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Absolute Centroid Location of Submarine Earthquakes from 3D Waveform Modeling of Water Reverberations

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Key Points:

- Seismic recordings of submarine earthquakes show strong arrivals that are caused by the reverberations of their P-wave energy in the ocean
- 3D hybrid simulations can be used to model water reverberations and extract valuable information about the earthquake's centroid location
- Resolution of current global bathymetry models is sufficient to model water reverberations in any oceanic fault system in the world

Supporting Information:

- Supporting Information S1

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Abstract Oceanic transform faults (OTFs) represent an attractive tectonic environment to investigate how slip is accommodated within the crust. However, as most of these fault systems grow in the deep ocean, where few local seismic observations are available, characterizing their earthquake behavior is complicated and remains a formidable challenge. Here we present a novel approach for retrieving precise centroid locations of submarine earthquakes that is based on the modeling of water phases in teleseismic records. Using a hybrid method for simulating far-field body waves with 3D source-side structures, we demonstrate that the scattered energy generated by the continuous bounces of an earthquake's P-wave trapped in the ocean is modelable and carries information about its source location. As a case study, we use a realistic bathymetry model of the Gofar transform fault on the East Pacific Rise and simulate the seismic wavefield at US Array stations for four of its moderate-sized (Mw5.0+) earthquakes. Our modeling results show that water phases are sensitive to a ~5 km change in the earthquake's horizontal location and that a remarkable agreement between observed and synthetic water phases is achieved when the location of an event is close to its true one. We then relocate three of these events by systematically computing their water phases in candidate locations until a satisfactory waveform fit is achieved. This analysis technique paves a new route for studying earthquake source properties in isolated marine environments and serves as a mean to investigate the seismic behavior of OTFs on a global scale.

1. Introduction

Understanding the influence of different tectonic parameters on the earthquake rupture process is key to anticipate fault behavior and assess seismic hazard. Over the past few decades, significant advancements in this field have been made by studying and monitoring large continental faults such as the San Andreas Fault in California or the Alpine Fault in New Zealand (e.g., Jolivet et al., 2015; Hamling et al., 2017; Savage & Burford, 1973; Scholz et al., 1969, 1973). However, owing to their obscure origin and complex geologic evolution, the formulation of a mechanical model that can accurately predict seismic behavior along these fault systems has remained one of the most outstanding problems in seismology. Oceanic transform faults (OTFs) on the other hand are among the simplest tectonic environments and, therefore, represent a more advantageous location to study how slip is accommodated within the crust (Froment et al., 2014; McGuire et al., 2005, 2012). Different from their continental analogues, OTFs have a simple geometric configuration, a close to homogeneous composition and a smooth thermal structure (Fox & Gallo, 1984; Turcotte & Schubert, 2014; Wilson, 1965). This relative simplicity leads to a somewhat predictable seismic behavior that is less dependent on the geologic history of the plate boundary, allowing us to dissect the different thermal-mechanical factors that control earthquake rupture along transform faults Boettcher (2005). However, despite the substantial attention that OTFs have received, there are still many of their seismological aspects that are not well understood, especially on a global scale.

A general view of the global distribution of oceanic earthquakes reveal a clear difference in seismic behavior between OTFs offsetting slow and fast spreading ridges, with small-to-moderate size earthquakes being common in slow-spreading centers and typically absent in fast spreading centers (Bohnenstiehl et al., 2008; Wiens & Stein, 1983). This commonality between ridge systems suggests that the thermal structure of the lithosphere controls the seismicity along OTFs (Abercrombie & Ekström, 2001; Boettcher et al., 2007; Rundquist & Sobolev, 2002). However, recent observations suggest that temperature cannot be the only control on seismicity since even the largest OTF earthquakes rupture only small segments of the thermally expected seismogenic zone (Boettcher & Jordan, 2004; Roland et al., 2010). This unusual rupture style results

in a dramatic slip deficit in which the majority of the lateral motion on OTFs is accommodated aseismically by mechanisms such as slow creeping events, and, on a smaller portion, seismically on fully coupled patches that are thought to develop above the 600 °C isotherm (Bird et al., 2002; Boettcher & Jordan, 2004; McGuire et al., 2012; Sykes & Ekström, 2012). To explain the nature of the slow transients, laboratory experiments suggest the presence of large amounts serpentinite along the faults, which under shallow crustal conditions have the potential to result in aseismic fault creep (e.g., Reinen et al., 1994; Reinen, 2000). This argument implies that there must exist some level of interaction between seawater and cold mantle rocks that is allowing serpentinization to occur and that fluid circulation along the fault may, therefore, be as an important control on OTF seismicity as temperature (e.g., McGuire et al., 2012). An unresolved issue, though, is the mechanism that determines the size and location of the coupled patches. Mapped surface traces of historical continental strike-slip earthquakes indicate that geometrical discontinuities in the form of fault step-overs and fault bends hold the potential to act as physical barriers to rupture propagation (Nielsen & Knopoff, 1998; Wesnousky, 2006). Conversely, high-resolution studies along a few OTFs suggest that along-strike variations in fault frictional properties dictate whether or not a rupture can propagate through a particular segment of the fault (Kuna et al., 2017; McGuire et al., 2012; Wolfson-Schwehr et al., 2014).

