

Short Note

Validation of 3D Velocity Models Using Earthquakes with Shallow Slip: Case Study of the 2014 M_w 6.0 South Napa, California, Event

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Abstract 3D velocity models constitute a key element in strong ground motion modeling, for example, earthquake-hazard assessment. Their validation is typically based on modeling weak earthquakes with foci limited to depths greater than ~ 5 km. However, ruptures during moderate and large earthquakes can propagate to shallower depths (and eventually reach the surface). For such shallow sources, velocity models may not be validated with sufficient accuracy. In this respect, we conduct a series of tests based on the 2014 M_w 6.0 South Napa earthquake, which was characterized by a very shallow slip asperity, to assess the performance of the U.S. Geological Survey 3D San Francisco Bay area velocity model within 20-km fault distance. Our study indicates that the velocity model performs generally well with some exceptions, in which large-amplitude surface waves not present in the observed data are systematically excited. We conclude that more complex fault geometries or slightly deeper slip would not result in a better fit of the observed data. Contrarily, we demonstrate that smoothing the velocity model (i.e., reducing the strong velocity contrasts between basin fill and bedrock) effectively attenuates the spurious oscillations.

Electronic Supplement: Description of the smoothing process, figures showing results for alternative fault geometry, other inversion tests and waveform comparisons, and a movie of wave propagation.

Introduction

An increasing number of strong ground motion modeling studies of past and future earthquakes rely on 3D velocity models (e.g., Graves *et al.*, 2008; Harmsen *et al.*, 2008; Aagaard *et al.*, 2010). The influence of 3D structures in modulating the ground-shaking levels at various source-distance ranges has been widely documented in a number of studies focused on major past earthquakes, such as Loma Prieta (e.g., Frankel and Vidale, 1992), Northridge (e.g., Graves *et al.*, 1998), or Kobe (e.g., Pitarka *et al.*, 1998).

These 3D velocity models are built upon data of heterogeneous nature (geologic studies, *in situ* and laboratory measurements, geophysical investigations, etc.), often incorporating significant simplifications and approximations that demand proper validation by means of real-data modeling (e.g., Hartzell *et al.*, 2006; Ma *et al.*, 2008; Tape *et al.*, 2010; Lee *et al.*, 2014).

For example, the U.S. Geological Survey (USGS) 3D velocity model versions 5.1.0 and 8.3.0 (the latter being used in the present article) have been validated by simulating waveforms of small and moderate earthquakes at low fre-

quency (< 1 Hz) in the greater San Francisco Bay region (Rodgers *et al.*, 2008; Aagaard *et al.*, 2010; Kim *et al.*, 2010). These studies have demonstrated that, although the 3D velocity model performs well in predicting peak ground velocity and P - and S -wave timing, observed waveforms can be matched either well or rather poorly, depending on the specific source–receiver path. Moreover, validations of this model have been based on events with hypocenters deeper than ~ 5 km because weak earthquakes do not typically nucleate at shallower depths. However, it has been observed that ruptures during moderate and large earthquakes can propagate to shallower depths, occasionally reaching the surface (e.g., the 1992 Landers and the 1999 Chi-Chi earthquakes). Thus, to reliably predict ground motions for such events, 3D velocity models must be validated also for shallow sources using recordings of larger earthquakes. We point out that such validation is not straightforward because large earthquakes require finite-extent source models, for which details are rather uncertain and possibly biased. Therefore, event-specific source models must be treated with care by

means of testing for various scenarios/possibilities, as illustrated in the present article.

Here, we assess the validity of the USGS 3D velocity model in a relatively small region of the northern Bay area based on the M_w 6.0 South Napa earthquake. This event struck the northern San Francisco Bay area on 24 August 2014 at 10:22:44 UTC, causing severe damage to buildings and one fatality in the Napa area (Brocher *et al.*, 2015). The mainshock has been well recorded by a relatively dense seismic network, resulting in 12 sites at less than 15 km from the epicenter. Several finite-fault models based on strong-motion recordings (Ji *et al.*, 2015; Wei *et al.*, 2015; Gallovič, 2016), Global Positioning System (GPS) data (Melgar *et al.*, 2015), GPS and Interferometric Synthetic Aperture Radar (InSAR) data (Barnhart *et al.*, 2015), and seismic, GPS, and InSAR data (Dreger *et al.*, 2015) were proposed. Although different in many details, all the models indicate a unilateral, prevalently up-dip, north-propagating rupture and a dominant slip asperity at shallow depths (< 5 km).

