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# Seismic Velocity Structure Along and Across the Ultraslow-Spreading Southwest Indian Ridge at 64°30'E Showcases Flipping Detachment Faults

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1 **Seismic Velocity Structure Along and Across the Ultraslow-Spreading Southwest**  
2 **Indian Ridge at 64°30'E Showcases Flipping Detachment Faults**

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15  
16 **Key Points:**

- 17     • First detailed 2D regional P-wave velocity models across and along an ultraslow-  
18     spreading Southwest Indian Ridge amagmatic segment
- 19     • A system of flipping detachments is imaged in the subsurface for the first time  
20     constrained by the velocity structure
- 21     • Lithosphere gradually transitions from highly fractured and fully serpentinized peridotites  
22     at the top to unaltered peridotites at depth

## 23 Abstract

24 We present two ~150-km-long orthogonal 2-D P-wave tomographic velocity models across and  
25 along the ridge axis of the ultraslow-spreading Southwest Indian Ridge at 64°30'E. Here,  
26 detachment faults largely accommodate seafloor accretion by mantle exhumation. The velocity  
27 models are constructed by inverting first arrival traveltimes recorded by 32 ocean bottom  
28 seismometers placed on the two profiles. The velocities increase rapidly with depth, from 3–3.5  
29 km/s at the seafloor to 7 km/s at depths ranging from 1.5–6 km below the seafloor. The vertical  
30 gradient decreases for velocities >7 km/s. We suggest that changes in velocity with depth are  
31 related to changes in the degree of serpentinization and interpret the lithosphere to be composed  
32 of highly fractured and fully serpentinized peridotites at the top with a gradual downward  
33 decrease in serpentinization and pore space to unaltered peridotites. One active and five  
34 abandoned detachment faults are identified on the ridge-perpendicular profile. The active axial  
35 detachment fault (*DI*) shows the sharpest lateral change (horizontal gradient of  $\sim 1 \text{ s}^{-1}$ ) and  
36 highest vertical gradient ( $\sim 2 \text{ s}^{-1}$ ) in the velocities. In the western section of the ridge-parallel  
37 profile, the lithosphere transitions from non-volcanic to volcanic over a distance of  $\sim 10$  km. The  
38 depth extent of serpentinization on the ridge-perpendicular profile ranges from  $\sim 2$ -5 km, with the  
39 deepest penetration at the *DI* hanging wall. On the ridge-parallel profile, this depth ( $\sim 2.5$ -4 km)  
40 varies less as the profile crosses the *DI* hanging wall at  $\sim 5$ -9 km south of the ridge axis.

## 41 Plain Language Summary

42 We investigate the Southwest Indian Ridge lithosphere at 64°30'E, where the Somalian and  
43 Antarctic plates move slowly away from each other at less than 14 mm/year. This is one of a  
44 limited number of places on Earth where mantle is currently being exhumed to the seafloor. We  
45 use seismic sensors, placed across and along the ridge axis, to analyze how seismic waves travel  
46 from the energy sources, through the lithosphere, to these sensors. Our results, in the form of  
47 two-dimensional velocity models, show that the rock velocities increase rapidly with depth.  
48 Lateral and vertical velocity changes delimit a system of detachment faults on the ridge-  
49 perpendicular profile, responsible for bringing mantle-derived rocks, peridotites, up to the  
50 seafloor. Based on the modeled velocities and velocity changes, and previous extensive seafloor  
51 sampling, we suggest that  $\sim 75\%$  of the lithosphere in the study area is composed of highly  
52 fractured and fully hydrothermally altered peridotites at the top with a gradual downward  
53 decrease in alteration and pore space to unaltered peridotites at depth. We also locate the  
54 transition to lithosphere with a magmatic component in the western section of the ridge-parallel  
55 profile.

## 56 1 Introduction

57 The global mid-ocean ridge system consists of spreading centres that greatly differ from  
58 each other in their spreading rate, spreading obliquity, melt supply and modes of seafloor  
59 accretion. Oceanic ridges are normally classified by their full-spreading rate as fast- ( $\sim 80$ –180  
60 mm/yr), intermediate- ( $\sim 55$ –70 mm/yr), slow- (20-55mm/yr), and ultraslow-spreading ( $< 20$   
61 mm/yr) (Dick et al., 2003). The scientific community has invested significant effort into studying  
62 the oceanic crust formed at fast- (e.g. Aghaei et al., 2014; Canales et al., 2003; Detrick et al.,  
63 1993; Dunn & Toomey, 2001; Grevemeyer et al., 1998; Han et al., 2014; Vera & Diebold, 1994),  
64 intermediate- (e.g. Canales et al., 2005; Carbotte et al., 2006, 2008; Nedimović et al., 2005,  
65 2008; Newman et al., 2011; Weekly et al., 2014) and slow-spreading rates (e.g. Arnulf et al.,

66 2012; Barclay et al., 1998; Christeson et al., 2020; Dannowski et al., 2010; Escartín & Canales,  
67 2011; Estep et al., 2019; Kardell et al., 2019; Seher et al., 2010, Xu et al., 2020). Less effort has  
68 been directed toward studying the lithosphere formed at ultraslow-spreading ridges, which thus  
69 remains relatively less well understood. Yet, the ultraslow-spreading centers constitute about  
70 35% of the global ridge system (Dick et al., 2003). The comparatively smaller effort put toward  
71 understanding the lithosphere at ultra-slow ridges is partially a consequence of their remoteness  
72 and inaccessibility, as these ridges are located in the Arctic (Gakkell Ridge, Knipovich Ridge,  
73 Mohns Ridge, and Kolbeinsey Ridge) and Indian Oceans (Southwest Indian Ridge - SWIR)  
74 (Argus et al., 2011; Bird, 2003; Kreemer et al., 2014; Müller et al., 2008).

75 Early controlled-source seismic surveys (e.g. Ewing & Ewing, 1959; Houtz & Ewing,  
76 1976; Peterson et al., 1974; Raitt, 1963; Talwani et al., 1965, 1971), in situ rock investigations  
77 and laboratory velocity measurements on rock samples (e.g. Carlson & Miller, 1997, 2003;  
78 Christensen, 1972; Christensen & Salisbury, 1975; Miller & Christensen, 1997), and ophiolite  
79 studies (e.g. Christensen, 1978; Christensen & Smewing, 1981) suggest an oceanic crustal  
80 structure composed of three seismically identifiable layers that often correspond, from top to  
81 bottom, to an extrusive basaltic lava layer (Layer 2A), a sheeted diabase dike layer (Layer 2B),  
82 and a Layer 3 comprised of isotropic gabbros at the top and layered gabbros at the bottom  
83 topping the uppermost mantle. However, the seismic boundaries do not necessarily correspond to  
84 lithological boundaries. For instance, the layer 2A/2B boundary, which is thought to be an  
85 alteration front or a major change in porosity, may or may not correspond to a change in  
86 lithology (e.g. Berge et al., 1992; Carbotte and Scheirer, 2004; Christeson et al., 2007; Wilcock  
87 et al., 1992). Igneous mafic layers 2A, 2B and 3 are often distinguished according to their  
88 commonly associated velocity and thickness ranges, as well as their vertical velocity gradients  
89 (Christeson et al., 2019; Grevemeyer et al., 2018a; White et al., 1992). However, identified  
90 outcrops of exposed mantle-derived peridotites on the seafloor at slow- and ultraslow-spreading  
91 ridges (e.g. Blackman et al., 2002; Cannat et al., 2006; Dick et al., 2008; Ildefonse et al., 2007)  
92 have challenged the 3-layer paradigm at slower spreading rates. Moreover, sampling of the  
93 seafloor at the SWIR has nearly exclusively (~90%) recovered serpentinized peridotites (Sauter  
94 et al., 2013) suggesting the total absence of continuous igneous mafic layers at the sampled  
95 locations.

96 Exhumed mantle seafloor exposures are generally associated with detachment faults (e.g.  
97 Canales et al., 2004; Cannat, 1993; Sauter et al., 2013; Tucholke & Lin, 1994 ) and Oceanic Core  
98 Complexes (OCCs) at slow- and ultraslow-spreading ridges (e.g. Dick et al., 2019; Escartín et  
99 al., 2003; Sauter et al., 2013; Smith et al., 2006; Zhao et al., 2013; Zhou & Dick, 2013).  
100 Widespread detachment faulting indeed plays a central role in oceanic lithosphere accretion and  
101 plate divergence accommodation at melt-poor ridge sections of slow- and ultraslow-spreading  
102 ridges (Cann et al., 1997; Cannat et al., 2006; Escartín et al., 2008; Sauter et al., 2013; Smith et  
103 al., 2006), particularly at segment ends as magma is focused toward the segment centers (Lin et  
104 al., 1990) and thus is scarce at the ends. At the slow-spreading Mid-Atlantic Ridge (MAR), the  
105 mode of asymmetrical accretion along detachment faults can last for 1–3 Myr (Tucholke et al.,  
106 1998) while at the SWIR continuous exhumation of mantle-derived rocks has been occurring  
107 during the last ~11 Myr in a flip-flop detachment faulting mode (Sauter et al., 2013). Steep long-  
108 offset normal faults rotate and flatten as footwall flexural bending occurs exposing the ultramafic  
109 mantle peridotites on the detachment surface (deMartin et al., 2007; Dick et al., 2010; Escartín et  
110 al., 2003; Ildefonse et al., 2007; Smith et al., 2006; Tucholke et al., 1998). These long-offset

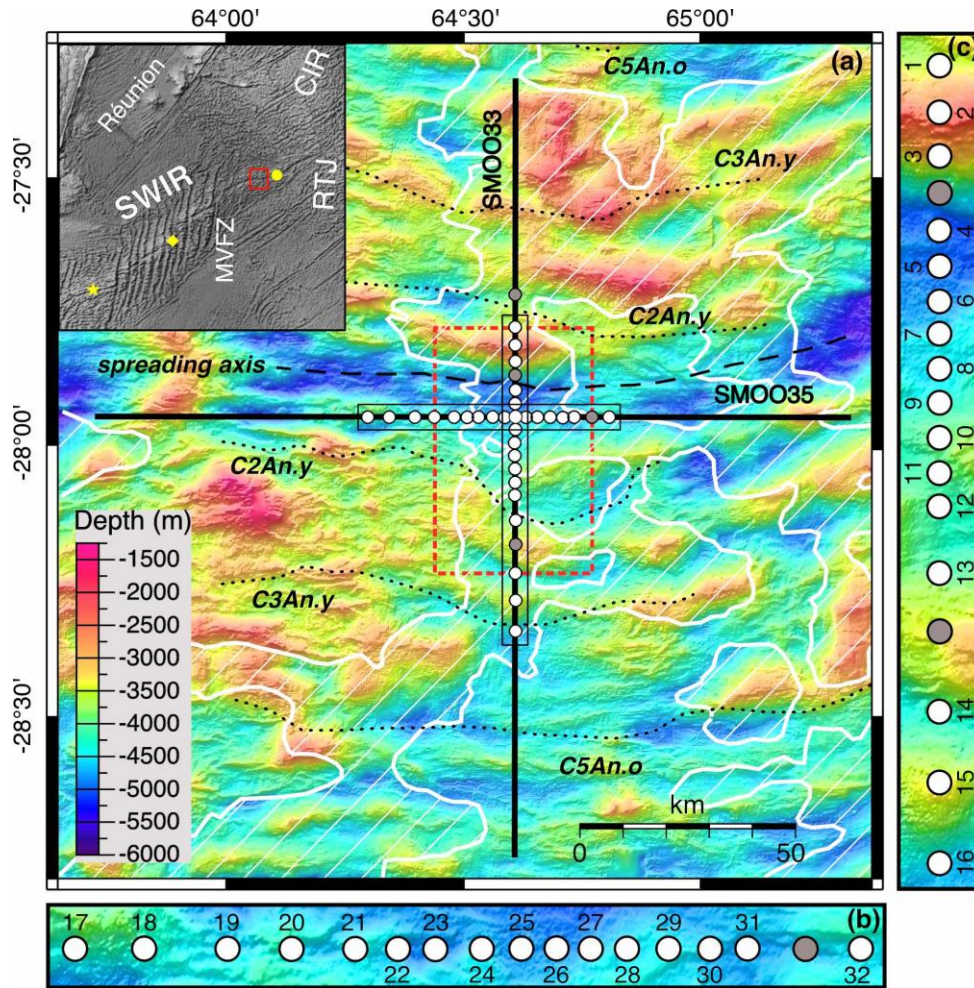
111 normal faults root on a steeply dipping ( $\sim 70^\circ$ ) interface (deMartin et al., 2007; Parnell-Turner et  
112 al., 2017) at depths up to 20 km (Bickert et al., 2020; Schlindwein & Schmid, 2016).

