The 2018 $M_w$ 6.8 Zakynthos, Greece, Earthquake: Dominant Strike-Slip Faulting near Subducting Slab

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Abstract

With different styles of faulting, the eastern Ionian Sea is an ideal natural laboratory to investigate interactions between adjacent faults during strong earthquakes. The 2018 $M_w$ 6.8 Zakynthos earthquake, well recorded by broadband and strong-motion networks, provides an opportunity to resolve such faulting complexity. Here, we focus on waveform inversion and backprojection of strong-motion data, partly checked by coseismic Global Navigation Satellite System data. We show that the region is under subhorizontal southwest–northeast compression, enabling mixed thrust faulting and strike-slip (SS) faulting. The 2018 mainshock consisted of two fault segments: a low-dip thrust, and a dominant, moderate-dip, right-lateral SS, both in the crust. Slip vectors, oriented to southwest, are consistent with plate motion. The sequence can be explained in terms of trench-orthogonal fractures in the subducting plate and reactivated faults in the upper plate. The 2018 event, and an $M_w$ 6.6 event of 1997, occurred near three localized swarms of 2016 and 2017. Future numerical models of the slab deformation and ocean-bottom seismometer observations may illuminate possible relations among earthquakes, swarms, and fluid paths in the region.

Introduction

Multiple faults acting during an earthquake have been generally well known, but on global scale less observations have been available for near-simultaneous ruptures of different faulting mechanisms, specifically in subduction zones. For example, Lay et al. (2013) reported a doublet of two $M_w \sim 7$ events below Japan trench, where a thrust faulting (TF) along the subduction interface was followed after 14 s by a shallower normal faulting in the overriding plate. A rare evidence of a thrust event on a plate interface, which triggered a normal-faulting event in the overriding plate was provided for an $M_w \sim 7$ earthquake in the Chile subduction zone (Hicks and Rietbrock, 2015). Particularly challenging in terms of strain partitioning are the regions where subduction terminates, and plate motions continue along transform faults. A good example is the 2016 $M_w$ 7.8 earthquake in New Zealand, in which a dominant strike-slip (SS) faulting occurred in the upper plate and possibly triggered minor slip on the underlying subduction thrust (Mouslopoulou et al., 2019; Ulrich et al., 2019). Resolving fault complexity for strong earthquakes ($M_w$ 6.0–6.9) in the shallowest parts of the subduction termination zones is even more challenging, and this article focuses on such a task in western Greece. The major ongoing convergent tectonic process in western Greece is subduction, imaged by seismic tomography and other structural studies (Spakman et al., 1993; Laigle et al., 2004; Suckale et al., 2009; Sachpazi et al., 2016; Halpaap et al., 2019). The active overriding of the Aegean plate over the subducting African plate, derived from Global Navigation Satellite System (GNSS) data, is oriented toward southwest, approximately perpendicular to the trench, being consistent with slip vectors of many earthquakes (Kiratzi and Louvari, 2003; Hollenstein et al., 2006; Shaw and Jackson, 2010).

Zakynthos (or Zante) Island is situated at a subduction-termination zone, a part of the Ionian Islands–Akarnania Block (IAB) (Pérouse et al., 2017). At the northwest edge, this block is separated from Apulian–Ionian microplate by the right-lateral Cephalonia transform fault. The southwest boundary of IAB is the Hellenic subduction backstop front. The northeast boundary of IAB is a mixture of SS and extensional structures (e.g., the Corinth Gulf). The southeast boundary of IAB has not been well known until the 2008 $M_w$ 6.3 Movri Mountain earthquake.

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Supplemental Material
enced three of the above-cited reflector depths. The latter study illuminated the spatial variation of the Moho depth, and provided a layered velocity model consistently used throughout this article. A weakly northeast-dipping (∼5°) seismic reflector has been detected at the depth of ∼10–15 km, supposedly mapping the top of the subducting plate in the area west and southwest of Zakynthos (Clément et al., 2000; Laigle et al., 2004), situated ∼10 km above Moho (Halpaap et al., 2019). Lacking more detailed information, in the following parts of this article we use the well-mapped Moho depth of Papoulia et al. (2014) and plot the slab top 10 km above the Moho. It provides an approximate (schematic) location of the slab top, at the depths 10–20 km, close to the above-cited reflector depths.

