High Resolution Seafloor Structure of the Gofar Oceanic Transform Fault

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High Resolution Seafloor Structure of the Gofar Oceanic Transform Fault

By

Paige Koenig

Accepted in Partial Completion of the Requirements for the Degree Master of Science

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Master's Thesis

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Paige Koenig

May 30, 2024
High Resolution Seafloor Structure of the Gofar Oceanic Transform Fault

A Thesis
Presented to
The Faculty of
Western Washington University

In Partial Fulfillment
Of the Requirements for the Degree
Master of Science

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Abstract

The Gofar oceanic transform fault (OTF) accommodates 12.5 cm/year of plate boundary motion through large earthquakes, microseismic swarms, and aseismic slip in distinct regions of the fault along strike. Local and teleseismic observations show that well-coupled segments of the fault tend to fail via M~6 earthquakes roughly every 5 years. These fully-coupled segments are bound by barrier zones, up to ~10 km-wide, that do not generate large-magnitude earthquakes, but instead host microseismic swarms, accompanied by aseismic slip (Shi et al., 2021). Geophysical modeling and observations provide evidence that hydrothermal fluid circulation and fault damage may influence slip behavior segmentation. Here I present the fine-scale surface morphology of the Gofar OTF as characterized by 1-meter-resolution autonomous underwater vehicle (AUV) multibeam bathymetry. These data reveal a complex fault system, with variations in fault strike, fault bends, sub-parallel strands, and a heterogeneous damage zone of variable width. The scale of fault complexity shown through the AUV-determined micro-bathymetry is not previously imaged with ship-based bathymetry (> 50 m resolution) and provides significant new insight into how surface fault morphology changes along strike. Newly mapped fault structural variation partially correlates with the along-strike rupture segmentation from local seismic observations, reflecting a possible change in material or structural fault properties related to changes in fault rheology and mechanisms of fault slip. Using new high-resolution information on surface morphology, we find that regions that sustain repeated, high magnitude ruptures are expressed as a relatively linear principal slip zones (PSZ) striking in a similar direction to global plate motions contained within a relatively thin (≤ ~ 500 m) margin of damage.
In contrast, regions that contain deep microseismicity and no high magnitude rupture are geomorphically complex, with changes to PSZ orientation within a relatively wide (≥ 500 m) region of damage. These observations imply that at Gofar, geometric bends in fault strike are also associated with a widened zone of subsidiary faulting and distributed fractures which could facilitate enhanced fluid flow, allowing more water to extend through the seismogenic zone, subsequently altering the material properties that contribute to slow slip.
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1 Introduction

Oceanic transform faults (OTFs) represent excellent environments to study the link between fault architecture and rupture dynamics because they have a simpler geometry than continental strike slip fault systems and slip rates that are well constrained by plate spreading rates. Several first order OTF characteristics are well understood. Brittle failure is limited to the 600°C isotherm in the oceanic lithosphere, as confirmed by thermal models (Abercrombie & Ekstrom, 2001; Behn et al., 2007; Roland et al., 2010), rock fabric (Warren & Hirth, 2006), and laboratory friction experiments (Boettcher et al., 2007). Characterizing thermal structure is thus foundational to understanding the rheology of the oceanic lithosphere (Abercrombie & Ekstrom, 2001; Boettcher & Jordan, 2004; Braunmiller & Nábělek, 2008; Wilcock et al., 1990). Assuming this thermally-dependent seismogenic area, along with plate spreading rates, teleseismic observations indicate that much of the slip at both fast and slow-slip OTFs (~ 80%) occurs aseismically (Boettcher & Jordan, 2004).

Observations of large magnitude earthquakes (~ M5.5 – 7.1) observed at OTFs over multiple earthquake cycles (Figure 1) indicate that large ruptures occur on roughly the same segments along-strike a given fault, and are separated by segments that do not sustain large-magnitude ruptures and are likely associated with stable creep or transient slow slip (Boettcher & McGuire, 2009; McGuire, 2008; Shi et al., 2021; Sykes & Ekström, 2012; Wolfson-Schwehr & Boettcher, 2019). This spatial variation in dominant slip behavior is also observed along continental strike slip boundaries such as the San Andreas fault and other tectonic plate boundaries such as Hikurangi and Nankai (Brudzinski & Allen, 2007; Kano et al., 2019; Liu et al., 2022; Wallace, 2020).
Along-strike variations in slip behavior at OTFs and other fault boundaries introduce important questions about how fault structure or material properties impact slip mechanisms and promote earthquake rupture and/or aseismic slip transients.

The Gofar OTF along the EPR is a natural laboratory for investigating the relationship between shallow fault mechanics and rupture patterns due to the breadth of local datasets available, including newly-acquired, high-resolution bathymetry presented in this thesis. The Gofar transform is the western-most OTF within the Quebrada-Discovery-Gofar (QDG) oceanic transform fault system located on the EPR (Figure 2), which accommodates spreading between the Pacific and Nazca oceanic plates. The Gofar OTF contains three fast slipping (12 – 15 cm/yr), left lateral transform faults linked with two intra-transform spreading centers (Searle, 1983).

In 2008, ocean bottom seismometers (OBS) deployed at Gofar captured the end of the earthquake-cycle along the western-most section of the G3 fault segment (Figure 3). This earthquake sequence included a M6 event, as well as a prolific foreshock and aftershock sequence in the days and months preceding and following the M6 event. Local seismic observations revealed a ~ 10 km section of the fault adjacent to the M6 rupture area that failed in a foreshock earthquake swarm in the several weeks prior to the ~ M6 rupture (McGuire et al., 2012). I will refer to this 10 km long foreshock zone as the “barrier” (Figure 3d). A 3% decrease in shear wave velocity through the barrier was observed during the foreshock swarm, indicating a time-dependent change in elastic properties during the swarm (McGuire et al., 2012). This finding suggests that some change in the physical properties of the barrier occurred right before the M6 mainshock. The change in elastic properties within the barrier is hypothesized to be caused by a
strain transient (or aseismic slip transient), which could have led to changes in rock porosity through dilatancy (Gong & Fan, 2022; Liu et al., 2023; Roland et al., 2012). This kind of slip transient could be consistent with rheologic conditions facilitated by elevated pore fluid pressure or the unique frictional properties of hydrated phases (McGuire et al., 2012). Earthquake swarms within the barrier region extend into the upper mantle (~ 8 km), whereas aftershocks from the M6 rupture are confined to the crust (< 6 km) (Gong et al., 2022; McGuire et al., 2012), which also provides evidence of different physical conditions (hydrothermal fluid flow and/or thermal structure) between the barrier and the 2008 rupture zone. Additionally, stress drops are higher within the 2008 rupture zone and lower in the barrier, indicating there is a change in fault strength between the barrier and 2008 rupture patch (Moyer et al., 2018). The findings from the 2008 experiment reveal Gofar as a highly variable fault zone both along strike and through time. Understanding the differences between the barrier zone and rupture area are key to understanding how fault properties influence fault rupture behavior.

Past work at Gofar implies that some change in material properties controlling slip behavior. Some proposed explanations for the persistent rupture/barrier regions at Gofar include increased crustal and mantle hydration caused by an increase in fluid filled pathways (Kohli et al., 2021; Leptokaropoulos et al., 2023). Models of P-wave velocity structure from seismic tomography and density derived from gravity anomalies at Gofar (Figure 4b) indicate that the barrier zone is associated with a significant low-velocity anomaly that extends throughout the crust, but shows only a slight decrease in density relative to ‘normal’ oceanic crust (Roland et al., 2012). This anomaly is consistent with increased porosity of 1.5 – 8% (Roland et al., 2012). Microseismicity
within the barrier, and a lack of large-magnitude earthquake ruptures could be controlled by either elevated fluid pressure in fluid-filled pores, or an increase in crustal and mantle hydrothermal alteration, although a larger gravity signature would be expected for the latter (Roland et al., 2012). The presence of metamorphically altered phases like serpentine and talc are also a proposed explanation for the aseismic creep observed along the Parkfield section of the San Andreas fault (Schleicher et al., 2010), and could be occurring in the barrier at Gofar. Increased fluid-filled porosity and any associated alteration in the barrier could be facilitated by increased fracturing or structural complexities that weaken the surrounding rock and allow for fluid pathways deep into the lithosphere (Froment et al., 2014; Rice, 1992; Roland et al., 2012) (Figure 4a). This is supported by the deep earthquakes within the barrier (Figure 3b), which are consistent with depressed isotherms (a cooler fault zone) facilitated by enhanced hydrothermal cooling (Behn et al., 2007; Boettcher et al., 2007; Roland et al., 2012; Warren, 2016). While increased fluid-filled porosity via fracturing and rock damage is supported by the combined velocity structure (Liu et al., 2023; Roland et al., 2012) and earthquake observations (Gong & Fan, 2022; McGuire, 2008), it remains unclear what material properties are controlling rheology there.

A potential strategy for identifying the source of lateral changes in material properties employs detailed mapping of the tectonic structure of the fault system. Detailed mapping is important because structural changes may indicate changes in the frictional conditions of a given area (Anderson, 1951; Byerlee, 1978), which in turn may be influenced by fault zone material properties such as rock type (Anderson, 1951; Byerlee, 1993; Fossen, 2010; Scholz, 1998; Sibson, 1986), porosity (Byerlee, 1993;
Flemings, 2021; Healy, 2008), temperature (Behn et al., 2007; Roland et al., 2010), and dilatancy (Escartín et al., 2001). However, the remote location of most OTFs, and certainly Gofar, has made it difficult to characterize the basic geomorphology and geology at the outcrop scale. Current bathymetric maps provided by shipboard surveys at Gofar (~ 25 – 75 m resolution) are insufficient to distinguish fine-scale features (Figure 5a, b), including first order surface expressions of fault zone complexity (i.e., damage, fracture patterns), and existing bathymetry is too coarse to be useful for identifying a primary slip zone (PSZ) or distinct regions of damaged or deformed rock (Fossen, 2010).

