and implementations implies that we can use the results from any of the modeling groups to characterize the ground shaking in our large scenario events. The differences in ground motion associated with imposing slightly different minimum shear-wave speeds and including or excluding topography and attenuation are relatively small compared to the agreement in the amplitude, duration, and shape of the waveforms.

Scenario Earthquakes

Table 2 summarizes the 39 events in our suite of earthquake scenarios, and Figure 5 shows the rupture lengths and epicenters. Aagaard et al. (2010) discuss each of the earthquake source parameters and the rationale for the choice of variation in detail. In this section, we discuss the general trends in the ground motions and the sensitivity in the shaking to variation of the earthquake source parameters for the long-period (T > 1–2 s) simulations. The following section discusses the broadband (T > 0.1 s) simulations in the context of the 1868 earthquake and the NGA ground-motion prediction equations.

Base Cases

With the modeling groups each examining a different subset of the suite of scenario earthquakes, we first demonstrate the consistency of the shaking intensity and velocity waveforms among the modeling groups for one of the scenarios. This extends the consistency we found for the Mw 5.45 Alum Rock and Mw 4.18 Oakland earthquakes to our larger scenarios, which have more complex rupture models. Figure 6 shows maps of the modified Mercalli intensity (MMI) for scenario HS G01 HypoH from the Graves, Larsen, Ma, and Rodgers modeling groups and residuals with respect to Aagaard’s MMI values; a map of MMI for Aagaard’s simulation is shown in Figure 7.

We compute the MMI values from our simulations using relationships between peak ground velocity (PGV), peak ground acceleration (PGA), and MMI developed for ShakeMap (Wald et al., 2005). We use this relationship for consistency with the ShakeMap generated for the 1868 earthquake by Boatwright and Bundock (2008). Refining this relationship with improved accuracy for large motions at long periods (T > 2 s) is an area of active research (Cua et al., 2010). For the broadband simulations,

\[
\text{MMI} = \begin{cases} 
\frac{1}{2}(\text{VII} - \text{MMI}_{\text{PGA}})\text{MMI}_{\text{PGA}} + (\text{MMI}_{\text{PGA}} - \text{V})\text{MMI}_{\text{PGV}} & \text{if } \text{MMI} < \text{V} \\
\text{MMI}_{\text{PGA}} & \text{if } \text{V} \leq \text{MMI} < \text{VII}, \\
\text{MMI}_{\text{PGV}} & \text{if } \text{MMI} \geq \text{VII}
\end{cases}
\]

(9)

\[
\text{MMI}_{\text{PGA}} = \begin{cases} 
2.20 \log(\text{PGA}) + 1.00 & \text{if } \text{MMI} < \text{V} \\
3.66 \log(\text{PGA}) - 1.66 & \text{if } \text{MMI} \geq \text{V}
\end{cases}
\]

(10)

and

\[
\text{MMI}_{\text{PGV}} = \begin{cases} 
2.10 \log(\text{PGV}) + 3.40 & \text{if } \text{MMI} < \text{V} \\
3.47 \log(\text{PGV}) + 2.35 & \text{if } \text{MMI} \geq \text{V}
\end{cases}
\]

(11)

where peak ground acceleration is in gs and peak ground velocity is in cm/s. For the long-period simulations, we only use the PGV to MMI relation (equation 11) because the simulations have artificially low PGA values due to the absence of short-period energy. In fact, because the intensities are above V at most locations, the intensities are relatively insensitive to the level of PGA. For this comparison among the modeling groups, we use the bandwidth of the deterministic long-period simulations (e.g., T > 1.0 s for Graves’s broadband simulations). This permits evaluation of the relative importance of including energy down to periods of 1.0 s, while simultaneously assessing the consistency among the modeling groups.

The amplitude and spatial variation of the shaking intensities are very similar among the groups with mean residuals less than 0.20 MMI units. The standard deviations in the residuals for Graves’s and Larsen’s simulations are about 0.50 MMI units because both of these simulations include periods down to 1.0 s, compared to Aagaard’s simulations, which include periods down to only 2.0 s. The standard deviations in the residuals for Ma’s and Rodgers’s simulations are smaller, with values of about 0.30 MMI units, because they use the same minimum period of 2.0 s as Aagaard’s simulations. The largest discrepancies among the modeling groups...
arises in the Great Valley east of the San Francisco Bay. The longer propagation distances for this region coupled with greater attenuation results in lower amplitudes of shaking for the Larsen and Graves modeling groups, which include intrinsic attenuation. Similar levels of agreement are obtained for the other scenarios, and the electronic supplement contains plots comparing the shaking intensities among the modeling groups for bilateral rupture of the Hayward South + North rupture length.

Velocity waveforms at sites throughout the San Francisco Bay area (plots for three sites are shown in Figure 8, with plots for 80 sites available in the electronic supplement) illustrate that, in addition to agreeing in amplitude, the modeling groups generate waveforms with the same features. The first arrivals are nearly identical. We find good agreement in the amplitude and duration of most later arrivals, but inclusion/exclusion of attenuation and topography and different minimum shear-wave speeds leads to small discrepancies in the arrival times for the surface waves. Larsen’s waveforms include large secondary arrivals at some locations, which appears to be related to simulating ruptures with period near the expression for the rise time, \( t_{\text{rise}} = \frac{C_s}{V_s} \) and the labels Tr10u and Tr20u correspond to nominally uniform rise times of 1.0 s and 2.0 s. The rupture speed labels Vr92, Vr82, and Vr141 correspond to the correlations between rupture speed and slip. The maximum local rupture speeds for Vr92, Vr82, and Vr141 are 0.92 \( V_s \), 0.82 \( V_s \), and \( \sqrt{2} V_s \), respectively. The modeling groups are Aagaard (A), Graves (G), Larsen (L), Ma (M), and Rodgers (R). We form the scenario names from abbreviations of the parameters but do not include parameters with significant redundancy (which are shown in italics).