In the instrumental era, the region near Zakynthos experienced three Mw 6–7 earthquakes, roughly every 20 yr (1959, 1976, 1997, and 2018), consistently with a high seismic coupling (Laigle et al., 2002; Chousianitis et al., 2015). No historical M > 7 event has been documented. On 25 October 2018, at 22:54 UTC, an Mw 6.8 earthquake occurred southwest of Zakynthos. It caused limited damage on the island and no casualties (Institute of Engineering Seismology and Earthquake Engineering [ITSAK], 2018). The event was observed globally, and its broad characteristics were soon outlined as follows. The Global Centroid Moment Tensor (Global CMT) project suggested a centroid depth of 12 km, scalar moment M0 = 2.3 × 1019 N · m, and strike/dip/rake angles of 13°/24°/165°, an oblique-TF mechanism. The Global CMT solution comprised a significant non-double-couple (non-DC) component, namely a compressional compensated linear vector dipole, CLVD = −44%. The European data centers reported centroid depths <20 km, with focal mechanisms ranging from the mixed SS and thrust type to an almost pure SS, often with a notable non-DC component. For example, the National Observatory of Athens (NOA) published the moment tensor (MT) with CLVD of −61%. Interestingly, an Mw 4.8 foreshock was a pure low-dip thrusting mechanism (strike/dip/rake ∼300°/10°/100°), similar to four major Mw 5+ aftershocks; however, smaller aftershocks were of both types, thrust and SS. Our preliminary analysis, reported to European Mediterranean Seismological Centre two weeks after the event, speculated about a segmented fault (Zahradník et al., 2018).

The goal of this article is to improve understanding of the complex faulting style taking place near Zakynthos, complementing our previous earthquake studies of the Ionian Sea islands of Lefkada and Cephalonia (Sokos et al., 2015, 2016). To this goal, we analyze source process of the 2018 mainshock and aftershocks using regional broadband, accelerometric, and GNSS data, considering also the 2011–2018 seismicity of the region. We interpret the mainshock in terms of a segmented source model, possibly related to trench-orthogonal fractures in the subducting plate and reactivated faults in the upper plate.

Source Modeling

Point-source models of mainshock and aftershocks

The mainshock nucleated ∼45 km southwest of Zakynthos, close to a local bathymetric low (the sea depth of 4 km; see Fig. 1). Exact hypocenter position is unknown because a small foreshock of an unknown position and magnitude preceded the mainshock by a few seconds, thus complicating the arrival-time picking. For the same reason, the first-motion polarities are problematic. We made a probabilistic location (Lomax et al., 2001), see Text S1, and Figures S1 and S2 (available in the supplemental material to this article), pointing to a shallow depth, and hereafter we use the epicenter (latitude/longitude 37.27°/20.43°) corresponding to the arbitrarily fixed source depth of 5 km, with origin time of 22:54:47.5 UTC. None of the following modeling methodologies relies on the particular hypocenter position.

A significant Mw 4.8 foreshock occurred 32 min before the mainshock (Fig. S1). The mainshock was followed by a standard exponentially decaying aftershock sequence that we first located with Hypoinverse (Klein, 2002), and then relocated with hypDD code (Waldhauser, 2001). Their median formal errors are ∼1 km, and a few hundred meters, respectively. The sequence included one Mw 5.1 event 15 min after mainshock, and four other events of Mw 5+ in the first week. After ∼80 days, another, smaller sequence of Mw < 4 appeared (see Fig. 1b), situated basically in the same region as the mainshock. In the first day after the mainshock, the activity occupied an area of ∼20 × 20 km, lacking any clear spatial pattern. Later, a much broader area (∼60 × 60 km) was activated; at least three aftershock streaks of the southwest–northeast orientation appeared near Zakynthos, gradually intensified with time, and another similarly oriented streak was created south of the mainshock since the beginning of the sequence (Fig. 1). This pattern is stable with respect to the used velocity models. However, relative to our reference model (Papoulia et al., 2014),
a systematic mean vertical shift of about 3 km downward and 5 km upward occurs in the models of Papadimitriou et al. (2012) and Haslinger et al. (1999), respectively. The activated volume has the 10–20 km depth range. Considering the location inaccuracies and the limited knowledge of the slab top, the events occurred at or near the slab top, but definitely above the Moho determined by Papoulia et al. (2014). The aftershock streaks were rather vertical fault structures in the upper plate (Fig. 2).