All the studies published so far considered 1D velocity models to image the source. However, Dreger *et al.* (2015) and Wei *et al.* (2015) noted that the earthquake occurred in an area of complex geologic structures, where path effects and site response may have played an important role. Their forward simulations up to 0.5 Hz using the USGS 3D seismic velocity model demonstrated dramatic differences in the ground motions with respect to the 1D case. Consequently, inversion studies based on 3D Green's functions (GFs) (e.g., Liu and Archuleta, 2004; Gallovič *et al.*, 2015) may lead to different source images with respect to the 1D case.

In this study, we conduct a series of tests including inversion and forward simulations to reproduce the observed data using the USGS 3D Bay area velocity model. Our results suggest that the velocity model performs generally well with some exceptions, such as toward the San Pablo Bay, where it systematically generates strong surface waves not present in the observed data. We exclude the possibility that more complex fault geometries or slightly deeper slip could result in a better fit of the observed data. We also demonstrate that smoothing the velocity model (i.e., reducing the strong velocity contrasts between basins fill and bedrock) effectively attenuates these spurious oscillations. Eventually, we raise warning that ground-motion modeling featuring near-surface ruptures (e.g., Aagaard *et al.*, 2010; CyberShake experiment in Graves *et al.*, 2011) may be potentially biased in terms of amplitude and duration at specific (basin) locations if the velocity models are not explicitly tested for such shallow events.

Inversion Technique and Wave Propagation Modeling

To image the source, we apply the linear slip-inversion technique introduced by Gallovič *et al.* (2015). The fault model is discretized to a dense spatial grid of subfaults and each subfault is characterized by a discretized slip-rate func-

tion lasting from the hypocentral time until the (assumed) end of the rupture process. Therefore, this generous inversion approach requires only minimal *a priori* constraints (such as slip-rate positivity) and has the potential to infer complex features of the rupture evolution, including slip reactivation and supershear rupture propagation. For further details, the reader is referred to Gallovič *et al.* (2015) and recent applications to synthetic (Gallovič and Ampuero, 2015) and real (Sokos *et al.*, 2015; Gallovič, 2016) data.

The imaging method requires GFs defined as displacement waveforms computed for each subfault–receiver pair, in which the subfault is characterized by a unit scalar seismic moment and (effectively) impulse slip-rate function. Here, the synthetic GFs are computed using SW4, a fourth-order finite-difference code based on the summation-by-parts principle (Petersson and Sjogreen, 2013). The numerical grid spans about 66 km in both horizontal directions and 35 km in the vertical one, including the absorbing boundaries (Ⓔ Fig. S1, available in the electronic supplement to this article). The grid, being curvilinear at shallow depth to accommodate topography, has a spacing of 125 m, ensuring correct waveforms modeling up to the maximum frequency of interest (0.5 Hz).

Data, Models of Fault, and Velocity Structure

We use the same set of 10 strong-motion stations (see Ⓔ Fig. S1) and data pre-processing procedure of Gallovič (2016). In particular, both observed and synthetic time series are transformed into displacement and band-pass filtered using a causal (minimum-phase) fourth-order Butterworth filter between 0.05 and 0.5 Hz. The horizontal components are rotated to the fault-parallel (FP) and fault-normal (FN) components.

We adopt the fault geometry and mechanism defined in Gallovič (2016): a 15-km-long 10-km-wide plane with strike/dip/rake values of 155/82/−172, discretized in 20×16 subfaults. In an attempt to verify whether another fault geometry would produce better waveforms fit using 3D GFs, we also considered the fault model proposed by Wei *et al.* (2015), characterized by two distinct planes dipping 82° eastward (see Ⓔ Fig. S1). Both fault planes have almost the same size ($11 \text{ km} \times 15 \text{ km}$ and $9 \text{ km} \times 15 \text{ km}$, discretized in 300 subfaults in total), slightly different orientation (strike values are 338 and 351, respectively), but identical rake (−172). We chose this geometry because, among all the alternatives proposed in the literature, it presents the largest differences with respect to the model of Gallovič (2016).