At Gofar, current 25 m resolution shipboard multibeam data reveal a 500 m seafloor depression within the western segment of Gofar that coincides roughly with the location of the barrier region (Figure 2) (Froment et al., 2014). This depression is hypothesized to be associated with an extensional bend in the primary fault trace, possibly associated with a pull-apart-basin, but high-resolution morphological data are required to map individual fault strands or detailed changes in geomorphic features along the different earthquake and barrier segments (Froment et al., 2014; Liu et al., 2020). Similar releasing geometries for continental faults are mapped in great detail using LiDAR data, and fine-scale variations have suggested relationships between fault structure and slip (Barth et al., 2012; DeLong et al., 2010; Scott et al., 2020). Autonomous underwater vehicles (AUVs) provide a new opportunity to collect equivalently high resolution (1 m) bathymetric maps of OTF fault boundaries using multibeam sonar technology (Figure 5c). Assuming physical properties of the fault zone influence slip behavior, high-
resolution imaging should help to illuminate relationships between fault rock properties and structural expression apparent at the surface, and specific slip mechanisms.

This thesis uses a new data set collected along the Gofar transform fault as a part of a multidisciplinary NSF-funded 4CastGofar Earthquake Experiment (Boettcher et al., 2023) to address first order scientific questions linking fault structure and slip at the Gofar OTF. From January – March 2022 we collected a series of new marine geophysical observations using the R/V Thompson (cruise number TN399) along the eastern (G1) and western most (G3) transform fault segments to investigate evidence for lateral changes in material properties along distinct segments at Gofar. As a part of this experiment, we conducted 14 AUV multibeam sonar surveys using Woods Hole Oceanographic Institution’s AUV Sentry (Figure 6) which mapped the seafloor within discrete along-strike fault segments and collected complementary datasets including side-scan and CHIRP sub-bottom profiles, conductivity, temperature, and depth (CTD) data, oxygen redux potential (ORP) data, magnetics and photo transects (~ 8 m of altitude).

In conjunction with AUV mapping, we conducted a 3-day electromagnetic experiment (Chesley et al., 2023; Chelsey 2024, in review) that crossed over distinct rupture segments and recovered ocean bottom seismometers (OBS) deployed the year prior. The multitude of datasets collected during TN399 in 2022 added to an existing body of work coordinated to understand fault mechanics at Gofar, including recent datasets also associated with the 4CAST Gofar Earthquake Experiment, as well as data sets from earlier experiments that included reconstructions of seismicity from OBS recordings of a M6 earthquake sequence on G3 (McGuire, 2012), high-precision
earthquake relocations (Gong & Fan, 2022), stress drops (Moyer et al., 2018), crustal Vp/Vs ratio modeling (Liu et al., 2023), wide-angle seismic tomography (Roland et al., 2012), rock sampling via dredging (Bahruth, 2023; Warren, 2016), electromagnetics (Chesley et al., 2023; Chelsey 2024, in review), and ship-based bathymetry and gravity observations (Morrow et al., 2021; Pickle et al., 2009). Collectively, ongoing work at Gofar is coordinated to identify the primary controls on changes in seismic behavior along-strike and with depth at oceanic transform faults.

The 4CastGofar experiment included the first AUV experiment to map an OTF at 1 m resolution, presenting an exciting opportunity to detail structural features never imaged before. In this thesis, I present the results from processing and interpretation of high resolution multibeam data collected by AUV Sentry that maps seafloor morphology of the western-most transform segment, G3, extending across several repeating earthquake rupture segments and barrier zones. I map first order fault features that may reflect or influence deformation and material controls on seismogenesis. I address the following questions:

(1) What is the fault structure and how does it change along strike?
(2) Does fault structure vary with along strike changes in slip behavior?
(3) Can surface morphology contribute to our current understanding of controls on slip behavior?
2 Background

Local studies along fast-slipping (~ 10 – 15 cm/yr) OTF boundaries such as Gofar identify rupture barriers between large (~ M6) rupture patches that seem to fail in microseismic earthquake swarms (Figure 3). Temporal and spatial patterns of microseismicity during these swarms, as well as temporal and spatial changes in elastic properties are used to suggest aseismic slip transients may occur within these barriers, potentially driving microseismicity and accounting for the majority of plate boundary strain in those segments (McGuire et al., 2012). This spatial variation in dominant slip behavior is also observed along OTF systems such as Blanco (Braunmiller & Nábělek, 2008), Discovery (McGuire, 2008; Wolfson-Schwehr et al., 2014), Eltanin (Shi et al., 2021; Wolfson-Schwehr & Boettcher, 2019) and Charlie-Gibbs (Aderhold & Abercrombie, 2016), along continental strike-slip boundaries such as the San Andreas fault (Liu et al., 2022) and at subduction boundaries such as the Hikurangi (Wallace, 2020) and Nankai (Kano et al., 2019).

Along-strike variations in slip behavior at OTFs and other fault boundaries introduce important questions about how material properties could promote aseismic and/or seismic slip. Some candidate changes in material properties that are expected to influence fault slip behavior include lateral variations in thermal structure (Roland et al., 2010), lithospheric material (Anderson, 1951; Scholz, 1998; Sibson, 1986), pore fluid pressure (Rice, 1992) and fault damage (Chester & Chester, 1998; Kato & Ben-Zion, 2020). In the following sections, I elaborate on how these material properties could impact slip at Gofar and other fault boundaries. Then I discuss local observations and models from the Gofar transform fault, and how the structural interpretations in this
thesis work can inform hypotheses on the link between fault structure and slip behavior at OTFs generally.

2.1 Thermal structure

The rheology of the oceanic lithosphere is thermally controlled (Abercrombie & Ekstrom, 2001; Boettcher & Jordan, 2004; Braunmiller & Nábělek, 2008; Wilcock et al., 1990). In the oceanic lithosphere, the maximum depth of earthquakes correlates to a 600°C isotherm based on estimates of conductive cooling of a half space. Various lines of evidence remarkably agree with 600°C being a critical thermal transition. Peridotite mylonites dredged from OTF boundaries show that viscous deformation occurs at ~600°C (Warren & Hirth, 2006). Laboratory friction experiments on olivine (e.g., Boettcher et al., 2007), a primary component of the oceanic lithosphere, indicate that olivine transitions from stable to unstable sliding at 600°C. While the 600°C isotherm predicted by half space cooling models is generally successful in determining the limit of earthquakes in the oceanic lithosphere, it does not include many processes known to influence temperature – such as hydrothermal cooling and frictional heating (Roland et al., 2010). Recent earthquake relocation data at both fast- (Kuna et al., 2019; Yu et al., 2021) and slow- slipping OTFs (Leptokaropoulos et al., 2023) have found deep earthquake swarms that originate below the half-space-predicted 600°C isotherm within distinct regions along-strike, suggesting there could be other mechanisms influencing temperature and/or brittle failure in the oceanic lithosphere. Additionally, global observations of earthquakes on OTF boundaries indicate low seismic coupling (~80% aseismicity) when the 600°C isotherm is used as the base of the seismogenic area (Boettcher & Jordan, 2004). Combined, earthquake observations and low seismic
coupling on OTF boundaries imply that there is some variation in fault zone properties that cannot be explained by temperature, or simplified thermal modeling alone.

Sophisticated numerical thermal models that include important rheologic feedbacks on the temperature structure of the oceanic lithosphere (i.e., Behn et al., 2007; Roland et al., 2010), find that hydrothermal circulation significantly increases the depth of brittle failure (Roland et al., 2010). Potential mechanisms for increased fluid circulation include intense fracturing and enhanced permeability. Fluids extending to the lower crust and upper mantle would also lead to mineral alteration to talc and serpentine. These phases, if present along the fault zone, would be expected to influence tectonic processes and slip because hydrated phases are frictionally weaker (Escartín et al., 2001) and in some cases, more frictionally stable (i.e., influence velocity strengthening behavior) (Moore & Rymer, 2007). Both enhanced cracking and porosity, and hydrothermal alteration would lead to a decrease in seismic velocities in the crust and upper mantle, which is observed in the barrier at Gofar (Roland et al., 2012) (Figure 4b). Elevated porosity is also supported by deep earthquakes within the barrier, which could be indicative of depressed isotherms (a cooler fault zone) facilitated by enhanced hydrothermal cooling associated with cracking, fluid-filled pore-space and enhanced permeability (Behn et al., 2007; Roland et al., 2010).

Aligned with the observation that a simple half-space cooling model-based temperature-dependent rheology overestimates the seismogenic area based on low observed seismic coupling, it seems likely that there are heterogeneities in thermal and/or frictional properties that are also controlling slip behavior on OTFs. Existing observations and models at Gofar including characterizations of temperature structure,
velocity structure, and earthquake observations, indicate that fluids and porosity/permeability structure also play a prominent role controlling rheology at OTFs.

### 2.2 Lithospheric material properties

The frictional properties of material in the crust and upper mantle influence fault mechanics and rheology (Anderson, 1951; Scholz, 1998; Sibson, 1986). Friction (μ) characterizes the resistance to slip, and experimental data indicates that the sliding friction between most rocks is independent of rock type and is constant at a given lithostatic pressure (\(0.6 < \mu < 0.9\)) (Byerlee, 1978). Serpentine, talc, and clays, which are hydrated minerals abundant in the oceanic lithosphere, have characteristically low coefficients of friction relative to their protolith (\(\mu \approx 0.1\)) and thus require less shear strength to begin stable sliding (Hirth & Beeler, 2015; Reinen et al., 1991). Due to the low coefficient of friction, clay, serpentine, and talc are considered to be weak rocks, while most other rocks in the brittle crust are mechanically strong (Byerlee, 1978; Escartín et al., 2001).

Rock strength alone does not determine if a rock will fail unstably (as an earthquake) with applied shear stress; once slip is initiated on a fault plane, the stability of slip is controlled by rate and state laws: \((a - b) = \Delta \mu_{ss} / \Delta \ln(V)\) where \(\Delta \mu_{ss}\) is the steady state friction coefficient and \(V\) is sliding velocity (Marone, 1998; Scholz, 1998). \((a - b)\) is a frictional parameter where \((a - b) > 0\) means as velocity increases so does the frictional resistance to slip, which promotes stable slip, also referred to as velocity strengthening. \((a - b) < 0\) means as velocity increases the frictional resistance to slip decreases, which promotes unstable slip and runaway rupture, also referred to as velocity weakening. Rocks such as granite are velocity weakening at temperatures < 350°C and olivine is
velocity weakening at temperatures < 600°C (Boettcher et al., 2007). Hydrated minerals like clays and serpentine are often attributed to velocity strengthening regimes \((a - b > 0)\) and can prohibit seismic rupture under certain pressure and temperature conditions (Reinen et al., 1991; Schleicher et al., 2010; UMass Claybox Experiments, 2017).