113 The exhumed peridotites become hydrothermally altered when in contact with seawater.  
114 Peridotite-seawater interactions release large amounts of methane and hydrogen to the water  
115 column (Alt et al., 2009; Mével, 2003) making serpentinization relevant to the emergence and  
116 thriving of unique microbial communities (e.g. Früh-Green et al., 2018; Kelley et al., 2005;  
117 Schrenk et al., 2004). Serpentinization also plays a crucial role in the detachment faulting and  
118 favors the development of large-offset low-angle detachment faults (Ildefonse et al., 2007;  
119 Lavier et al., 1999; Tucholke et al., 1998). New steep normal faults may initiate and become the  
120 new master detachment faults with the same or reverse polarity (e.g. Reston & McDermott,  
121 2011; Sauter et al., 2013). Successive detachment faults that change polarity develop a flip-flop  
122 fault mode (Bickert et al., 2020; Reston, 2018; Sauter et al., 2013) and reveal exhumed mantle  
123 domains in this process on both sides of the spreading axis (Cannat et al., 2019; Reston, 2018;  
124 Sauter et al., 2013). What causes the abandonment of a fault and the initiation of a new master  
125 fault with opposite polarity is still under debate.

126 Coincident multichannel seismic (MCS) reflection and wide-angle ocean bottom  
127 seismometer (OBS) refraction surveys at ultraslow-spreading ridges are critical to understanding  
128 the crustal and uppermost mantle structure, the faulting dynamics associated with divergence,  
129 and the mechanisms of seafloor accretion. Seismic surveys at ultraslow-spreading ridges beyond  
130 the SWIR include work done at the Knipovich (Kandilarov et al., 2008, 2010), Mohns  
131 (Klingelhöfer et al., 2000), Gakkel (Jokat & Schmidt-Aursch, 2007), and Mid-Cayman (Van  
132 Avendonk et al., 2017; Grevemeyer et al., 2018b) spreading centers. At the SWIR, large efforts  
133 have been focused on the Dragon Flag OCC at  $49^\circ 39' E$  (e.g. Zhao et al., 2013), the anomalously  
134 thick magmatic crust inferred at  $50^\circ 28' E$  (e.g. Jian et al., 2017; Li et al., 2015), the Atlantis II  
135 Fracture Zone at  $57^\circ E$  (e.g. Muller et al., 1997, 2000), the non-volcanic seafloor exposures at  
136  $64^\circ 30' E$  (e.g. Momoh et al., 2017), and the mafic crustal structure at  $66^\circ E$  (e.g. Minshull et al.,  
137 2006).

138 In the Fall of 2014, French and Canadian scientists collaborated on the marine  
139 geophysical project SISMO-SMOOTH (Leroy & Cannat, 2014) aboard the R/V Marion-  
140 Dufresne to carry out a major 2D and 3D MCS and OBS survey across the SWIR at  $64^\circ 30' E$   
141 (Fig. 1), one of the geologically most sampled areas of the ultra-slow spreading seafloor. This  
142 location was also selected because, albeit remote, it does not have ice floes like the ridges in the  
143 Arctic Ocean, which allows for low-risk use of long MCS streamers and large groups of OBSs.  
144 The main goal of the SISMO-SMOOTH project was to determine the seismic reflection and  
145 velocity structure of an ultraslow-spreading ridge to investigate the geophysical fingerprints of  
146 variably serpentinized mantle peridotites, map the lithospheric fabric, and better understand the  
147 mantle exhumation dynamics, all in an area where the mode of oceanic lithosphere accretion has  
148 been interpreted to form broad exposures of exhumed mantle (Cannat et al., 2006). The first  
149 results from the analysis of the collected data were focused on the 2D and 3D MCS reflection  
150 imaging (Momoh et al., 2017, 2020). Here, we present the results from a tomographic analysis of  
151 two orthogonal  $\sim 150$ -km-long OBS profiles, the north-south (NS) profile that is perpendicular to  
152 the spreading axis (SMOO33; Figs. 1a and 1c) and the east-west (EW) that is subparallel to it  
153 (SMOO35; Figs. 1a and 1b). The seismic velocity structure imaged along these two profiles  
154 provides new information on the subsurface expression of the detachment fault system, including  
155 the distribution of the detachment faults, the degree of exhumed mantle serpentinization and its

156 anisotropy, and the east-west transition from predominantly exhumed mantle to more magma-  
 157 rich lithosphere.



158  
 159 **Figure 1.** (a) Two regional orthogonal OBS profiles NS (SMOO33) and EW (SMOO35) from  
 160 the SISMO SMOOTH Survey (Leroy et al., 2015) overlay a bathymetry map (color background;  
 161 Cannat et al., 2006; Momoh et al., 2017). Thick black lines outline shot locations, and white and  
 162 gray circles with black outlines are the positions of OBS instruments that did and did not provide  
 163 useful data, respectively. The shooting distance inside the dashed red rectangle is 150 m and  
 164 outside 300 m. Dashed black line shows the spreading axis location. Dotted black lines indicate  
 165 the magnetic anomalies isochrons C2An.y (2.581 Ma), C3An.y (5.894 Ma), and C5An.o (10.949  
 166 Ma) (Cande & Kent, 1995; Cannat et al., 2006; Sauter et al., 2008; Reston, 2018). Areas filled  
 167 with inclined thin white lines and bounded by thick white lines delimit the smooth non-volcanic  
 168 seafloor (Cannat et al., 2019). Inset in the top left shows the location of the Southwest Indian  
 169 Ridge (SWIR) relative to the Réunion Island, the Central Indian Ridge (CIR), the Melville  
 170 Fracture Zone (MVFZ), and the Rodriguez Triple Junction (RTJ). Red rectangle shows the limits  
 171 of the study area presented in the main figure. The yellow star, diamond and circle indicate the  
 172 locations of earlier SWIR investigations at 50°E, 57°E and 66°E, respectively. (b, c)  
 173 Magnifications of the main map within thin black rectangles show positions of the OBS

174 instruments along the EW (b) and NS (c) profiles. Only the OBS with useful data are  
175 sequentially numbered, first in NS and then EW direction.

## 176 **2 Study area**

177 The SWIR is an ultraslow-spreading center (Dick et al., 2003; Sauter & Cannat, 2010)  
178 with a full spreading rate of <14 mm/year (Kreemer et al., 2014). Variations in melt-supply,  
179 mantle thermal and compositional heterogeneities, and changes in spreading obliquity along this  
180 ridge result in significant large-scale variations of the accreted lithosphere (Cannat et al., 2008).  
181 For instance, the easternmost SWIR, east of the Melville Fracture Zone (61°E) and west of the  
182 Rodriguez Triple Junction (70°E), is considered to be an endmember in the global ridge system  
183 where the interplay between plate- and mantle-driven processes results in a thin or absent mafic  
184 igneous crust with a complex relationship between intermittent volcanic edifices, corrugated  
185 volcanic seafloor and extensive exhumed mantle domains (Cannat et al., 2003, 2006). At 66°E  
186 the mafic igneous crust is estimated to be 2.2-5.4 km thick (Minshull et al., 2006), less than the  
187 global average of ~6 km (Chen, 1992; Christeson et al., 2019), while the SWIR at 50°28'E shows  
188 an anomalously thick (~9.5 km) crust (Jian et al., 2016). At 64°30'E, which is at the center of our  
189 study area (Fig. 1), the SWIR exhibits the widest non-volcanic seafloor documented thus far  
190 (Cannat et al., 2006) with on-axis volcanic centres inferred to the east and west (Cannat et al.,  
191 2003; Schlindwein & Schmid, 2016).

192 In the last two decades, several surveys have focused on studying the SWIR at 64°30'E  
193 and have used gravity (Cannat et al., 2006), magnetics (Sauter et al., 2008) and side-scan sonar  
194 (Sauter et al., 2013) to identify and map the extension of these non-volcanic seafloor domains  
195 (Fig. 1a) known as “smooth-seafloor” (Cannat et al., 2006). Variably serpentinized mantle-  
196 derived rocks, peridotites, are the dominant lithology at the “smooth-seafloor” (Sauter et al.,  
197 2013) and coincide with high residual mantle Bouguer gravity anomaly (RMBA; 30–50 mGal)  
198 (Cannat et al., 2006). The seafloor topography is characterized by rounded broad ridges with a  
199 height ranging from 500 to 2000 m and a length ranging from 15 to 90 km (Cannat et al., 2006,  
200 2019). Sixteen dredges collected across-axis in the amagmatic corridor of the SWIR at 64°30'E  
201 nearly exclusively recovered variably serpentinized peridotites with a minor amount (<5%) of  
202 basalts and gabbros (Sauter et al., 2013). Oxygen isotope analyses on these samples suggest  
203 relatively high serpentinization temperatures (271–366°C) and in-situ and bulk-rock analyses of  
204 the samples support seawater as the serpentinizing fluid, ruling out leaching of basalts or gabbros  
205 (Rouméjon et al., 2014). Moreover, the scarcity of melt products is expected as a result of the  
206 very low melt supply inferred in the area based on large axial depth and high mean basalt sodium  
207 content (Cannat et al., 2008; Meyzen et al., 2003; Seyler et al., 2003).

208 Momoh et al. (2017) studied the 3D seismic reflection structure at the SWIR at 64°30'E  
209 and suggested that the uppermost lithosphere consists of a 4–5 km thick layer mostly composed  
210 of serpentinized peridotites with a small proportion of igneous rocks derived from occasional and  
211 incipient magmatism. Two main packages of seismic reflectors are imaged across the ridge axis:  
212 (1) a group of subparallel reflectors dipping south at 50–60° in the footwall and (2) a group of  
213 north dipping reflectors in the hanging wall of the active detachment fault (Momoh et al., 2017).  
214 The former are interpreted to be related to the damage zone of the active axial fault, and the latter  
215 are suggested to represent either the damage zone of a previously active fault's footwall or to be  
216 related to recent tectonic extension occurring on the hanging wall. Similarly, Momoh et al.  
217 (2020) proposed that the crustal-type seismic velocities are related to extensive tectonic damage

218 and hydrothermal alteration of both peridotites and occasional intruding gabbros.  
219 Serpentinization and incipient magmatism are thought to occur in two successive phases: first  
220 when mantle-rocks are exhumed on the active detachment fault footwall and later when these  
221 rocks constitute the hanging wall of the next detachment fault (Cannat et al., 2019; Momoh et al.,  
222 2020).

223 Microseismicity studies have constrained a thick (20–25 km) brittle lithosphere in the  
224 vicinity of our study area (Schlindwein & Schmid, 2016). As a consequence of the virtually-zero  
225 melt supply, the seafloor is largely created by successive, flipping polarity, detachment faults  
226 that form broad unroofed mantle domains both north and south of the spreading axis (Cannat et  
227 al., 2006, 2019; Sauter et al., 2013; Reston, 2018). Active and abandoned detachment fault  
228 surfaces have been imaged with side-scan sonar at the SWIR at 64°30'E (Sauter et al., 2013) and  
229 seismic reflectors associated with the currently active axial detachment fault have been observed  
230 in 3D MCS data (Momoh et al., 2017). Numerical models have demonstrated that a combination  
231 of serpentinization and grain size reduction in thick brittle lithosphere can generate flip-flop  
232 detachment faulting (Bickert et al., 2020). Detailed bathymetric and kinematic analysis have  
233 been carried out to investigate the emergence and breakaway of the interpreted faults and explain  
234 the mantle exhumation dynamics at detachment-dominated spreading ridges (Cannat et al., 2019;  
235 Reston, 2018). Still lacking, however, is a well-resolved regional-scale velocity model capturing  
236 these detachment faults, active and abandoned, to back up or rebut the proposed lithospheric  
237 accretion models.

### 238 **3 Data acquisition and analysis**

#### 239 **3.1 Seismic experiment**

240 The MD 199 - SISMO-SMOOTH Cruise 2014 (Leroy & Cannat, 2014; Leroy et al.,  
241 2015) in the easternmost SWIR collected a variety of datasets including pseudo-3D MCS  
242 (Momoh et al., 2017), 2D MCS (Momoh et al., 2020), 3D wide-angle OBS, and 2D wide-angle  
243 OBS. The 3D MCS and 3D OBS data are focused in narrow (1.8x24 km and 20x30 km,  
244 respectively) rectangles at the ridge axis, while the 2D profiles extend ~150 km across and along  
245 the spreading axis. In this paper, we show and interpret the results from analysis of the regional  
246 2D OBS wide-angle data set (Fig. 1). Shots for the 3D MCS survey recorded by OBSs 4-13 on  
247 the NS profile (Fig. 1) were used by Momoh et al. (2017) to form a simple velocity model for  
248 migration of the reflection signal. However, none of the regional 2D OBS shots recorded by the  
249 32 OBSs on the NS and the EW profile (Fig. 1), which provide a far greater source-receiver  
250 offset range and crossing ray area that are needed for extracting high-quality detailed velocity  
251 information, have been analyzed prior to this study.