For each subfault–receiver pair, we calculate two sets of 40-s-long GFs using the GIL7 1D model (as slightly modified by Gallovič, 2016) and the USGS 3D geologic and seismic velocity model v. 8.3 (USGS BayAreaVM-08.3.0. etree; see Data and Resources). For the sake of brevity, we refer to these two sets as 1D and 3D GFs, respectively. We limit the minimum shear wavespeed to 500 m/s, resulting in 8 points per wavelength at the highest frequency (0.5 Hz). Topography and intrinsic attenuation are included in the 3D calcu-

lations, although both factors influence the waveforms only marginally in the present frequency and distance ranges. We note that the 1D GFs of the SW4 code match perfectly with the 1D GFs computed by Gallovič (2016) using the Axitra code (Coutant, 1989) based on the discrete wavenumber technique by Bouchon (1981).

1D GF Source Inversion and Forward 3D Modeling

Slip distribution and slip-rate functions for each individual subfault retrieved using the 1D velocity model are shown in Figure 1a (the source image is the same as in Gallovič, 2016). In agreement with other studies, the model shows a major asperity located at shallow depth (<5 km), rupture propagating mainly up-dip during the first 3–4 s, and along strike (south-to-north) at later times (see also slip-rate snapshots in Fig. S2a). Hereafter, we refer to this source model as the 1D GF source model.

In order to evaluate the effects of the lateral heterogeneities on the resulting wavefield, in Figure 1b we show a comparison between observed and synthetic seismograms at FP components of selected stations (for all seismograms, see Fig. S3) computed using the 1D GF source model and the 1D and the 3D GFs. The depth to bedrock is included in the figure to delimit the major basins in the area under study. The waveform comparison suggests that in the 3D case, we observe either a similar (e.g., 68150 and N019B), worse (e.g., NGVB and NTO), or even better fit of the main pulse (e.g., 1765 and N002) with respect to the 1D case. On the other hand, in the 3D case a few stations (NSP, NTO, and N016) present high-energy late arrivals, whose amplitudes are much larger than observed in actual recordings. These late arrivals are mainly constituted by surface waves trapped inside basins or excited in proximity of their edges, as visible in the animation of wave propagation in Movie S1. We point out that these waves do not appear in the 1D case.

The slip-inversion result based on the fault model of Wei *et al.* (2015) and 1D GFs is characterized by a shallow slip patch and a similar evolution of the rupture front (see Fig. S4a,c). The waveform fit is qualitatively similar to that for the Gallovič (2016) fault geometry (Fig. S5). Moreover, when the USGS 3D velocity model is used in the forward calculation, high-amplitude oscillations can be observed at the same receivers as discussed above (and, to some extent, also at the 6831 and 1765 stations), see Figure S6.

During the inversion procedure, we considered only those stations closer to the fault, where the source signature is strongest, providing a good azimuthal coverage. To further test the source model, we also predict waveforms at those stations inside the simulation domain that were not utilized in the inversion (see Fig. S1 for their location). Results for the Gallovič (2016) fault geometry in the 1D and 3D velocity models are shown in Figure S7. The waveform fit is slightly worse than for the receivers used in the inversion, although still being acceptable. Interestingly, also in this case we note occasional large-amplitude oscillations in the 3D