### Table 2

<table>
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</tbody>
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Figure 5 shows the rupture lengths and epicenters. The rise time labels Tr10, Tr15, and Tr20 correspond to values of 1.0, 1.5, and 2.0 for C in the expression for the rise time, \( t_{\text{rise}} = \frac{C_s}{V_s} \) and the labels Tr10u and Tr20u correspond to nominally uniform rise times of 1.0 s and 2.0 s. The rupture speed labels Vr92, Vr82, and Vr141 correspond to the correlations between rupture speed and slip. The maximum local rupture speeds for Vr92, Vr82, and Vr141 are 0.92 \( V_s \), 0.82 \( V_s \), and \( \sqrt{2} V_s \), respectively. The modeling groups are Aagaard (A), Graves (G), Larsen (L), Ma (M), and Rodgers (R). We form the scenario names from abbreviations of the parameters but do not include parameters with significant redundancy (which are shown in italics).
primarily at longer periods associated with $M_w$ 7.8 earthquakes on the northern San Andreas fault (Aagaard, Brocher, Dolenc, Dreger, Graves, Harmsen, Hartzell, Larsen, McCandless, et al., 2008).

The close agreement in the amplitude, duration, and features of the waveforms across the modeling groups means that we can use scenarios HS G01 HypoH and HS + HN G04 HypoO to tie the results of the different modeling groups together. We use different groups to characterize the sensitivity of the ground motions to different source parameters, with Aagaard’s simulations for rupture length and slip distribution, Graves’s simulations for rupture length and hypocenter at broadband frequencies, Larsen’s simulations for rise time, and Ma’s simulations for rupture speed. While the results of each modeling group provide independent characterization of the sensitivity of the ground motions to a single parameter or a small subset of the parameters, we rely on the consistency among the modeling groups to establish the relative sensitivity of the ground motions to variation of the rupture length, slip distribution, hypocenter, rise time, and rupture speed.

Rupture Length, Slip Distribution, and Hypocenter

Aagaard’s ($T > 2$ s) simulations include 25 earthquake scenarios that provide a comprehensive view of the ground motions for the five different rupture lengths with one to three hypocenters per rupture length and two to three slip distributions per hypocenter. The electronic supplement for part I includes plots of the slip distribution and rupture time in each scenario. The rupture length (which correlates with magnitude) exerts the greatest influence on the amplitude of shaking, with longer rupture lengths (larger magnitude earthquakes) generating stronger shaking as evident in Figure 7 and Figure 9. Scenario HS G01 HypoH causes shaking greater than or equal to MMI VII over about 24% of the San Francisco Bay urban area. MMI VII corresponds to the approximate shaking intensity when modern structures begin to suffer damage (Wald et al., 2005). The fraction of the urban area experiencing MMI VII increases to 33% in scenario CC + HS G03 HypoH, 58% in scenario HS + HN G04 HypoO, and 60% in scenario HS + HN + RC G06 HypoSPB. We attribute these high levels of shaking experienced by such large fractions of the urban area to the Hayward fault running directly through the urban corridor along the eastern edge of the San Francisco Bay.

The slip distribution has less influence on the overall distribution of shaking, but changes in location of large slip patches (see the electronic supplement of part I [Aagaard et al., 2010] for plots of the slip distributions) affect the shaking close to the rupture. The features in the maps of MMI in Figure 7 closely resemble each other for corresponding rupture lengths and hypocenters with different slip distributions (one uses a background slip distribution with a vertical gradient in slip to account for creep paired with one stochastic distribution, and the other uses a background slip distribution with the slip-predictable approach for accounting for creep paired with another stochastic distribution). These same observations hold for the other hypocenters for the Hayward South, Hayward South + North, and Central Calaveras + Hayward South rupture lengths and changing the random seed in the stochastic slip distribution with the same vertical gradient in slip for the Hayward South rupture length (see the electronic supplement to this article for plots of the shaking intensities and velocity waveforms for these scenarios with other hypocenters, rupture lengths, and slip distributions).

For the Hayward South, Hayward South + North, and Central Calaveras + Hayward South rupture lengths, we consider three hypocenter locations. This yields cases with north-to-south rupture, bilateral rupture, and south-to-north rupture. As indicated in Table 2, the various rupture lengths use different slip distributions; this means the locations of large and small slip differ. As shown in Figure 10, rupture directivity along the strike of the fault causes the ground motions to
Figure 6. Comparison of modified Mercalli intensity (MMI) across the modeling groups for scenario HS G01 HypoH. The lower two rows show the residuals with respect to Aagaard’s MMI values (see Fig. 7). The black line in each map indicates the rupture, and the black star identifies the epicenter. Aagaard, Ma, and Rodgers use the same minimum period and similar minimum shear-wave speeds without intrinsic attenuation, resulting in very small residuals. Larsen and Graves include shorter periods and intrinsic attenuation, resulting in lower amplitudes in the Great Valley. The color version of this figure is available only in the electronic edition.
be much smaller in the San Jose area for south-to-north rupture compared with north-to-south or bilateral rupture. For example, the peak horizontal velocity at site CT06085501400 in San Jose (location is shown in Fig. 5) is 0.13 m/s in scenario HS G01 HypoH with south-to-north rupture compared with 0.31 m/s in scenario HS G01 HypoH with bilateral rupture. The difference is even more extreme for the Hayward South rupture; south-to-north rupture results in a peak velocity of only 0.05 m/s compared with 0.37 m/s for both bilateral and north-to-south rupture. Likewise, the ground motions around San Pablo Bay are much smaller for north-to-south rupture compared with bilateral or south-to-north rupture. These trends are consistent with previous studies of rupture directivity (Somerville et al., 1997; Aagaard et al., 2001; Spudich and Chiou, 2008). In our discussion of Graves’s broadband ground motions in the Comparison with NGA Models section, we will examine the spatial variation in spectral acceleration for the different hypocenters in the context of the Spudich and Chiou (2008) directivity corrections to the NGA ground-motion prediction equations.