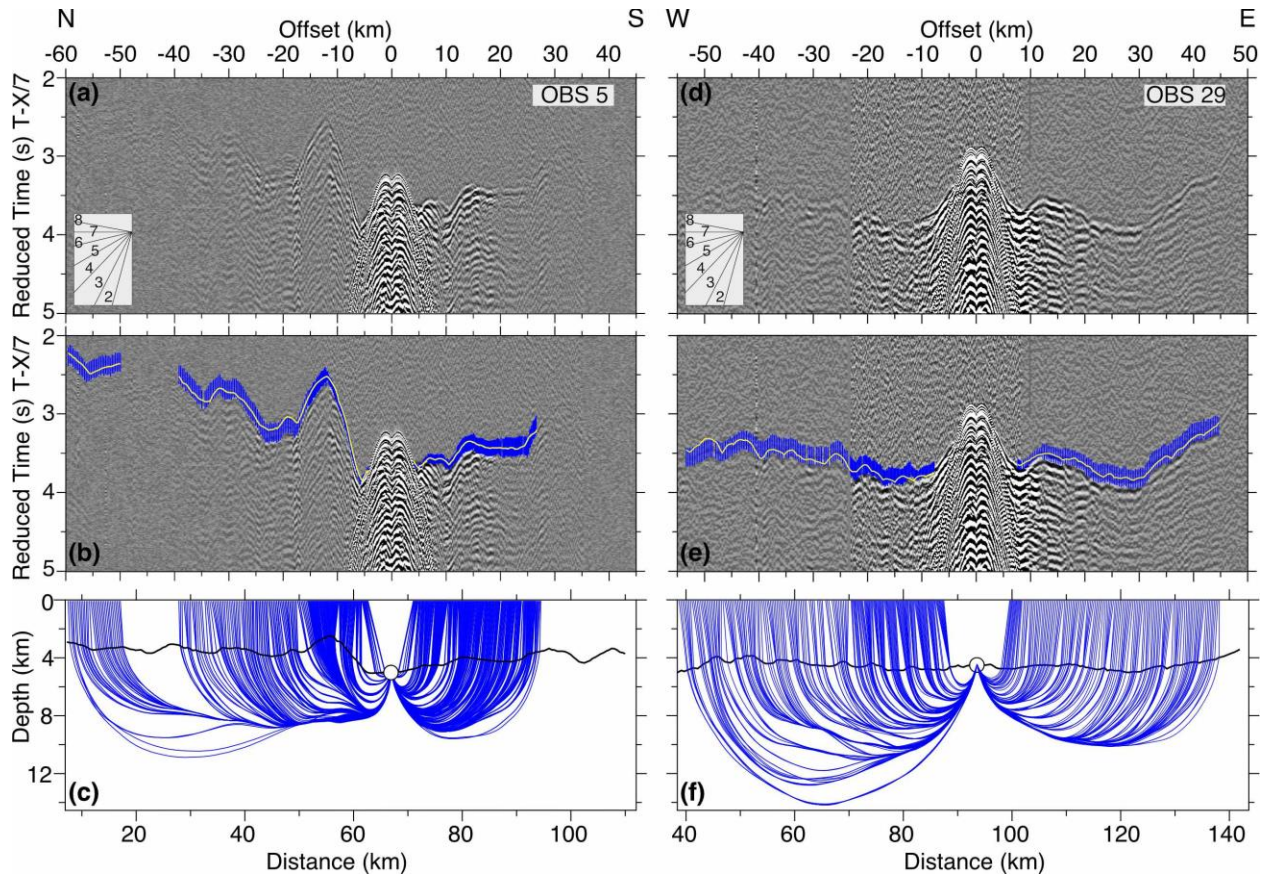
252 The seismic source consisted of two linear arrays of 7 air guns towed at an average depth  
253 of 14 m with a total nominal volume of 6790 in<sup>3</sup>. Three different types of short-period OBS were  
254 used to record the wide-angle data: 16 Canadian OBSs from Dalhousie University, 7 French  
255 OBSs from the Institut National des Sciences de l'Univers du CNRS, and 9 micro OBSs from the  
256 National Central University of Taiwan. All the OBSs recorded continuously at a sampling rate of  
257 250 Hz (4 ms) with one hydrophone and a 3 orthogonal component geophone. The OBS spacing  
258 ranges from 3 to 10 km. For presentation purposes, OBS names used during the survey were  
259 converted to sequential numbers (Fig. 1; Table S1). OBSs north of OBS 1, and between OBSs 3  
260 and 4, OBSs 13 and 14, and OBSs 31 and 32 were lost during the survey or recorded unusable



261 data. The white circles in Figure 1 are the 32 OBS instruments used for the modeling. These  
262 OBSs form the two ~150 km long wide-angle seismic profiles. The EW profile (SMOO35) lies  
263 in the spreading axis direction, and the NS profile (SMOO33) is orthogonal to and crosses the  
264 spreading axis (Fig. 1a). The NS profile cuts through the inferred detachment faults and  
265 practically all of its OBSs lie within the previously mapped smooth seafloor. The EW profile is  
266 presumed to transition from volcanic seafloor in the west to exhumed mantle at the seafloor in  
267 the east.

### 268 3.2 Data processing

269 The OBSs were relocated to their true positions on the seafloor (more information in  
270 Supporting Information) and the data were bandpass filtered using a minimum phase trapezoidal  
271 band-pass filter with corner frequencies 1-5-18-25 Hz. PASTEUP software (Fujie et al., 2008)  
272 was used to manually pick the first break of arrivals on the OBS records. Two examples of OBS  
273 gathers for the NS and the EW profiles, with and without picks, are shown in Figure 2. More  
274 examples are shown in Supporting Information (Figs. S1-S4). The picking was carried out on  
275 unfiltered data as much as possible, with the filtered data used only to extend the picks to further  
276 offsets. Where first arrivals are not clear at long offsets, arrivals of the first water multiple were  
277 picked where possible and then time-shifted until picks from the multiple for near and/or mid  
278 offsets coincided with equivalent picks for the first arrivals. The 2D bathymetry from a previous  
279 multibeam survey (Cannat et al., 2006) was plotted in a separate window above the OBS data  
280 window as a function of model distance and source-receiver offset to help guide the picking  
281 process and identify seafloor diffractions. First arrivals were picked to offsets of up to ~90 km  
282 along the NS profile, and ~60 km along the EW profile. Assigned pick uncertainty is offset  
283 dependent as follows: 30 ms for high-quality waveforms at offsets <6.5 km; 60 ms for offsets  
284 between 6.5 km and 12 km; 100 ms for offsets between 12 km and 20 km; 120 ms for offsets  
285 >20 km; and 150 ms for time-shifted multiple arrival picks. A total of 6,523 and 4,193 first  
286 arrival picks were made for the NS and the EW profile, respectively. Secondary arrivals, such as  
287 Moho reflections, were not found in the OBS records, and sediment arrivals were negligible.



288

289 **Figure 2.** Example OBS gathers for the NS and EW profiles (a and d) are shown together with  
 290 related first arrival traveltime picks (b and e) and calculated ray paths (c and f). (a) Hydrophone  
 291 data for OBS 5 on the NS profile and (d) vertical geophone data for OBS 29 on EW profile after  
 292 application of a reduction velocity of 7 km/s and band-pass filtering (1-5-18-25 Hz). Insets show  
 293 the linear moveout of different phase velocities in km/s. (b and e) Same as a and b but with  
 294 picked (centers of blue error bars) and modeled (yellow curves) first arrival traveltimes  
 295 superimposed. Synthetic traveltimes and raypaths were computed by ray tracing through the final  
 296 velocity model. (c and f) Raypath diagram for the modeled first arrival traveltimes in b and e.  
 297 Black thick line is the seafloor and white circles with thin black outlines show the OBS  
 298 locations.

299

### 3.3 Traveltime tomography

300

We performed P-wave traveltime tomography using TOMO2D (Korenaga et al., 2000).  
 301 Traveltimes of P-wave first refracted arrivals and later Moho reflection arrivals (*PmP*) are  
 302 commonly used for joint inversion of the regional 2D velocity structure in a sheared mesh model  
 303 hung from the seafloor (e.g. Watremez et al., 2015). We, however, do not model the Moho  
 304 reflector because of the lack of *PmP* arrivals in the data. In the TOMO2D method, forward  
 305 modeling is first applied to find the shortest raypath from the shot to the receiver for each arrival,  
 306 followed by a least-squares regularized inversion, in which the starting velocity model is  
 307 perturbed and updated until the targeted chi-squared ( $\chi^2$ ) or the set number of maximum  
 308 iterations is reached (Korenaga et al., 2000). Application of smoothing and optional damping  
 309 constraints is used to regularize the iterative inversion process (Korenaga et al., 2000). Thus,

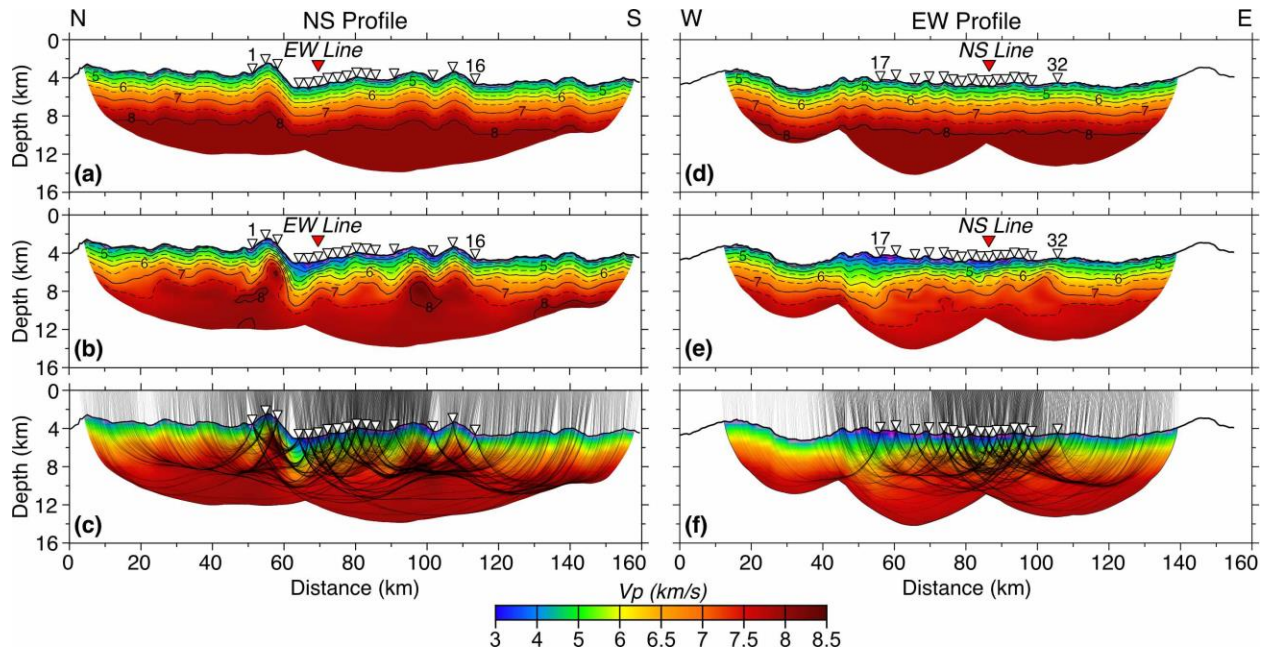
310 damping and smoothing weighting factors control the data fitting and the smoothness of the  
311 model, and similarly, the correlation lengths for the velocity nodes control the inversion stability.  
312 The final TOMO2D product is a minimum-structure smooth velocity model, meaning that  
313 minimum a priori information is used to resolve the minimum or simplest structure needed to  
314 explain the data. This approach reduces subjective input from the interpreter in the development  
315 of the final tomographic model.

316 Cells in our models, which are 160 km long and 30 km deep, are 1 km wide and 500 m  
317 high. The starting 1D velocity model is based on the 1D average velocity of Momoh et al. (2017)  
318 and the average velocity structure reported at other ultraslow-spreading centers bearing  
319 serpentinized mantle domains (e.g. Grevemeyer et al., 2018b; Van Avendonk et al., 2017). We  
320 opted for simplicity and, after taking into consideration the expected geology, settled on a 1D  
321 starting velocity model with 3 velocity-depth points: 4 km/s velocity at the seafloor, 6.5 km/s at 2  
322 km depth below the seafloor (dbsf), and 8.0 km/s at 5 km dbsf. For consistency, the same 1D  
323 starting velocity model extended in 2D by hanging it off the seafloor was used for both  
324 orthogonal profiles. Figures 3a and 3d show the starting velocity model extended in 2D for both  
325 the NS and EW profile, respectively. For the NS profile, the starting velocity model produced a  
326  $\chi^2$  of 11.97 and an RMS travelttime residual of 208 ms. For the EW profile, it produced a  $\chi^2$  of  
327 15.37 and an RMS travelttime residual of 261 ms. We set our inversion to run 5 iterations and use  
328 the same parametrization values for both profiles to prevent modeling inconsistencies (more  
329 information in Supporting Information).