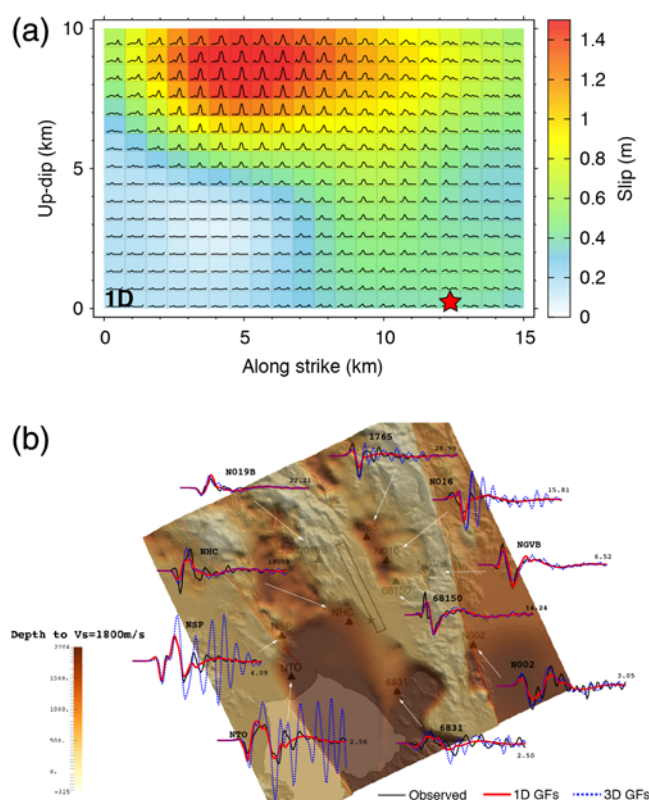


Figure 1. (a) Slip distribution and slip-rate functions as inferred from inversion based on 1D Green's functions (GFs). The slip rates start at the hypocenter time and last for 8 seconds. Star denotes the hypocenter (not considered in the inversion). For the snapshots of the rupture propagation, see Fig. S2a (available in the electronic supplement to this article). (b) Depth (relative to mean sea level) to shear-wavespeed of 1800 m/s for the U.S. Geological Survey (USGS) 3D velocity model and comparison between observed and synthetic displacement waveforms (fault-parallel [FP] component) computed using the 1D or 3D GFs and the same 1D GF source model at selected stations (for all stations and components, see Fig. S3). Note the large oscillations in the synthetics at stations NTO, NSP, and N016 in the 3D GF case not observed in the real data, consisting of surface waves trapped inside basins, or excited and guided along the proximity of their edges (see also Movie S1). The maximum amplitudes of the observed waveforms in centimeters are shown as numbers. The color version of this figure is available only in the electronic edition.

synthetics not present in the data and 1D synthetics (e.g., station 68367).

To further clarify the response of the medium, we determine the relative energy content of late arrivals for each of the 3D GFs computed on the Gallovič (2016) fault. The relative energy content is defined as the energy arriving between 4 s past the direct *S* wave and the end of the time series, normalized by the total energy. The 4-s-long time window is meant to account for the duration of the main pulse in the frequency range under investigation. We summarize the results, expressed in percent, in Figure 2. Although in principle controlled by the relative source–receiver position, our analysis reveals that the excitation of surface waves is strongly affected by local factors, being thus quite heterogeneous for many receivers. For example, stations 1765, N016, and 68150, even