Serpentine phases have a subsequent impact on slip, where the decrease in overall rock strength can potentially facilitate localized weak fault zones and the velocity-dependence can allow for aseismic creep (Hirth & Beeler, 2015). Along the Parkfield section of the San Andreas fault, clays may contribute to the fault's weakness and dominantly creeping behavior (Schleicher et al., 2010). The formation of weak hydrothermally altered rocks is particularly likely at OTFs, where fluids may easily access the seismogenic lithosphere.

### 2.3 Pore fluid pressure

Pore-fluid pressures lower the effective normal stress (effective normal stress \(\sigma'_n\) = normal stresses \(\sigma\) – pore fluid pressure \(u\)) which allows for slip at lower shear stresses (Bedford et al., 2021; Faulkner & Rutter, 2001). Pore fluid pressure changes can also change the degree of fluid saturation by altering pathways for fluids to migrate (Flemings, 2021), subsequently changing stress conditions at depth. As described above, fluid saturation can lead to metamorphic alteration (Faulkner & Rutter, 2001). The geometry of pores is also important; the crack aspect ratio can impact the availability of fluid to migrate by influencing permeability (Flemings, 2021). Additionally, in terms of their impact on elastic properties (i.e., seismic velocity) crack-like pores can modify 2D velocity structures more than spherical pores if they are in the right orientation (Kuster & Toksöz, 1974). High pore fluid pressures are suggested as a
possible mechanism for relative fault weakness (Rice, 1992) and are also thought to promote creep (Liu & Rice, 2005). Slow slip is likely controlled by variations in pore fluid pressure (Liu & Rice, 2005; Rice, 1992), as documented along subduction zone boundaries including Nankai (Bedford et al., 2021) and Cascadia (Brudzinski & Allen, 2007).

Increased fluid-filled porosity in the barrier is the leading hypothesis for causing variations in slip at Gofar (Liu et al., 2023; Liu et al., 2020; Roland et al., 2012). Local tomographic and gravity models show a low velocity zone within the barrier region, consistent with enhanced fluid-filled porosity (Roland et al., 2012). Enhanced fluid-filled porosity could explain the velocity-strengthening nature of the barrier through several processes. If a heterogeneous distribution of trapped fluids is present at depth, this would lead to heterogeneous stress conditions (Fagereng & Sibson, 2010; Roland et al., 2012). Additionally, connected fluid pathways could lead to metamorphic alteration, lubricating the fault zone with weak material lithologies, and promoting velocity-strengthening behavior (Rice, 1992; Roland et al., 2012). Either mechanism could be occurring in the barrier and contribute to slow slip. In fluid-saturated material undergoing shear, dilatancy strengthening, which refers to the tendency for pore space to increase and pore fluid flow to be restricted under shear, could suppress dynamic slip even in velocity-weakening conditions (Segall et al., 2010). Dilatancy can increase in highly fractured damage zones (Segall et al., 2010), which could be consistent with the structure of the barrier at Gofar (Froment et al., 2014) and the low velocity zone in the barrier (Roland et al., 2012). If dilatancy strengthening in zones with high porosity leads to heterogeneous slip behavior observed at Gofar, characterizing how these high-
porosity zones are formed, and how they could be identified from surface fault
expressions would be useful for characterizing fault zone slip conditions across entire
fault systems.

### 2.4 Fault structure controls

First order structural observations, including the location and orientation of the PSZ and
the width of damage at the surface of a fault zone boundary could give important insight
into the physical controls on slip at a given fault boundary.

Slip along seismically-active strike slip faults is typically concentrated in a thin
region of shear referred to as the PSZ bordered by a damage zone of fractured,
brecciated, and pulverized rock (Figure 4a) (Chester & Chester, 1998; Dor et al., 2006;
Sibson, 1986). The PSZ is a narrow, focused fault core that accommodates slip over
many rupture episodes (Chester & Chester, 1998). The remainder of inelastic strain is
distributed through the damage zone, a volume of warped and fractured rock that
exhibits the highest degree of damage close to the PSZ (Shelef & Oskin, 2010). Based
on experiments of damage zones (Nelson & Jones, 1987; Shelef & Oskin, 2010), the
distributed shear displacement of damage zones on strike slip faults can account for up
to 60% of the total displacement across the fault zone (Nelson & Jones, 1987; Shelef &
Oskin, 2010). Additionally, fault zones with wide margins of pulverized material along
the PSZ are shown in lab experiments to promote aseismic slip (Marone, 1998).

The extent of damage along strike may influence fault rheology and the style of
fault slip by creating more pathways for fluid to migrate through brittle fractures that
extend through the crust (Figure 4a) (Byerlee, 1993; Chester & Chester, 1998;
Kanamori & Brodsky, 2004; Rice, 1992; Sibson, 2003). Heterogeneous variations in
fault damage can lead to heterogeneous fluid-mechanical rock properties, and can subsequently be associated with variations in slip behavior (Froment et al., 2014; Roland et al., 2012; Shi et al., 2021). Heterogeneous fluid-mechanical rock properties are a proposed explanation for slip distribution along both fast- and slow-slipping OTFs, such as the Gofar (McGuire, 2008) and the Chain fracture zone on the Mid-Atlantic ridge (Leptokaropoulos et al., 2023).

2.5 Gofar OTF structure and seismicity

2.5.1 Gofar Geomorphic Segmentation

Shipboard multibeam (~25 m for G3 and G1 segments; ~75 m for all of Gofar) reveals transform segments that are defined by relatively linear sections of fault with small-scale median valleys, as well as segments with depressions where faulting patterns are more complex (note the changes in seafloor depth in Figure 2). The Gofar OTF contains three, fast slipping sinistral east-west striking transform segments separated by two intratransform spreading centers (ITSC) (Figure 2) (Searle, 1983). The western-most transform strand (G3) is ~90 km long and connects with the EPR spreading ridge in an uplifted “J-shaped” ridge structure (Gong et al., 2022; Grevemeyer et al., 2021; Searle, 1983). The center of the G3 fault strand contains a 4 km deep depression that runs 15 km along strike. G2 connects the western-most segment G3 to the eastern-most segment G1. There is a deep (~4.8 km, as compared to ‘normal’ seafloor at ~3.5 km) ITSC connecting G1 and G2. G1 is ~45 km long and connects to the EPR spreading ridge to the east. G1 contains a similar central depression to G3 that is also deep (~4 km depth) and wide (~5 km). This thesis will primarily focus on linking the
geomorphology and fault slip behavior at G3, though observations made here may be applicable to other segments.

2.5.2 Gofar Seismicity Distributions

The western-most transform segment at the Gofar OTF, G3, contains distinct seismic and aseismic zones along strike and with depth that exhibit different styles of slip (Boettcher & McGuire, 2009; Gong et al., 2022; McGuire et al., 2012). Recent work by Gong & Fan, (2022) uses earthquake relocations from the 2008 M6 Gofar earthquake dataset (McGuire et al., 2012) to identify five distinct seismic patches at G3 (Figure 3). These five seismicity-determined zones appear to be related to three dominant slip styles, including: repeated, high magnitude ruptures zones (Figure 3 - Zones 1, 3 and 4), a barrier zone that contains deep microseismicity and no high magnitude rupture (Figure 3 - Zone 2) and the December swarm region (Figure 3 - Zone 5). Below I will outline the along-strike changes in seismicity and how they relate to slip segmentation.

Starting from the eastern portion of the fault that connects G3 to the ITSC, Zone 1 (Gong & Fan, 2022) presents as a ~ 45 km long, presumably seismically-well-coupled region marked by frequent (~ 5 – 6 yrs) M5 – 6 earthquakes (Shi et al., 2021) and abundant microseismicity (Figure 3; note that the Eastern extent of Zone 1 is not represented in Figure 3). West of this rupture zone is the 10 km barrier (as described by McGuire et al., 2012) where foreshocks propagated for several weeks leading up to the 2008 M6 rupture before seismicity rates dropped to close to background levels. In the barrier zone, there are distinct clusters of shallow (2 - 6 km) and deep (7 - 8 km) microearthquakes that could correspond to different fault strands given the increased fault width in shipboard bathymetry data (Zone 2 - Gong & Fan, 2022). Directly west of
the barrier zone is the >15 km 2008 M6 mainshock rupture area; the mainshock initiated on the western end of this zone at ~6 km depth (Zone 3 - Gong & Fan, 2022; Shi et al., 2021). The majority of the seismicity in the mainshock region occurred from 4 - 7 km depth (Gong & Fan, 2022; McGuire et al., 2012). Adjacent to the mainshock zone is a ~12 km transition zone that is marked by shallow (4 - 7 km) earthquakes that cluster in three distinct fault patches along strike that could correspond to active fault strands and a potential step-over (Zone 4 - Gong & Fan, 2022). Fast seismic slip during the 2008 rupture may have extended partially through this zone, as most of the earthquakes included in the Gong and Fan (2022) analysis are aftershocks of the 2008 M6 event. The western-most zone (Zone 5 – Gong & Fan, 2022) connects the fault to the EPR. This section of the fault hosted a swarm of ~20,000 earthquakes that followed the 2008 M6 mainshock in December 2008 (Gong & Fan, 2022; McGuire et al., 2012). This region also hosted distinct repeating swarms prior to the M6 mainshock with 24.4 day recurrence intervals, which Gong and Fan (2022) refer to as pulsating events. These may be related to magmatism close to the ITSZ (Figure 3). I will note that in this study I define the rupture zone as the entire boundary from 105.98°W to 106.2°W, but Gong & Fan (2022) limit the rupture zone to 105.98°W to 106.1°W (Zone 3 - Figure 3) and define 106.1°W to 106.2°W as a transition zone (Zone 4 - Figure 6). In the absence of a finite fault inversion, I consider the full aftershock extent to be the extent of the rupture zone (Figure 3d).