Graves’s broadband velocity waveforms (Fig. 11) further illustrate these trends in rupture directivity. In San Francisco, the ground motions are largest for south-to-north rupture (Fremont epicenter) for both the Hayward South and Hayward South + North rupture lengths with peak horizontal velocities of 0.14 m/s and 0.20 m/s, respectively. These peak velocities are about 50% larger than those for north-to-south rupture and bilateral rupture. This is consistent with larger ground motions for ruptures propagating toward the city. The waveforms in Livermore tend to be largest for north-to-south rupture because Livermore lies north of the more rigid rock underneath the hills east of San Jose. This more rigid material tends to trap energy in the Livermore basin in north-to-south ruptures and shield Livermore from energy radiated in south-to-north ruptures. The waveforms in Livermore are also sensitive to the slip distribution. For the Hayward South + North rupture length and Fremont epicenter, the velocities are less than about 0.2 m/s with a duration of only about 15 s, whereas the amplitudes reach 0.8 m/s with 50 s of significant shaking for the San Pablo Bay epicenter; however, the ground-motion amplitudes and duration of shaking for the three epicenters and the Hayward South rupture length are quite consistent, with amplitudes of about 0.15 m/s. We attribute these different sensitivities to the hypocenter to the different stochastic portions of the slip distribution in the two sets of scenarios. For the Hayward South + North rupture length, there are no large slip patches south of Livermore, so rupture starting in Fremont does not radiate significant energy until it is further north. In the case of the Fremont epicenter for the Hayward
Figure 8. Velocity waveforms at three sites (see Fig. 5) for scenario HS G01 HypoH for each of the five modeling groups. We have low-pass filtered the waveforms to a common bandwidth of $T > 2.0$ s using two passes of a two-pole Butterworth filter with a corner frequency of 0.5 Hz. The waveforms demonstrate consistency in the amplitude and duration of shaking with nearly identical initial arrivals and some secondary arrivals. The color version of this figure is available only in the electronic edition.
South rupture length, there is a large slip patch at the southern end of the rupture, so northward propagating ruptures radiate energy into the Livermore area.

Some features in the distributions of shaking persist as we vary the rupture length, slip distribution, and hypocenter. These features are related to geologic structure as opposed to source features. This includes higher intensity shaking extending 20–40 km east of the Hayward fault due to deeper soft material east of the fault compared to west of the fault. We also find higher intensity shaking in the sedimentary basins, such as the Livermore basin, the San Pablo basin under San Pablo Bay, the Evergreen basin east of San Jose, the Cupertino basin west of San Jose, and the San Leandro basin west of Hayward. The shaking intensities in the Evergreen and Cupertino basins reach values 1–2 MMI units higher than locations several kilometers outside the basins. Similarly, the river valleys north of San Francisco Bay (e.g., Napa River valley running northwest from Napa) tend to have intensities 1–2 MMI units higher than the surrounding areas.

Larsen et al. (2000) also found amplification of ground motions in the San Pablo, Evergreen, and Livermore basins for magnitude 7.0 Hayward fault scenario earthquakes using a very simple 3D seismic velocity model. Using seismic velocity models defined nearly identically to the USGS Bay Area Velocity Model version 08.3.0, Harmsen et al. (2008) observed persistent patterns of shaking very similar to those in this study for scenario earthquakes involving the Hayward and Calaveras faults. The ruptures excite surface waves that are amplified in the Livermore, Evergreen, and Cupertino basins. Furthermore, Harmsen et al. found high intensities extending south along the east side of the Santa Clara Valley from the Evergreen basin, as we do in this study.

Rise Time and Rupture Speed

Larsen’s simulation of eight scenarios with four different rise times for Hayward South and Hayward South + North bilateral ruptures characterize the sensitivity of the ground
Figure 10. Modified Mercalli intensities from Graves's broadband ($T > 0.1$ s) simulations of six scenario earthquakes with different hypocenters for the Hayward South rupture length (top row) and the Hayward South + North rupture length (bottom row). The black line indicates the rupture, and the black star identifies the epicenter. Changing the hypocenter has a strong impact on the distribution of shaking by altering the rupture directivity. The Hayward South scenarios use slip distribution G01, which differs significantly from the slip distribution G04 used in the Hayward South + North simulations. The color version of this figure is available only in the electronic edition.
motions to the rise time in the slip-time history (duration of slip at a point). For each rupture length, we consider three different scaling factors of 1.0, 1.5, and 2.0 in the expression for the rise time as a function of slip,

\[
\frac{t_{95}}{t_0} = C \frac{D_f}{D_0},
\]

Figure 11. Velocity waveforms from Graves’s broadband \((T > 0.1 \text{ s})\) simulations of six scenario earthquakes with different hypocenters. The top two rows show waveforms for the Hayward South rupture length (scenarios HS G01 HypoO, Oakland epicenter; HS G01 HypoH, Hayward epicenter; and HS G01 HypoF, Fremont epicenter) and the bottom two rows show waveforms for the Hayward South + North rupture length (scenarios HS + HN G04 HypoSPB, San Pablo Bay epicenter; HS + HN G04 HypoO, Oakland epicenter; and HS + HN G04 HypoF, Fremont epicenter). The value on the right side of each trace indicates the peak velocity. The color version of this figure is available only in the electronic edition.
on values of 1.0, 1.5, or 2.0 (denoted by Tr10, Tr15, and Tr20 in the scenario labels). This provides a factor of 2 difference in rise times across the scenarios with the median rise time in each scenario similar to the value proposed by Somerville et al. (1999) for self-similar rupture. For example, with $C$ equal to 1.5, a slip of 1.0 m yields a rise time of 1.5 s ($M_w$ 6.7 for Somerville et al., 1999) and a slip of 1.5 m yields a rise time of 1.8 s ($M_w$ 7.2). We also consider uniform rise times, $t_{95}/0.136$: 0 s for the Hayward South rupture length (denoted by Tr10u in the scenario name) and $t_{95}/0.135$: 2 s for the Hayward South/ North rupture length (denoted by Tr20u in the scenario name). Even though Larsen’s simulations include periods down to 1.0 s, the limited bandwidth likely inhibits their ability to completely resolve differences in the ground motions for these rise times.