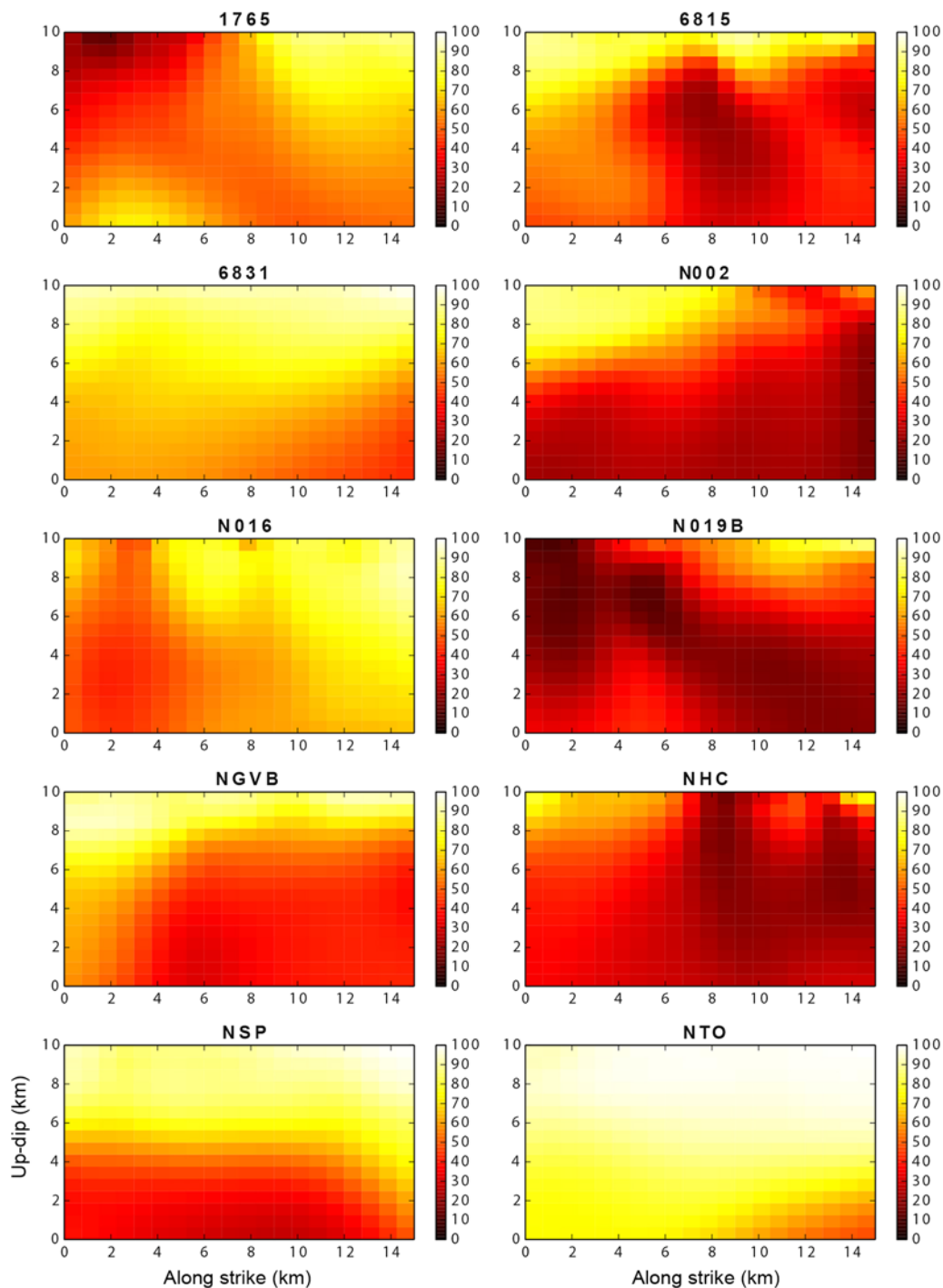


Figure 2. Relative energy content (in percent) of late arrivals for each 3D GFs (i.e., each receiver–subfault couple) used in the inversion. Late arrivals are defined as waves occurring at least 4 seconds after the direct *S* wave. See the [1D GF Source Inversion and Forward 3D Modeling](#) section for details. The color version of this figure is available only in the electronic edition.

though sharing the same distance from the fault plane, present different pattern and excitation values. The same holds for stations N019B and 1765, whose positions are almost symmetrical with respect to the fault. At the same time, we notice that for stations NSP and NTO, a large area of the fault plane is able to generate a significant amount of late arrivals (> 80%). This is not the case for station N002, located at about the same

distance from the fault. By comparison, a similar analysis on the 1D GFs shows that for stations NSP and NTO only the uppermost 1.5 km of the fault plane could generate similar amounts of late arrivals. We note that small perturbations of mechanism (20° for strike) and fault position (1500 m along the horizontal plane) do not alter significantly the energy pattern shown in Figure 2.