Given that the western-most December swarm segment and the foreshock barrier segment can sustain significant sequences of microearthquakes but do not host large magnitude ruptures, it seems likely that physical properties along the fault zone
differ in these ~ 10 km-long segments from those in the rupture zones (McGuire et al., 2012). Based on high-precision earthquake locations and fault zone imaging (Gong & Fan, 2022), the transition from foreshock activity to large seismic rupture in the main shock zone may correspond with a slight change in fault strike. Although difficult to resolve in low-resolution (ship-based) multibeam bathymetry maps, this could be the location of an extensional bend in the fault, which would also potentially explain the bathymetric relief in the central depression if this zone is similar to an extensional pull-apart-basin. If the transition in earthquake behavior corresponds with a geometrical bend in fault strike, extension around this area could lead to a wide zone of normal faulting and distributed fractures, which could facilitate enhanced fluid flow and hydrothermal circulation, allowing water to extend throughout the seismogenic zone. Many studies have hypothesized a slight change in fault strike (Froment et al., 2014; Guo et al., 2023; Liu et al., 2023; McGuire et al., 2012; Roland et al., 2012), however, visualizing this kind of fracture pattern is not possible with ship-based bathymetry data. Based on seismic imaging and modeling work (Liu et al., 2023; Roland et al., 2012), dilatancy strengthening in discrete zones of high-porosity and pore-fluid pressures is the current leading hypothesis for explaining variations in slip behavior at Gofar.
3 Data and Methods

3.1 Multibeam Data Collection

A total of 14 AUV Sentry dives were conducted from R/V Thompson (TN399) along the eastern (G1) and western most (G3) transform fault segments in 2022 as part of the 2019 - 2022 4Cast Gofar Earthquake Experiment (Figure 6 and A1). Multibeam data were collected at an average altitude of ~ 70 m above bottom with a Kongsberg EM2040 Multibeam Sonar system operating at 400kHz along track lines spaced ~ 160 m apart. After multiple rounds of data and navigation processing, these data yield bathymetry at 1 m resolution. Some surveys were conducted with near-bottom video using a 24 Megapixel still image camera, which required a survey altitude closer to 8 m, and could not be acquired at the same time as the multibeam data. Camera transects thus altered the line spacing and line distribution on dives. Sidescan and chirp sub-bottom data were collected simultaneously with both bathymetry and camera data using a Edgetech 4 – 24 kHz Sub-Bottom Profiler and Edgetech 220 multibeam echosounder with 120 kHz/540 kHz side-scan enabled. Sentry was also equipped with an APS 1540 magnetometer, National Oceanic and Atmospheric Association’s (NOAA) Pacific Marine Environmental Laboratory Oxygen Redux Potential Sensor (PMEL ORP), and conductivity, temperature, and depth (CTD) sensor.

Average speed, duration, and depth for each dive survey varied with the vehicle flight path, vehicle altitude, and obstacle interference, but in general, each G3 dive survey was conducted at an average speed of ~0.9 m/s for an average duration of ~ 19 hrs at a mean depth of ~ 3400 m. The entire mapped region covered ~ 90 km$^2$ from 106°16.5´W to 105°25´W, with emphasis on mapping along-strike structures of interest.
While a total of 14 AUV *Sentry* dives were conducted along two transform segments (G1 and G3) (Figure A1), this study focuses on the 10 multibeam bathymetry dives collected from the western-most transform, G3 (Figure 6).

### 3.2 Overview of multibeam processing steps

I conducted four distinct rounds of processing for AUV *Sentry* multibeam data (Figure 7):

1. Initial shipboard processing
2. Post-shipboard processing
3. Feature-based individual dive processing
4. Feature-based merged dive processing

Initial shipboard processing of multibeam data includes applying predetermined vehicle attitude offsets and determining vehicle positioning by merging the navigation collected on the AUV (referred to as dead reckoning) and navigation collected relative to the research vessel position using an ultra-short baseline (USBL) acoustic positioning system co-registered with the vessel’s satellite GPS positioning system (Figure 7 - “initial shipboard processing”) (Caress et al., 2008; Chadwick Jr. et al., 2001). The initial-processed data were then post-processed onboard the ship for early post-dive interpretations of seafloor morphology and for determining subsequent dive plans. I refer to this as post-processed shipboard multibeam (Figure 7). Feature-based individual dive processing was conducted once the cruise was completed (Figure 7) in order to improve the relative navigation and remove noise associated with extraneous sonar pings. As a final data processing step, each individual AUV Sentry dive was merged into one complete map – and in doing so, I determined the final navigation
solution that incorporated agreement between overlapping dive footprints (Figure 7). All AUV Sentry multibeam processing was completed utilizing MB-Systems open source software (Appendix Figure B1; Caress & Chayes, 1996).

### 3.3 Navigation corrections

The navigation of raw AUV multibeam data needs to be processed in order to correct for errors in both the dead-reckoning and USBL-based navigation tools. Since electromagnetic waves needed for global navigation satellite system (GNSS)-based ship navigation do not penetrate through the water column, GNSS cannot be used to determine absolute vehicle positioning on a submerged AUV. Instead, AUV positioning is determined using two positioning methods that are combined in post-processing to determine a navigation solution. These include: (1) dead reckoning, or point to point vehicle fixes based on estimated vehicle heading and speed (recorded with the doppler velocity log, or DVL located on the vehicle) (Saksvik et al., 2021), and (2) ultra-short baseline acoustic positioning system, or USBL which, using acoustic ranging, provides a position of the vehicle relative to the research vessel where GNSS positioning is available. The USBL positioning system can determine absolute AUV positioning to ~10 m when in constant communication over the entire survey duration (Caress et al., 2012). The USBL transceiver and transducer pair must be calibrated to one another prior to the mission in what is referred to as a calibration of attitude sensors in USBL Systems (CASIUS) routine. This routine determines the pitch, roll, and heading misalignments between the USBL transceiver and vessel motion for the duration of the entire mission. The CASIUS routine was conducted on the R/V Thompson prior to the TN399 mission.
Shipboard processed AUV navigation positioning for TN399 was determined by merging USBL and DVL fixes to obtain an absolute vehicle navigation solution that is closer to the true vehicle flight path (Appendix Section B). If present, outlying USBL positions are filtered out using a maximum range cutoff to eliminate unrealistic navigation data. The maximum range cutoff value can be manipulated when there are issues with the USBL dataset, but is often unchanged during a given research cruise. Then the two separate navigation datasets (USBL and dead reckoning) are combined using an applied complementary filter. These filter parameter values are often unchanged during a given research cruise, but can be modified in post-processing if there are issues with the USBL and/or DVL datasets. The combination of dead-reckoning and USBL fixes determine the initial shipboard processed vehicle positioning, which is closer to the true flight path than either of the raw navigational data sets (Caress et al., 2012; Wu et al., 2022).

3.4 Navigation errors

Despite the combination of dead reckoning and USBL fixes in shipboard processing, navigational errors will still persist in this shipboard processed dataset. Navigational accuracy is an essential component of multibeam data to produce a high quality bathymetry grid, but due to the submerged nature of data collection, 3-D vehicle navigation positioning contains inherent uncertainties that accumulate over the duration of the survey. Two sources of error in this study include (1) obstacles in the vehicle terrain and (2) loss of USBL connection (between the ship and AUV).

Navigational error can be caused by changes in the AUV flight path due to obstacles in rugged terrain. When AUV nears the seafloor or encounters obstacles, the AUV
enters a correction procedure that abandons the maintained altitude and shuts off the propeller before maintaining safe altitude and resuming the mission coordinates. In these cases, DVL positioning is lost, resulting in poorer mission navigation that relies on USBL connection and/or prior DVL adjustments. Altitude aborts and steep terrain can lead to gaps in multibeam coverage and loss of absolute navigation position (Caress et al., 2008; Caress et al., 2012; Clague et al., 2020). Such a correction procedure was used once during this survey during Dive 621 when the Sentry AUV ran into a fault scarp, resulting in a loss of multibeam coverage and navigational accuracy for this dive (refer to Appendix Figure B 17 for collision location). Additionally, obstacle interference during Dive 621 caused the vehicle to experience a slipped clutch, where the controllers that command the servos to drive to precise navigational positions fail. A slipped clutch creates greater uncertainty for the vehicle’s speed, pitch, and roll because the vehicle is more difficult to drive and control. This mechanical issue was fixed on deck after vehicle recovery as to not impact subsequent dives, but the navigational errors still persist in the dataset for Dive 621.

Loss of the USBL connection between the AUV and research vessel results in navigation that relies solely on DVL measurements and thus increases navigational uncertainty (refer to Appendix B for navigation adjustments for each dive). During TN399, several other experiments were run in conjunction to Sentry AUV dives, including recovery of ocean bottom seismometers (OBS) and deployment and recovery of ocean bottom electromagnetic (OBEM) receivers. When the research vessel’s USBL transducer system is out of range from the AUV transceiver (> 11 km), the USBL navigation is lost and AUV positioning is determined using DVL measurements. Since
dead reckoning relies on the previous time-step, solely relying on dead reckoning can diminish the quality of navigation data and position accuracy can drift with time as small errors in position based on velocity and heading accumulate. Thus, when USBL connection is lost, the navigation uncertainty increases, and absolute positioning is diminished.

### 3.5 Feature-based multibeam processing procedure

Multibeam sonar data were post-processed after the TN399 cruise in steps I refer to as feature-based multibeam processing (Appendix Figure B1) using the open source MB-System software package (Caress & Chayes, 1996). I based our multibeam processing scheme (Figure 7) on both the shipboard post-processing workflow and MB-systems multibeam processing scheme (Appendix Figure B1) used by the Sentry engineering team on board. Given the noisiness of the individual AUV Sentry dives from post shipboard processing, I conducted a more elaborate navigation processing scheme that includes shifting the navigation via identifying navigation ties across overlapping dive tracks. This was completed first for individual dive tracks and then subsequently merged dive data for overlapping dive tracks (Figure 7). This produced a complete, high resolution image of Gofar’s G3 transform segment.

A processing flow is begun by converting multibeam data to MB-System data format following the MB-Systems processing scheme (Appendix Figure B1). When converted, ancillary files are created and can be read and manipulated by the different functions within the MB-Systems program. For example, an MB-System’s “fast bathymetry” or .fbt file contains each ping number and the associated absolute positioning which is used by the MB-System's terminal based gridding program.
“mbeditviz”. MB-System converted files were then processed by applying AUV vehicle configuration, attitude, sensor depth, navigation, and tides, following the processing procedure conducted by the on board engineering team. Multibeam sonar files underwent a round of a terminal-based program “mbclean” which removed abnormally short along track beams, pings outside of the determined depth range, and pings outside a maximum acceptable rate of change in heading (deg/sec). These sonar files were then processed within MB-Systems using “mbprocess” to incorporate those parameters prior to ping-editing. Manual ping editing was then conducted using the MB-System’s program “mbedit” in order to remove anomalous pings that could not be removed with “mbclean”. These edits were then incorporated into the files using “mbprocess” prior to navigation processing to ensure that I adjust the navigation of sonar files with clean multibeam sounding data.

The navigation position for each AUV Sentry dive went through two rounds of feature-matching navigation adjustments, including adjustment within overlapping segments of individual dives and feature-based adjustment between overlapping dives (Figure 7). Navigation adjustments were conducted using the MB-Systems “mbnavadjust” program which uses systematic feature matching to invert for a navigation model that is consistent with the original navigation data (Caress & Chayes, 1996). Adjustments for this dataset were conducted for both horizontal (x and y) and vertical (z) offsets (see Appendix Section B for individual dive adjustments). The “mbnavadjust” program automatically identifies overlapping segments, then the user can manually adjust the offset between the overlapping segments by creating navigational “tie-points”. These tie points are navigation control points that represent the
corrected, manually adjusted navigation offset between two overlapping swaths and are further used in the navigation inversion. Tie points are set when there is a clear feature imaged by two overlapping swaths, such as a volcanic mound or fault. Once tie points for well-matched, crossing features are set, an inversion is run that solves for navigation changes using each tie point. The solution for the optimal navigation model is determined as sparse overdetermined least square matrix problems, following the least squares algorithm by Paige & Saunders, (1982). If there are conflicting tie-points, such that one requires a control point to move west, while an adjacent one requires the same control point to move east, then the offsets will not fit the inversion model exactly, an inherent limitation of remotely navigated datasets. After the final navigation inversion is conducted, the adjusted multibeam data is inspected using the “mbeditviz” grid viewer, which temporarily grids and plots all data. If overlapping swath segments appear clear with little noise, the entire dive file is processed using “mbprocess” and the navigation model fit by the inversion is applied to all sonar data files. If there is still noise present, the inversion can be run again and iterated until the final product is consistent and less noisy. All ten, individual 1 m resolution G3 AUV Sentry dives are gridded and plotted in Appendix Section A.

The navigation of separate, overlapping dives were combined and corrected in reference to shared overlapping features using the “mbnavadjust” program (Figure 7). First, the navigation projects of separate, overlapping dives were combined using MB-Systems “mbnavadjustmerge” terminal based command. This ensures that the internal navigation of any individual dive is contained within the merged navigation project. Then, the merged navigation project is modified within “mbnavadjust” by creating tie
points between crossing dive segments. Once the final navigation inversion is achieved and the gridded product is inspected within the “mbeditviz” grid viewer, the navigation solution of the merged dives is applied to the dive files using “mbprocess”. Once the final navigation solution of individual and overlapping Sentry dives is corrected, inverted, and applied, the final, merged multibeam bathymetry dataset was gridded at 1m resolution using a weighted sonar footprint with slope gridding algorithm and spline interpolation.

3.6 Bathymetric Interpretations

Merged, navigation-corrected 1 m bathymetry grids were interpreted by tracing primary fault features geolocated using ArcGIS Pro’s polyline function. Given the high resolution of this dataset, this study focused on identifying previously-unmapped fault-related features such as the location of the main fault strand (here referred to as the principal slip zone, or PSZ), changes in main strand geometry, the width of fault-induced damage, and extent of subsidiary fractures.

I defined the PSZ as a linear, continuous fault trace that is > 1 km in length and parallel to relative Pacific and Nazca plate motion (~101.4°; DeMets et al., 2010). Any fault segments < 1 km are defined as subsidiary fractures for this study. PSZ interpretations were ranked by confidence as either high confidence (denoted by solid lines in Figure 12) and low confidence (denoted by dotted lines). High confidence refers to a fault trace that is continuous, > 1 km, and contains no gaps in high resolution bathymetry. Low confidence refers to an observable fault trace > 1 km, but with gaps in the data that limit the ability to interpret it as a continuous feature. I recognize that the
PSZ boundary interpreted from the surface morphology may not represent the true slip surface at all depths.

Fault-induced deformation refers to the volume or area of brittlely deformed rock that surrounds the PSZ, often including a region of increased shear and tensile fractures that are more concentrated than in the surrounding surface (Figure 4a) (Fossen, 2010; Sibson, 2003). The width of fault-induced damage is interpreted in this dataset using several attributes, including rugosity and fracture density (Rodriguez Padilla et al., 2022). Rugosity was calculated using NOAA’s Benthic Terrain Modeler (BTM) Vector Ruggedness Measure (VRM) in ArcGIS, which measures terrain heterogeneity using a tape-chain measurement, which is the ratio of the length of a profile distance to the length of a chain draped over the 2D terrain profile (Walbridge et al., 2018). I calculated rugosity using an 8 m sensitivity kernel which calculates rugosity over a set of 8 m wide neighborhood cells (Walbridge et al., 2018). This sensitivity was chosen because it best visualized the variations in rugosity within the 1 m-resolution G3 dataset. Interpretations of the width of the fault damage zone were ranked by confidence as either high confidence (denoted by solid lines) and low confidence (denoted by dotted lines). High confidence refers to damage margin that is continuous, clear, and contains no gaps in high resolution bathymetry. Low confidence refers to when the width of damage goes outside the bounds of high resolution bathymetry, or the margin is unclear. Additional data products include slope, aspect, and rugosity attributes which were calculated and gridded at 1 m resolution in ArcGIS (Appendix Figures C6, C7, C8). A combination of these data products in combination with the merged, 1 m-resolution bathymetry map were used for interpretations. Where possible, I also utilized seafloor observations from...
camera-transects to, for example, confirm the location of highly-brecciated seafloor rocks (Appendix Figures D2 and D3; Bahruth, 2023) and some subsidiary fault scarps that were well-imaged in seafloor photos (Appendix Figure D1; Tracy et al., 2023).
4 Results

The goal of this study is to address what fault features (geometry, damage) are present along strike the Gofar G3 OTF and consider if these fault features change with respect to changes in distinct rupture segments to better understand the lithospheric properties that could be controlling rupture behavior. To this end, I processed, mapped, and interpreted a 1 m resolution image of G3 (Figure 8; Appendix Figures C1- C5). The entire mapped region covered ~90 km$^2$ and a depth range of 2817 to 4104 m below sea level. It extends in longitude from 106.25°W to 105.5°W. In this section I elaborate on first order surface observations in 1 m resolution bathymetry and evaluate how prominent surface structures change laterally from east to west. I interpret bathymetry in the context of 5 prominent geomorphic regions from 105.81°W to 106.25°W (Figure 9):

(1) Eastern G3 (Figure 10; 105.81°W to 105.92°W)

(2) Eastern Central Depression (Figure 12; 105.92°W to 106.01°W)

(3) Western Central Depression (Figure 15; 106.01°W to 106.07°W)

(4) Western G3 (Figure 17; 106.07°W to 106.16°W)

(5) EPR ridge transform intersection (Figure 18; 106.16°W to 106.23°W)

Note that multibeam coverage from 105.5°W to 105.81°W included a single swath, and was difficult to interpret (Appendix Figure A8). Prominent first order observations include the location of the PSZ at the surface and the extent and morphologic expression of damage. While the width of deformation is difficult to constrain in some locations given the limited extend of high-resolution multibeam coverage, general variations are highlighted below, with an emphasis on relative and not absolute widths.
4.1 Eastern G3

The morphology within the eastern mapped region of G3 is marked by a localized, linear PSZ (Figure 10). The PSZ strikes at an angle (~ 100°) similar to the relative plate motions of the Nazca and Pacific plates (101.4°) observed in 25 m resolution bathymetry and calculated with MORVEL models (DeMets et al., 2010). The PSZ splits into two sub-parallel fault strands at ~105.86°W (Figure 10) located about 70 m apart from one another, but otherwise the PSZ remains narrow and linear throughout this region (105.8°W to 105.92°W).

The PSZ here is surrounded by a ~ 100 - 500 m wide valley that I interpret to be the zone of fault-related damage, as it exhibits more concentrated deformation features, and is distinct from relatively smooth seafloor. This zone contains faults and folds, including en echelon conjugate and antithetic faults and folds (Figure 10) and is also more rugose than the surrounding seafloor (Figure 11). The interpreted width of damage changes from ~ 100 m wide in the east to ~ 500 m to the west (Figure 11).

4.2 Eastern Central Depression

The morphology of the eastern portion of the central depression is marked by a discrete change in PSZ orientation (Figure 12). At 105.93°W, the average strike of the PSZ changes from ~ 100° to ~ 69° from east to west, creating a 3 km long bend in the PSZ fault trace that is co-located with a ~500 m-deep basin (Figure 12). There are also several long (~ 2 km), sub-parallel anastomosing fault strands that initiate at the start of the bend (Figure 12). Given the left-lateral motion of the fault, this 31° change in PSZ orientation creates a releasing bend that is promoting localized extension, which may be related to the abrupt increase in seafloor depth (Figure 12). Further west from the
location of the releasing bend (from 105.95°W westward), PSZ orientation fluctuates moderately, promoting short regions (~ 700 m) of contractional and releasing bends (Figure 12). At 105.97°W, PSZ orientation changes from ~ 100° to ~ 110° within ~ 500 m and is accompanied by several long (1.5 km) anastomosing fault strands striking obliquely to the continuous PSZ trace. Given left lateral plate motions, this bend is compressional. Further west at 105.98°W, PSZ orientation bends again from ~ 100° to ~ 95° within ~ 500m, promoting a slight extensional bend (Figure 12). There are two sub-parallel PSZ strands between these contractional and releasing bends from 105.97°W to 106.0°W (Figure 12).

Subsidiary faulting in the eastern central depression (Figure 12) extends as far as 1 km north of the PSZ trace and I characterize the broad extent of subsidiary faulting as the extent of fault-related damage. Faulting is concentrated to the slope north of the PSZ and few subsidiary fractures extend from the PSZ trace. The orientation of subsidiary fractures varies with no observable trend in directionality. The seafloor south of the PSZ trace contains few subsidiary faults and rugosity measurements suggest that it is relatively smooth (Figure 12). This interpretation of the damage zone is supported by findings from prior dredge survey locations which sampled the northern slope in this region (Figure 13). These dredge surveys recovered ~ 60% fault breccia ~ 250 m north of the location of the PSZ, suggesting that the rocks present within the steep slopes are deformed via fault related damage (Bahruth, 2023) (Figure 13). Therefore, the extent of fault related damage here extends at least 1 km wide, which varies from the 0.5 km wide margin of damage located just 500 m east (Figure 14). This nearly 2-fold increase
in fault related damage width in this region coincides with the location of the releasing bend (Figure 14).

### 4.3 Western Central Depression

The PSZ trace within the western central depression is difficult to determine due to the high degree of sedimentation and mass wasting deposits that cover large portions of the seafloor (Figure 15). At 106.01°W the surface expression of the PSZ trace is lost and reemerges at 106.07°W (Figure 15). Between these two locations, there are linear, continuous features that could resemble the PSZ located at the bottom of some steeply sloped features, but it is unclear whether these locations are simply the base of the steep slopes or an active fault trace (Figure 15). There is a linear feature present in the southern region of the multibeam survey that could possibly be the location of the PSZ, but is unclear based on limited multibeam coverage.

Where the PSZ trace is lost at the surface (106.01°W), the trace is replaced by a 2 km long, obliquely angled (~ 135°), steeply sloped (~ 100 m of gain) feature that could be a mass wasting deposit given the individual landslides present within the majority of the slope (Figure 15). At the base of this deposit to the southwest is a wide (~ 300 m), sedimented valley bounded on either end by steep hill slopes. This region could be interpreted as a widened PSZ valley, but it is unclear in surface bathymetry if this is a fault-related feature (Figure 15). At 106.03°W, this wide valley bends at ~ 135° around an elevated block that is about ~ 200 m above the surrounding surface; this feature is composed of a parallel series of hills and valleys that extends southwest (Figure 16). From 106.03°W to 106.04°W, the wide valley is covered by a 1 km long mass wasting deposit (Figure 15 and Figure 16). To the northwest of this deposit, a linear feature
strikes at ~ 45° but it is unclear whether this is a part of the PSZ or a subsidiary fracture since it appears to continue outside of the bounds of the multibeam survey (Figure 15 and Figure 16). From 106.04°W to 106.06°W, there is a linear trace that is located at the base of a small valley ~ 100 m wide. At 106.06°W, this linear feature cuts evenly through a mass wasting deposit that continues to propagate to 106.075°W, indicating that this could be an active fault trace. However, at 106.07°W there is a more defined linear PSZ that resembles the morphology of the PSZ in other locations along strike that emerges just south, which could be the true PSZ fault trace (Figure 15 and Figure 16).

There are several indications that this region is a compressive regime. The PSZ east of this highly sedimented region (at ~106.01°W) strikes ~ 98°, while the strike of the PSZ located to the west is ~ 92°, causing a right lateral discontinuity and thus promoting a localized region of compression (Figure 16). There are also several compressive structures, including the ~ 200 m tall block (Figure 16).

The width of fault-related damage is difficult to determine in this region. The concentration of identifiable subsidiary fractures is lower compared to other regions along strike, which could be due to an increase in sedimentation and mass wasting deposits (Figure 15). There is a clear cluster of subsidiary faults at 106.045°W striking at both ~ 45° and ~135° (Figure 15), but otherwise subsidiary fracturing is not apparent at the surface. Despite the lack of fractures apparent at the surface, the width of fault related damage likely exceeds 1 km and could exceed the bounds of the survey. Rugosity measurements in this region are difficult to interpret (Appendix Figure C7); however, the increase in sedimentation, abrupt changes in depth, and compressive fault
structures suggest that the width of fault related damage is sustained throughout this wide compressional zone (Figure 15 and Figure 16).

### 4.4 Western G3

The PSZ within western G3 includes a continuous PSZ trace with a right lateral stepover located at 106.15°W (Figure 17). From 106.07°W to 106.14°W, there is a clear, linear, and concentrated PSZ striking at ~ 103°, following the general direction of plate motion (Figure 17). At 106.09°W there is a long (1 km) anastomosing fault strand that extends from the PSZ trace, striking at an angle of ~ 85° (Figure 17). Riedel shears extend from the PSZ fault trace throughout this region indicating sustained left-lateral strike slip motion (Figure 17). At 106.15°W, there is a 520 m wide, right-stepping bend in the PSZ fault trace where the PSZ trace striking at ~ 103° to the east, alters orientation to an obliquely angled valley striking at ~ 124° and reemerges to the north as a set of short (~ 100 - 300 m), parallel faults striking at ~ 106° (Figure 17). In some locations, these fractures cut through large (some as wide as 240 m), hummocky, volcanic mounds (Figure 17). The right step-over is transpressional given the left lateral plate motions and accompanying ~ 300 m of seafloor uplift (Figure 17).

The location of the compressional step over (at 106.15°W) marks a change in PSZ fault expression and change to the surrounding fault fabric. East of the compressional step over, the PSZ is localized, linear, and contains associated Riedel shears indicative of strike slip faulting surrounded by folded seafloor fabric that extends north (Figure 17). West of the compressional step over, subsidiary faults transition to normal faults oriented similarly to the orientation of plate motions (~ 103°W) (Figure 17). Additionally, the initiation of the step-over marks the transition from folded seafloor fabric to
hummocky volcanic mounds (Figure 17). These volcanic mounds increase in size laterally from east to west, and some are cone shaped (Figure 17). The location of these volcanic mounds along strike is interesting given the distance from the EPR ridge transform intersection (~ 15 km west) (Figure 17).

The width and expression of fault related damage fluctuates in this region (Figure 17). From 106.07°W to 106.09°W, damage is sustained to ~ 500 m between the PSZ trace and anastomosing fault strand that extends from the main trace (Figure 17). From 106.09°W to 106.14°W, damage is sustained to ~ 100 m wide and includes the PSZ and associated subsidiary faulting including Riedel shears that extend from the PSZ trace (Figure 17). There is also a substantial gap in multibeam bathymetry coverage in this location, which masks the true extent of Riedel fractures to the south of the PSZ, and the seafloor fabric located to the north of the PSZ. From 106.14°W to 106.16°W located at the compressional step-over, damage extends to ~ 1 km wide and includes an increase in the number of brittle fractures located at the surface (Figure 17).

4.5 EPR ridge transform intersection

The location of the PSZ near the ridge transform intersection between western G3 and the EPR is less localized, and is replaced by short, normal faults (Figure 18). There is a long (4 km), steep (~ 250 m of relief) slope striking at about the same angle to the PSZ and dipping to the south (Figure 19), suggesting that this could be a normal fault.

The location of the long fault slope marks a distinct change in seafloor morphology from north to south (Figure 18). North of the exposure is dominated by a dense quantity of short, parallel faults that cut through hummocky volcanic mounds (Figure 18 and Figure 19). South of the exposure contains little faulting and instead
contains an obliquely angled (45° to the PSZ trace) line of volcanic mounds which are unaltered by faulting (Figure 19). From 106.22°W to 106.23°W, there are a dense quantity of faults that are located south of the exposure and strike ~ 89°, oblique to the short faults located to the north of the exposure (Figure 19).

The width of fault related damage corresponds with the limits of multibeam survey within this region, ~ 2 km wide, except from 106.19°W to 106.22°W, where there are apparently undeformed volcanic landforms located to the south of the steep slope, limiting off-fault damage to ~ 1 km wide (Figure 19). The quantity, density, and orientation of off-fault damage in this region (Figure 19) differs from other regions along strike.

4.6 Summary results of seafloor morphology

Processed 1 m resolution bathymetry of Gofar’s western-most segment G3 reveals a complex fault zone boundary that contains discrete changes in seafloor depth, fault fabric, PSZ orientation, and width of fault related damage. Changes from a distinct, linear PSZ to a change in the orientation or localization of the PSZ occurs within distinct regions along strike (Figure 20). Most notable is the central depression, which is distinct from the relatively linear and concentrated PSZ present to both the west and east (Figure 12Figure 15). Tracing the location of the PSZ at the surface reveals that G3 likely accommodates localized compressional and extensional motions which are collocated with abrupt changes in seafloor depth, seafloor morphology, and the width of fault related damage. The width and surface morphology of fault related damage changes sharply and drastically, in some regions doubling in width in the span of hundreds of meters (e.g., Figure 14).
5 Discussion

The goal of this study is to generate a high-resolution bathymetric map covering Gofar's western-most G3 segment and interpret prominent fault features (including fault geometry and damage). I consider if these prominent fault features change with respect to slip to better understand potential changes in physical fault zone properties that could influence rupture segmentation along strike. These data image the seafloor within distinct fault segments with different earthquake behavior, and will be used to answer the following research questions:

(1) What is the fault structure and how does it change along strike?
(2) Does fault structure vary with along-strike changes in slip behavior?
(3) Can surface morphology contribute to our current understanding of controls on slip behavior?

In the results section, I outlined how major geomorphic structures change along strike (addressing question 1). Here, I will put those geomorphic observations in the context of the distinct rupture segments observed at Gofar (Figure 21), in order to address research questions (2) and (3). These research questions were motivated by past work at Gofar, where seismic observations indicate transitions in fault slip behavior (i.e., rupture zones to barrier zones) may occur (1) at locations where there are slight changes in fault strike (Froment et al., 2014; Gong & Fan, 2022) and (2) between fault zones with narrower or wider low velocity zones (Liu et al., 2023; Roland et al., 2012; Shi et al., 2021). The second of these observations supports the hypothesis that an increase in hydrothermal circulation within a wide damage zone in the barrier may allow for elevated pore fluid pressures there to facilitate slow- or aseismic slip behavior (Liu et
In this section, I use interpretations from the ultra-high-resolution seafloor morphology to evaluate this hypothesis and associated consequences for along-strike changes in slip behavior.

The Gofar G3 study area contains three distinct zones of rupture behavior (Boettcher & Jordan, 2004; Gong & Fan, 2022; McGuire et al., 2012), including the 2008 and 2007 M6 rupture zones, the barrier, and the 2008 December swarm zone (Figure 3d). The 2008 and 2007 M6 rupture zones are regions marked by frequent (~ 5 – 6 yr repeat time) large magnitude earthquakes (~ M5 - 6) (Figure 1) (Shi et al., 2021). In between the 2007 and 2008 M6 rupture zones is the 10 km long barrier, where abundant, deep (7 - 8 km) microseismicity propagated for several weeks leading up, and terminated at the time of the 2008 M6 rupture (Figure 3) (Gong & Fan, 2022; McGuire et al., 2012). The December swarm region, which connects G3 to the EPR, experienced a swarm of ~ 20,000 earthquakes that followed the 2008 M6 mainshock; it also hosted reoccurring periodic swarms prior to the M6 mainshock, referred to as pulsating events (Figure 3) (Gong & Fan, 2022; McGuire et al., 2012). These distinct segments with distinct earthquake behavior will frame my interpretations for considering how structural changes influence slip at Gofar.

The two major structural trends I identify include: (1) the location, orientation, and degree of localization of the PSZ and subsidiary fractures, and (2) the width of fault related damage. Changes in PSZ and subsidiary fault orientations are important for determining fault zone kinematics and hint at stresses that control fault zone architecture at depth (Fossen, 2010). Additionally, faults are structural manifestations of the frictional conditions of a given area (Anderson, 1951; Byerlee, 1978). Frictional
conditions are in turn influenced by physical fault zone properties such as rock type (Anderson, 1951; Byerlee, 1993; Fossen, 2010; Scholz, 1998; Sibson, 1986), porosity (Byerlee, 1993; Flemings, 2021; Healy, 2008), temperature (Behn et al., 2007; Roland et al., 2010), and dilatancy (Escartín et al., 2001). The extent of damage at the surface can be related to the porosity and permeability structures at depth, which in turn influence physical fault zone properties (Healy, 2008; Sibson, 2003). Physical fault zone properties influence the mechanics of slip: for example, elevated pore fluid pressures prime the fault zone for slow slip (Roland et al., 2012). Thus, it is important to map faults (PSZ and subsidiary) and damage width because these changes ultimately point to variations in physical fault zone properties, which in turn influence the mechanics of slip.

I find that at Gofar, regions that sustain repeated, high magnitude rupture (including the 2007 eastern rupture and 2008 western rupture zones) are expressed as a relatively linear PSZs striking in a similar direction to global plate motions contained within a relatively thin (≤ ~ 500 m) margin of damage (Figure 20a and Figure 22). In contrast, regions that contain deep microseismicity and no high magnitude rupture (including the barrier and December swarm regions) are geomorphically complex with changes to PSZ orientation (including contractional and extensional bends and step-overs) within a relatively wide (≥ 500 m) region of damage (Figure 20b and Figure 22). Additionally, most of these geomorphic structures change abruptly along strike (in some regions within 100s of meters). Collectively, these observations imply that geometric bends in fault strike, particularly the extensional bend within the barrier and compressional step-over in the December swarm region, lead to a widened zone of normal faulting and distributed fractures which could facilitate enhanced fluid flow, allowing more water to
extend through the seismogenic zone, subsequently altering the material properties that contribute to slow slip. This is different from what is observed in the rupture zones, which have relatively simpler geometries and no slow slip.

Along-strike segmentation in slip behavior does not perfectly co-located with geomorphic changes to PSZ orientation and damage width. The 2008 rupture zone contains a complicated, compressional push-up structure, obliquely oriented landslides, and loss of PSZ in the east that appears geomorphically different from the western region which contains a linear PSZ and thin margin of damage (Figure 15 and Figure 22). This could be because the width of fault-related damage and location of the PSZ at the surface does not represent the true morphology, location, and extent of damage at depth. There also could be overprinting of geomorphic structures in this area that may be influencing the visual extent of fracturing and fault damage at the surface. With the exception of the eastern-most extent of the 2008 rupture zone, however, the clearest correlation between fault structure and slip behavior is the degree of heterogeneity of fault zone geomorphology along-strike. The barrier and December swarm regions contain heterogenous fault zone morphology, while the 2008 and 2007 rupture zones contain simple fault morphology (Figure 22).

5.1 Linking geomorphic structure to physical fault zone properties

The width of the damage zone could play an important role in influencing fault-zone properties at Gofar by changing the degree of hydration at depth. As mentioned in the background, the increase in fracture density within damage zones can allow for increased fluid filled pathways, which can decrease the overall strength of the host rock and facilitate microseismicity (Froment et al., 2014; Roland et al., 2012; Shi et al.,
Increased fluid filled pathways can facilitate hydrothermal circulation, and through lithospheric cooling, increase the depth extent of brittle failure, which is potentially why we observe microseismicity past the thermally-predicted limit for brittle failure in the barrier and December swarm regions at Gofar (Figure 3 and 4b) (Roland et al., 2010). Increased porosity via fault related damage could also explain the low velocity zone resolved through the barrier from the active source tomographic profile (Figure 4b) (Roland et al., 2012). Image changes in fault-induced damage at the same location where the orientation of the PSZ changes at the surface, including locations where the PSZ is oblique to plate motions (Figure 22); it is possible extensional stresses here have further contributed to the extent of deformation.

If hydration is altering the mechanical properties of the upper crust down to the mantle transition, why was there no direct hydrothermal vent signal over the barrier, or anywhere on G3, from AUV Sentry? AUV Sentry was equipped with an oxygen redux potential sensor, which can identify hydrothermal plumes in the water column (German & Seyfried, 2014). The 2008 December swarm region is the only region mapped during TN399 where there was an anomalous temperature and ORP spike from AUV Sentry water column data (Appendix Figure D4) which could indicate an increase in localized hydrothermal flow (German & Seyfried, 2014). Seafloor bathymetry and photo transect data also lack strong evidence for hydrothermal vents or associated plumes, further indicating that hydration is not expressed through hydrothermal vent systems apparent at the surface (Tracy, 2023). Instead, bathymetric data at G3 supports the idea that efficient, potentially diffuse hydrothermal circulation could be promoted by an increase in fault damage.
5.2 Stress heterogeneity and fault-zone kinematics

Changes in the geometry of the PSZ and subsidiary fractures occur where there are changes in slip: areas that host deep, abundant microseismicity, including the barrier and December swarm region, are aligned with releasing bends and normal subsidiary faults, while areas that sustain large ruptures are aligned with a linear PSZ (including the 2007 rupture zone and western portion of the 2008 rupture zone) and compressional structures (eastern portion of the 2008 rupture zone) (Figure 22). Collectively, these observations imply that releasing bends and associated extensional structures could promote microseismicity and slow slip. Perhaps distributed cracking through localized extension facilitates enhanced permeability and fluid flow, leading to associated material variations that promote slow slip (as discussed earlier in section 5.1). Although difficult to determine conclusively from surface expression alone, stress perturbations apparent at the surface seem to support a straight or slightly restraining orientation of the PSZ and prevalence of compressional subsidiary faults where large ruptures occur, and extensional structures where earthquake swarms and barrier zones occur (Figure 22).

5.3 Potential Slow Slip Mechanisms

Lateral variations in the width of damage, such as highly-damaged regions next to an intact fault zone, and possibly changes in associated normal or effective normal stress (linked to pore fluid pressures) likely contribute to the distribution of slip mechanisms at Gofar. Damage zones reduce the shear rigidity of the surrounding rock, which can further alter elastic strain distribution during interseismic events (Shelef & Oskin, 2010) and can allow for a wide distribution of microseismic activity. Distributed shear through
broad zones of damage on strike slip faults can account for up to 60% of the total displacement across the fault zone (Healy, 2008), therefore, perhaps the 80% earthquake deficit at Gofar and other OTF boundaries (Boettcher & Jordan, 2004) is controlled by lateral variations in the width of damage and degree of distributed displacement. A prime example of this at Gofar is within the eastern central depression boundary, where the width of damage nearly doubles to at least 1 km along strike in the span of ~ 500 m (Figure 14). This drastic change in damage width has implications for the rheology of the surrounding rock, where highly-fractured material contains more fluid pathways for water to migrate at depth, further changing the frictional and material properties of the rock, as mentioned in section 5.1.

Work on subduction zone lithologies indicate that the dominant seismic style is controlled by fault strength (weak versus competent fault lithologies) and material heterogeneity (Fagereng & Sibson, 2010). This observation could potentially explain changes in along-strike slip distribution at Gofar. Laboratory work on the frictional characteristics on serpentine indicate that serpentine can act in a velocity strengthening manner and could promote the propagation of unstable slip under some conditions (Moore & Rymer, 2007; Reinen et al., 1991). Widespread serpentinization could result in similar rupture behavior to the barrier region of Gofar, where there is an abundance of microseismicity and no high magnitude rupture (Liu et al., 2020).
6 Conclusions

Findings from this study support the following conclusions related to possible mechanisms that control slip:

(1) Gofar’s G3 segment is a geomorphically complex fault zone boundary, including along-strike variations in fault orientations, depth, damage width, and volcanics.

(2) Major geomorphic structures include a continuous PSZ, subsidiary fractures and pervasive fault-related damage.

(3) The orientation of the PSZ varies indicating that the fault boundary likely undergoes varying degrees of transpression and transtension.

(4) Damage is pervasive throughout the fault boundary, and the width of damage varies abruptly along strike.

(5) Regions that sustain repeated, large ruptures (including the 2007 eastern rupture and 2008 western rupture zones) are expressed as a relatively linear PSZ striking in a similar direction to global plate motions contained within a relatively thin (≤ ~ 500 m) margin of damage.

(6) Regions that contain deep microseismicity and no high magnitude rupture (including the barrier and December swarm regions) are geomorphically complex with changes to PSZ orientation (including contractional and extensional bends and step-overs) within a relatively wide (≥ ~500 m, up to ~1 km) region of damage.

(7) Variations in the style of slip at Gofar is likely controlled by lateral changes in fault zone material properties and mechanical behavior, specifically highly
damaged fault zones in regions that limit large magnitude earthquakes next to an intact fault zone that sustains rupture.

Surface observations from this study indicate that fault structures may be associated with a change in material or structural fault properties at depth within the seismogenic zone that contributes to rheology and fault slip behavior. Slow slip at Gofar is likely controlled by a heterogenous distribution of fault zone material, influenced by variations in the width of fault-related deformation (i.e. width of the damage zone) and associated changes in PSZ orientations. Collectively these structural features, many of which are visible at the surface, alter fault zone properties and control the distribution of large earthquake ruptures at Gofar.
References


Large magnitude earthquakes over time along the Quebrada-Discovery-Gofar (QDG) transform segments within the equatorial East Pacific Rise. a) $\geq$ M 5.3 earthquakes recorded teleseismically since 1990 as a function of distance along strike. Different colors denote the discrete segments where high magnitude earthquakes occur. The red box outlines the location of Gofar G3, the study area for this work. b) Map view of high magnitude earthquakes located along QDG. Note the regularity of earthquakes both along strike and over time. Updated from Wolfson-Schwehr et al. (2014) (courtesy of Dr. Margaret Boettcher).

Figure 1. QDG seismicity over time.
Figure 2. Gofar OTF Study Site Location

a) regional map showing the transform offset along the EPR. b) Bathymetric map of the QDG OTF systems. White lines denote the boundaries of the Pacific and Nazca plates as determined by side-scan backscatter (Pickle et al., 2009). c) Zoomed in bathymetric map of the Gofar OTF three transform segments. The yellow rectangle marks the transform segment focused on in this study. The G3 central depression, as mentioned in the text, is labeled. Note the complex topographic variations present along G3.
Along strike variations in rupture behavior along Gofar’s G3 transform segment using relocated earthquake data from the 2008 M 6 catalog (Gong and Fan, 2022). Note that since the seismic data is from the 2008 seismic array, microseismicity close to the ITSC might not be recorded. (a) Map view of seismicity distributions. Zones are colored based on the distinct rupture segments that display different slip behaviors. (b) Down dip extent of earthquake seismicity. Note that the brittle-ductile transition predicted by thermal models is ~ 7km depth. Black dotted lines denote 95% of earthquake depths. (c) Spatiotemporal evolution of seismicity. Note the shutdown of microseismicity in Zone 2 (orange) at the time of the mainshock. (d) Seismicity patches defined by Gong and Fan, 2022 and patches defined by this study. Zones 3 and 4 from Gong and Fan, 2022 are defined as the entire 2008 M6 rupture zone in this study.
Figure 4. Descriptions of damage zone.

a) Schematic representation of brittle fault zones in adjoining members of the crust, including the principal slip zone (PSZ), fault core, and damage zone, where the coarse dots represent conduits for hydrothermal alteration (Copied from Sibson, 2003). b) Preferred velocity model for Gofar’s G3 barrier zone (right) (from Roland et al., 2012). The figure on the left shows the location of the wide angle refraction lines (yellow line) and OBS (red dots) over 75 m shipboard resolution bathymetry. The figure on the right is the velocity model for G3’s barrier region. Note the extent of the low velocity anomaly through the G3 barrier region, potentially due to hydrothermal alteration. The extent of hydrothermal alteration at depth at Gofar is still relatively unknown.
Figure 5. Comparison of shipboard and AUV Sentry bathymetry resolution
Comparison of shipboard (75m and 25m resolution) and AUV (1m resolution) bathymetry, all imaging the same region.
**Figure 6. G3 AUV Sentry dive locations**

All 10 Sentry AUV dive tracks (numbered 614 – 627) along the G3 transform segment. Note that there were 4 dives conducted on the G1 transform segment (north) and are not included in this plot but can be referenced in Appendix Figure B1.
Workflow document used in this study to process multibeam data acquired by AUV Sentry. Processing is segmented into four steps: initial shipboard processing, post shipboard processing, feature based individual dive processing, and feature based merged dive processing.

**Figure 7. AUV Sentry workflow document**

Workflow document used in this study to process multibeam data acquired by AUV Sentry. Processing is segmented into four steps: initial shipboard processing, post shipboard processing, feature based individual dive processing, and feature based merged dive processing.
**Figure 8. Final merged navigation solution**

Final merged navigation solution for the 1 m resolution bathymetry acquired from the 10 AUV Sentry dives plotted over 25 m resolution shipboard bathymetry. b) 2008 M 6 relocated earthquake catalog plotted over the 1 m resolution bathymetry, including foreshocks (yellow), aftershocks (red), and swarm (blue) (Gong et al., 2021). Refer to Appendix Section C for supplemental, zoomed in 10 km long segments of the 1 m resolution dataset. Grids of the bathymetric dataset are available in Supporting Data.
Figure 9. Five geomorphic segments evaluated in Methods

Five geomorphic segments (red outlined rectangle) evaluated in the methods section, starting from the eastern most box westward. Geomorphic segments were determined from distinct changes in PSZ orientation and/or damage widths.
Figure 10. Eastern G3 Bathymetry
Zoomed in, 10 km segment of 1 m resolution bathymetry including a) the location along strike, b) the 1 m bathymetry map, c) 1 m bathymetry map with the relocated 2008 M 6 earthquake catalog plotted on top to show relative earthquake distributions (Gong et al., 2022), and d) 1 m bathymetry map with manual interpretations including subsidiary fractures (grey) and the PSZ (black lines). The locations of the distinct rupture zones from the 2008 catalog are plotted at the bottom for reference.
Rugosity attribute of eastern G3, where red value indicate more rugose terrain. The interpreted margin of damage is outlined in blue and the PSZ annotations are in black.

Figure 11. Eastern G3 damage zone rugosity

Rugosity attribute of eastern G3, where red value indicate more rugose terrain. The interpreted margin of damage is outlined in blue and the PSZ annotations are in black.
Figure 12. Eastern central depression bathymetry
Zoomed in, 10 km segment of 1 m resolution bathymetry including a) the location along strike, b) the 1 m bathymetry map, c) 1 m bathymetry map with the relocated 2008 M 6 earthquake catalog plotted on top to show relative earthquake distributions (Gong et al., 2022), and d) 1 m bathymetry map with manual interpretations including subsidiary fractures (blue) and the PSZ (black lines). The locations of the distinct rupture zones from the 2008 catalog are plotted at the bottom for reference.
The location and direction of the two dredge lines along the northeastern slope of the central depression. D02 recovered ~ 60% brecciated rock while D03 recovered ~ 10% brecciated rock despite being located ~ 250 m of the main PSZ trace at the surface, indicating that fault related damage is likely sustained outside of the location of the PSZ here (Bahruth, 2023).

**Figure 13. Eastern central depression dredge survey lines.**

The location and direction of the two dredge lines along the northeastern slope of the central depression. D02 recovered ~ 60% brecciated rock while D03 recovered ~ 10% brecciated rock despite being located ~ 250 m of the main PSZ trace at the surface, indicating that fault related damage is likely sustained outside of the location of the PSZ here (Bahruth, 2023).
Figure 14. Eastern central depression changes in damage margin width

Eastern Damage zone at the eastern edge of the central depression. The top figure indicates the region along strike with an extensional bend. The middle plot is 1 m resolution bathymetry of this region, including annotations for the damage margin width and PSZ location. Note the drastic change in damage margin width from East to West. The bottom plot is the gridded rugosity attribute from the 1 m bathymetry, where red and yellow indicate a more rugged surface.
Figure 15. Western central depression bathymetry

Zoomed in, 10 km segment of 1 m resolution bathymetry including a) the location along strike, b) the 1 m bathymetry map, c) 1 m bathymetry map with the relocated 2008 M 6 earthquake catalog plotted on top to show relative earthquake distributions (Gong et al., 2022), and d) 1 m bathymetry map with manual interpretations including subsidiary fractures (blue) and the PSZ (black lines). The locations of the distinct rupture zones from the 2008 catalog are plotted at the bottom for reference.
Figure 16. Western central depression slope and rugosity attributes.

a) 1m bathymetry of Western section of the central depression where the PSZ trace is lost, including b) slope and c) aspect attribute grids to better visualize the compressional surface features. Note the change in orientation of the structures, from a linear PSZ to an obliquely angled hill and valley sequence.
Figure 17. Western G3 bathymetry
Zoomed in, 10 km segment of 1 m resolution bathymetry including a) the location along strike, b) the 1 m bathymetry map, c) 1 m bathymetry map with the relocated 2008 M 6 earthquake catalog plotted on top to show relative earthquake distributions (Gong et al., 2022), and d) 1 m bathymetry map with manual interpretations including subsidiary fractures (blue) and the PSZ (black lines). The locations of the distinct rupture zones from the 2008 catalog are plotted at the bottom for reference.
Figure 18. Ridge transform intersection bathymetry
Zoomed in, 10 km segment of 1 m resolution bathymetry including a) the location along strike, b) the 1 m bathymetry map, c) 1 m bathymetry map with the relocated 2008 M 6 earthquake catalog plotted on top to show relative earthquake distributions (Gong et al., 2022), and d) 1 m bathymetry map with manual interpretations including subsidiary fractures (blue) and the PSZ (black lines). The locations of the distinct rupture zones from the 2008 catalog are plotted at the bottom for reference.
Figure 19. Interpretations of western most G3

1 m resolution bathymetry of the Western G3 transform near the ridge transform intersection with the EPR. Notice the change in strike and overall expression of the PSZ surrounding the compressional step over. This location marks a stark change from primarily strike slip regime to normal stress regime. a) Zoom in of 1m resolution bathymetry of the steep, normal fault that separates brittle faulted fabric to the normal and a line of volcanic mounds to the south, including a cross section to illustrate the depth gradient. b) Zoom in of the 1 m resolution bathymetry of the step-over, which includes faults that cut through volcanic mounds at oblique angles to the overall orientation of the PSZ.
Figure 20. Examples of PSZ expression.

Location of the PSZ along strike including examples of a) uncertain PDZ trace within the central depression and b) certain PDZ, where the trace is easily identifiable at the surface.
Figure 21. Rupture zones in reference to 1 m bathymetry

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Figure 22. Summary of interpretations

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c) Moving average of the orientation of the PSZ minus the G3 plate motion average (101.4°) determined by MORVEL plate motions (DeMets et al., 2010) to illustrate along strike areas of compression (positive value; shaded red) and extension (negative values; shaded blue). The black line denotes the plate motion average (y=0; angle=101.4°). PSZ orientations were sampled using a moving average every 1 km and each sample segment overlapped by 0.5 km.

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Appendix A: Dive track and final 1 m resolution bathymetry for individual AUV Sentry dives

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a) 1m resolution multibeam bathymetry of the G3 OTF and EPR spreading center boundary. There is an abrupt (~3 km) change in morphology where the primarily strike slip regime through volcanics to the East changes to normal faulting to the West near the EPR. The orientation of volcanics is ~45° to the orientation of slip. b) The location of 2008 M6 foreshocks (red circles), aftershocks (yellow circles), and swarm event (blue circles). Normal faulting to the North is dominated by foreshock and aftershock activity while the ~45° volcanics to the south are dominated by the pulsating swarm events. The green dot denotes the location of the temperature spike and ORP anomaly near 18:34.

Figure D4. Sentry dive 627 ORP anomaly
Supporting Data

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