The MT of the mainshock, particularly the strike/dip/rake angles (hereafter, s/d/r) are much better determined than the hypocenter because we calculate these parameters by inverting relatively low-frequency seismic waveforms (Sokos and Zahradník, 2013; Zahradník and Sokos, 2018). Using broadband stations in Greece and Italy at distances 270–630 km and frequencies 0.01–0.02 Hz, our centroid appeared at latitude/longitude = 37.39°/20.63°; it was an event of the “odd” (mixed) type (Frohlich, 1992), with a large negative CLVD component (s/d/r = 12°/41°/165°, $M_0 = 1.7 \times 10^{19}$ N·m, DC ~ 40%, CLVD ~ −60%; see Fig. 1 and Table S1).

MTs of the strongest aftershocks were also calculated by full-waveform inversion (Table S2). We obtained a variety of focal mechanisms, ranging from TF, to SS, and mixed types. Their spatial distribution indicates that in the western part of the sequence we observe mostly TF, whereas in the streaks we have rather SS. The focal mechanisms provide an estimate of stress field (Vavryčuk, 2014). The $\sigma_3$ axis of the maximum compression is well resolved featuring plunge $\sim 20^\circ$ and azimuth of $\sim 230^\circ$ (see Fig. 1b). This azimuth is close to the plate-motion vectors (Hollenstein et al., 2008). The eigenvalues corresponding to the stress axes $\sigma_2$ and $\sigma_3$ have similar magnitudes, the shape ratio is $\sim 0.75$, and hence the orientation of these two axes is rather ambiguous, explaining the coexistence of the TF and SS events in the region.

Before constructing a finite-fault model, it is useful to understand the shear (pure-DC) part of the rupturing process. To this goal, we made multiple-point source (MPS) modeling in a DC-constrained mode (Zahradník et al., 2005, 2017; Zahradník and Gallovič, 2010; Zahradník and Sokos, 2014; Sokos et al., 2016). Full waveforms of 11 near-regional strong-motion accelerometer stations were inverted, using a variety of trial source positions, velocity models, and frequency ranges (e.g., Figs. S3 and S4). They robustly indicate that the largest DC point-source contribution (major subevent), situated near the centroid, has a moderately dipping SS mechanism, $s/d/r \sim 10^\circ/40^\circ/180^\circ$, hereafter simply called SS. Although the other significant point-source subevents were less stable in terms of their position and mechanism, many featured a low-dip TF type, for example, $s/d/r \sim 300^\circ/10^\circ/60^\circ$, acting almost simultaneously with the SS subevent (within $\sim \pm 5$ s).
The idea of the mainshock source process involving the TF and SS contributions is plausible for several reasons: (1) the TF mechanism is an obvious candidate for a large event in subduction environment. (2) Both TF and SS mechanisms were observed in the aftershock sequence. (3) The TF + SS combination could explain the large negative CLVD of the mainshock. (4) Despite the different fault types, the indicated TF and SS subevents share similar orientation of slip vectors, if both rupture their low-dip nodal planes. Then, the latter indicates that the moderately dipping SS fault is right lateral.

**Segmented fault model of the mainshock**

We used the linear slip inversion (LSI) method (Gallovič et al., 2015; Gallovič, 2016; Mai et al., 2016; Pizzi et al., 2017), in which the rupture process is discretized in space and time along the assumed fault planes. Model parameters are the spatial–temporal slip-rate samples, spanning the whole rupture duration. The inversion result is thus presented in terms of full slip-rate functions instead of derived kinematic rupture parameters such as rise time, rupture speed, and so on. The same accelerometer stations as in the MPS modeling were used (see Fig. S5). Importantly, the hypocenter position is not used in the LSI method, and thus its only rough knowledge, for the Zakynthos earthquake, does not affect the performance of the inversion. Nevertheless, slip inversion is uneasy, possibly containing artifacts, because the epicentral

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**Figure 2.** Aftershocks. (a) Activity in the 120 days after mainshock and map view of the northwest–southeast (NW-SE) and SW-NE profiles. Thick black lines refer to major tectonic structures (as in Fig. 1). (b) Vertical cross sections along the NW-SE (left) and SW-NE (right) profiles with events situated a few kilometers off the profiles, as indicated by the ± symbols. The surface topography (black), schematic slab top (blue dashed line), and the Moho (red line) are also shown. Note that toward Zakynthos Island the activity is gradually shifted from the slab top to the upper plate, best seen in the NW3-SE3 section. The same plot also shows that the NW-SE streaks observed in the map view close to Zakynthos are associated with subvertical structures.
distances are relatively large (~50–180 km) and the azimuthal coverage is poor (Gallović and Zahradník, 2011). The inversion needs to be regularized by smoothing, ensuring slip-rate positivity and preserving total seismic moment.