The ground motions exhibit very little sensitivity to the rise time. Figure 12 shows that varying the rise time without any corresponding changes in any of the other source parameters has no effect on the shape of the waveforms. Nevertheless, the velocity amplitudes can vary up to about 60% for variation in rise times by a factor of 2. These variations are much less dramatic than the changes we observed for the variations in rupture length and hypocenter. The electronic supplement contains waveforms at 80 sites and maps of the shaking intensity that demonstrate that all of the sites display similarly weak sensitivities to the rise time. Aagaard et al. (2001) arrived at the same conclusion through variation of the peak slip rate in simpler rupture models in a 1D structure.

We characterize the sensitivity of the ground motions to the rupture speed using Ma’s six simulations with three rupture speeds for Hayward South and Hayward South/North bilateral earthquake ruptures. The scenarios include local rupture speeds in high slip regions at 82% of the local shear-wave speed (denoted by Vr82 in the scenario name) and supershear rupture (denoted by Vr141 in the scenario name). Part I (Aagaard et al., 2010) discusses the details of the local rupture speed variation. The two sub-shear cases (Vr82 and Vr92) encompass the range of rupture speed variation in these scenarios.

Figure 12. Velocity waveforms from Larsen’s long-period ($T > 1.0$ s) simulations of four scenario earthquakes with different rise time distributions. The scenarios are identified by the rise time and include HS + HN Tr10, HS + HN G04 HypoO Tr15, HS + HN Tr20, and HS + HN Tr20u. The ground motions exhibit a relatively weak sensitivity to the variation in the rise time distribution. The value on the right side of each trace indicates the peak velocity. The color version of this figure is available only in the electronic edition.
speeds for many crustal strike-slip events. The supershear case includes locally supershear rupture where the slip exceeds the average slip with a maximum rupture speed of \( \sqrt{2} \) times the local shear-wave speed at the location with the maximum slip. Most observations of supershear rupture involve large strike-slip rupture of nearly planar faults (Bouchon and Karabulut, 2008, 2009). Thus, the relatively short length of the Hayward fault, slightly complex geometry (e.g., variations in strike and dip along its length) as well as the presence of creep (i.e., stable sliding regions) suggest that supershear rupture on the Hayward fault is unlikely. Thus, the scenarios with subshear rupture have substantially higher probabilities.

The ground motions at most sites exhibit a weak sensitivity to the variation in rupture speed. The amplitude of the velocity waveforms display roughly the same variation across the range in rupture speeds as they do for the range of rise times considered earlier. However, ruptures with faster rupture speeds radiate energy in a shorter time period, which results in sharper arrivals and shorter duration velocity pulses. This gives rise to some small changes in the shape of the waveforms as evident in Figure 13 for bilateral Hayward South + North ruptures. The waveforms in San Francisco and Livermore, which lie well off the strike of the Hayward fault display less variation in shape than those in San Jose, which lies close to the southern end of the rupture and is more sensitive to the rupture duration. The waveforms in San Francisco and Livermore have only subtle shifts in the duration of the velocity pulses with differences in peak velocity of less than 15%. On the other hand, in San Jose the different rupture speeds give rise to variations in the envelope of the waveform with fluctuations in amplitude of about 50–100%.

Graves et al. (2008) found a similar strong sensitivity to rupture speed in ground motions within the deep sedimentary basins of southern California for magnitude 7.8 events on the southern San Andreas fault. Thus, while most sites exhibit a weak sensitivity to variations in rupture speed, some sites

Figure 13. Velocity waveforms from Ma's long-period \((T > 2.0 \text{ s})\) simulations of three scenario earthquakes with different rupture speed distributions. The scenarios are identified by the rupture speed and include HS + HN Vr82, HS + HN G04 HypO Vr92, and HS + HN Vr141. The ground motions exhibit a relatively weak sensitivity to the variation in the rupture speed distribution. The value on the right side of each trace indicates the peak velocity. The color version of this figure is available only in the electronic edition.
prone to basin effects may exhibit a greater sensitivity; the degree of sensitivity depends on the locations and geometries of both the basin and rupture.

At sites very close to the rupture, we do not find evidence for rotation of the peak motion from fault-perpendicular to a more fault-parallel orientation (see velocity waveforms in the electronic supplement) as predicted in theoretical models of supershear rupture (Aagaard and Heaton, 2004; Dunham and Archuleta, 2004). This is likely due to the few relatively short bursts (about 15 km long) of supershear rupture in our rupture models and complexity of the velocity waveforms associated with the 3D geologic structure.

Creep and Coseismic Slip

As discussed in part I (Aagaard et al., 2010), in places the Hayward fault accommodates part of its long-term slip via aseismic creep. Creep generally occurs at shallow depths (from the ground surface down to about 3–5 km depth), but Funning et al. (2007) image a deep creeping region beneath Berkeley. Most of the earthquake scenarios use a vertical gradient in slip in creeping regions to account for how creep may affect the coseismic slip distribution. The vertical gradient decreases the slip in creeping regions as the rupture propagates into shallower regions. We chose the vertical gradient of −0.12 m/km for consistency with the paleoseismic record and the reduced area factor developed by the WGCQE (2003). Although we expect creep, which accommodates some of the long-term fault slip-rate, to exert some influence on the coseismic slip distribution, its effect could be minimal in large earthquakes. At the other end of the spectrum, perhaps very little or no coseismic slip occurs in the creeping regions. Neglecting the influence of creep corresponds to a vertical gradient in slip of zero, whereas preventing coseismic slip in creeping regions corresponds to an infinite vertical gradient in slip. As the vertical gradient in slip increases, slip in the creeping regions decreases, which reduces the average slip and earthquake magnitude. We consider both of these end-member cases for bilateral rupture of the Hayward South and Hayward South + North rupture lengths.