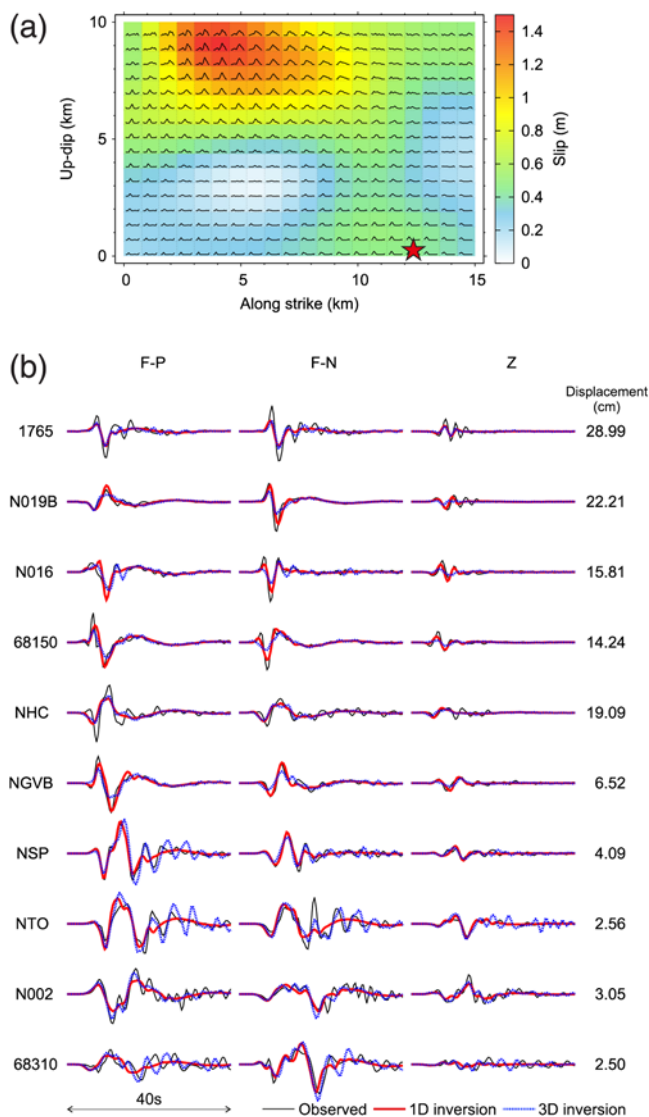


Figure 3. (a) Slip distribution and slip-rate functions as inferred from the inversion based on 3D GFs (for the snapshots of the rupture propagation, see ⑤ Fig. S2a). (b) Comparison between observed and synthetic displacement waveforms (0.05–0.5 Hz) for the inverted source model based on the 1D and 3D GFs. In general, we note that in the latter case the waveform fit for the later arrivals at distant stations improves, whereas it worsens for the first main pulses at the nearest stations. The color version of this figure is available only in the electronic edition.

As a further test, we arbitrarily taper the uppermost 3 km of the 1D GF source model and compute synthetic seismograms using 3D GFs (see ⑤ Fig. S8). In this case, the late reverberations at stations N016, NSP, and NTO experience a modest amplitude drop, still exceeding the observations. Similar oscillations can be observed in the 3D simulations of Wei *et al.* (2015), whose source model has smaller shallow slip than our 1D GF source model. This suggests that their excitation is effectively related to the whole subsurface asperity, and not just to its shallowest part. A comparison between our synthetics, those of Wei *et al.* (2015), and recorded

time series at stations N016, NSP, NTO, 1765, and 68150 is reported in ⑤ Figure S9.

We emphasize that so far the USGS 3D velocity model has been validated in the greater San Francisco Bay region by simulating low-frequency waveforms of small and moderate earthquakes with focal depths always larger than 5 km (Rodgers *et al.*, 2008; Aagaard *et al.*, 2010; Kim *et al.*, 2010), whereas our source releases most of its energy well above this level. Our analysis has shown that the USGS 3D velocity model may be significantly inaccurate in some of its parts, because it becomes apparent when shallow sources are considered. Later, we demonstrate that smoothing the velocity model effectively attenuates the spurious oscillations. Nevertheless, we must also keep in mind that the rupture model presented in this section might be biased due to the use of simplified 1D GFs in the inversion. Therefore, in the following section we show that the inversion does not improve when 3D GFs are employed.

3D GF Source Inversion

Slip distribution and slip-rate functions for each individual subfault retrieved using the 3D velocity model (hereafter denoted as the 3D GF source model) are shown in Figure 3a. Both source images based on 3D and 1D GFs present consistent basic features, but also a number of subtle differences. In particular, both models show a major asperity roughly located in the same area on the fault (being slightly more confined at depth in the 3D case) and a similar evolution of the rupture front, indicating a unilateral, up-dip, and northward-propagating rupture propagation (see also the snapshots in ⑤ Fig. S2b). On the other hand, the 3D GF source image features longer rise time and lower peak slip rates, particularly toward the free surface, resulting in lower final slip values. The slip-rate time histories are definitely more complex in the 3D case and several secondary energy-release episodes can be observed, expressed also as pronounced ghost features in the slip-rate snapshots in ⑤ Figure S2b. This notable complexity of the slip-rate functions and the rupture front as retrieved during the 3D inversion may raise some doubts on the reliability of the source model, especially when considering spontaneous rupture propagation studies (e.g., Ripperger *et al.*, 2008; Shi and Day, 2013). Gallovič *et al.* (2015) conducted a series of synthetic tests and found that incorrect GFs may result in such artificial slip-rate peaks during source imaging.

Interestingly, as shown in Figure 3b, the waveform fit for the 3D inversion does not improve with respect to the 1D case. In fact, a slight drop in variance reduction occurs (0.62 vs. 0.59 for the 1D and 3D case, respectively). Although we observe a generally better fit of secondary phases (see, e.g., stations NTO and NSP), occasionally first main pulses are less well reproduced; this is clearly evident on the FP and FN components at stations NGVB and N016, respectively. We explain this peculiar behavior by noting that unlike the 1D case, the 3D GFs are more complex and the