Recent studies have also suggested that OTF seismicity has a high level of predictability over different spatial and temporal scales. On short time scales, OTFs show a clear increase in foreshock activity just hours before any mainshock occurs (McGuire et al., 2005). These foreshock sequences are believed to result from the increased stressing rates caused by preceding slow transients (Lohman & McGuire, 2007; Roland & McGuire, 2009). On long time scales, OTFs exhibit a quasi-periodic seismic behavior with interevent times that are dependent on their slip rates (McGuire, 2008). Furthermore, OTF mainshocks appear to be rupturing on the same overlapping fault patches, which in turn seem to be acting in phase with each other (McGuire, 2008). This type of rupture style suggests that there must exist some type of along-strike variations in the rheology of OTFs that is allowing the synchronization of large earthquakes (Lynch et al., 2003). However, it is still obscure if both seismic and aseismic slip can occur on the same fault patches or if seismic slip persistently occurs on the coupled patches over many earthquake cycles, while the rest of the fault always slips aseismically (Boettcher & Jordan, 2004).

Elucidating any of the enigmas listed above will tremendously improve our understanding of the behavior of OTFs, which could then be extended to their more complicated and hazardous continental counterparts. Such task requires a detailed characterization of their seismic behavior on a global scale. However, owing to their remoteness, OTF earthquakes are generally located with an uncertainty that is a few times larger than their actual rupture dimensions, thus limiting our capability of probing their fine-scale structure and investigating their spatial variations in seismic properties (McGuire, 2008; Wolfson-Schwehr et al., 2014). This obstacle has resulted in substantial efforts towards developing methodologies that will allow us to accurately constrain the location of mid-oceanic earthquakes in general (e.g., Jordan & Sverdrup, 1981). However, the majority of these techniques can only solve for relative locations and require a small number of events located with insignificant uncertainty, which in most cases are not available (Cleveland & Ammon, 2013; McGuire, 2008; Waldhauser & Ellsworth, 2000). Attempts to determine the absolute locations of OTF earthquakes have been made by associating earthquake focal mechanisms with seafloor bathymetry (Pan et al., 2002). Nonetheless, as the resolution of bathymetric data continues to improve, it is being revealed that the morphology of OTFs can be complicated to interpret and that constraining earthquake locations by seafloor bathymetry alone is not always possible.

More direct attempts to investigate OTF behavior have been made by deploying ocean bottom seismometers (OBSs; e.g., Horning et al., 2018; McGuire et al., 2012; Rowlett, 1981). However, these studies have been limited to a small number of faults around the world due to the high cost of placement, maintenance, and retrieval of the seismic equipment. This limitation calls for a more efficient yet comparably robust scheme to study OTFs worldwide. In this paper, we present a novel approach for determining accurate centroid locations of oceanic earthquakes at unprecedented resolutions using land-based seismic stations alone. First, we demonstrate that the waveform complexity that is introduced by the continuous water reverberations of any submarine earthquake is modelable and carries valuable information about the absolute location of the rupture within the fault. We then simulate four moderate-sized ($M_w5.0+$) earthquakes along the Gofar transform fault (GTF) on the equatorial East Pacific Rise (EPR) and show that we can improve their reported locations by modeling their water phases in a set of candidate locations. Finally, we test the sensitivity of

Figure 1. Schematic diagram showing the ray paths of the direct P-wave, the depth phases (pP , sP) and the first four water multiples (pw_nP and sw_nP ; left). Synthetic velocity waveforms for a 20-km deep strike-slip earthquake under a homogeneous water layer (right). For simplicity, the synthetic seismogram is shown without water phases (top) and with water phases (bottom). The water multiples alone are presented in the middle. The waveforms were calculated for a station at a distance of 84° using the Preliminary Reference Earth Model (PREM; Dziewonski & Anderson, 1981) with a 3-km-thick water layer on the source side. Note how the ray parameter of the water phases is the same as the one of the direct P-wave.

water phases to bathymetry changes and show that the current resolution of publicly available bathymetric data is sufficient to model them in any oceanic environment in the world.

2. Waveform Modeling of Water Reverberations

Teleseismic recordings of earthquakes occurring directly beneath the seafloor show prominent arrivals caused by the reverberations of P-wave energy trapped in the water column (Figure 1; Herrmann, 1976; Ward, 1979). These arrivals, however, do not often appear as the systematic set of reflections that are expected to result from an acoustic wave propagating in a conventional flat seafloor model (i.e., with decreasing amplitude and alternating polarity). In a series of investigations, Wiens (1987, 1989) examined the effects of a planar dipping water-crust interface on teleseismic waveforms and determined that the high variability of their P-wave coda amplitude results from changes in the take-off angle of water phases. Moreover,

Figure 2. Example illustrating the effect of seafloor roughness on teleseismic waveforms. The synthetic waveforms show a noticeable increase in amplitude and complexity as the ocean bottom roughness increases (a). The dashed green line on top of the elevation profiles depicts the event's epicenter. The high resolution bathymetry profile used in the simulations is extracted from ETOPO1 global relief model (<https://www.ngdc.noaa.gov/mgg/global/global.html>) along profile P-P' in (b). The low resolution bathymetry profiles are built from applying running averages of different lengths to the high-resolution one. Note that we only show the first 1,000 km of each profile to accentuate the near-source structure but the models extend all the way to the inland station. Focal mechanism and takeoff angle of the direct P wave are shown in (c).