For the Hayward South + North rupture length, the moment magnitude of the scenario without coseismic slip in creeping regions (fully creeping) is 6.82, the moment magnitude of the scenario with a gradient of −0.000012 m/km is 7.05, and the moment magnitude of the scenario neglecting creep is 7.12. This variation in the magnitudes of the scenarios as we vary the vertical-slip gradient arises from maintaining consistency with the magnitude-area relation. Figure 14 displays maps of the MMI for these three scenarios, and Figure 15 shows the velocity waveforms at three sites. The shaking intensity and velocity amplitudes follow the variation in magnitude. The values for the \(M_w\) 6.82 scenario are similar to those for the bilateral \(M_w\) 6.76 Hayward South scenario, and the values for the \(M_w\) 7.12 scenario are slightly higher than those for the \(M_w\) 7.05 scenario. Because we use the same random seed in the stochastic portion of the slip distribution, the differences are limited to the amount of slip and the relative distribution between the creeping and locked portions. In the fully creeping scenario, very little slip occurs near the surface. Consequently, the rupture generates smaller amplitude surface waves, so the velocities in Livermore are about one-third to one-half of those in the scenarios in which creep exhibits less influence on coseismic slip. Similarly, the velocities are about 50% smaller in San Jose...
for the fully-creeping scenario. The corresponding ground motions for the Hayward South rupture length display similar trends.

The simulations suggest that the ground motions are only moderately sensitive to the presence of the creeping regions if creep has a moderate to minimal impact on the coseismic slip distribution. This would likely be the case in regions with only very shallow creep or regions where the creep rate is a small fraction of the long-term fault slip rate. On the other hand, if creeping regions have little or no coseismic slip with rise times comparable to locked regions, the expected magnitudes of Hayward fault events are about 0.1–0.2 units smaller, with a corresponding decrease in the intensity of the shaking with even smaller excitation of surface waves.

Broadband Simulations

Graves extended his simulations of the six Hayward South and Hayward South + North scenarios to shorter periods using the hybrid procedure described in Graves and Pitarka (2004) and Graves and Pitarka (2010). This simulation technique combines a stochastic approach at short periods (0.1 s < T < 1 s) with the 3D deterministic approach described earlier at long periods (T > 1 s) to produce broadband ground-motion synthetics. The procedure has been validated using the 1989 Mw 6.9 Loma Prieta earthquake (Aagaard, Brocher, Dolenc, Dreger, Graves, Harmsen, Hartzell, Larsen, Zoback, et al., 2008; Graves and Pitarka, 2010) and for the southern California region using the 1979 Mw 6.5 Imperial Valley earthquake, 1992 Mw 7.2 Landers earthquake, and 1994 Mw 6.7 Northridge earthquake (Graves and Pitarka, 2010). We also employed this methodology to calibrate the wavenumber at which we crossover from the nominal, background slip distribution to the stochastic slip distribution in part I (Aagaard et al., 2010).

In the short-period simulations, we sum the response for each subfault assuming a random phase, an omega-squared source spectrum, and simplified Green’s functions calculated
Figure 16. Comparison of MMI from Graves's broadband simulations of Hayward South scenario earthquakes with the compilation of Boatwright and Bundock (2008) from the 1868 earthquake, with the residuals for the synthetics shown in the bottom two rows. The bilateral rupture gives the lowest mean residual. The color version of this figure is available only in the electronic edition.
for a specified 1D velocity structure. This approach follows from Boore (1983) with the extension to finite faults given by Frankel (1995) and Hartzell et al. (1999). Each subfault ruptures with a moment proportional to the final slip of the subfault given by the original source model, and the values are scaled uniformly so that the moment matches that of the original source model.

As discussed in the previous section (Creep and Coseismic Slip), the creeping portion of the fault requires special attention when developing the kinematic rupture model. The simulation of high frequency motions using the semistochastic approach (Graves and Pitarka, 2004; Graves and Pitarka, 2010) also must account for this effect. In determining the effective area and magnitude of the rupture, we use the area reduction factor \( R \) developed by the WGCEP (2003). In the semistochastic simulation, the moment release of each subfault scales with the high frequency stress parameter, \( \sigma_p \) (Boore, 1983). Following self-similarity, the moment also scales as Area\(^{1/2}\), or \( R^{3/2} \). Thus, in order to properly account for the creeping portions of these ruptures, the stress parameter must also be scaled by \( R^{3/2} \). Our default value for the stress parameter is 50 bars. For the Hayward South ruptures, \( R = 0.79 \), which yields a stress parameter for these ruptures of 35 bars; for the Hayward South + North ruptures, \( R = 0.86 \), which yields a stress parameter for these ruptures of 40 bars.

The formulation requires specification of a 1D velocity model in calculating simplified Green’s functions and impedance effects. In this study, we create a 1D velocity model that roughly follows the average depth variations in the 3D structure, and we include both direct and Moho-reflected rays, which are attenuated by \( 1/R_p \), where \( R_p \) is the total path length traveled by the particular ray. For each ray, we compute a radiation pattern coefficient by averaging over a range of slip mechanisms and takeoff angles. Anelasticity is incorporated via a travel-time weighted average of the \( Q \) values for each of the material layers and a generic rock site spectral decay operator, \( \kappa = 0.04 \) (Anderson and Hough, 1984). Finally, gross impedance effects are included using quarter wavelength theory (Boore and Joyner, 1997) to derive amplification functions that are consistent with the specified 1D velocity structure.

To account for site-specific geology in the broadband motions, we apply frequency-dependent, nonlinear amplification factors based on \( V_{\text{s30}} \), the travel-time-weighted shear speed in the upper 30 m at the site. The site-specific \( V_{\text{s30}} \) values were taken from the map of Wills et al. (2000). The form of the amplification factors were developed using equivalent linear site response analysis (Walling et al., 2008) as implemented in the NGA ground-motion prediction equations of Campbell and Bozorgnia (2008).
Comparison with the 1868 Earthquake

Although we are not attempting to simulate the 1868 Hayward fault earthquake in detail (because little is known about its source parameters), several of the scenarios are designed to have source parameters that might be similar to this event. The Hayward South scenarios are consistent with the rupture length (Yu and Segall, 1996; Bakun, 1999) and magnitude (Bakun, 1999) of the 1868 earthquake. Boatwright and Bundock (2008) suggest that the north–south symmetry of the intensities is consistent with bilateral rupture compared with either predominantly north-to-south or south-to-north rupture. Our selection of three hypocenters permits further analysis to identify which rupture propagation pattern is most consistent with the shaking intensities from the 1868 event.

Figure 16 compares modified Mercalli intensity from Graves’s broadband ($T > 0.1$ s) simulations of three $M_w$ 6.76 Hayward South ruptures (HS G01 HypoO, HS G01 HypoH, and HS G01 HypoF) with the intensities of the 1868 earthquake compiled by Boatwright and Bundock (1999) of the 1868 earthquake. Boatwright and Bundock (2008) suggest that the north–south symmetry of the intensities is consistent with bilateral rupture compared with either predominantly north-to-south or south-to-north rupture. Our selection of three hypocenters permits further analysis to identify which rupture propagation pattern is most consistent with the shaking intensities from the 1868 event.

Comparison with NGA Models

Comparison of the broadband ground motions with ground-motion prediction equations, such as the NGA relations (Abrahamson and Silva, 2008; Boore and Atkinson, 2008; Campbell and Bozorgnia, 2008; Chiou and Youngs, 2008), provides an additional perspective from which to assess the ground motions for our scenario events. We will use AS08, BA08, CB08, and CY08 to refer to these four NGA relations, respectively. In calibrating the earthquake source parameters via comparison of broadband synthetics from a 1D velocity model with the NGA relations, we focused on minimizing the mean residual, not the variance or spatial variation. The broadband synthetics for the six scenarios that incorporate variability in the hypocenter (rupture directivity) and magnitude permit a much more detailed comparison,
including examination of effects due to basin response, local site conditions, and rupture directivity.

Figure 17 compares spectral accelerations (SA) at a period of 1.0 s from Graves’s broadband simulation of the $M_w 6.76$ Hayward South bilateral rupture (scenario HS G01 HypoH) with those predicted by the AS08, BA08, and CB08 NGA relations. The mean residuals correspond to event terms in the ground-motion prediction models and express how the average ground motions from the 3D simulations differ from the median of the ground-motion prediction model for the specified earthquake magnitude.

The mean residual for each of the three NGA relations for this bilateral Hayward South rupture is small, with the peak in the histogram within about 0.2 log$_2$ units (15%) of zero. The maps of the residuals clearly show that the 3D ground-motion simulations predict stronger shaking off the ends of the rupture than the NGA ground-motion prediction equations. The 3D ground-motion simulations include strong along-strike directivity which is not explicitly included in the NGA relations. The NGA relations incorporate the distance from the rupture, so that the spectral values average the directivity effects along the fault strike. In the next section, we examine this issue in more detail using Graves’s broadband simulation of three Hayward South + North ruptures.

We examine the spatial variation in the residuals of the spectral acceleration with the period for the BA08 NGA relation using the Graves’s Hayward South + North ruptures (Fig. 18). The electronic supplement contains similar plots for peak horizontal ground acceleration (PGA), peak horizontal ground velocity (PGV), and spectral acceleration at periods of 0.3 s, 1.0 s, and 3.0 s for each of Graves’s broadband simulations and the AS08, BA08, and CB08 NGA ground-motion prediction equations.

At shorter periods (0.3 s), the mean bias is 0.43 log$_2$ units (34%) with most of the positive residual values in the region east of the Hayward fault. The variance is relative small (0.44 log$_2$ units, or 36%). The mean residual is smaller at 1.0 s (0.27 log$_2$ units, or 21%) but increases at 3.0 s (0.81 log$_2$ units, or 75%). The variance increases with period and is 0.71 log$_2$ units, or 64%, at 1.0 s and 1.06 log$_2$ units, or 110%, at 3.0 s. At longer periods the largest residuals occur east of the Hayward fault, in the river valleys near Santa Rosa and Napa, and in the Santa Clara Valley around San Jose. A similar increase in residuals with period was seen in the analysis of the $M_w 7.8$ San Andreas ShakeOut simulations for southern California (Graves et al., 2008). In that study, as in the current study, the large variances at the longer periods are primarily due to the effects of rupture directivity and amplification within relatively low shear-wave velocity material, such as sedimentary basins. These are robust features of the 3D long-period deterministic ground-motion simulations that we associate with the 3D geologic structure and source characteristics. The NGA models incorporate such effects via very simple approximations as we discuss in the following two sections.

We summarize the consistency of the simulations with the AS08, BA08, and CB08 NGA relations in Figure 19 by computing the median residual and its variance averaged over the three NGA relations for Graves’s six broadband simulations (scenarios HS G01 HypoP, HS G01 HypoH, HS G01 HypoF, HS + HN G04 HypoSPB, HS + HN G04 HypoP, and HS + HN G04 HypoF). In general, the simulated motions fall about one standard deviation above the median value, suggesting that, on average, the simulations are within the expected range of event-to-event variability observed in recorded earthquakes of the same magnitude.

The average median residuals for the three $M_w 7.05$ Hayward South + North scenarios tend to be slightly larger than those for the three $M_w 6.76$ Hayward South scenarios. This may indicate that the ruptures associated with the G01...
and G04 slip distributions radiate slightly more coherent energy than real earthquakes. Without similar analysis for broadband simulations of the rest of the suite of scenarios, we hesitate to conclude that this bias extends to the entire suite of scenarios. The smallest average median residuals generally occur for the southernmost epicenter (Fremont) for both the $M_{w}$ 6.76 and $M_{w}$ 7.05 scenarios. We suspect this results from trade-offs between the assumed hypocenters, the kinematic slip distribution, and the interaction of the seismic waves with the 3D geologic structure. Drawing any broad conclusions about the variations in the average median residuals would require applying the broadband simulation methodology to the entire suite of scenarios which incorporates greater variability in the rupture parameters. Furthermore, it is difficult to assess how one might adjust the NGA models to account for the effects of creep on coseismic slip beyond the effect on the magnitude for a given rupture area; creep also affects secondary features, such as the along-strike and down-dip distribution of slip.

**Accounting for Directivity.** As discussed earlier, at the longer periods the simulated motions generally exceed the
empirical ground-motion relations in regions with strong forward directivity and fall slightly below the empirical relations in regions with backward directivity; consequently, the PGV and SA at 1.0 s and 3.0 s at most sites are highly sensitive to the hypocenter. Figure 18 clearly illustrates this effect by comparing the $M_{w} 7.05$ bilateral Hayward South / North rupture with the BA08NGA relation for SA at 1.0 s. While the overall mean of the residuals for these cases is in the range of 10%–20%, sites located in the forward rupture direction have simulated motions up to 2–3 times larger than the empirical relation, whereas sites in the backward rupture direction can have motions 2–3 times smaller than the empirical relation.

Somerville et al. (1997) was the first to develop a directivity model that could be applied as a correction to ground-motion prediction equations. Two additional directivity models have been developed in conjunction with the NGA program. Spudich and Chiou (2008) proposed a model based on isochrone theory and Rowshandel (2010) proposed a model based on rupture heterogeneity and source-site geometry. The Spudich–Chiou and Rowshandel corrections give similar results, although the Rowshandel model generally predicts stronger directivity effects, particularly for ruptures containing strong slip asperities.

Figure 20 shows the Spudich–Chiou directivity corrections for the three $M_{w} 7.05$ Hayward South + North scenarios applied to the BA08NGA relation for 1.0 s SA. Although some of the locations with higher residuals lie in sedimentary basins, the pattern of these corrections corresponds quite well to the residuals shown in Figure 20, but the absolute level is smaller, with the maximum correction not exceeding about 25%. Consequently, applying these corrections to the NGA relation only reduces the standard deviation of the residuals by a few percent. Similar results are found for PGV and 3.0 s SA, as well as for the other NGA relations (see the electronic supplement).

We attribute the differences in the strength of the rupture directivity to several factors, all of which arise from the fact that there are relatively few ground-motion recordings close to large strike-slip earthquake ruptures. We developed our rupture models using information gleaned from source inversions of past earthquakes as well as theoretical and laboratory analyses of rupture dynamics (see Aagaard et al., 2010). We calibrated the models to match, on average, existing ground-motion records. However, the sparsity of data can leave some details of the rupture process rather poorly constrained. For example, it is generally accepted that ruptures tend to propagate at a speed of about 80%–85% of the local shear-wave speed; however, it is not uncommon for ruptures to propagate slower than this, and there are several cases where supershear rupture has been proposed (Olson and Apsel, 1982; Archuleta, 1984; Spudich and Cranswick, 1984; Anderson, 2000; Bouchon et al., 2000; Bouchon et al., 2001; Sekiguchi and Iwata, 2002; Bouchon and Vallee,
In the development of their ground-motion model, Boore and Atkinson (2008) noted the strong correlation between $V_{S30}$ and basin depth in the NGA data set and argued that $V_{S30}$ can be used as a proxy for basin depth in the empirical regression. While this is true for the NGA data set in general, it may not hold for the greater San Francisco Bay area, where the basins are relatively shallow compared to other regions (Fig. 21). In this context basin depth is defined as the depth to the 1.5 km/s shear-wave isosurface, hereafter referred to as Z1.5. For the San Francisco Bay area, we measured Z1.5 in the 3D USGS Bay Area Velocity Model version 08.3.0 on a grid of sites at 1 km spacing and estimated $V_{S30}$ for each site using Wills et al. (2000). The broadband simulations also use $V_{S30}$ values from Wills et al. (2000) in site corrections associated with imposing a minimum shear-wave speed in the simulations. Because Wills et al. classify sites into discrete $V_{S30}$ bins, we totaled the number of observations within each bin and scaled the symbols in Figure 21 by that number. While the NGA data set as a whole shows a clear and strong increase of Z1.5 with decreasing $V_{S30}$, the Bay area sites (both within the NGA data set and those for the 3D model) exhibit a very weak correlation. This could reflect differences in the tectonic environments between the San Francisco Bay area and the sites in the NGA data set.

We explore the implications of this difference in correlation to determine how well it explains larger amplitude ground motions in the 3D simulations compared with the NGA ground-motion prediction equations. We believe this difference in correlation dominates other factors, such as the details of the attenuation model. We derive an approximate amplification correction to the Boore and Atkinson NGA model for our San Francisco Bay sites. For linear site response the BA08 amplification term is given by

$$A_{site} = \left(\frac{V_{S30,site}}{V_0}\right)^x,$$  \hspace{1cm} (13)

where $V_{S30,site}$ would be the site-specific $V_{S30}$ value (from Wills et al., 2000, for example). For periods longer than about 1 s, $x$ is approximately constant and equal to $-0.725$. From this relation we define an approximate amplification correction to the BA08 NGA relation for our sites as

$$A_{cor} = \frac{A_{pred}}{A_{site}} = \left(\frac{V_{S30,pred}}{V_{S30,site}}\right)^x,$$  \hspace{1cm} (14)

where $V_{S30,pred}$ is the $V_{S30}$ value predicted by the correlation between Z1.5 and $V_{S30}$ noted by Boore and Atkinson (2008). The amplification correction has the effect of replacing the $V_{S30}$ values from Wills et al. (2000) with $V_{S30}$ values ($V_{S30,pred}$) predicted by Z1.5 and the correlation between $V_{S30}$ and Z1.5 in the NGA data set. We determine the correlation using the following relational form:

$$\log_{10}(V_{S30}/V_0) = A + B \log_{10}(Z1.5/Z_0).$$  \hspace{1cm} (15)
where $V_0 = 760 \text{ m/s}$, $Z_0 = 1 \text{ km}$, $A = -0.375$, and $B = -0.211$.

Figure 21 shows the spatial distribution of $A_{ox}$ for our model region. Comparing this to the residuals for SA at 1.0 s for the $M_w$ 7.05 Hayward South + North scenario shown in Figure 18, we see many similarities both in terms of spatial pattern and amplitude (keeping in mind that the residuals in the figures also contain the effects of rupture directivity). The residuals indicate amplification of motions along the eastern side of the Hayward fault, which extends north into the San Pablo basin and south toward Gilroy. The region immediately east of the Hayward fault has relatively high $V_{530}$ but also relatively deep $Z1.5$; thus, the NGA relations (without the amplification correction) tend to underpredict the simulated motions in this region. Likewise, regions surrounding the margins of the San Francisco Bay and the Sacramento River delta have relatively low $V_{530}$ but a relatively shallow $Z1.5$. Consequently, the BA08 NGA relations (and likely the others as well) overpredict the simulated motions in these regions. Other regions where $V_{530}$ and $Z1.5$ tend to be correlated, such as the Cupertino and Evergreen basins near San Jose and the Great Valley, the NGA relations are similar to the simulated motions. This suggests that refinement of the ground-motion prediction models may be required in order to adequately account for the effects of amplification across the diverse range of tectonic environments, including shallow basins.

Conclusions

These ground-motion simulations demonstrate that large Hayward fault earthquakes generate strong shaking throughout the San Francisco Bay area, with about 50% of the urban area experiencing MMI VII or larger for magnitude 7.0 earthquakes. The details of the shaking are strongly dependent on the rupture length (earthquake magnitude), hypocenter (rupture directivity), and slip distribution. The ground motions exhibit a relatively weak sensitivity to variations in the rise time, consistent with results from a previous study using a generic 1D variation in material properties (Aagaard et al., 2001). The ground motions at most sites also display less sensitivity to the rupture speed; we do find some evidence in the Evergreen basin east of San Jose for greater sensitivity to the rupture speed and comparable to that found in the Los Angeles basin for northwest rupture of the southern San Andreas fault (Graves et al., 2008). The close proximity of the Evergreen basin to the Hayward fault appears to facilitate strong interactions between the rupture and basin with a strong sensitivity to the speed rupture toward the basin. Thus, these simulations provide further evidence for significant variability in ground motions associated with earthquake source parameters that are difficult to forecast, such as the rupture extent and hypocenter, while highlighting the ability to capture the effects of 3D sedimentary basins, which can be constrained a priori; the simulations also emphasize the difficulty in constraining rise times and rupture speed due to the relatively weaker sensitivity of ground motions to these parameters.

The simulations predict ground motions consistent with the Abrahamson and Silva (2008), Boore and Atkinson (2008), and Campbell and Bozorgnia (2008) NGA ground-motion prediction equations with two areas of departure. The 3D simulations generate stronger rupture directivity than that predicted by the Spudich and Chiou (2008) directivity correction to the NGA relations, although the spatial variation in ground motion in the simulations associated with rupture directivity closely matches the spatial variation in the Spudich and Chiou model. Similar discrepancies exist with respect to individual events used to construct the Spudich and Chiou model, suggesting that the simulations may be realistic and the accuracy of the model could be improved by incorporating a period dependence on the areal extent of directivity effects. Analysis of ground-motion amplification in sedimentary basins from our simulations indicates that amplification in shallow basins (e.g., the Cupertino and Evergreen basins near San Jose, the San Pablo and San Leandro basins near Oakland, and the Cotati and Windsor basins near Santa Rosa) and regions with deep soft material but relatively fast $V_{530}$ values (e.g., the region immediately east of the Hayward fault between Hayward and Richmond) may exceed that predicted by the NGA relations. This appears to arise from the strong correlation between $V_{530}$ and basin depth for the sites recording ground motions used to construct the NGA relations. In these same areas, peak amplitudes from our simulations of two moderate earthquakes agree quite well with the observed amplitudes. Verifying this discrepancy and the greater accuracy of the 3D simulations requires further testing, especially in areas outside of southern California with shallow sedimentary basins.

Our ground-motion simulations include a reduction of the coseismic slip in creeping regions through either a slip-predictable approach or a vertical gradient in slip in the creeping regions. Both of these approaches reduce the earthquake magnitude for a given rupture area. Consideration of the end-member cases for the vertical-gradient approach (creep having no affect on coseismic slip and creep preventing any slip in creeping regions) demonstrates that considering creep when computing ground motions for Hayward fault scenario earthquakes reduces the amplitude of the ground motions compared to when creep is neglected. In the extreme case of no coseismic slip in creeping regions, the ground motions in Livermore and San Jose are about 50% smaller as a result of reduced excitation of surfaces waves associated with the limited amount of shallow slip. This highlights the important role of continued and improved characterization of the spatial extent and rates of creep along the Hayward fault for accurate assessment of the seismic hazard associated with the Hayward fault.

Data and Resources

The 2000 census is available at http://www.census.gov/geo/www/census2k.html (last accessed April 2009). Observed ground motions for the 2007 $M_w$ 5.45 Alum Rock...
earthquake and the 2007 $M_w$ 4.18 Oakland earthquake can be obtained from the Incorporated Research Institutions for Seismology Data Management Center at http://www.iris.edu/ (last accessed October 2009) and the U.S. Geological Survey (USGS) National Strong Motion Program at http://nsmp.wr.usgs.gov (last accessed October 2009). The USGS Bay Area Velocity Model version 08.3.0 can be obtained from http://earthquake.usgs.gov/regional/ncfa/3Dgeologic/ (last accessed November 2009). All other data used in this paper came from published sources listed in the references. Many of the figures were generated using Generic Mapping Tools (Wessel and Smith, 1998), and the low-pass filtering of the waveforms was performed using SAC2000 (Goldstein et al., 2003).

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