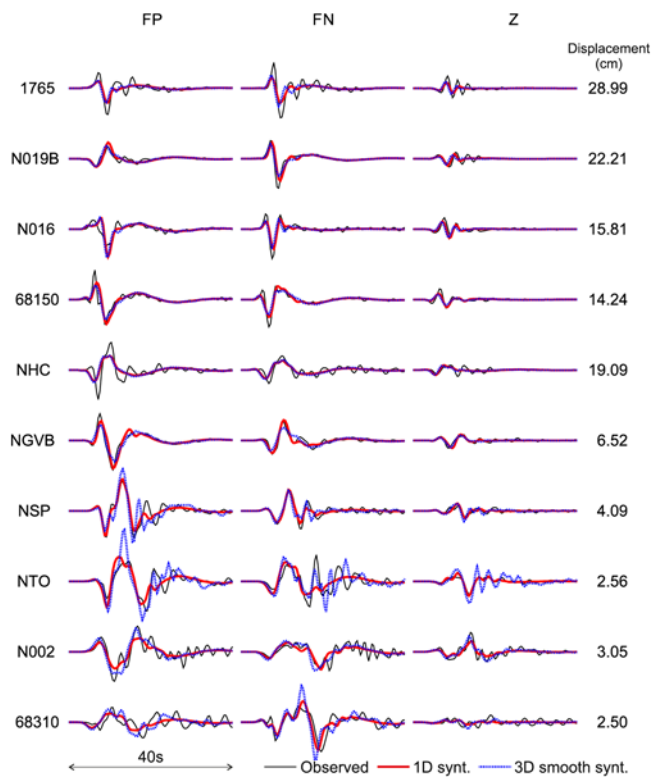


Figure 4. Comparison between observed and synthetic displacement waveforms (0.05–0.5 Hz) based on the 1D GF source model and the 1D or the smoothed 3D velocity model (for details on the smoothing process and sample cross sections, see ⑤ the electronic supplement). The smoothing efficiently suppresses the large oscillations in the 3D synthetics (compare with Fig. 1b). The color version of this figure is available only in the electronic edition.

inversion algorithm, in principle, is able to fit more fine features of the observed waveforms. However, fitting minor wiggles and later arrivals trades off with the simplicity of the rupture model and, at the same time, puts less emphasis on the main waveform pulses.

We carry out an analogous test considering the fault geometry of Wei *et al.* (2015). As before, both source images present consistent basic features (⑤ Fig. S4b,d), although the model based on the 3D GFs is more complex with several secondary rupture episodes. Similarly, ⑤ Figure S5 shows that the spurious oscillations decrease in amplitude, whereas variance reduction does not change significantly (0.60 vs. 0.62, for the 1D and 3D inversion, respectively).

Excluding stations NTO, NSP, and N016 (characterized by the largest late oscillations) from the inversion based on the 3D GFs leads to a source model characterized by less complex slip-rate functions (at least in correspondence with the large slip patch) and a more regular rupture front (see ⑤ Fig. S10a,b). It should be noted that this new model is closer to the 1D GF source image in terms of rise time and final slip (Fig. 1). The variance reduction in this case amounts to 0.68. We note that synthetic seismograms forward simulated at receivers excluded from the inversion (i.e., NSP, NTO,

and N016) are still characterized by large-amplitude surface waves (⑤ Fig. S10c). This confirms that the complexity of the 3D GF source model retrieved by inverting all the data is induced by the necessity of the inversion algorithm to minimize as much as possible the large oscillations characterizing the 3D GFs at these three receivers.

Discussion and Conclusions

We imaged the source responsible for the 2014 M_w 6.0 South Napa earthquake using the linear-inversion technique of Gallovič *et al.* (2015) considering 1D and 3D GFs. The latter were calculated using the USGS 3D velocity model (v. 8.3). We considered single (Gallovič, 2016) and double (Wei *et al.*, 2015) fault-plane geometries, both of which provided a similar fit with the observed data.

Forward simulations in the 3D medium based on source models derived using the 1D GFs (denoted as the 1D GF source models) exhibit strong basin-induced surface waves inside the Napa Valley, within the western side of the Sonoma Valley, and in the area northwest of San Pablo Bay, characterized by a sharp basin edge (see ⑤ Movie S1). These waves are much larger than those observed at stations NSP, NTO (located northwest of San Pablo Bay), and N016 (located within the Napa Valley).