## 330 **4 Results**

### 331 4.1 Velocity models

332 Figure 3b shows the final velocity model for the NS profile. The model is shown over the  
333 areas with ray coverage and the calculated raypaths are shown in Figure 3c overlaying the final  
334 velocity model. The model converges to an RMS travelttime residual of 53 ms and a  $\chi^2$  of 1.10.  
335 Similarly, the final velocity model of the EW profile is shown in Figure 3e and the calculated  
336 raypaths are shown in Figure 3f. The EW model converges to an RMS travelttime residual of 55  
337 ms and a  $\chi^2$  of 0.86. Figure S5 shows the evolution of  $\chi^2$  as a function of iteration number for  
338 both models. Most of the lateral and vertical velocity variations are found in the central parts of  
339 the models, below the areas covered by the OBSs, and therefore in the areas where there are  
340 crossing raypaths (Figs. 3c and 3f). Toward the profile ends, which are sections covered by shots  
341 but no OBSs, or sections with no crossing raypaths, the velocity structure in the final models  
342 mostly follows the seafloor topography and the starting velocity models.



343

344 **Figure 3.** Results from first arrival traveltimes tomographic inversion of the NS and EW profiles:  
 345 (a, d) Starting velocity models; (b, e) final tomographic models; (c, f) raypaths (black lines) of  
 346 the first arrivals traced through the final velocity models (b, e), for the NS and EW profiles,  
 347 respectively. Iso-velocity contours are shown every 0.5 km/s in (a, b, d and e); solid black at  
 348 every km/s and dashed in between. White inverted triangles show the positions of the OBSs on  
 349 the seafloor. Red triangles show the location at which the profiles cross each other.

350 Traveltimes are well fitted at all model distances (Figs. S6a and S6b) with the majority of  
 351 traveltimes residuals, calculated as the difference between the observed and calculated  
 352 traveltimes, reduced by the inversion to  $\pm 60$  ms (Figs. S6c and S6d).

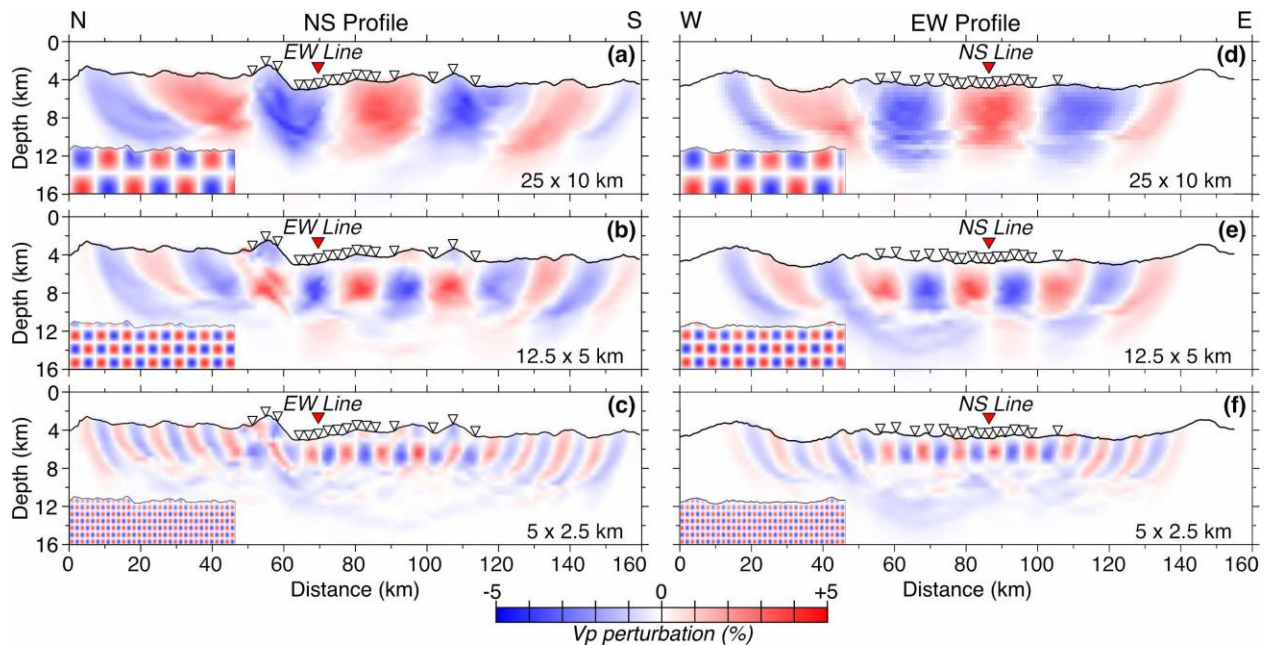
353 The final velocity models (Figs. 3b and 3e) show that seismic velocities increase rapidly  
 354 with depth with velocities ranging from  $\sim 3.5$  km/s at the seafloor to 7 km/s at 1.5–5.5 km dbfsf  
 355 at the NS profile, and from  $\sim 3$  km/s at the seafloor to 7 km/s at 2–6 km dbfsf at the EW profile. The  
 356 NS profile reaches velocities in the range of 7.8–8.4 km/s, while the EW profile only reaches  
 357 velocities in the range of 7.6–7.8 km/s. The NS profile shows greater lateral changes in the  
 358 velocities than the EW profile, including a sharp lateral change (horizontal gradient of  $\sim 1$  s $^{-1}$  at  
 359  $\sim 62$  km model distance) at the highest topographic feature. Within the area of OBS coverage, the  
 360 EW profile also shows a smooth trend of increasing velocities toward the East, which is  
 361 accentuated between OBSs 30 and 32. Similarly, a distinct increase in the velocities is observed  
 362 between OBSs 17 and 18. Both profiles show high vertical velocity gradient (velocity contours  
 363 closely spaced) in upper sections of the models (velocities  $< 7$  km/s) and a considerably lower  
 364 vertical gradient (sparse velocity contours) in deeper sections (velocities  $> 7$  km/s).

## 365 4.2 Uncertainty assessment

### 366 4.2.1 Checkerboard Tests

367 We assess the resolution of our models with checkerboard tests as follows. Checkerboard  
 368 patterns for a set of different horizontal and vertical cycle lengths with a  $\pm 5\%$  periodic velocity  
 369 perturbation are added to our two final velocity models. The chosen cell widths and heights are:

370 25x10 km (Figs. 4a and 4d), 12.5x5 km (Figs. 4b and 4e), and 5x2.5 km (Figs. 4c and 4f). The  
 371 perturbed velocity models are used as the starting velocity models for the inversions. The  
 372 recovered perturbations are obtained by subtracting the input velocity models from the  
 373 corresponding final inverted velocity models for different cell sizes. The extent of perturbation  
 374 recovery at any particular section of our models is a measure of resolvability of corresponding  
 375 velocity anomalies for that model area. The input and inverted perturbation models are compared  
 376 for the two profiles in Figure 4. Full-size checkerboard test figure is included in the SI (Fig. S7).  
 377 Large structures are resolved across the profiles where there is any ray coverage (Figs. 4a and  
 378 4d). Structures 12.5 km wide and 5 km high are resolved in the areas below the seafloor  
 379 encompassing the first and last OBS location to depths of  $\sim 5$  km. In these central areas,  
 380 structures as small as 5x2.5 km are also resolved but the depth or resolvability below the seafloor  
 381 is reduced to  $\sim 2$ – $3$  km. In all the checkerboard tests, the resolved cells are smeared toward  
 382 profile ends and with increasing depth, with the resolution progressively declining.



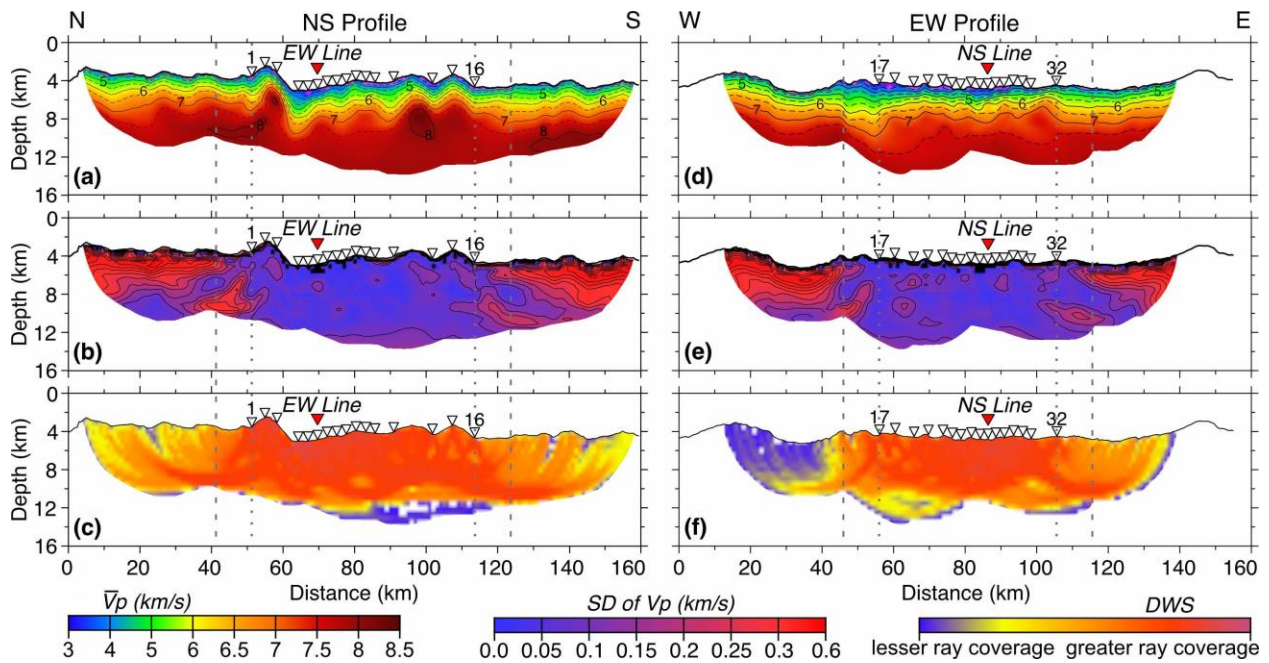
383

384 **Figure 4.** Checkerboard resolution tests for the NS (left column) and EW (right column) profiles  
 385 for perturbation cells: 25 km wide x 10 km high (a and d), 12.5 km wide x 5 km high (b and e), 5  
 386 km wide x 2.5 km high (c and f). The input perturbed model is shown in the bottom left inset and  
 387 the recovered perturbed model is the full-size figure. White and red inverted triangles as in  
 388 Figure 3 caption.

#### 389 4.2.2 Monte Carlo Analysis

390 We run a nonlinear Monte Carlo analysis (e.g. Tarantola, 1987) to estimate velocity  
 391 uncertainty across our models following the strategy of Korenaga et al. (2000). For this analysis,  
 392 the input velocity model (or starting velocity model) is randomized to create a set of 100  
 393 different input models (Fig. S8). This is done by randomizing the velocities of the three nodes of  
 394 the starting velocity model within a  $\pm 6\%$  range and by randomizing the depths to the two sub-  
 395 seafloor layer interfaces or inflection points within a  $\pm 1.1$  km range. The 100 randomized 1D  
 396 starting velocity models are used to form 100 randomized 2D starting velocity models and run

397 100 inversions (for both NS and EW profiles) using the same inversion parameters applied to  
 398 calculate the final velocity models (Figs. 3b and 3e). The resulting tomographic models are  
 399 averaged to produce average velocity models for the NS and EW profiles (Figs. 5a and 5d) and  
 400 compute standard deviations of the P-wave velocities across these models (Figs. 5b and 5e). The  
 401 results show that the inversion process is stable as the average velocity models (Figs. 5a and 5d)  
 402 are very similar to the final velocity models (Figs. 3a and 3d) for both profiles. The standard  
 403 deviation of seismic velocities for most of the lithosphere at central parts of the models is  $< \sim 0.1$   
 404 km/s (Figs. 5b and 5e). Larger standard deviations are observed in the areas that are less well  
 405 resolved as indicated by the checkerboard patterns (Fig. 4), in areas with lower ray density (Figs.  
 406 5c and 5f) and especially where there are no crossing rays (Figs. 3c and 3f), and below the  
 407 seafloor where no instruments were deployed. The ray density is presented by the derivative  
 408 weight sum (DWS; Toomey & Foulger, 1989), a nondimensional relative indicator of ray  
 409 coverage.



410

411 **Figure 5.** Results from the Monte Carlo analysis. Averaged final velocity models for the NS (a)  
 412 and EW (d) profiles. Iso-velocity contours are shown every 0.5 km/s in (a and d); solid black at  
 413 every km/s and dashed in between. Standard deviation of the P-wave velocity calculated via the  
 414 Monte Carlo analysis for the NS (b) and EW (e) profiles. Solid black contours are shown every  
 415 0.05 km/s in (b and e). Derivative weight sums (DWS) indicating the ray coverage for the NS (c)  
 416 and EW (f) profiles. In all panels, the dotted light grey vertical lines mark the end of the best  
 417 resolved areas, and the dashed light grey vertical lines mark the ends of the well resolved areas.  
 418 White and red inverted triangles as in Figure 3 caption.

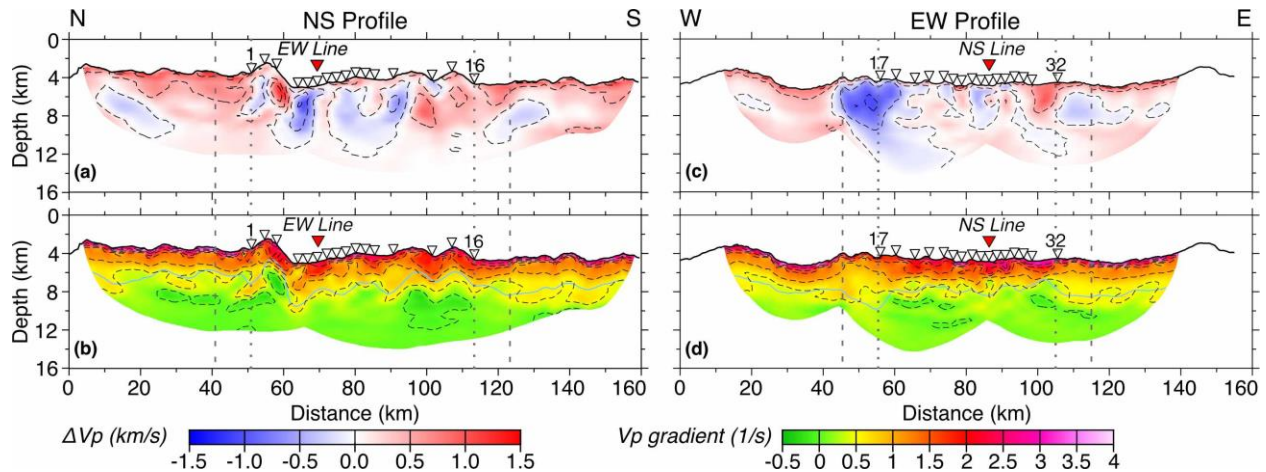
419

#### 4.3 Derivatives of the velocity models

420

To aid the discussion, we plot 1D velocity-depth functions (Fig. S9) extracted at every 1  
 421 km distance within the best resolved areas of the two average velocity models (Figs. 5a and 5d).  
 422 We use these functions to determine the average 1D velocity-depth functions and the extent of  
 423 the velocity-depth envelopes for both profiles (Fig. S9). We further augment our interpretation

424 by computing and plotting velocity anomalies (Figs. 6a and 6c) and vertical velocity gradients  
 425 (Figs. 6b and 6d) for both the NS and EW profile. The 2D velocity anomalies are calculated as  
 426 the difference between the average velocity models (Fig. 5a and 5b) and the respective average  
 427 1D velocity-depth functions (Fig. S9). The vertical velocity gradients are calculated by  
 428 computing the central first derivative of the average velocity models (Fig. 5a and 5b).



429

430 **Figure 6.** Velocity anomaly (top panels) and velocity gradient (bottom panels) results for the NS  
 431 and EW profiles, respectively. Dashed black contours are shown every 0.5 km/s in (a and c) and  
 432 every 0.5  $s^{-1}$  in (b and d). The depth to the 7 km/s velocity contour is shown in (b) and (d) in a  
 433 solid light blue line. White and red inverted triangles as in Figure 3 caption. Vertical dotted and  
 434 dashed light grey lines as in Figure 5 caption.

## 435 5 Discussion

436 The final and average P-wave tomographic velocity models for the NS and EW profiles  
 437 give insight into the subsurface structure of the SWIR at 64°30'E. In the following subsections,  
 438 we discuss (1) the ridge structure and the distribution of the active fault and the older and now  
 439 inert detachment faults; (2) the inferred lithospheric composition and its anisotropy; and (3) the  
 440 velocity structure in our study area in the context of the known velocity structure elsewhere at  
 441 the SWIR. We limit our discussion only to the geological structures that can be resolved as  
 442 indicated by the checkerboard tests, and the areas of the velocity models that show standard  
 443 deviations of  $< \sim 0.1$  km/s (Figs. 5b and 5e) and/or high ray coverage (Figs. 5c and 5f). This  
 444 effectively limits our detailed interpretation to the best resolved areas (within the dotted light  
 445 grey vertical lines in Fig. 5) with crossing rays and OBSs on the seafloor (51 to 114 km model  
 446 distance for the NS profile and 55 to 106 km for the EW profile), with the regional interpretation  
 447 extended to include the well resolved areas (within the dashed light grey vertical lines in Fig. 5)  
 448 found up to  $\sim 10$  km away from the first/last OBS on both profiles, as indicated by the Monte  
 449 Carlo analysis results (Fig. 5).

### 450 5.1 Ridge structure

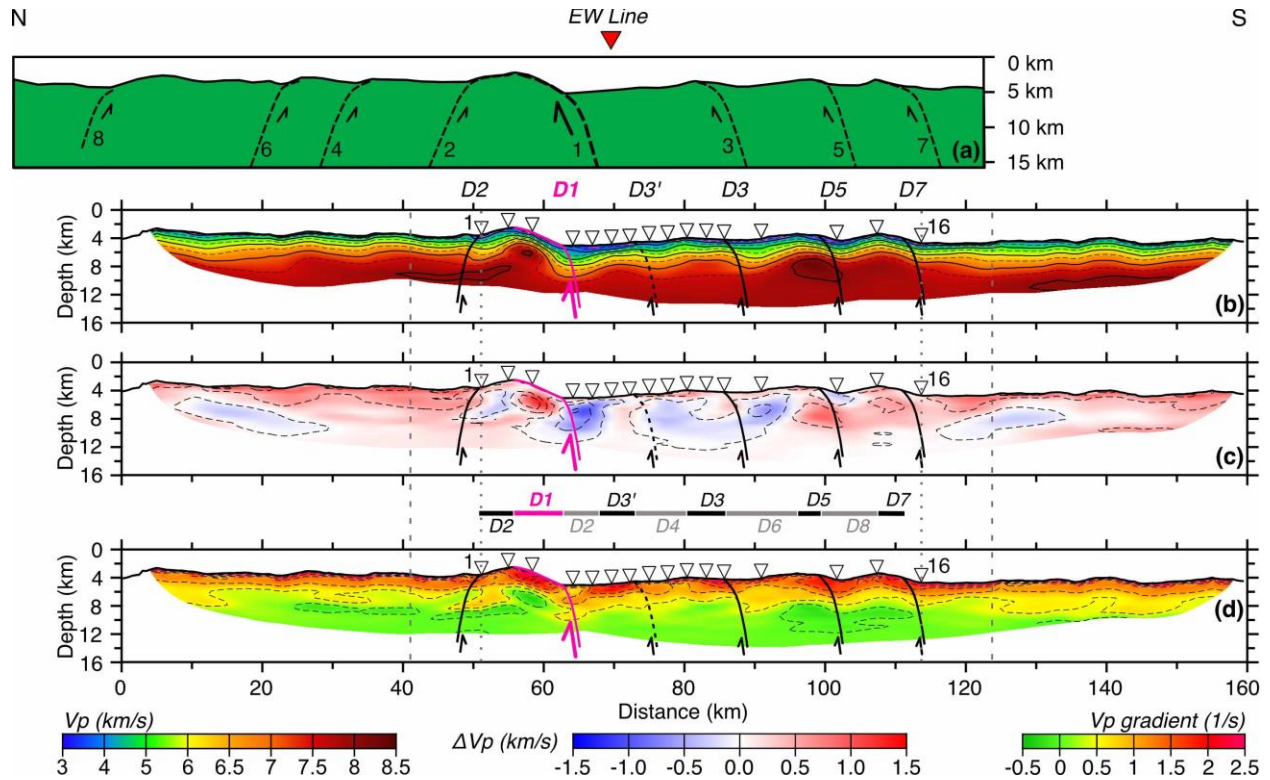
#### 451 5.1.1 Delineation of detachment faults on the NS profile

452 The information extracted from the NS profile (Figs. 5a, 6a and 6b), which cuts across  
 453 the SWIR at 64°30'E, is particularly useful for subsurface identification of the active and

454 abandoned detachment faults. We first interpret a sharp lateral change in the velocities (a  
455 horizontal gradient of  $\sim 1 \text{ s}^{-1}$  at  $\sim 62 \text{ km}$  model distance; Figs. 5a and 6a) and a high vertical  
456 velocity gradient ( $1.5\text{--}2.5 \text{ s}^{-1}$ ; Fig. 6b) coincident with the shallow section (top 2 km, i.e. depth 0-  
457 2 km below seafloor) of the highest topographic feature ( $\sim 51\text{--}62 \text{ km}$  model distance) as the  
458 seismic expression of an active axial detachment fault (hereafter *D1*; Fig. 7). This bathymetric  
459 high is characterized by higher velocity than the surrounding regions (Fig. 5a) and its top surface  
460 has previously been interpreted as an active axial detachment fault based on side-scan sonar  
461 (Sauter et al., 2013), bathymetric and kinematic analysis (Cannat et al., 2019; Reston, 2018), and  
462 seismic reflection data (Momoh et al., 2017, 2020). Therefore, our velocity model supports these  
463 earlier interpretations as it shows the footwall exhuming (or bringing up to shallow depth) rocks  
464 of high velocities, those corresponding to lithologies typically found at greater depths, which  
465 become superimposed by the lower velocity rocks of the hanging wall at the topographic low  
466 immediately south. This north-south transition at  $\sim 62 \text{ km}$  model distance from the detachment  
467 footwall to the detachment hanging wall is characterized by an abrupt decrease in the vertical  
468 gradient, from a high of  $1.5\text{--}2.5 \text{ s}^{-1}$  to a low of  $0.5\text{--}1.0 \text{ s}^{-1}$ , as well as by a switch in polarity of  
469 the largest velocity anomaly, from  $1.5 \text{ km/s}$  to  $-1.5 \text{ km/s}$ .

470 Elsewhere, the NS profile exhibits smoothly varying low to moderate vertical velocity  
471 gradient values ( $<1.5 \text{ s}^{-1}$ ), except for several locations that show similar lateral changes in the  
472 vertical velocity gradient to *D1*, with higher vertical gradients ( $1.5\text{--}2.0 \text{ s}^{-1}$ ) juxtaposed with lower  
473 gradients ( $0\text{--}1.0 \text{ s}^{-1}$ ) at alike depths ( $\sim$ top 2 km). These are found at model distances of about 50  
474 km, 74 km, 87 km, 101 km, and 113 km (Fig. 6b). The velocity anomalies (Fig. 6a) at these  
475 model distances also show a similar change to *D1*, with a switch in the polarity from positive to  
476 negative anomaly, except at 50 km where the switch in polarity is reversed. Additionally, the  
477 velocities (Fig. 5a) across the profile show a repeat drop pattern that coincides with the changes  
478 observed in the vertical velocity gradient and velocity anomalies. While the identified vertical  
479 velocity gradient, velocity, and velocity anomaly changes are not as pronounced as for the  
480 interpreted active detachment fault *D1*, they are clearly recognizable and we interpret them to  
481 indicate the subsurface location of the abandoned detachment faults (*D2* at  $\sim 50 \text{ km}$ , *D3'* at  $\sim 74$   
482 km, *D3* at  $\sim 87 \text{ km}$ , *D5* at  $\sim 101 \text{ km}$ , and *D7* at  $\sim 113 \text{ km}$ ; Fig. 7b).





483

484 **Figure 7.** Comparison of a (a) conceptual model based on previous interpretations (Cannat et al.,  
 485 2019; Reston, 2018; Sauter et al., 2013) depicting the sequence of flipping rolling-hinge faults  
 486 (black dashed lines) along the NS profile and (b) the average velocity model from Figure 5a, (c)  
 487 velocity anomalies from Figure 6a, and (d) vertical gradient from Figure 6b, with interpreted  
 488 locations of corresponding faults (solid lines) for the best resolved area (51-114 km). In (a),  
 489 previously interpreted faults are numbered 1–8 from youngest to oldest. In (b, c, and d) the  
 490 identified faults are named *D2*, *D1*, *D3'*, *D3*, *D5*, *D7*, and are equivalent to faults 2, 1, 3, 5, and 7  
 491 in (a) omitting *D3'* that is not inferred in (a). *D* stands for detachment. *D1* is the active  
 492 detachment fault (solid pink line) and the fault numbers increase sequentially to describe older  
 493 abandoned faults (solid black lines). Inset between (c) and (d) shows the interpreted extension of  
 494 each detachment footwall surface located on the seafloor. The thin rectangle in pink limits the  
 495 extension of *D1* footwall surface, the black rectangles limit the extension of *D2*, *D3'*, *D3*, *D5* and  
 496 *D7* footwall surfaces, and the grey rectangles limit the extension of *D2*, *D4*, *D6*, and *D8* hanging  
 497 wall surfaces. Iso-velocity contours are shown every 0.5 km/s in (b); solid black at every km/s  
 498 and dashed in between. Dashed black contours are shown every 0.5 km/s in (c) and every 0.5 s<sup>-1</sup>  
 499 in (d). White and red inverted triangles as in Figure 3 caption. Vertical dotted and dashed light  
 500 grey lines as in Figure 5 caption.

501

### 5.1.2 Comparison with earlier interpretations

502 In Figure 7, we compare our interpretation on the location of the detachment faults with  
 503 that of previous interpretations (Cannat et al., 2019; Reston, 2018; Sauter et al., 2013), which are  
 504 primarily based on observed bathymetric features, i.e. the location of the breakaway and  
 505 emergence points on the fault surfaces. For simplicity, we list the detachment faults previously  
 506 inferred from 1 to 8, with 1 being the youngest and 8 being the oldest fault (Fig 7a). Our velocity

507 model (Fig. 7b), the velocity anomalies (Fig. 7c), and the vertical velocity gradient (Fig. 7d),  
 508 provide supporting evidence for the existence of faults 1–3, 5 and 7. The reverse polarity in the  
 509 pattern observed in the velocity anomalies at ~50 km model distance (Fig 7c), at the interpreted  
 510 location of *D2*, coincides with the previously interpreted location of fault 2 (Cannat et al., 2019;  
 511 Reston, 2018; Sauter et al., 2013), only offset by ~1 km south. Fault *D2* was active prior to *D1*  
 512 and thus it shows opposite polarity. Similarly, the interpreted location for fault *D1* is comparable  
 513 to the location of the previously interpreted fault 1, only offset by ~1 km north (Fig. 7).  
 514 Previously interpreted locations of faults 5 and 7 are identical to the interpreted locations of *D5*  
 515 and *D7* in this work, respectively, while fault 3 is offset by ~4 km north from *D3* (Fig. 11). The  
 516 locations of faults *D1* and *D3* are also consistent with previously identified south dipping  
 517 reflectors in MCS data (Momoh et al., 2017, 2020) interpreted to be related to fault damage in  
 518 the footwall. The results presented here also demarcate an abandoned detachment *D3'* that has  
 519 not been inferred previously. Detachment *D3'* may have been missed in other models (Cannat et  
 520 al., 2019; Reston, 2018; Sauter et al., 2013) because its seafloor expression shows a smoother  
 521 emergence topography in comparison to the other interpreted faults based on bathymetry. One  
 522 possible cause for this is a relatively short lifespan of the *D3'* fault, which precluded full  
 523 development of the characteristic seafloor geometry of a detachment fault. Alternatively, the  
 524 velocity signature observed at ~74 km model distance may possibly be related to recent  
 525 extensional damage occurring in the hanging wall of *D1* as proposed by Momoh et al. (2020),  
 526 since no south dipping reflectors related to this fault are imaged in MCS data (Momoh et al.,  
 527 2017). However, given the resolution of our model, the presence of fault *D3'* provides an  
 528 explanation that is more consistent with the velocity, velocity anomaly, and vertical gradient  
 529 patterns observed at the locations of other interpreted faults (Figs. 7b, 7c, and 7d).

530 For our study area, Cannat et al. (2019) estimated the duration of the active deformation  
 531 period for each detachment fault. The active life for these detachments ranges from 0.6 Myr to  
 532 2.8 Myr, with an average life of 1.35 Myr. Our interpretation of an additional detachment fault  
 533 (*D3'*; Fig. 7) combined with the adjustment in the location of *D3* calls for a reevaluation of these  
 534 age numbers. However, we can only estimate total fault longevity for detachments *D1* and *D2*,  
 535 and partial longevity for faults south of *D1* since our NS velocity model does not resolve the  
 536 location of the faults north of *D2* (Table 1). Like Cannat et al. (2019), we estimate the partial and  
 537 total fault lifespans based on the horizontal distance between faults' emergence and breakaway  
 538 points (Table 1). The inset shown in between Figures 7c and 7d illustrates the interpreted  
 539 extension of each detachment footwall surface on the seafloor used for our calculations. Portions  
 540 of each abandoned fault footwall are located both north and south of the spreading axis, except  
 541 for the active detachment (*D1*) that has not yet been cut off by a new master fault and carried  
 542 away from the axis. Portions of *D4*, *D6* and *D8*, the faults inferred further north of *D2* but not  
 543 resolved by our velocity model, are also considered based on earlier interpretations of seafloor  
 544 data (Cannat et al., 2019; Reston 2018; Sauter et al., 2013) to estimate the ages in Table 1. Our  
 545 calculations point to a duration of ~0.7 Myr for *D2*, half the previously suggested age, and ~0.5  
 546 Myr for *D1*, larger than the previously suggested age (0.3 Myr).

547

<b>Detachment #</b> <i>(Youngest to oldest)</i>	<b>Dip</b>	<b>Horizontal distance from fault emergence to</b>	<b>Horizontal distance from fault breakaway</b>	<b>Estimated partial fault duration</b>	<b>Estimated total fault duration (Myr)</b>
----------------------------------------------------	------------	----------------------------------------------------------------	-------------------------------------------------------------	-----------------------------------------------------	---------------------------------------------------------

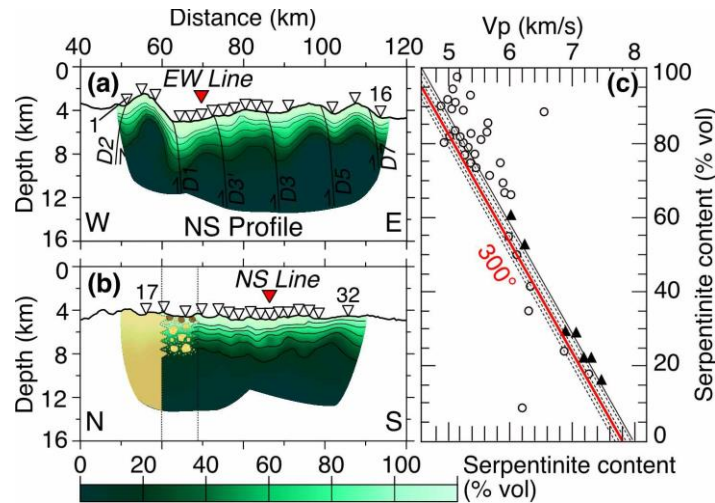
		<b>adjacent fault breakaway (km)</b>	<b>to adjacent fault emergence (km)</b>	<b>based on distance (Myr)</b>	
D1 ( <i>active</i> )	South	7.1	0	0.51	0.51
D2 ( <i>North of D1</i> )	North	5.0	-	0.36	0.72
D2 ( <i>South of D1</i> )	North	-	5.1	0.36	
D3'	South	5.1	<i>Unknown</i>	0.36	<i>Unknown</i>
D4	North	<i>Unknown</i>	7.3	0.52	<i>Unknown</i>
D3	South	5.5	<i>Unknown</i>	0.39	<i>Unknown</i>
D6	North	<i>Unknown</i>	10.2	0.73	<i>Unknown</i>
D5	South	3.4	<i>Unknown</i>	0.24	<i>Unknown</i>
D8	North	<i>Unknown</i>	8.0	0.57	<i>Unknown</i>
D7	South	3.9	<i>Unknown</i>	0.28	<i>Unknown</i>

548 **Table 1.** Estimated detachment fault longevity based on the horizontal distance between faults'  
549 emergence and breakaway locations of the inferred faults across the profile SMOO33 (NS  
550 profile).

## 551 5.2 Lithospheric composition

### 552 5.2.1 Exhumed mantle area

553 In order to relate the modeled velocities to the subsurface lithology, we assume that the  
554 investigated area is composed of exhumed mantle rocks ranging from variably serpentinized and  
555 fractured peridotites at shallower depths, to unaltered peridotites at greater depths. This is  
556 followed by conversion of the seismic velocities along the NS and EW profiles (Figs. 5a and 5d)  
557 to degree of serpentinization (Figs. 8a and 8b) using a linear relationship for partially-  
558 serpentinized peridotites (Fig. 8c) from Carlson & Miller (2003). These authors used empirical  
559 data (Christensen, 1978, 1996; Miller & Christensen, 1997) to evaluate how the relationship  
560 between P-wave velocities and the degree of serpentinization changes with different confining  
561 pressure and temperature. We choose the relationship for a temperature of 300°C, in agreement  
562 with the high serpentinization temperatures suggested for the rock samples analyzed along the  
563 NS profile (Roum  jon et al., 2014). Albeit converting seismic velocities to degree of  
564 serpentinization is a simplification of the causes for the velocity heterogeneities, especially at the  
565 top of the model, it allows us to discuss how deep serpentinization may extend and what may  
566 control such depth.



567

568 **Figure 8.** Estimated degree of serpentinization for the NS (a) and EW (b) profiles based on their  
 569 tomographic velocities (Figs. 5a and 5d) and the linear relationship (c) of serpentinite content  
 570 with P-wave velocity from Carlson and Miller (2003) for a temperature of 300°C, in agreement  
 571 with oxygen isotopes studies in the area (Rouméjon et al., 2015). Black triangles in (c) are  
 572 laboratory measurements of P-wave velocities in serpentinized peridotites at 25°C and at 200  
 573 MPa and open circles are velocity measurement at other confining pressure (up to a 1000 MPa).  
 574 Carlson & Miller (2003) collected data points from several studies (Christensen, 1978, 1996;  
 575 Miller & Christensen, 1997). Dashed lines show the approximate relationships at temperatures of  
 576 100, 200, 300, 400, and 500°C corrected from the best fitting relationship measured at 200 MPa  
 577 (solid black line). In (a) the solid black lines show the detachment faults' locations interpreted in  
 578 Fig. 7. In (b) the dotted lines show the location at which there is a change in the lithosphere  
 579 (volcanic before 59 km, non-volcanic after 69 km, and transitional in between) as discussed in  
 580 the text. Solid black contours in (a and b) show serpentinite content at every 20% vol. White and  
 581 red inverted triangles in (a) and (b) as in Figure 3 caption.

582 While our assumption of the exhumed mantle domains throughout the study area is a  
 583 simplification of the true geology, it is supported by several lines of evidence and justified by our  
 584 intent of carrying out a first-order interpretation of the subsurface geology. First, earlier rock  
 585 sampling of the seafloor in our study area (Rouméjon et al., 2014; Sauter et al., 2013) has  
 586 predominately retrieved peridotites and the extent of the mapped smooth seafloor (Cannat et al.,  
 587 2006) (Fig. 1), interpreted to represent exhumed mantle domains, covers majority ( $\sim 3/4$ ) of the  
 588 area along the NS and EW profiles contained by the OBS instruments. Second, no reflection  
 589 Moho has been identified in the pseudo-3D MCS dataset by Momoh et al. (2017), which  
 590 indicates the lack of a distinct and seismically-well characterized Moho in our study area.  
 591 Reflection Moho is often well imaged in upper oceanic lithosphere formed at faster spreading  
 592 mid-ocean ridges where a 3-layer mafic oceanic crust tops the ultramafic uppermost mantle (e.g.  
 593 Aghaei et al., 2014; Bouhalanis et al., 2020). Third, no wide-angle Moho reflected arrivals  
 594 (*PmP*) are identified in the OBS data examined in this work, and these arrivals are routinely  
 595 observed and picked in data collected worldwide in oceanic lithosphere formed at faster  
 596 spreading rates (e.g. Canales et al., 2000; Horning et al., 2016). Furthermore, the lack of *PmP*  
 597 arrivals in our study area is consistent with observations from other ultraslow spreading ridges  
 598 dominated by magma starved spreading conditions (Grevemeyer et al., 2018b).

599 We interpret the uppermost lithosphere to be composed of highly fractured and fully  
600 serpentinized peridotites at the top with a gradual decrease in pore space volume and  
601 serpentinization degree to unaltered peridotites at depth (Figs. 8a and 8b). The velocities lower  
602 than 4.5 km/s at the top of the model may be representative of peridotites that are fully  
603 serpentinized and heavily fractured. Geologically, these low velocities could also be  
604 representative of basalts. However, this possibility is highly unlikely because, whereas scattered  
605 low-volumetric basalts may be found across the study area (Sauter et al., 2013), the results of the  
606 extensive seafloor studies and the lack of Moho reflections in both wide-angle OBS and MCS  
607 data (e.g. Cannat et al., 2006; Sauter et al., 2013; Momoh et al., 2015) clearly indicate that a  
608 continuous basaltic top layer is not present.

609 While our first order interpretation is comparable to that of Momoh et al. (2017), the  
610 higher resolution and lateral variability of the new results presented in this work make it possible  
611 to carry out a detailed interpretation. Serpentinization seems more extensive and penetrates  
612 deeper along the NS profile than the EW profile. For example, 20% serpentinization extends to  
613 up to ~5 km dbf on the NS profile and up to ~3–4 km dbf on the EW profile, with the anomaly  
614 being the location of the active detachment fault (*DI*) footwall where the 20% serpentinization  
615 contour is found at the shallowest depths of ~1.5–2 km dbf (Figs. 8a and 8b).

616 The change from relatively large and quickly decreasing vertical velocity gradients (1.5-  
617  $2.5 \text{ s}^{-1}$ ) at velocities <7 km/s, to relatively low and gradually decreasing vertical velocity  
618 gradients (<0.5  $\text{s}^{-1}$ ) at velocities >7 km/s (Figs. 6b and 6d) likely illustrates a change in the  
619 porosity and permeability regime that impacts the serpentinization process for the upper and  
620 lower sections of the profiles. These low and gradually decreasing vertical velocity gradients  
621 within deeper sections of the models also indicate that the Moho transition is not an abrupt  
622 serpentinization or alteration front, as has been suggested for the study area and elsewhere along  
623 the SWIR (e.g. Dick et al., 2019; Mével, 2003; Minshull et al., 1998, 2006), but rather a gradual  
624 transition from hydrated peridotites to unaltered peridotites. The smoother gradient of seismic  
625 velocities above 7 km/s in the EW profile suggests a more uniform upper mantle in comparison  
626 to the NS profile.

627 From Figure 8a, we further decipher the following: (i) Despite the similar penetration  
628 depth of serpentinization on both sides of individual detachment faults, for similar depths on  
629 both sides of the detachment, serpentinization is more pervasive in the hanging walls than in the  
630 footwalls. This is possibly due to the continuous exhumation of the footwalls that leaves hanging  
631 walls longer exposed to water at temperatures that are more optimal for the serpentinization to  
632 take place (e.g. *DI*, Fig. 8a); (ii) The depth and degree of serpentinization depends on the length  
633 of time the detachments were active. For example, amongst the abandoned detachments south of  
634 *DI*, the 20% serpentinization contour reaches deeper levels around faults *D3'* and *D3* than  
635 around *D5* and *D7*, which agrees with the estimated longer active life for faults *D3'* and *D3*  
636 (Table 1). Therefore, we suggest that the bulk of the serpentinization occurs while detachment  
637 faults are active, with much diminished alteration after the lithosphere migrates off-axis. The  
638 longer a particular detachment is active, the deeper serpentinization reaches due to both more  
639 extensive deformation of the fault walls resulting in greater porosity and permeability and longer  
640 exposure to seawater. The overall increase in velocity and, therefore, decrease in estimated  
641 degree of serpentinization from the active to abandoned detachment faults is possibly caused by  
642 a reduction in porosity by closure of cracks and fractures once fault activity stops and the fault  
643 moves away from the ridge axis.

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### 5.2.2 EW transition from exhumed mantle to volcanic seafloor

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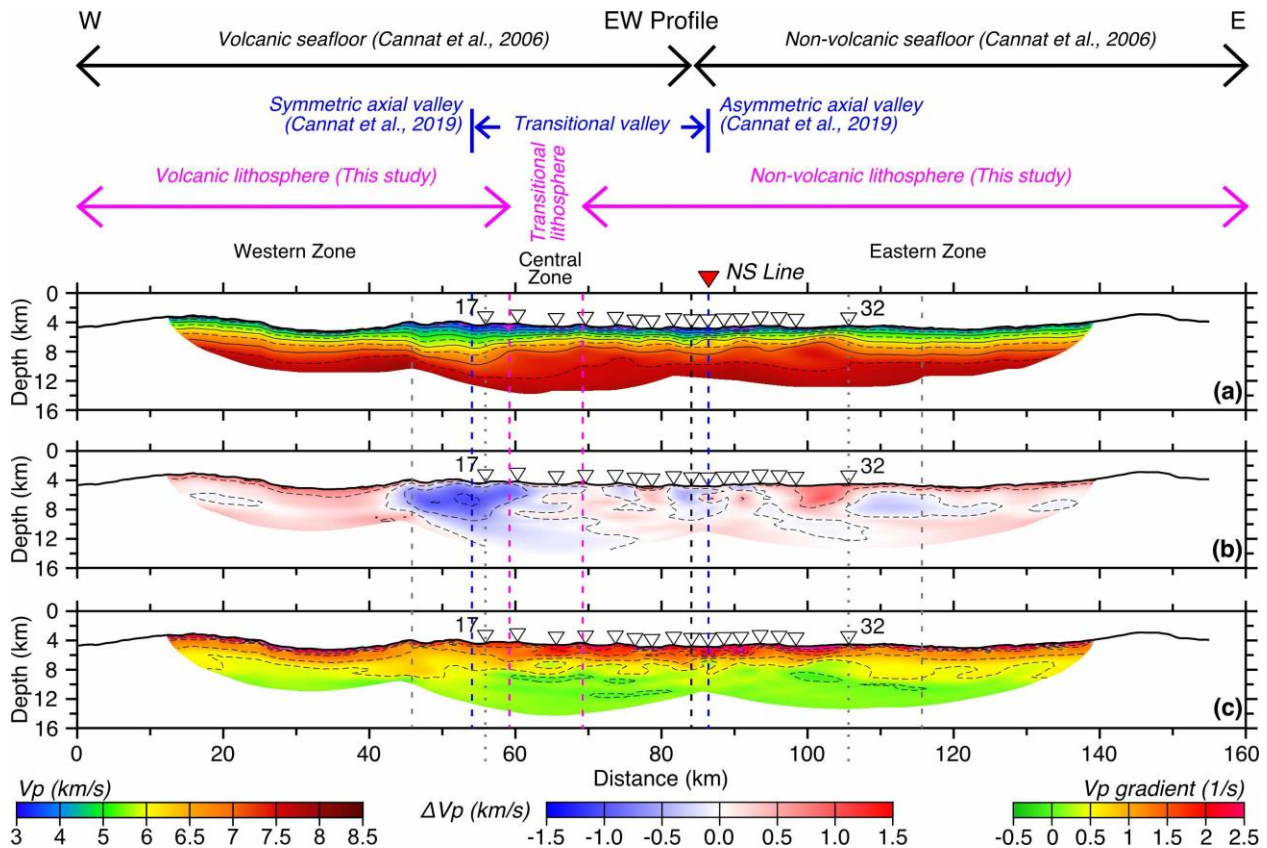
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Earlier work on seafloor mapping in the study area (Cannat et al., 2006) indicated that the EW profile crosses a transition within the uppermost lithosphere from exhumed and serpentinized ultramafic mantle domains in the east to volcanic and magmatic mafic domains in the west (Fig. 1). This transition was initially interpreted to occur at the OBS 25 location (model distance ~84 km; Fig. 1) (Cannat et al., 2006). Cannat et al. (2019) later proposed a wider transitional domain from the detachment-dominated asymmetric topography, characteristic of non-volcanic seafloor, to the magmatic symmetric topography, characteristic of volcanic seafloor. The authors examined four ridge-perpendicular bathymetric profiles spread over a distance of ~50 km and observed a change in the seafloor topography across the profiles from ridge-asymmetric topography, indicative of non-volcanic seafloor, to ridge-symmetric topography, indicative of volcanic seafloor. The distance between the profile showing asymmetric topography and the closest profile showing ridge-symmetric topography is 32.6 km. A profile showing transitional bathymetry is found in between these two profiles, which led the authors to interpret that the transitional region occurs over a 20-30 km distance.

Our velocity, velocity anomaly, and vertical velocity gradient models (Fig. 9) also suggest a major change in the lithospheric structure and composition along the profile. This change occurs over a distance of ~10 km between OBSs 18 and 20 (model distance ~59-69 km; Fig. 9). In Figure 9 we plot the limits of the interpreted volcanic and non-volcanic seafloor as proposed by Cannat et al. (2006, 2019) and in this study. The recent interpretation by Cannat et al. (2019) put the transitional area at model distance of ~54-86 km, which is wider but in broad agreement with the results from this work.



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667 **Figure 9.** Comparison of the (a) average velocity model from Figure 5d, (b) velocity anomalies  
 668 from Figure 6c, and (c) vertical gradient from Figure 6d, along the EW profile. Arrows in the top  
 669 of the figure show the different interpretations of the proposed transition from volcanic to non-  
 670 volcanic lithosphere: Cannat et al. (2006) in black, Cannat et al. (2019) in blue, and this study in  
 671 pink. The blue short straight lines indicate the location of Cannat et al. (2019) bathymetric  
 672 profiles that show contrasting symmetric and asymmetric axial valleys and constrain the  
 673 transitional zone. The black, blue, and pink dashed lines extend the limits of the different  
 674 interpretations across (a, b, and c). The pink arrows also delimit the three zones in the EW  
 675 model: western, central, and eastern. White and red inverted triangles as in Figure 3 caption.  
 676 Vertical dotted and dashed light grey lines as in Figure 5 caption.

677 The observed changes in velocity, velocity anomaly, and vertical velocity gradient divide  
 678 the EW profile into three distinctive zones: eastern (model distance >69 km), central (model  
 679 distance from 59 to 69 km), and western (model distance <59 km). The eastern zone shows high  
 680 and laterally variable seismic velocities at shallow depths, reaching 4–4.5 km/s within 0.5 km  
 681 dbsf and 7 km/s at depths 2–3.5 km dbsf (Fig. 9a), an overall positive velocity anomaly including  
 682 a larger positive anomaly (0.5–1 km/s) at model distances 99–106 km (Fig. 9b), and a high  
 683 vertical velocity gradient of 1–2 s<sup>-1</sup> (Fig. 9c). These characteristics are consistent with exhumed  
 684 and variably serpentinized mantle peridotites, where serpentinization extent diminishes as a  
 685 function of depth. A low in the velocities that corresponds with a low velocity anomaly (Fig. 9b)  
 686 and a more moderate vertical gradient than for the neighboring areas (Fig. 9c) is observed within  
 687 model distances 81–86 km (below OBSs 24–26; Fig. 9a). We speculate that this is indicative of a  
 688 ~5km-wide volcanic dike injection, a feature resolvable in our model (Fig. 4f), that could be  
 689 responsible for the lower RMBA values (10–0 mGal) that led to the interpretation of this area  
 690 and the area further west as volcanic seafloor (Cannat et al., 2006). Other smaller low velocity  
 691 anomalies, e.g. below OBSs 21 and 22 (model distances 73–76 km; Fig. 9b) and below OBS 28  
 692 (model distances 90–92 km; Fig. 9b) may suggest the presence of additional smaller dikes that  
 693 are not fully resolvable by our data. Detailed seafloor mapping in our study area with side-scan  
 694 sonar shows small sparse lava patches on top of the exhumation surfaces (Sauter et al., 2013).  
 695 Our results bolster the argument that the lava patches are erupted directly onto the exhumed  
 696 surface by small offset high-angle normal faults (Cannat et al., 2019; Sauter et al., 2013), as  
 697 opposed to being allochthonous rafted volcanic blocks transported to the surface off-axis by  
 698 successive detachments (Reston, 2018). Furthermore, the presented evidence of volcanic dike  
 699 injections within the “smooth-seafloor” favors the interpretation by Sauter et al. (2013) that the  
 700 abandonment of the active axial detachment fault and consequent activation of the successive  
 701 detachment fault may be a consequence of increased diking.

702 The western zone shows sharply lower seismic velocities that reach 3.5 km/s at 0.5 km  
 703 dbsf and 7 km/s at 4–5.3 km dbsf (Fig. 9a), a large negative velocity anomaly of -1.5–1 km/s  
 704 (Fig. 9b), and a moderate to low vertical velocity gradient of 0.5–1 s<sup>-1</sup> (Fig. 9c). These  
 705 characteristics are consistent with top of the lithosphere being partially constructed by mafic  
 706 magmatic rocks. This interpretation is further supported by analysis of the ray coverage or the  
 707 DWS (Fig. 5e). Although the ray density at all ends of the seismic profiles is gradually reduced  
 708 with increasing shot distance from the last OBS (Figs. 5c and 5e), the reduction at the west end  
 709 of the EW profile (Fig. 5f) is considerably greater than at the three other profile ends, which is  
 710 indicative of a major change in the nature of the lithosphere. This more rapid drop in ray density  
 711 also coincides in an apparent westward velocity increase at the western limit of the well resolved  
 712 area (Fig. 9a), but this is an artifact. The inverted velocity model follows the starting velocity

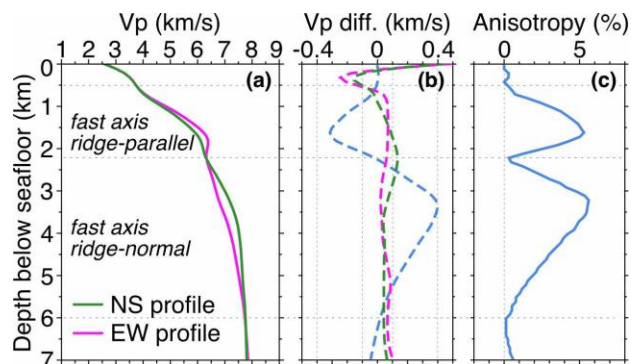
713 model in areas of low ray coverage (i.e. outside the well resolved area), and in the well resolved  
 714 section of the western zone seismic velocities are considerably slower than the starting velocity  
 715 model. This forces a gradual lateral change from the well resolved area. On the contrary, in the  
 716 east the starting and the average velocity models have similar velocities and no lateral change is  
 717 observed across the limit of the well resolved area.

718 In between the eastern and western zones is the central zone, which displays gradual  
 719 westward changes in the velocities (decreasing; Fig. 9a), velocity anomaly (from low positive to  
 720 low negative Fig. 9b), and vertical velocity gradient (vertically less variable, Fig. 9c). We  
 721 interpret this as indicative of a transition from the amagmatic upper lithosphere of the eastern  
 722 zone to the mafic magmatic rocks topping the lithosphere of the western zone and also likely an  
 723 indicator of a transitional lithosphere that is heterogeneous in its composition with layers of  
 724 mafic extrusive and intrusive rocks laterally intertwined with layers of fully and partially  
 725 serpentinized ultramafic rocks (Fig. 8b).

### 726 5.2.3 Anisotropy

727 We extract 1D velocity-depth functions at the crossing point between the NS and EW  
 728 velocity profiles (Fig. 10a) to evaluate if there is directional dependence in seismic velocities or  
 729 seismic anisotropy (Fig. 10). The difference between these two 1D velocity-depth functions is  
 730 shown in Figure 10b (dashed blue line), and the corresponding anisotropy is shown in Figure  
 731 10c. Both velocity functions are nearly coincident for the first 0.5 km dbsf. At greater depth,  
 732 from  $\sim 0.5$  to  $\sim 2.2$  km dbsf, velocities on the EW profile are faster (up to  $\sim 5\%$  difference) than on  
 733 the NS profile. The velocity relationship reverses from  $\sim 2.2$  to  $\sim 6$  km dbsf, with the NS profile  
 734 being faster (up to  $\sim 5\%$  difference). At depths greater than 6 km dbsf, the two velocity-depth  
 735 functions are again nearly coincident.

736 Figure 10b also compares the difference in the two velocity-depth functions to the  
 737 standard deviation (SD) estimated for the NS and EW velocity models at their crossing point.  
 738 The SD-depth functions are similar for both profiles at the crossing point, with most SD values  
 739 being  $< \pm 0.1$  km/s. The P-wave velocity difference near the function peaks reaches or approaches  
 740  $\pm 0.4$  km/s, which is at least several times greater than the SD of the both velocity models  
 741 indicating that the difference in velocities we observe falls well within the estimated uncertainty.  
 742 While some of the differences in velocities at the two crossing profiles may be due to the  
 743 limitations of modeling wave propagation in 2D, most of the observed differences can be  
 744 attributed to seismic anisotropy thus rendering our results a useful first-order approximation.



745



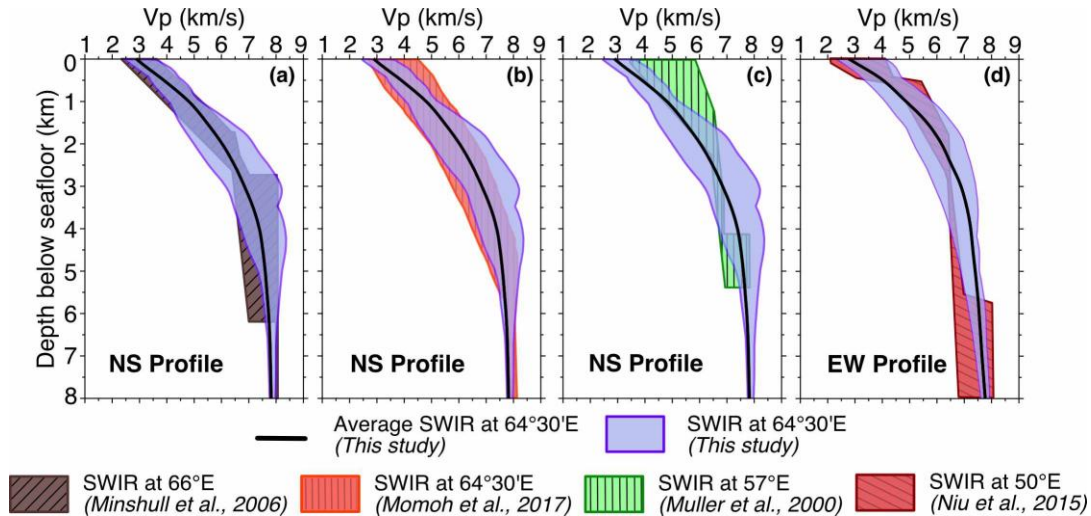
746 **Figure 10.** (a) Comparison of the 1D velocity-depth functions extracted from the NS (green) and  
 747 EW (pink) profiles at their crossing point. (b) Dashed blue line shows the difference between NS  
 748 and EW velocities presented in (a), and green and pink dashed lines show the standard deviation  
 749 of the velocities presented in (a) (from Figs. 5b and 5e). (c) Functional fit of % of anisotropy  
 750 with depth in light blue. The horizontal dashed grey lines in (a, b and c) indicate the depths for  
 751 the anisotropy consistent with a fast axis aligned in a ridge-parallel direction ( $\sim 0.5$  to  $\sim 2.2$  km  
 752 dbsf) and for the anisotropy consistent with a fast axis aligned in a ridge-normal direction ( $\sim 2.2$   
 753 to 6 km dbsf).

754 The velocity differences in the shallower zone ( $\sim 0.5$  to  $\sim 2.2$  km dbsf) are consistent with  
 755 the anisotropy reported at other ridges near the spreading axis of 1%–12% at shallow to  
 756 intermediate depths (0–3 km) and with the fast axis aligned in a ridge-parallel direction (e.g.  
 757 Seher et al., 2010; Weekly et al., 2014). This anisotropy is associated with the alignment of  
 758 vertical cracks within the crust in the ridge axis direction (Dunn & Toomey, 2001). Christeson et  
 759 al. (2019) synthesized the oceanic crustal structure formed at spreading ridges with half-  
 760 spreading velocities greater than 5 mm/yr from 2D seismic profiles and documented that  
 761 anisotropy may be restricted to the upper crust in areas near the ridge axis. Cracks induced by  
 762 stresses related to the footwall exhumation and bending at shallow to intermediate levels both in  
 763 the footwall and the hanging wall, as well as extensive extensional damage on the hanging wall,  
 764 have been suggested in our study area (Cannat et al., 2019; Momoh et al., 2017, 2020).  
 765 Therefore, we suggest the uppermost anisotropy is due to the preferential distribution of cracks  
 766 parallel or subparallel to the axis at depths from  $\sim 0.5$  to  $\sim 2.2$  km dbsf. We expect that the top 0.5  
 767 km dbsf are also characterized by axis parallel or subparallel cracks and the resulting anisotropy  
 768 but, due to the discussed limitations of our data and velocity models, this anisotropy was not  
 769 possible to resolve.

770 With greater depth, the increasing lithospheric pressure gradually closes the cracks thus  
 771 removing the source of the ridge-parallel anisotropy. Serpentinization also diminishes with  
 772 increasing depth leading to increased ratio of olivine minerals versus serpentine or other  
 773 alteration minerals, thus generating anisotropy with the ridge-normal fast direction that starts to  
 774 prevail at  $\sim 2.2$  km dbsf. Our data can resolve the ridge-normal fast anisotropy to 6 km dbsf, after  
 775 which the model resolution is insufficient for this purpose. Velocities from 6– $\sim 7$  km dbsf have  
 776 diminished resolution and are already influenced by the starting velocity model through  
 777 smoothing. This uppermost mantle, ridge-normal anisotropy is related to the lattice-preference  
 778 orientation of olivine minerals in the direction of lithospheric strain (Hess 1964), which is  
 779 consistent with the near-orthogonal spreading direction attributed to our study area (Cannat et al.,  
 780 2008), and has been reported for older oceanic crust (e.g. Ismail and Mainprice, 1998;  
 781 VanderBeek et al., 2016).

### 782 5.3 Comparison of velocity-depth fields

783 We compare the average 1D velocities and velocity field envelopes of the NS and EW  
 784 profiles (Fig. S9) with corresponding results from previously published ridge-normal and ridge-  
 785 parallel seismic profiles, respectively, at other locations along the SWIR.



786

787 **Figure 11.** Comparison of the average 1D velocity-depth profiles (thick black line) and 1D  
 788 velocity-depth fields (light purple areas bounded by dark purple lines) from this study with the  
 789 velocity fields from earlier studies of the SWIR at the following locations: (a) 66°E (Minshull et  
 790 al., 2006); (b) 64°30'E (Momoh et al., 2017); (c) 57°E (Muller et al., 2000); and (d) 50°E (Niu et  
 791 al., 2015). Note that the locations of the earlier work done at 66°, 57° and 50°E are shown in the  
 792 Figure 1 inset with yellow circle, diamond and star, respectively, while the work done at 64°30'E  
 793 coincides with our study area.

794 Figure 11 displays a comparison of our results at 64°30'E with velocity fields from earlier  
 795 work at the SWIR. From East to West, shown are velocity fields at 66°E (Minshull et al., 2006)  
 796 (Fig. 11a), 64°30'E (Momoh et al., 2017) (Fig. 11b), 57°E (Muller et al., 2000) (Fig. 11c), and  
 797 50°E (Niu et al., 2015) (Fig. 11d).

798 Our NS profile velocity envelope and that of Momoh et al. (2017) (Fig. 11b), both from  
 799 64°30'E, are mostly in general agreement. Momoh et al. (2017) shows little structure with  
 800 velocities smoothly increasing with depth, from  $\sim 2.7$  km/s at the seafloor to  $\sim 7$ – $8$   
 801 km/s at 5 km dbfs. The NS velocity model constrains the velocities at the seafloor to a narrower  
 802 and slower range ( $\sim 2.5$ – $3.7$  km/s), increasing to a similar range by about 1.4 km dbfs, and  
 803 becoming overall higher for depths up to  $\sim 5$  km dbfs. Momoh et al.'s (2017) velocities show  
 804 smooth vertical velocity gradient changes with depth, while the NS velocities indicate a more  
 805 complex structure for how the velocity gradient changes with depth and show a higher vertical  
 806 velocity gradient at the top that it is reduced to a lower velocity gradient for depths greater than  
 807  $\sim 2$  km dbfs. While our interpretation is broadly similar to that of Momoh et al. (2017), the NS  
 808 velocity model has recovered deeper and more detailed velocity information thus providing more  
 809 constraints on the lithospheric structure. This is likely because the NS profile is longer (150 km  
 810 vs. 43 km) and has more OBSs (16 vs. 8) than the profiles used by Momoh et al. (2017),  
 811 resulting in denser and deeper ray coverage with a larger range of source-receiver offsets.

812 Minshull et al. (2006) at 66°E (Fig. 11a) show velocities at the seafloor ranging from  
 813  $\sim 2.3$ – $3.5$  km/s and increasing at a high vertical velocity gradient up to  $\sim 6.4$ – $7$  km/s at  $\sim 2.7$  km  
 814 dbfs. At depths greater than  $\sim 2.7$  km dbfs, the velocities increase at a low-velocity gradient and  
 815 range from  $\sim 6.5$ – $7$  to 8 km/s. The average velocities for the NS profile fit well with the velocity  
 816 field of Minshull et al. (2006). The NS velocity field also presents a similar range and vertical  
 817 gradient for depths up to 1 km dbfs but it has a lower low and a higher high velocity for depths

818 of 1–3.5 km dbf. At greater depths, the NS profile velocities are overall higher than the  
819 velocities at 66°E. Despite the significant similarities in velocities between the two models, large  
820 differences exist on the lithospheric structure interpretation in these two study areas that are only  
821 ~150–200 km apart. Minshull et al. (2006) suggested a crustal structure composed of mafic  
822 oceanic layers 2 and 3, with a mean crustal thickness of 4.2 km, and constrained the Moho with  
823 *PmP* arrivals and complementary gravity data. They interpreted that serpentinized peridotites do  
824 not form the dominant lithology in the seismic lower crust and instead they suggested a Layer 3  
825 with a variable thickness (0.5–3 km) governed by melt focusing toward segment centers.

826 The presence of Segment-8 volcano at ~65°40'E (Cannat et al., 2006; Schindwein &  
827 Schmid, 2016) could explain some of the differences between the two results and interpretations.  
828 Velocity structure of Minshull et al. (2006) shows a higher gradient for the top and lower  
829 gradient for the bottom of the model than the NS profile, which is consistent with a high-gradient  
830 Layer 2 on top of a low-gradient Layer 3, while the SWIR at 64°30'E has a gradual decrease in  
831 the vertical velocity gradient consistent with a gradual decrease in serpentinization and pore  
832 pressure with depth. However, