Source characteristics of the 2015 $M_w$6.5 Lefkada, Greece, strike-slip earthquake

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Abstract

We present a kinematic slip model from the inversion of 1 Hz GPS, strong motion, and interferometric synthetic aperture radar (InSAR) data for the 2015 $M_w$6.5 Lefkada, Greece, earthquake. We will show that most of the slip during this event is updip of the hypocenter (10.7 km depth) with substantial slip (>0.5 m) between 5 km depth and the surface. The peak slip is ~1.6 m, and the inverted rake angles show predominantly strike-slip motion. Slip concentrates mostly to the south of the hypocenter, and the source time function indicates a total duration of ~17 s with peak moment rate at ~6 s. We will show that a 65° dipping geometry is the most plausible due to a lack of polarity reversals in the InSAR data and good agreement with Coulomb stress modeling, aftershock locations, and regional moment tensors. We also note that there was an ~20 cm peak-to-peak tsunami observed at one tide gauge station 300 km away from the earthquake. We will discuss tsunami modeling results and study the possible source of the amplitude discrepancy between the modeled and the observed data at far-field tide gauges.

1. Introduction

The Lefkada region (Ionian Sea, Greece) is considered among the most active tectonic areas in Europe and one of the most active zones in the eastern Mediterranean region [Papadopoulos et al., 2003; Chousianitis et al., 2016; Illeiva et al., 2016; Ganas et al., 2016; Sokas et al., 2016]. Lefkada is situated close to the Cephalonia Fault Zone (CFZ), a well-known plate boundary with the highest seismicity rate in Greece [e.g., Scordilis et al., 1985; Kiratzi and Langston, 1991; Louvari et al., 1999; Sachpazi et al., 2000; Papadimitriou, 2002], separating Aegean crust to the east from the African plate to the west. Lefkada is located at the end of this major, dextral strike-slip system from where deformation northward is expressed by crustal shortening across NW Greece and Albania (Figure 1) [Anzidei et al., 2001; Ganas et al., 2013].

The most recent significant earthquake related to the northern CFZ, with magnitude of $M_w$6.2, occurred on 14 August 2003 offshore the western coast of Lefkada Island. It caused substantial damage to the island [Papadopoulos et al., 2003; Karakostas et al., 2004; Papathanassiou et al., 2005]. The 2003 earthquake was a shallow strike-slip event; however, no surface rupture was observed in the field [Papathanassiou et al., 2005]. From seismological data [Benetatos et al., 2005, 2007] it was confirmed that this was a NNE-SSW oriented right-lateral event, consistent with the aftershock sequence [Papadopoulos et al., 2003; Pavlides et al., 2004; Karakostas et al., 2004]. The inferred source of the 2003 event is a 20 km long segment of the Cephalonia Transform Fault (CTF) striking NNE-SSW to NE-SW, dipping to the east [Pavlides et al., 2004; Illeiva et al., 2016].

At 07:10:07 UT on 17 November 2015, a $M_w$6.5 event occurred onshore on Lefkada Island (Figure 1a). The hypocenter was determined by the Institute of Geodynamics at the National Observatory of Athens (NOA, http://www.gein.noa.gr) to be ~4 km east of the inferred trace of the Cephalonia Transform Fault (CTF) at 10.7 km depth (hypocentral uncertainty is ±1.5 km). The NOA centroid moment tensor (CMT) solution from waveform inversion of broadband regional stations (http://www.gein.noa.gr) shows a main nodal plane with a strike of 203° and near vertical dip of 88° with predominantly right-lateral motion (rake of 160°). Strong ground shaking was widely felt, and there was substantial landsliding on the steep terrain of the western flank of the island [Ganas et al., 2016]. Although there were numerous environmental impacts from the event, field mapping by Ganas et al. [2016] revealed no evidence of surface faulting.
In this study we present a detailed analysis of the source characteristics of the earthquake. There are numerous regional observing stations part of the geophysical monitoring networks in Greece (strong motion and high-rate GPS (HR-GPS)), which we supplement with interferometric synthetic aperture radar (InSAR) line of sight (LOS) measurements from the Sentinel-1A satellite. From these geophysical observables we invert for a kinematic rupture model. Each data set used has different sensitivities to different parts of the fault geometry, and as will be shown in section 4, the best overall resolution is achieved by combining many different observations. We use the NOA CMT strike for the inversion geometry; however, due to the lack of a polarity reversal in the InSAR line of sight measurements we favor an eastward dipping strike-slip fault rather than a vertical one. In this proposed configuration the fault intersects the surface just offshore Lefkada Island. An analysis of this model shows good agreement with the relocated aftershocks [Ganas et al., 2015, 2016]. We also show that the Coulomb stress transfer model that results from the distributed slip model and inversion geometry agrees well with aftershocks and regional moment tensors in the historical catalogue.

Finally, we note that an ~20 cm peak-to-peak amplitude sea-surface height anomaly was registered at Crotone, Italy, 300 km northwest of the epicentral region (Figure 1b), 1 h after the origin time (OT). This is unusual for a predominantly strike-slip source. We explore the origin of the signal by using the coseismic deformation predicted by the source to model tsunami propagation. Tsunami modeling with kinematic coseismic sources inferred from land-based geophysical observables can accurately reproduce observed tsunami propagation [Melgar et al., 2016]. We find that the model predicts the amplitude for the tide gauge at Crotone to be 2 orders of magnitude smaller and its arrival to be 30 min earlier than what was observed during this event. We thus investigate potential causes for this disagreement, such as landsliding and local amplification effects at Crotone.

### 2. Data and Methods

We use local and regional geophysical data (high-rate GPS and strong motion) supplemented by InSAR observations to model the kinematic characteristics of the earthquake rupture. We also rely on far-field observations at tide gauges across the Mediterranean to study the tsunami.
2.1. Observations and Processing

Geophysical data consist of three component recordings at three 1 Hz GPS and four strong motion stations (Figure 1). Twenty-one waveforms in total constrain the event kinematics (Figure S1 in the supporting information).

2.1.1. GPS and Strong Motion Waveforms

The GPS stations are located in the near-field and are part of the national network of NOA [Ganas et al., 2011]. GPS epoch-wise positions are determined with the precise point positioning with ambiguity resolution (PPP-AR) method of Geng et al. [2013]. Final GPS satellite orbits, Earth rotation parameters (ERPs), and 1 Hz GPS satellite clocks were obtained from Center for Orbit Determination in Europe (ftp://ftp.unibe.ch/aiub/CODE/2015/) and are fixed to estimate 1 Hz satellite clocks and 30 s fractional-cycle biases with 43 reference stations located over 1000 km from Greece. We then fix the orbits, clocks, and ERPs to enable precise point positioning with ambiguity resolution (PPP-AR) to estimate epoch-wise positions along with zenith troposphere delay parameters, receiver clocks, and integer-cycle phase ambiguities for the high-rate stations. All position estimates are with respect to the International Terrestrial Reference Frame 2008. Solid Earth tides, ocean tide loading, and pole tide are applied, and antenna phase center variations are also corrected. The strong motion data are instrument-corrected to convert from digital counts to physical units; we remove the preevent DC-offset, integrate the data to velocity, and band-pass filter the waveforms between 0.05 Hz and 0.5 Hz to eliminate baseline offsets [Boore and Bommer, 2005].

2.1.2. InSAR Line of Sight

We also generated InSAR images of the event. We processed two pairs of Copernicus Sentinel-1A satellite (operated by the European Space Agency) synthetic aperture radar (SAR) images acquired in TOPS (Terrain Observation by Progressive Scans) mode by using a prototype processor added to the InSAR Scientific Computing Environment software [Rosen et al., 2012]. The ascending and descending preevent images were collected on 5 and 11 November 2015 respectively. The repeat, postevent ascending and descending scenes are from 17 and 23 November 2015, respectively. We use the 1 arcsec Shuttle Radar Topography Mission (SRTM) version 3 digital elevation model to coregister and resample the TOPS mode data. We then produce line of sight (LOS) deformation maps by phase unwrapping the interferograms using SNAPHU [Chen and Zebker, 2002]. The final LOS maps are geocoded to the same digital elevation model. Figure 2a shows the ascending interferogram; the results are contaminated by a residual trend. In order to better image the deformation at the epicentral area we mask out a region close to the source (black rectangle in Figures 2a and 2b) and fit a second-order curved surface to the far-field LOS map. We remove this surface to produce a corrected LOS map (Figures 2b and 2c) that more accurately measures the coseismic deformation. We then subsample the LOS map by using the QuadTree algorithm (Figure 2d) [Lohman and Simons, 2005]. We only utilize LOS measurements from the ascending interferogram, being very close to by comparison between the measured LOS offsets and the ones calculated by projecting the GPS offsets from the three high-rate stations onto the LOS direction (Figure 2e). The near-source LOS offsets are devoid of a residual trend and agree well with the GPS measurements. However, the LOS offsets to the south do not agree with the GPS offsets at station VLSM; thus, we have disregarded LOS measurements further afield in the inversion. The same procedure was followed for the descending interferogram (Figure S2). We do not observe good agreement between the LOS offsets from interferometry and the GPS static offsets projected to the satellite look direction. This could be due to remaining long-period noise in the interferogram or early afterslip since the repeat-pass was collected 6 days after the event. Whatever the cause, due to the large discrepancies between the GPS and the LOS offsets, we consider that the descending interferogram is not suitable for modeling the coseismic source process.