In preliminary runs, we considered just a single SS fault, whose position and orientation were suggested by the major point-source subevent. In such a case, some seismic stations were poorly fitted. Then, we tested several combined SS + TF models, as suggested by the MPS modeling. Finally, we propose a source model with two fault segments, providing a very good fit of waveforms, overall consistency with the distribution and focal mechanisms of aftershocks, and explaining the observed non-DC part of the overall single-point-source MT.

The proposed model, consisting of two segments (TF and SS), is described in Table S3 and shown in Figure 3. It fits seismic data in the 0.02–0.15 Hz range with variance reduction (VR) of 0.66 (see Fig. 4 and Fig. S5), better than the single-fault model (VR = 0.52, see Fig. S6). The space–time evolution of the slip rate (see Fig. S7) can be characterized as follows. The moment release started on the TF segment near the shallow hypocenter and propagated to northeast along both faults. The major slip-rate episodes were identified; two of them occurred on the TF segment (~6–9 s and 16–19 s after origin time, centered at the depths of 6 and 16 km, respectively), and one appeared on the SS segment (~9–13 s, centered at the depth of 13 km). The SS fault ruptured a single compact patch, with peak slip of 0.8 m, situated between the two TF episodes. The entire thrust process developed on or near the slab top, but due to limited data and location inaccuracies we cannot be more specific. The dominant SS segment, likely situated in the upper plate, cannot be easily associated with any aftershock cluster. The ~6 s delay between the observed shallow nucleation and the deeper significant slip can be attributed to weak nucleation as observed, for example, in dynamic inversion of the 2016 Amatrice, Italy, earthquake (Gallović et al., 2019). We note that these main aspects of the source model are preserved even when removing station LTHK from the inversion as it is the closest one, supposedly affecting the inversion significantly (Fig. S8).

Importantly, if summing the DC-constrained MTs of the TF and SS faults (shown in Fig. 3c), we obtain a significant CLVD value (~56%), thus explaining the observed non-DC mechanism of our single-point MT solution (CLVD ~60%), in agreement with other agencies, for example, Global CMT and NOA (Fig. S2), reporting CLVD of ~44% and ~61%, respectively. Obviously, even more complex models cannot be ruled out.

Although we believe that major features of the presented source models are realistic, some details cannot be taken at face

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Figure 3. Segmented fault model. (a) Map view of the slip distribution on the modest thrust (left) and dominant strike-slip (SS; right) faults is shown by color. Overprinted are aftershocks (dots) and space–time high-frequency energy release from backprojection (small circles, shade-coded according to time; time 0 corresponds to origin time). Star marks the epicenter; rectangles correspond to the respective fault segments, and their top edges are marked by thick lines. Slip vectors are shown by arrows at the black focal mechanism plots. Subevents of the two-point model (blue, denoted Sub 1 and Sub 2), taken from Figure S3, are included for comparison. More details are in Figures S5–S11. (b) Moment-rate time functions of the two fault segments. (c) Comparison of the CMT (left) and the MT obtained by summing the double-couple tensors of the thrust faulting (TF) and SS segments from the slip inversion (right).
value due to limited azimuthal coverage. For example, we cannot strictly rule out that the source models as a whole are shifted by \(\sim 10 \text{ km}\) in the northwest–southeast direction, as indicated by the location uncertainty (Fig. S1b). Also, some linear source features (e.g., the aftershock streaks or slip-distribution patches) could be somewhat lengthened in the northwest–southeast direction (see Gallovič and Zahradník, 2011). The latter warning is particularly relevant for the backprojection, which—due to involvement of high frequencies—might be less robust than slip models.

To further validate the source model, we considerably increased the frequency band (2–8 Hz) and performed backprojection of strong-motion records at local distances, basically of the \(S_g\) wave group (Evangelidis and Kao, 2014; Evangelidis, 2015). For methodical details, see Text S2 and Figure S9. As shown in Figure 3, we observe a well-focused backprojection pattern, systematically moving northeast away from epicenter with increasing time. These independent data agree very well with the finite-fault inversion, confirming the dominant space–time progression of the source process in the southwest–northeast direction.

Finally, we considered also coseismic static GNSS data (see Text S3 and Table S4; Kostelecký and Karský, 1987; Kouba and Héroux, 2001). The nearest permanent stations (ZAKY and STRF) were situated at similar epicentral distances (\(\sim 44 \text{ km}\)) on Zakynthos and Strofades islands. Both recorded \(\sim 5 \text{ cm}\) horizontal displacements, pointing toward south-southwest and south-southeast, respectively. An \(\sim 2 \text{ cm}\) motion toward southwest was recorded on station PYRG. Our seismic source model agrees with ZAKY and PYRG (Fig. S5). For STRF, we fit the displacement orientation but underestimate the amplitude by a factor of \(\sim 4\).

The GNSS amplitude fit at STRF could be remarkably improved in a joint inversion of the seismic and GNSS data; such an approach adds a significant slip patch near STRF station (Fig. S10), similar as if we invert only GNSS data (Fig. S11). Nevertheless, we consider this patch as rather unrealistic for the following reasons. There are no indications for such a slip patch from the backprojections (Fig. 3). There are also no early aftershocks close to this patch. Duration of this patch is about 15 s, starting as soon as 2 s after the origin time (although being \(\sim 40 \text{ km}\) far from the hypocenter). The consequence of such long duration is that this patch does not effectively generate any seismic signal in the frequency range considered. Therefore, we tentatively associate the observed STRF amplitude with a triggered rapid aseismic movement, perhaps similar to that proposed for previous earthquakes by Stiros (2005) to explain unusually large displacements on the Strofades island. For example, very large motion (\(\sim 12 \text{ cm}\)) was recorded at STRF during the \(M_w 6.6\) earthquake of 1997 (Hollenstein et al., 2006).

**Discussion**

Although the low-dip TF part of the source process is naturally acceptable in the shallow part of the subduction zone, the
dominant SS fault slip of the 2018 earthquake is a significant finding deserving more discussion (Fig. 5). A frequently evoked explanation of SS faulting near the trench as transpressional strain partitioning due to oblique plate convergence (Wesnousky, 2005) is not likely in our case because (1) our region does not feature a clearly oblique plate motion relative to the trench, and (2) such SS faults should be largely trench-parallel, with strike N45°W, while our SS fault is striking at N10°E. Nevertheless, shear strain and rotation, possibly generating SS faulting, have been documented near Zakynthos by GNSS measurements of plate velocities (Hollenstein et al., 2008; Chousianitis et al., 2015). Moreover, shearing strain was proposed to explain the right-lateral SS faulting on a vertical buried fault in western Peloponnese in the 2008 $M_w$ 6.3 Movri Mountain earthquake (Fig. S5; Serpetsidaki et al., 2014). The offshore continuation of the Movri fault was also hypothesized (Pérouse et al., 2017). Strain is likely accommodated by trench-orthogonal (southwest–northeast) fractures created in response to segmentation of the subducting slab and piecewise rollback (Sachpazi et al., 2016).

Instead of the transpressional strain partitioning, we propose two alternative interpretations: our SS segment striking at $\sim$10° could be generally explained as a fault offset, for example, a restraining bend or a part of a SS duplex, in our case possibly connecting one of the aftershock streaks near Zakynthos (streak number 3 in Fig. 1) with the southern streak (number 4). These streaks represent vertical faults in the upper plate, and, interestingly, they have a geometrical similarity with the major trench-orthogonal, right-lateral fault system in the slab (Sachpazi et al., 2016).

Another possible interpretation, independent of the trench-orthogonal fractures and the aftershock streaks, is that our 10°-striking right-lateral SS occurred on an approximately

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**Figure 5.** Faulting geometry: An overview. Slip on the thrust and SS segments are presented together with aftershocks (see legend). Two selected snapshots are shown here; (a) view to NE from an elevation of 50°, and (b) roughly the same view direction from an elevation of 20°. Snapshots are taken from the MATLAB code provided as supplemental material.
north–south-oriented reverse fault or normal fault that was reactivated as SS fault (Naliboff et al., 2013) and bounds a relay ramp (Rotevatn and Peacock, 2018). Relay zones are highly fractured, so this interpretation seems to be supported by the dispersed aftershocks. The moderate 40° dip of our SS fault seems quite acceptable in regions where previous thrust faults or normal faults were reactivated as SS to accommodate the present-day strain (Avouac et al., 2014; Sato et al., 2015).

All the mentioned fault types exist in the region (Stiros, 2005; Kokkalas et al., 2013; Papoulia et al., 2014; Avramidis et al., 2017). We also note that some features of the mainshock, that is, the shallow nucleation, the dominant SS motion (situated likely in the upper plate), and the minor TF on (or near to) the plate interface resemble other top-to-bottom breaking systems of subduction terminations of the world (e.g., New Zealand; Mouslopoulou et al., 2019).

In a more general context, we attempt to interpret how large events in the region are connected to long-term, ~8 yr lasting deformation processes (Fig. 6). The earthquake sequence that occurred near Zakynthos in April 2006 comprised six events of $M_w$ 5.0–5.7 (Global CMT; Serpetsidaki et al., 2010; Papadimitriou et al., 2012). Their total moment is $M_w$ 6.7, thus the 2006 sequence is a long-term segmented analogy of the short-term segmented 2018 mainshock. Similar to 2018, the 2006 sequence “activated a single sub-horizontal fault zone at a depth of about 13 km, corresponding to the inter-plate subduction boundary” (Serpetsidaki et al., 2010). Contrary to the 2018 event, all the 2006 earthquakes were remarkably similar in terms of their pure TF mechanisms, strike 345°–21°, dip 10°–39°, rake 118°–128° (Global CMT), not including any SS process. Another notable event to be discussed here is the 1997 $M_w$ 6.6 earthquake that occurred within ~20 km from the 2018 event. Considering the centroid inaccuracies, we can assume that they occurred at the same place. Focal mechanism ($s/d/r = 8°/31°/162°$, Global CMT) is a mixed TF-SS type with negative CLVD (~27%). Two (TF) subevents, 9 s delayed, were indicated by teleseismic inversions (Kiratzi and Louvari, 2003). Thus, the 2018 event seems to be a (larger) analogy of the 1997 event.

Figure 6. Seismicity 2011–2018. (a) Space distribution of earthquakes 2011–2018 (until the 2018 mainshock). The centroids of significant earthquakes southwest of Zakynthos, in 1997 (Kiratzi and Louvari, 2003), 2006 (Serpetsidaki et al., 2010), and 2018 (this article), occurred near two localized clusters. (b) Magnitude–time distribution in 2016–2017, demonstrating the swarm character of the clusters near the 1997 and 2018 $M_w > 6$ earthquakes. More details are in Figure S12.
The area is characteristic for its noteworthy long-term background seismicity at depths <40 km. Figure 6 and Figure S12 show our relocation (Klein, 2002) based on phase data of the National Observatory of Athens for the period 2011–2018. In Figure 6, clustering is obvious: (1) one cluster occurred in 2011 and 2015 at the place of the 2006 Zakynthos sequence, and (2) one cluster was repeatedly activated three times (twice in 2016 and once in 2017) very near the 1997 and 2018 events. A closer look at the latter cluster in Figure 6b shows that its activity was of swarm type (lacking any magnitude-dominant event). Similar swarms recur in time (irregularly), as shown in a broader time window of Figure S12. To explain the spatial correlation of the swarms with the significant earthquakes, we can speculate that the region is affected by fluids of deep origin.

The localized “chimney” of small events detected down to a depth of >80 km, and reported southeast of Zakynthos, out of the region covered by Figure 6 (Papoulia et al., 2014), could track a possible fluid channel from the upper mantle, as suggested in previous studies (Konstantinou et al., 2011; Giannopoulos et al., 2012). Our shallower events rather suggest that fluids could be channeled up-dip in the slab (Halpaap et al., 2019), and this flow is enhanced by the heavy fracturing of reactivated relay ramps here (Rotevat and Peacock, 2018). Alternatively, faults existing in the seismically coupled brittle crust can be loaded by the partially and episodically creeping of the topmost serpentinitized part of the upper mantle; then creep in the mantle could activate earthquake swarms, some of which may represent foreshocks of large earthquakes (Kuna et al., 2019). It is matter of future study to distinguish between these explanations by geodynamic modeling (Čižková and Bina, 2013).

**Conclusion**

Fault interactions, specifically in subduction zones, belong to key problems of earthquake mechanics and seismic hazard assessment. Following previous rare studies of Lay et al. (2013), Hicks and Rietbrock (2015), and Mouslopoulo et al. (2019), we aimed at resolving fault complexity of a strong $M_w$ 6.8 earthquake in the shallowest part of the subduction-termination zone in western Greece.

The studied region near Zakynthos is under subhorizontal compression of azimuth $\sim$230°, and both thrust (TF) and SS events are possible. The $M_w$ 6.8 event in 2018 was of a mixed TF + SS type, as was also the whole aftershock sequence. The mainshock consisted of two fault segments: a TF (dip $\sim$10°) and a right-lateral SS (dip $\sim$40° and strike $\sim$10°). Both slip vectors have the same southwest-orientation, consistent with the GNSS-derived major plate motion. The two-segment fault model explains the observed non-DC radiation. The low dip of the TF segment (similar to many aftershocks) suggests the interplate origin. Relatively distant aftershocks (seemingly occurring off the mainshock fault) mapped several southwest–northeast streaks, or faults. As a whole, the mainshock progressed also in the southwest–northeast direction. This direction closely follows the trench-orthogonal fractures related to differential slab motions. The 10°-striking SS segment of the mainshock could be a fault bend connecting such offshore structures, or it could be explained by reactivation of approximately north–south-trending fault relay zone in the upper plate. Although the 2018 sequence involved a thrust fault, the SS faulting was dominating the mainshock; this result should be considered in future seismic and tsunami hazard assessment in the region. The observations could also apply to other subduction-termination zones worldwide.

The previous earthquake of similar focal mechanism (1997 $M_w$ 6.6) occurred almost at the same place as the 2018 event. This place, coinciding with a large bathymetric deep, hosted three earthquake swarms in 2016 and 2017. Precise investigation of the swarms would deserve future deployment of an ocean-bottom seismometer network; it could confirm or rule out possible triggering of the major $M$ 6+ earthquakes in the area by deep-seated processes such as fluid flows or the loading effect of the episodically creeping upper mantle. Future observations should be complemented by numerical modeling of the slab deformation, especially in the shallow depths of the subduction-termination zone.

**Data and Resources**

Waveform data were sourced from the corresponding Observatories and Research Facilities for European Seismology (ORFEUS) European Integrated Data Archive (EIDA) node (http://www.orfeus-eu.org/eida) operated by the National Observatory of Athens (http://eida.gein.noa.gr). We used data from the HL (Institute of Geodynamics, National Observatory of Athens, doi: 10.7914/SN/HL), HP (University of Patras, doi: 10.7914/SN/HP), which is cooperating certain stations jointly with the Charles University in Prague, HT (Aristotle University of Thessaloniki, doi: 10.7914/SN/HT), HA (National and Kapodistrian University of Athens, doi: 10.7914/SN/HA), and HI (Institute of Engineering Seismology and Earthquake Engineering, doi: 10.7914/SN/HI) networks. Global Navigation Satellite System (GNSS) data for STRF station were provided by Istituto Nazionale di Geofisica e Vulcanologia (INGV), Rete Integrata Nazionale Global Positioning System (RING) Working Group (2016), (doi: 10.13127/RING). PYRG and ZAKY GNSS station data were provided by METRICA S.A. The bathymetry maps were based on European Marine Observation and Data Network Digital Terrain Model (EMODNET DTM) (1/16 arcmin grid resolution) obtained and modified from the Hellenic Centre for Marine Research (file was provided by D. Sakellariou). Figures were created using Generic Mapping Tools (GMT) v.5.4.5 (Wessel et al., 2013). Software ISOLA (developed by J. Z. and E. S.) can be downloaded from https://github.com/esokos/isola and with recent updates from http://geo.mff.cuni.cz/~jz/for_Cuba2018/. All websites were last accessed May 2019. The supplemental material for this article includes four tables and 12 figures, illustrating details of seismicity, focal mechanisms, GNSS data, multiple-point source modeling, and slip inversions. In addition, it includes a zip file with text files and a MATLAB script enabling 3D view of aftershocks and slip on two fault planes.
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