Figure 3. Waveform modeling results for the 2013 Mw6.1 Gofar earthquake using different structural models (see Figure 4 for the event location). Black traces are observed data, and red traces are synthetics (0.01–0.2 Hz). Station network and name is shown in panel (a) only but is the same for all four panels. Synthetics in (a) are computed using a 1D Earth (Model 1). Synthetics in (b) are computed using a 2D Earth with a flat ocean bottom on the source side (Model 2). Synthetics in (c) are computed using a 2D Earth with realistic topography (Model 3). Synthetics in (d) are computed using the hybrid method with 3D source side structures (Model 4). The average waveform misfit of each set of synthetics is shown in the upper part of the panels. Examples of the velocity models (for the 1D and 2D simulations) and the bathymetry mesh used in the 3D simulations are shown on the right. The elevation data for Model 4 are extracted from the Global Multi-Resolution Topography synthesis (GMRT; Ryan et al., 2009).

Okamoto and Miyatake (1989) used 2D numerical simulations to demonstrate that the multipathing effects on far-field body waveforms due to realistic seafloor topography are remarkable and cannot be approximated by a planar interface model. Taken together, these studies suggest that most of the water phase complexity derives from the wavefield's interaction with the seafloor's irregular structure. To exemplify this connection, Figure 2 shows a set of 2D finite-difference seismograms (Li et al., 2014) from a strike-slip source calculated with a two-domain (i.e., fluid and solid) velocity model that includes realistic bathymetry with different levels of smoothness. The synthetic waveform coda shows a systematic increase in amplitude and complexity

Figure 4. Topographic map of the Gofar transform fault system. G1, G2, and G3 mark the different segments of the fault that are separated by intratransform spreading centers. Beachballs show the focal mechanisms and centroid locations from the Harvard GCMT catalogue. The circled numbers represent candidate locations for the 2002 Mw5.1, 2007 Mw6.1 and 2016 Mw5.7 earthquakes. The average intersource spacing between neighboring candidate locations is 5 km. The small inset map is a zoom of the 2016 Mw 5.7 event epicentral region. The dashed purple line marks the extension of a potential subparallel fault (from surface bathymetry). The red circled numbers mark the centroid locations determined by this study. Elevation data are extracted from the Global Multi-Resolution Topography synthesis (GMRT; Ryan et al., 2009).

as the ocean bottom roughness increases that results from the intrinsic relationship between the source radiation pattern and the irregular near-source structure.

Even though water phases in realistic geological settings appear to be overwhelmingly complicated, the causality between seafloor geometry and waveform intricacy presents the unique opportunity to associate particular P-wave coda patterns to both earthquake source properties and the Earth's structure. Such task, however, heavily relies on our ability to model water reverberations adequately which, in turn, depends on the level of accuracy to which we can capture the bathymetry effect on teleseismic waveforms. To this end, we employ a hybrid method for calculating teleseismic waveforms with 3D source side structures (Wu et al., 2018). This method uses the spectral element method (Komatitsch & Tromp, 1999, 2002) to compute a complex 3D wavefield in the source region and then propagates that wavefield to large epicentral distances using the direct solution method (Geller & Ohminato, 1994; Geller & Takeuchi, 1995). The coupling of these two methods allows us to capture the full 3D effects of complex source side structures while reducing the heavy computational cost of calculating high-frequency waveforms on a global scale (supporting information, Figure S1). To show the relevance of considering 3D wave propagation effects in oceanic environments, Figure 3 shows a direct comparison of waveform modeling results for the 2013 Mw6.1 GTF earthquake using different structural models (1D, 2D, and 3D) and the reported Global Centroid Moment Tensor solution (GCMT; Dziewonski et al., 1981; Ekström et al., 2012). From this comparison, four critical points should be noted. First, the majority of the P-wave coda of a submarine earthquake is composed of water reverberations that cannot be modeled by synthetics computed for a 1D Earth. Second, for a shallow strike-slip earthquake, a flat seafloor model is insufficient to match the true amplitude of water phases, as the take-off angle of the water multiples is nearly vertical and therefore close to the P and SV nodal planes. Third, an irregular ocean bottom allows larger regions of the focal sphere to be sampled by different rays and causes a larger energy distribution in the P-wave coda. However, 2D modeling is insufficient to match the later part of the waveforms at a desirable level. Fourth, the inclusion of the 3D structure in the epicentral region at an adequate resolution allows us to match the full complexity of the observed waveforms and is, therefore, required to model water reverberations. Note that an even better waveform fit might not have been achieved due to the

Figure 5. Waveform modeling results for the 2002 Mw5.1 (a), 2007 Mw6.1 (b), 2008 Mw6.0 (c), and 2016 Mw5.7 (d) Gofar earthquakes using the reported Global Multi-Resolution Topography (GCMT) solutions and a 5-km source depth. Black traces are observed data, and red traces are synthetics (0.01–0.2 Hz). Station name and network are shown above each trace pair.

inherent location error of the GCMT catalog or the assumption of a 1D crust and mantle structure. For more information regarding the meshing and validation of the 3D numerical simulations, the reader is referred to supporting information Text S1.

3. Earthquake Relocation by Waveform Modeling of Water Reverberations: A Case Study in the GTF

Once the bathymetry effects on teleseismic P-waveforms are properly accounted for, water phases should be modeled with high levels of accuracy. Any significant waveform disagreement between observed and synthetic water reverberation phases can, therefore, be attributed to an error on the earthquake's location, focal mechanism, and/or to the presence of any near-source heterogeneity that is not included in our velocity model. However, given our current knowledge of plate-motion direction and the structural simplicity of OTFs, it is less likely that the two latter factors are the main causes of poor waveform fitting. Moreover, because the impedance contrast is largest at the fluid-solidus interface, the teleseismic P-wave coda of these type of events will be dominated by acoustic waves scattered at the ocean bottom rather than by other type of complex waves scattered at the fault's and lithosphere irregular structure. Receiver-side structures can also introduce some complexity into the observed waveforms, but here we base our analyses on data from dense seismic networks to minimize their effect. Under this set of assumptions, and for a given faulting geometry, we can then generate a library of water phase templates along and around any mid-oceanic fault system to relocate earthquakes whose current catalog location results in poorly modeled coda waveforms. To demonstrate the effectiveness of this approach, we use the fast-slipping ($\sim 14\text{cm year}^{-1}$) GTF as a test case and simulate four of its Mw5.0+ events at all available US Array stations, including the 2008 Mw6.0 Gofar earthquake which was captured by a nearby OBS array (Figure 4). The location of the 2008 Mw6.0 event was previously determined relative to its Mw5.1 aftershock and assuming that the aftershock's centroid location is equal to its own epicenter location (McGuire et al., 2012). The earthquake depth, on the other hand, is not well-constrained but suggested to be above the 600°C isotherm, which given the warm

Figure 6. Record section of 3D synthetic waveforms generated for one teleseismic station using source locations 7–27 in Figure 4a. All waveforms have been band-passed between 0.01 and 0.2 Hz. The blue shaded region delimits the traces for the fault-perpendicular sources, whereas the yellow shaded region delimits the traces for the fault-parallel sources. The mean azimuth and distance to the receiver is of 0.70° and 51.76° , respectively. Beamforming output for the three time windows that are enclosed by the red dashed lines in the record section (b). Slow and prominent arrivals can be observed being generated from multiple azimuths. Collection of all maximums in every beamforming output for the full station synthetic array (c). The size of the markers represent the beam power, whereas the color represent the apparent wave speed. The presence of slow arrivals concentrated in the azimuth along the fault's strike suggests that the fault itself is behaving as a strong scatterer.

thermal structure of the GTF is ~ 5 km for the center of the transform fault (Roland et al., 2010). Using the reported GCMT moment tensor solutions and centroid locations, our initial modeling results show that a remarkable waveform fit is achieved for the 2008 Mw6.0 earthquake (most-likely due to its proximity to its revised and probably true location; Figure 5c), whereas the water reverberation phases for the remainder three earthquakes are poorly matched by the synthetic seismograms (Figures 5a, 5b, and 5d). The large waveform discrepancy between the observed and synthetic traces for the 2002 Mw5.1, 2014 Mw5.5 and 2016 Mw5.7 events is therefore likely to be caused by severe earthquake mislocation.

Before relocating the poorly modeled events, it is an interesting exercise to analyze the regional bathymetric effect of the GTF on the teleseismic wavefield. To do so, we simulate two orthogonal lines of equi-spaced strike-slip sources placed at a constant 5 km depth along and across the G3 segment of the fault (locations 7–27 in Figure 4) all the way to a 207 station array uniformly spread across the United States (supporting information, Figure S3). We then use the full synthetic data set to analyze the wavefield's sensitivity to source location and perform a source-side beamforming to detect the direction and apparent wave speed at which the multiple arrivals in the P-wave coda are traveling outward from the source array (Figure 6). Our analysis reveals that, for the 0.01- to 0.2-Hz frequency band, water phases are sensitive to a ~ 5 -km change in the earthquake's horizontal location and that a large portion of the water multiples are composed of arrivals that are generated from multiple azimuths. However, our most relevant finding is that the fault topographic relief appears to behave as a strong scatterer, as is confirmed by the large concentration of slow arrivals that are coincident with the fault's strike (Figure 6c). This observation supports our hypothesis that water phases can be used to retrieve valuable information about the earthquake's centroid location as the timing and amplitude of these scattered waves can be used to constrain the event's position along and across the fault.

3.1. Main-Fault Events

In the simplest of cases, and for earthquakes whose initial reported location is the vicinity of prominent bathymetric features, we can correlate the water reverberation signals generated by several equi-spaced strike-slip sources along any obvious candidate locations until an optimal solution is found. For this case,

Figure 7. Waveform modeling results for the 2002 Mw5.1 using source locations 1–6 in Figure 4. Black traces are observed data, and red traces are synthetics (0.01–0.2 Hz). Station name and network are shown above each trace pair in the first panel only. The green contoured panel shows the waveform comparison of the observed and synthetic traces for the location that results in the best fit (location 1), whereas the blue contoured panel shows the waveform comparison for the location that is closest to the Global Multi-Resolution Topography (GCMT) solution (location 5).

we limit the depth of all simulated sources to 5 km due to the warm temperature of the GTF and separate all test locations by 5 km on account of the frequency band of analysis. An example of such search process is presented in Figure 7, which shows the water phase variations along the G2 segment of the fault and their comparison with the 2002 Mw5.1 seismic recordings. Apart from exhibiting significant water phase changes as the source location is shifted along the fault, our computed seismograms are, at a particular location, able to fit almost every single arrival that is present in the observed P-wave codas. By analyzing the goodness-of-fit between the observed and synthetic traces in a total of 99 US Array stations via a shape-fitting criteria, which is defined as follows:

$$\text{Misfit} = 1 - \frac{[\text{Max}(R_{XY})]^2}{\sum X_i^2 \sum Y_i^2}, \quad (1)$$

where X_i is the observed waveform, Y_i is the synthetic waveform, and R_{XY} is the cross-correlation of the traces, we are able to decrease the average waveform misfit by 22% by taking location 1 from Figure 4 as the earthquake's true centroid location. The difference between our solution and the reported GCMT for this event is over 19 km, which is approximately twice the average rupture length of an Mw5.0 earthquake. Our preferred location displaces the current catalog location away from the extensional pull-apart basin that is between the G1 and G2 fault segments, which is tectonically more consistent with the expected faulting geometry of an oceanic strike-slip event. It is important to note that our best-fitting location is at the western end of the source array, and that the true hypocenter of the event may still be further west from our proposed site. However, given the small room for improvement in terms of waveform fit, it is highly unlikely that this location will deviate substantially from our current solution.

3.2. Low SNR Events

Given the moderate magnitudes of OTF events, as well as their large distance from any seismic station, one difficulty that commonly arises when analyzing their teleseismic waveforms is the low signal-to-noise ratio (SNR) of the earthquake recordings. This complication can bias the interpretation of water phases and limit their use in relocating events. Nevertheless, with the availability of dense seismic networks, it is possible to apply beamforming to enhance the coherent signals buried within the traces and suppress the incoherent noise (Rost & Thomas, 2002). Here, given that the ray-parameter of the pw_nP and sw_nP phases is substantially close to the one of the direct P-wave arrival (Figure 1), we can apply a systematic delay-and-stack using local apparent P-wave slownesses at different groups of stations to extract coherent sequences of water phase arrivals. This process increases the overall SNR of the seismic array at the elevated cost of resolution and, therefore, should only be applied when the seismic traces are uninterpretable. An example of a set of records

Figure 8. Illustration exemplifying the beamforming process applied to the 2007 Mw6.1 recordings (a). The map shows the stations used to produce a beamed trace at station CI.ARV (red inverted triangle). The blue inverted triangles represent the 19 closest stations used to form the subarray. The inset plot shows the beamforming output for a time window around P-wave arrival of the 20 traces. The subarray and beam information is shown in the bottom white box. The record section in (b) shows the raw waveforms of the 2007 Mw6.1 event at all available US Array stations (gray inverted triangles in a). The record section in (c) shows the waveforms at the same stations after applying the spatio-temporal beamforming. Both set of records have been aligned with the P-wave onset time and band-pass filtered between 0.01 and 0.2 Hz. The red dashed lines in (b) and (c) mark the P-wave and PP-wave arrival time. The green arrows in (c) mark the arrival time of two prominent water phase arrivals.

that can benefit from this SNR enhancement are the ones from the 2007 Mw6.1 event. To beamform the signals, we first divide the entire US Array into multiple small-aperture 20-station subarrays such that, for each individual seismic station, we find the optimal stacking velocity using the 19 closest stations and generate a single beamed trace (Figure 8a). To ensure waveform coherency among stations, and maintain the incident plane wave approximation for teleseismic arrivals (Rost & Thomas, 2002), we only apply beamforming if the radius of a given subarray is smaller than 400 km. Although the choice of stacking parameters is somewhat subjective, one should attempt to use the least amount stations to allow the subtle waveform changes of the water phases as a function of distance and azimuth be preserved and, at the same time, still be below the alias limit of the subarray (Breguier et al., 2015; Roux et al., 2008). The outcome of implementing this type of spatio-temporal beamforming results in a remarkable increase in the SNR of the traces such that coherent water phases can now easily be observed (Figures 8b and 8c).

Once the records have been beamed, and the water reverberations signals are discernible, we take the same approach as the one in the 2002 Mw5.1 event relocation and place multiple strike-slip sources at 5 km depth along the G3 segment of the fault. The waveform modeling results for the 2007 Mw6.1 event are presented in Figure 9. A simple visual inspection of the synthetic waveforms reveals that not all candidate locations are able to replicate the water phase patterns that are observed in the recorded seismograms. In particular, there exist two prominent and persistent water phase arrivals at 35 and 70 s after the direct P-wave (green arrows in Figures 8c and 9) that are only generated when the earthquake source is placed within a small section of the fault. Based on the source-side beamforming results, we believe that these arrivals are produced somewhere at the edges of the fault probably as a result of seismic guided waves trapped in the fault's low topographic relief. For obvious reasons, this type of waveform features is useful to investigate near-source seismic structures and, in this case, allow us to tie down the absolute location of the source to a particular point in space. From a goodness-of-fit analysis in over 460 beamed traces, our preferred solution displaces the earthquake's location 11.6 km east from the reported GCMT location by taking location 15 in Figure 4 as the event's true centroid location.

Figure 9. Waveform modeling results for the 2007 Mw6.1 earthquake. Black traces are observed data, and red traces are synthetics (0.01–0.2 Hz). The leftmost panel shows the synthetic waveforms of earthquake sources 7–17 in Figure 4 for station CI.SWS. The black waveform plotted on top corresponds to the observed beamed trace around the same station. The top green arrows mark the arrival time of the two prominent water phase arrivals. Note how source locations 13–16 are the only ones capable to produce the first prominent water phase arrival (blue-circled numbers). The three rightmost panels correspond to the waveform simulation results for a strike-slip source placed at the GCMT reported location at 5 km depth, at source location 15 at 5 km depth, and at source location 15 but at 15 km. Station name and network are shown above each trace pair in the first panel only. The green contour marks the waveform fit of our preferred solution.

Different from the other events analyzed in this study, the direct P-wave of the 2007 Mw6.1 earthquake has a double-pulse feature that could easily be interpreted as complexity in the rupture process. However, through our waveform analysis, we find that a single point source is sufficient to explain most of the information present in the seismograms, and that, to fully match the direct P-wave and the entire water phase packet simultaneously, the earthquake depth must be close to 15 km (Figure 9). This observation suggests that the secondary pulse in the direct P-wave corresponds to a depth phase reflecting from the bottom of the seafloor rather than be the result of a complicated source time function. If the latter is correct, then this would imply that either not all earthquakes along the GTF are thermally controlled, or that the thermal structure of the fault cannot be explained by a simple half-space cooling model. This claim is reasonably consistent with McGuire et al. (2012) OBS observations, which revealed large along-strike variations in the material properties of the GTF such that there might be regions of enhanced fluid circulation that are affecting the thermal state of the fault. Although investigating the mechanics of this particular earthquake is out of the scope of this study, our analysis highlights the potential of water phases to place valuable constraints on both the earthquake rupture and the fault structure from teleseismic observations alone.

3.3. Off-Main-Fault Events

Off-main-fault events such as the 2016 Mw5.7 earthquake present a greater challenge to relocate since there is no primary surface feature to which we can initially tie the earthquake's epicenter to. This requires a broader exploration of the solution space and can therefore become computationally demanding, but not unfeasible. An example of such search process is shown in Figure 10, where we first place a total of

Figure 10. Search space for the 2016 Mw5.7 event optimal location. The circles depict the earthquake candidate locations (see magnified map in Figure 4) and the traces show the waveform fit between observed (black) and synthetics (red) at station TA.L02F for four different locations (0.01–0.2 Hz). The color of the markers represent the waveform misfit for the same station. The waveform modeling results for this earthquake best-fitting location are shown on the right. Station name and network are shown next to each trace pair. The dashed purple line marks the extension of a potential subparallel fault (from surface bathymetry). GCMT = Global Centroid Moment Tensor.

10 strike-slip sources in a 5-km radius around the 2016 Mw5.7 GCMT catalog location and quantify the waveform similarity at each point. We then use the position with the minimum average misfit as our new starting location and simulate 10 new strike-slip-sources around it. This process is applied iteratively until a well-established waveform misfit minimum is found. By quantifying the goodness-of-fit between observed and synthetic records in over 114 US Array stations, we are able to decrease the average waveform misfit by 11% by taking position 42 from Figure 4 as the event's true centroid location. Our new solution is over 10 km east from the current catalog location and is, once again, able to fit almost every single arrival present in the observed P-wave codas. Interestingly enough, our solution places the 2016 Mw5.7 event in close proximity to the 2013 Mw6.1 earthquake, whose current GCMT location also results in a remarkable waveform fit (Figure 3d). Both of these events are in close alignment with a linear surface feature that appears to be composed of a series of subparallel faults that appear to capable of hosting Mw5.0+ earthquakes (Figure 4). It is important to mention, however, that since these type of events are not in the immediate vicinity of a prominent bathymetric feature, the water phase sensitivity to horizontal location drops dramatically (as evidenced by the smooth misfit surface exhibited by the candidate locations in Figure 10). This limitation increases the uncertainty with which these type events are relocated and may, therefore, require higher frequency simulations to determine their accurate location. Examples of modeling attempts for two other off-fault events (including the 2013 Mw6.1 earthquake) are presented in supporting information Figures S8–S11.