2.1.3. Sea Level Data

We also analyzed tide gauge data from stations around the Mediterranean at regional distances from the event (Figures 1b and S1). To observe and model the tsunami signal we have removed tides from 6 h long time series at each station. There are 10 tide gauge stations, of which 8 were operational on the day of the earthquake.

2.2. Kinematic Inversion and Tsunami Modeling

We use the MudPy open source code [Melgar and Bock, 2015] to jointly invert the high-rate GPS waveforms, strong motion seismograms, and InSAR LOS offsets to produce a kinematic rupture model. We consider the 1-D regional velocity model of Haslinger et al. [1999] and the NOA hypocenter as the nucleation point of the
rupture. We define the fault as an 80 km long planar surface extending bilaterally from the hypocenter. We use a strike of 20° and a dip of 65° to the south-east. This dip is consistent with relocated aftershocks [Ganas et al., 2015]. Furthermore, this dip value is compatible with the absence of a polarity change in the InSAR LOS measurements onshore (Figure 2), suggesting that the surface projection of the fault plane is situated off the western coast of Lefkada (Figure 1). Further discussion on the dip angle will be offered in section 4.

The fault is discretized into 2 km long by 1.5 km wide elements for a total of 480 subfaults. We compute elastodynamic Green’s functions (GFs) for every subfault/station pair by using the frequency wave number

Figure 2. Descending raw unwrapped ascending InSAR line of sight (LOS) measurements. (a) The area in the black rectangle is masked out for correction. (b) Corrected LOS map after fitting a second-order surface to the area outside the black rectangle and removing it. (c) Close-up of the epicentral area after correction and location of nearby GPS stations. (d) LOS measurements after QuadTree subsampling. (e) Comparison between LOS measurements at the locations of the GPS stations and the LOS-predicted displacements from projecting the GPS offsets onto the satellite look direction.
integration technique of Zhu and Rivera [2002]. Three-component (north, east, and up) static GFs are computed with the same code and projected onto the satellite look direction to obtain LOS GFs at the location of every pixel in the InSAR scene. To allow variability in rupture onset times, and thus rupture speed, we use a multitime window approach and allow slip on five 50% overlapping triangles with 1 s risetimes. Kinematic inversion is an ill-posed problem; thus, spatial regularization is achieved though minimum norm smoothing of the model parameters (total slip at each subfault), and temporal smoothing on the amplitude of successive time windows at each subfault is achieved with first-order forward finite differences. Moment tensor solutions from regional and global data indicate predominantly right-lateral strike-slip motion. Thus, we constrain the rake of our solutions to be in the 135° to 225° range and use nonnegative least squares to ensure positivity.

Data weights are critical to the resulting model. First we weigh each of the time-dependent waveforms (GPS and strong motion) by the standard deviations of 60 s of pre-event noise. For the HR-GPS the standard deviations are 0.2–2 cm, while the vertical standard deviations about 3–5 times larger than the horizontal values. For the InSAR we calculated the standard deviation of the far-field signal (4 cm) after removing the second-order surface. After removing this trend only short-length-scale noise remains [Lohman and Simons, 2005]. These weights are an expression of the quality of each measurement and reflect its trustworthiness. Measurements with a smaller standard deviation will be preferentially fit.

Aiming at assessing the relative contributions of each type of data, a second set of weights, is determined by trial and error. First, we normalized each data set by dividing it by its particular L2 vector norm; otherwise, one type of data can dominate the inversion. For example, the norm of the data vector of HR-GPS waveforms is 2.6, while the norm of the vector of strong motion waveforms is 1.2; this is simply due to the fact that the GPS waveforms have larger relative amplitudes, with peak values as high as 0.5 m (Figure S1); meanwhile, the filtered strong motion data (after integration to velocity) have smaller peak amplitudes (~0.2 m/s); thus, if no normalization is applied, then the inversion will preferentially fit one data set (GPS) over another. After normalizing, each data set is graded as equally important in the inversion. However, this might not yield the optimal resolution or fit to each individual data set. It is important to try different relative data weights and strive to attempt the highest possible fits to each data type. To do so, we perturb the relative weight of each data type, perform the inversion, and examine the changes to the misfit of each data type. This trial and error procedure leads to adjustments in the weights such that each data set is independently fit to the highest possible level. Thus, our inversion scheme has two sets of weights; the first is meant to represent the quality of each measurement and the second one is selected to maximize the fit to each individual data set.

We note that this is a cumbersome process, and more work is necessary to develop automated and objective procedures of determining the weights that should be assigned to each data type. A good discussion of this difficulty and sometimes ambiguous part of the inversion procedure as it pertains to multidata slip inversions can be found in Yue et al. [2015]; there is an opportunity here for substantial improvement. For example, one potential avenue would be to study the impacts of changing relative weights on the model resolution matrix; this issue remains the subject of future research.

After the kinematic slip inversion we model the ensuing tsunami at a regional scale (Figure S2). To do so, we calculate the seafloor deformation predicted by the slip model on a regular grid with nodes spaced every 250 m around the fault. We account for the horizontal advection of bathymetry as well as for coseismic uplift or subsidence [Melgar and Bock, 2015]. This is then the initial condition that is used in a nonlinear shallow water code Geoclaw [LeVeque et al., 2011] to model tsunami propagation. We collect the tsunami model output at 60 s intervals and model a total propagation time of 6 h. We also collect synthetic sea surface height time series at the locations of the tide gauges in Figure S3. The simulation is carried out by using the SRTM15+ (~450 m sampling) topography and bathymetry data set [Becker et al., 2009].

3. Results

In the following sections we show results from the kinematic slip inversions and tsunami modeling.

3.1. Kinematic Slip Inversion

The total slip from the inversion and event source time function is shown in Figure 3. The inversion has a moment of \(8.36 \times 10^{18}\) N·m (\(M_w6.54\)), and the majority of the slip is updip of the hypocenter (10.7 km
depth) with substantial slip (＞0.5 m) between 5 km depth and the surface. The peak slip is ~1.6 m, and the inverted rake angles show predominantly strike-slip motion. The source is compact, with slip concentrating mostly on the south of the hypocenter in an area of roughly 30 km by 10 km. The source time function indicates a total duration of ~17 s with peak moment rate at ~6 s.

Snapshots of the time dependence of rupture (Figure 4) display interesting behavior. After nucleation, between 0 and 5 s the rupture propagates updip toward the surface. Upon reaching the surface it spreads bilaterally with the bulk of slip to the south of the hypocenter. The rupture propagation pattern shows that the large slip regions are consistent in space with areas where there are subsequently numerous aftershocks. Waveform fits for the HR-GPS and strong motion data are shown in Figure S1, and InSAR residuals are shown in Figure 5. The fits to the data sets are good; variance reductions to the HR-GPS, strong motion, and InSAR are 72%, 68%, and 87%, respectively. We note that despite the good fits to the InSAR data (Figure 5) there is still a visible residual pattern to the south. We ran the inversion at seven different rupture speeds between 2.0 and 3.2 km/s and found that the best data fits occur at 2.6 km/s.

3.2. Tsunami Modeling

In order to model tsunami propagation we calculate the vertical coseismic deformation predicted by the slip inversion in Figure 3. The result is shown in Figure 6. There is modest peak vertical deformation of ~0.12 m.

![Figure 3](image-url)

**Figure 3.** (a) Total slip from the inversion. The green star is the hypocenter, the black dots are relocated aftershocks [from Ganas et al., 2015] projected onto the fault plane, and the green arrows denote the direction of slip. (b) Event source time function.

![Figure 4](image-url)

**Figure 4.** One-second snapshots of rupture propagation. The green star is the hypocenter, the black dots are aftershocks relocated by NOA projected onto the fault plane, and the grey circles are reference rupture fronts moving out at 1 km/s, 2 km/s, and 3 km/s.
Because the surface trace of the fault is offshore and substantial slip beneath the open-sea region south of Lefkada has been implicated implied by the current work, the area contributing to tsunami generation is substantial. The peak tsunami amplitude in the first 6 h is fairly modest (Figure S3) with the largest modeled tsunami amplitude of ~0.1 m arriving at the coastal regions adjacent to the source. Modeled tsunami amplitudes farther away are substantially smaller.

Figure 7 shows the comparison between the observed tide gauge record at three tide gauges in Italy (locations on Figure S3), Crotone (Figure 1b) (CR08), Otranto (OT15), and Taranto (TA18), at distances between 200 and 300 km from the source, and the prediction of the tsunami model at each tide gauge. The observed tide gauge amplitude is larger than the modeled tsunami amplitude at the Crotone tide gauge. At the Otranto and Taranto tide gauges there is no easily discernible tsunami signal in the observations; this is consistent with the model which predicts amplitudes smaller than 0.2 cm, well below the ~1 cm noise level of the instruments. At Crotone we note that the arrival time of the tsunami, as predicted by the model, is at ~30 min after the origin time, while the observations do not show a signal above the tide gauge noise level until ~60 min after OT. At the remaining operational tide gauges in Sicily, Malta, and Western Greece, further afield from the source, there is no visible tsunami signal in the tide gauge records.

4. Discussion

The kinematic rupture model shows most of the slip concentrating at shallow depths (<5 km); however, the bulk of the aftershocks are deeper than that. This suggests that there are different material properties between the deep and shallow segments of the fault. The best fitting rupture speed (2.6 km/s) is close to the shear-wave speed of the elastic model; however, it is notable that rupture is divided into a first episode where it is mostly updip of the hypocenter before a second episode where it expands bilaterally along the shallow portion of the fault; a similar source processes was noted by Sokos et al. [2016] from multiple point source analysis.

In order to determine the reliability of our interpretations we performed a checkerboard test of the inversion in two steps: (a) using GPS, strong motion, and InSAR data independently and (b)
using the joint data set. We generated synthetic slip distributions with 4 km slip patches, forward modeled the data, and added Gaussian noise with the same standard deviations as the data estimates (section 3). We then inverted this synthetic data set and studied how well the input checkerboard is recovered. The results are shown in Figure 8. There, it can be observed that all data sets present good resolution for the shallow fault patches (0–3 km). The strong motion data has somewhat lower resolution for the remaining parts of the model. The HR-GPS has substantial smearing between 5 and 13 km depth but high resolving power for the deep portion of the model (>13 km). Meanwhile, the InSAR data have diminishing resolution with depth, and low resolution to the north and south where the fault is furthest from the measurements. The joint inversion shows improved resolution over the individual data set inversions. We note that low resolution does not mean that the inversion will not place slip in these areas, rather any actual slip during the earthquake in an area of low resolution will present in the inversion as being smeared over several subfaults instead of being precisely located.

In order to further determine which features of the inversion are robust we conducted a jackknife test. We randomly removed 20% of the data and ran the inversion with the same parameters on the reduced data set. We repeated this process 200 times; the mean model, the standard deviation, and the coefficient of variation (the standard deviation divided by the mean) are shown on Figure 9. The standard deviation indicates the absolute variability of slip on any given patch. It shows, for example, that the peak slip of 1.6 m has a standard deviation of ~7 cm. To determine which features are persistent and required by the data, the coefficient of variation (CV) is a better measure. Areas with a high CV vary wildly about their mean value, while areas with low CV have a persistent value of slip and are thus required by the data. Figures 8 and 9 argue that the overall pattern of slip, concentrated predominantly between 10 and 20 km depth along a NNE-SSW strike, and with large shallow slip (<3 km), is reliable since none of the slip is an area of large smearing or high CV.

Figure 7. Five hours of observations at the Crotone (CR08), Taranto (TA18), and Otranto (OT15) tide gauges in Italy (locations are in Figure 1b). The earthquake origin time is indicated by the black dashed line. The red tide gauge record is the synthetic predicted by the tsunami propagation model of Figure 1b. Note that the vertical scales are different for observed and synthetic waveforms.
Chousianitis et al. [2016] inverted regional seismic and GPS as well as teleseismic data. They find a slip pattern that broadly agrees with the results shown here (Figure 4), but note that the main slip segmented into two smaller patches of ~5 km separated by ~5 km. We do not observe this in our inversion, and we note that given the results of the checkerboard analysis it would be difficult with the available data and our model parameterization to robustly identify two patches of this dimension.

We also computed the static stress changes due to variable slip across the fault plane (Figure 10). An isotropic elastic half-space is assumed to represent crustal rheology. The static stress transfer is calculated on optimal strike-slip receiver faults relative to the ENE-WSW directed compression [Ganas et al., 2013] at 6, 8, and 10 km depths on a 100 × 100 km grid surrounding the epicenter using the Coulomb 3 code [Toda et al., 2011]. Our results correlate well with the distribution of relocated aftershocks [Ganas et al., 2016] and indicate that optimally oriented strike-slip faults to the north and south of the epicenter have been loaded with stresses up to 2 bars (0.2 MPa), while faults in other areas (mostly across-strike) have been subjected to stress decrease (Figure 9). The stress increases we report could also trigger events to the south of Lefkas where active strike-slip faults are favorably oriented to the regional compression Kassaras et al. [2016].

Our modeling suggests strongly that the fault that ruptured during this event has a substantial dip (65°–70°). This is a key difference with previous findings from Chousianitis et al. [2016] and Sokos et al. [2016], who employ the steeper dipping fault (~80°). This shallow dip, although somewhat unusual for a strike-slip fault, is supported by the data. It is compatible with the arrangement of hypocenters and focal mechanisms of aftershocks (Figure 10) and has been observed elsewhere, e.g., for the 2014 M₉.6 Napa, California, earthquake [Dreger et al., 2015; Melgar et al., 2015] and M₉.7 Balochistan earthquake [Jolivet et al., 2014]. The InSAR observations strongly advocate for a dipping geometry; the hypocenter is well onshore (even considering the ±1.5 km uncertainty) at 10.7 km depth, and the LOS measurements have no polarity change. For a subvertical strike-slip fault striking at 20° the projection of the fault plane to the surface should delineate an obvious change in polarity of measured LOS across the surface projection. This can be more clearly seen in Figure 11. We have modeled a simple M₉.6.5 rectangular 30 km by 10 km fault with uniform slip dipping at

Figure 8. Checkerboard test of the inversion resolution. The input pattern has alternating patches of 1 m of slip and no slip. The results of inverting each data set individually and the joint inversion of all three data sets are shown. The green star is the hypocenter.
65°, 70°, and 75° and calculated the LOS offsets after projecting the predicted coseismic deformation field onto the satellite look direction. In all cases the surface trace marks an obvious change from positive to negative LOS offsets. This is simply not observed on Lefkada Island; thus, while there is leeway in the modeling for different geometries it is unlikely that a steeply dipping near-vertical fault ruptured in this event. While the geographic extent of our slip pattern is broadly in agreement with other studies [Chousianitis et al., 2016; Sokos et al., 2016] we conclude that the fault must have a substantial southeastward dip and its surface trace be offshore. Furthermore, fault plane solutions from a new catalogue compiled by Kassaras et al. [2016] (Figure 10) favors a >60°–70° eastward dipping fault geometry in this region.

As noted, the predicted tsunami at Crotone (Figures 1b and 7) from the coseismic deformation produced by the earthquake (Figure 6) is far too small and there is no clear tsunami observation at the neighboring tide gauges in Otranto and Taranto. There are two main possibilities for the amplitude discrepancy. First, that there is significant contribution from secondary tsunami sources such as landslides, either aerial or subaerial. These can be important tsunamiogenic contributors [i.e., Tappin et al., 2014]. Indeed, during the Lefkada earthquake there was substantial landsliding on the western coast of the island [Ganas et al., 2016]. This is reflected in the InSAR observations (Figure 2c), and the areas on the western coast of Lefkada with no LOS data are areas of decorrelation in the interferogram produced by mass wasting events. Generally, the smaller areal extent of a landslide tsunami source and strong dispersive effects from a reduced source area leads to faster attenuation than for the broader coseismic uplift portion of the source [Ward, 2001]. Thus, in order for the

Figure 9. Jackknife testing. Results of removing 20% of the data 200 times. (top) The mean model, (middle) the standard deviation, and (bottom) the coefficient of variation.
tsunami to reach the South coast of Italy ~300 km away a substantial contribution from landsliding is necessary. Furthermore, the onset of the tsunami at CR08 is delayed by ~30 min. This could be explained by delayed subaerial landsliding triggered by one of the numerous aftershocks, but there are no reports of this and the lack of a discernible tsunami signal at the other tide gauges in Italy seems to preclude this possibility. Future marine geophysical surveys of the region will elucidate the matter.

The second potential explanation for the large amplitude observed at CO08 is that the tsunami is due to the local effect of bathymetry and coastline shape at Crotone; it has been shown that local features can substantially amplify tsunamis [i.e., Horillio et al., 2008]. Figure 12 shows an analysis of 30 days of the 1 min sampled preevent data at the CR08 tide gauge. The short-period power spectral density (PSD) of this time series has peaks at 6–9 min and 11–12 min periods. This suggests that due to local effects, sea level oscillations at these short periods are preferentially amplified. Figure 12 also shows the spectral analysis for a 6 h record around the time of the earthquake. The spectrum shows a large peak between 6 and 7 min periods, and there is some energy at the 11–12 min periods as well suggesting preferential amplification. It is possible that the small-amplitude tsunami coseismically generated by the Lefkada earthquake was, over the first few tens of minutes after its arrival at Crotone, which corresponds to 1–5 oscillation periods, amplified due to the natural resonant periods of that portion of the coastline, until it reached its peak amplitude of 10 cm. This also accounts for the delayed onset, as the resonance grew there was some delay time until the signal was large enough to overcome the noise level of ~1 cm at the tide gauge.

We consider this second explanation more likely for the tsunami observed at Crotone, but we note that we cannot, with the observations available, unequivocally attribute the tsunami to this mechanism; the observed signal at the Crotone tide gauge station is intriguing, but the issue remains unresolved and the topic of future

Figure 10. (a) Coseismic Coulomb stress changes at 8 km depth. The black dots are the 30 day aftershock sequence of Ganas et al. [2015]. The green line denotes the surface projection of the fault trace, and the yellow star denotes the epicenter. AB and CD denote cross sections across the model. (b) Cross sections along strike (AB). The black dots denote aftershocks, and the dashed line is the calculation depth used in Figure 10a. (c) Cross-dip section (CD). The yellow line is the fault geometry used for the slip inversion (Figure 1). (d) CD cross section with focal mechanisms [Kassaras et al., 2016]. The dashed red line denotes an ~62° dipping fault plane inferred from the focal mechanisms [Kassaras et al., 2016]. The green and red beach-balls are the projected revised moment tensor solution from NOA and the one derived from the fault geometry used for slip inversion, respectively (Figure 1).
research. We note that understanding the tsunamigenic behavior of offshore strike-slip fault systems is important for a complete assessment of hazard. There is vertical coseismic motion during strike-slip faulting, however modest, which could generate small-amplitude tsunamis that can be preferentially amplified if local features favor it. There are other offshore strike-slip fault systems in close proximity to heavily populated coastlines, for example, in the California borderlands [Sahakian et al., 2017], and should there be more evidence that strike-slip faulting can contribute to tsunami hazard in this way then these sources should be considered in hazard assessments.

5. Conclusions

We present a kinematic slip model for the 17 November 2015 Lefkada earthquake from the inversion of GPS, strong motion, and InSAR data. Our results show that most of the slip is updip of the hypocenter (10.7 km depth) with substantial slip (>0.5 m) between 5 km depth and the surface. The peak slip is ~1.6 m, and the inverted rakes show predominantly strike-slip motion. Slip concentrates mostly to the south of the hypocenter, and the source time function indicates a total duration of ~17 s with peak moment rate at ~6 s. We have also discussed the small-amplitude tsunami observed at Crotone, Italy, ~300 km away from the event. There is a 2 order of magnitude discrepancy between the modeled amplitudes and the observed ones. We suggest that the tsunami observed at Crotone is due to local amplification from the coastline geomorphology and local bathymetric features.
Figure 12. Time series analysis of 1 min sampled data at the Crotone, Italy tide gauge (CR08). (top) A 30 day record before the earthquake and its short-period power spectral density (PSD). (bottom) Six hours of the tide gauge recorded around the time of the earthquake with its short-period PSD.

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