Our tests demonstrate that the large long-period oscillations are excited by a shallow (less than ~ 5 km deep) slip patch, which characterized the Napa earthquake. Hypothesizing a deeper fault that may lead to weaker oscillations does not seem reasonable, because observations of surface slip reaching 40–46 cm (Brocher *et al.*, 2015) indicate that the fault responsible for the earthquake was effectively shallow. Moreover, artificially damping the slip distribution in the shallowest part of the fault does not fully suppress these reverberations. Slip inversion excluding stations NSP, NTO, and N016 leads to a source model that still excites very strong surface waves at these sites.

We found two ways to effectively decrease the amplitude of the large reverberations: (1) by performing inversion using all stations and 3D GFs, and (2) by smoothing the 3D velocity model over a 2-km-wide averaging window. In the former case, the obtained source model is characterized by longer rise times, lower peak slip rates, slightly lower slip values at very shallow depths, and by complex rupture propagation with unrealistic ghost features. However, the fit of the first main pulses worsens with respect to the 1D inversion at several stations.

Contrarily, smoothing the 3D velocity model leads to much better results (details on the smoothing process and sample cross sections are reported ⑤ in the electronic supplement). Indeed, as shown in Figure 4, plugging the 1D source model into the smoothed 3D velocity model results in much weaker reverberations at stations NSP and N016 but still slightly larger than those observed at the NTO station. We note that less satisfactory results are obtained for narrower averaging windows (0.5 and 1 km). Interestingly, tests

based on anisotropic averaging windows (2 km wide, 0.5 km long, and vice versa) have revealed that stronger vertical smoothing is required to damp the reverberations at all stations; smoothing in the horizontal plane leads to weaker damping, particularly at stations NSP and NTO. We note that simply clipping the shear-wave velocity to 1 km/s in the original 3D velocity model does not attenuate the reverberations as efficiently. This suggests that strong surface waves are caused by either too sharp velocity discontinuities in proximity to the basin edges and/or an incorrect shape of the basins themselves.

Although source imaging based on the 3D GFs could, generally speaking, lead to potentially more reliable results, especially in case of source models characterized by multiple rupture patches, we stress that this is only possible if the velocity model is highly accurate (Galović *et al.*, 2015). In particular, our investigations suggest that the USGS 3D velocity model should be improved at least in the area northwest of the San Pablo Bay and in some parts of the Napa Valley.

Our study emphasizes other aspects that go beyond mere source imaging in laterally heterogeneous media. Correctly modeling wave propagation incorporating 3D velocity models is of utmost importance for scenario simulations and hazard assessment, because it is well known that related 3D effects (e.g., basin resonance and channeling of seismic waves) can dramatically impact pattern and level of the estimated ground motion (e.g., Koketsu and Kikuchi, 2000; Olsen *et al.*, 2006; Wang *et al.*, 2008; Ramirez-Guzman *et al.*, 2015). We stress that any 3D velocity model may potentially include some unrealistic features such as exaggerated velocity contrasts or inaccurate basin geometries, which can give rise to strong yet artificial ground shaking. Unfortunately, classic velocity model validations based on waveform modeling of small events are rather limited, because these events do not nucleate at shallow depths. This may represent a critical factor in numerical simulations of large earthquakes characterized by significant shallow slip (e.g., Aagaard *et al.*, 2010; CyberShake experiment in Graves *et al.*, 2011), leading to overestimated seismic hazard at many (basin) locations. Here, we have shown that modeling of moderate-to-strong earthquakes reaching the surface may represent a valuable tool for velocity model validation in the near-source region, provided the finite-extent source models are carefully investigated.

Data and Resources

Recorded seismograms were obtained from the online repository Center for Engineering Strong Motion Data 459 (CESMD, <http://strongmotioncenter.org>, last accessed January 2016). The U.S. Geological Survey (USGS) 3D seismic velocity model v. 8.3 is available at <http://earthquake.usgs.gov/data/3dgeologic/download.php> (last accessed April 2016). The SW4 software package is available at www.geodynamics.org/cig/software/sw4 (last accessed January 2016).

Acknowledgments

W. I. was supported through a contract with the Swiss Federal Nuclear Safety Inspectorate (ENSI). F. G. was supported by the Czech Science Foundation project 14-04372S. Calculations were carried out at the Swiss Supercomputing Center (SCSC) and the King Abdullah University of Science and Technology (KAUST) Supercomputing Laboratory. Constructive reviews by Shengji Wei and an anonymous reviewer led to improvements in the article.

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Manuscript received 8 February 2016;
 Published Online 17 January 2017