4. Discussion

The modern-day availability of high-resolution multibeam bathymetric data, well-documented slip rates, and comprehensive thermomechanical models of mid-ocean ridges shed light on some of the most relevant factors that control seismicity in OTFs. However, the lack of well-determined earthquake source parameters along these fault systems hinders the accurate characterization of the degree of influence of different tectonic parameters (e.g., offset length, spreading rates, and lithosphere temperature) on the distribution of earthquake slip. A thorough characterization of OTF seismicity at a global scale is, therefore, necessary to place strict constraints on the primary controls on earthquake behavior in oceanic plate boundaries. To evaluate the limits of our analysis technique, and whether it can be applied to any oceanic setting on Earth, we generate a series of 2D synthetic seismograms computed with different levels of bathymetric resolution

Figure 11. Heatmap showing the waveform sensitivity to bathymetry resolution as a function of frequency. The thick lines depict the P-wave wavelength in the water ($\lambda = 1$ equals to one wavelength assuming a constant water velocity of 1.5 km/s). The white dashed lines mark the 1-arc minute base resolution of current global bathymetry models, the resolution limit of the resampled SRTM30 global bathymetry model Becker et al. (2009), and the resolution limit of the resampled SRTM15+ global bathymetry model (Tozer et al., 2019). The thick black arrow marks the bathymetry resolution for the Gofar transform fault (61 m; Ryan et al., 2009). The topographic profile used in these calculations is cut from the East Pacific Rise (EPR) to the center of the United states (from $-4.75^{\circ}/-105.40^{\circ}$ to $41.38^{\circ}/-119.17^{\circ}$ lat/lon). The misfit is quantified via the shape-fitting criteria (equation 1) in which the record generated in the highest resolution run is taken as the “true” waveform. Synthetics are calculated for a station at a distance of 48° using a 15-km deep strike-slip source.

and quantify the waveform similarity as a function of frequency (Figure 11). More in detail, this exercise entails computing numerous synthetic waveforms with the same bathymetry profile but sampled at different resolutions and meshed with variable grid sizes so that the wavefields are calculated at different frequencies. Then, for each frequency, we take the synthetic recording that was simulated with the highest resolution velocity model and use it a “true” waveform to be compared with the lower resolution simulations. Our analysis reveals that significant waveform changes begin to occur when bathymetric features with scales larger than the P-wave wavelength in the water are modified. Now, given that water phases propagate in a medium bounded by an elastic half-space and a free surface, the superposition of opposed reverberations leads to a standing wave with one-quarter of a period. Thus, for an average 3-km deep ocean, the fundamental mode of a sequence of water phases will have a central frequency of 0.125 Hz (assuming a constant water velocity of 1.5 km/s). This result suggests that numerical simulations accurate down to ~ 8 s period are required to model the simplest form of water phases in an typical mid-oceanic setting, and that bathymetric features of a size comparable to the ocean depth need to be adequately represented in our velocity models. Current publicly available bathymetry models constructed from both satellite data and ship soundings have a 1-arc minute base global resolution and can go as high as 15-arc seconds in regions where more modern ship sounding are available (Becker et al., 2009; Tozer et al., 2019). This type of global resolution technically enable us to accurately model ~ 6 s water reverberations in any OTF system in the world.

To this point, our methodology has only been applied to relocate events occurring in the equatorial EPR, which has a high spreading rate and, hence, a relatively smooth seafloor geometry. In such tectonic setting, and because of the frequency band of analysis, our numerical simulations provide us with enough resolution to capture waveform changes resulting

from a 5-km location difference. However, in a marine environment with a more chaotic seafloor, such as the mid-Indian ridge, the waveform changes that result from a 5-km source shift are likely to be more notable. As a result, the centroid locations obtained from modeling the water reverberations of earthquakes occurring in slow spreading ridges should be of higher resolution, and the uncertainties associated with those solutions should be lower. To illustrate this concept, Figure 12 shows a comparison between the recordings of two pairs of similar-sized transform earthquakes occurring in two different marine environments. Both pair of events have the same intersource distances, as determined from the cross-correlation of their Rayleigh wave packets (supporting information, Figures S12–S13 Cleveland & Ammon, 2013), and for each pair, their waveforms are shown for the same seismic stations. From this analysis, it is easily observed that, despite the fact that both pairs have almost the same intersource distances (~ 25 km), the water phases from the events occurring in the fast-spreading EPR are more similar between each other than the water phases from the events occurring in the slow-spreading Aden ridge. Consequently, the location resolution that one may achieve by modeling the water phases of the events in the Aden Ridge is technically higher than the one that is achievable in the EPR. This particularity causes that both the sensitivity and the uncertainties of our technique to vary between regions, and therefore, a careful evaluation of the wavefield’s sensitivity to location changes should always be performed before relocating submarine events in different study areas.

Here we have also restricted the analysis of water phases to mid-oceanic settings due to their relatively simple velocity structure and lack of sediments, which have been shown to have a direct effect on the amplitude of teleseismic water reverberations (Wiens, 1989). Nonetheless, our method of analysis can also be applied to extract source characteristics of earthquakes occurring in subduction zones where offshore seismic data has been collected, and high-resolution 3D velocity structure models are available (Qian et al., 2019). The analysis of subduction-related water phases may help us to constrain important source characteristics such as the up-dip extent of the earthquake rupture and whether any breakage of the seafloor occurred,

Figure 12. Waveform comparison of two similar-sized pair of earthquakes occurring in a fast spreading oceanic environment (top) and a slow spreading environment (bottom). Both pair of events are separated by the same distances, as determined the time shifts of their fundamental-model Rayleigh waves for a full azimuth range (supporting information, Figures S12–S13). Event names and focal mechanisms are shown on the left size of each panel. Station name, distance, and azimuth are shown above each pair of waveforms. Distances and azimuths are approximate and calculated from the center of the fault. Fault name and slip rate are shown in the top part of each panel together with the estimated intersource distance. The normalized cross-correlation coefficient of each pair of waveforms is shown in blue next to the traces. Note that, despite having the same intersource distances, the water phases for the events occurring in the fast spreading ridge are more similar than the ones occurring in the slow spreading ridge. All waveforms have been band-passed between 0.01 and 0.2 Hz. For each panel, the focal mechanism and waveforms are color-associated.

which would generate unusually large amplitude water reverberations (Lay & Rhode, 2019; Lay et al., 2019). Understanding the water phase signature of complicated rupture process in subduction zones has important implications for probabilistic tsunami hazard assessments and may one day be used to identifying tsunamigenic events in real time.

Although so far we have solely focused in improving earthquake locations along the GTF, we have shown that our technique can be optimally applied to investigate the seismic behavior of different mid-oceanic environments. An analysis as the one presented here but at a global scale should allow us to answer some of the fundamental open questions regarding the mechanics of OTFs and earthquake faulting in general. For instance, the distribution of seismically active patches along multiple oceanic fault strands around the globe will provide us with sufficient observational information to resolve whether structural discontinuities affect the size and location of the seismogenic zone, or if along-strike variations in fault frictional properties are more likely to limit rupture propagation. Moreover, with an accurate global catalog that spans over a few earthquake cycles, we will be able to determine where the transition from aseismic to seismic takes place along a given fault and whether the frictional properties of OTF segments evolve temporally between stable

and unstable slip. Understanding the relationship between seismic and aseismic processes is essential to hazard analyses given that major continental fault systems have been observed to accommodate part of their cumulative strain by aseismic slip (e.g., Irwin & Barnes, 1975; Rolandone et al., 2008) and because slow transients have been observed to occur minutes before devastating earthquakes (e.g., Kanamori & Cipar, 1974). The formulation of a reliable global OTF seismic catalog will also serve as an important foundation for numerous future investigations that go well-beyond earthquake source studies. For example, once that accurate absolute centroid locations are available, it will become possible to use wavefield methods to probe the fine-scale structure of OTFs and explore the physical conditions within oceanic seismogenic zones. Consequently, with better constraints on the architecture of OTFs, we will be able to investigate how these fault systems serve as pathways for the hydration of oceanic plates and how they interact with the underlying mantle (e.g., Eakin et al., 2018). This last topic is of great interest not only to communities studying subduction processes but also to biogeochemical groups investigating the chemical reactions associated with aqueous alteration of peridotite to serpentine minerals, as the interaction between seawater and the cold mantle lithosphere is an important mechanism in the formation of hydrogen, methane, and small organic molecules (Cannat et al., 2010; Francis, 1981).

5. Conclusion

We showed that the water reverberation signals of submarine earthquakes are both interpretable and modelable and can be used to constrain the absolute location of the rupture within a given fault. By computing synthetic water phases using a hybrid method for modeling body wave arrivals with 3D source-side structures, we relocated three OTF events with a ~5-km uncertainty along the GTF in the EPR using land-based stations alone. The solutions we obtain are within the area of the average location error of the GCMT catalog for OTF earthquakes (~20 km) and are consistent with the tectonics of the region. We also demonstrated that current global bathymetry maps have enough resolution to allow our methodology to be applied in any oceanic setting in the world. Our approach of using water reverberation phases as fingerprint of earthquake location is a first step towards a reliable global OTF earthquake catalog and opens the door to a new and unexplored way of studying OTFs in general.

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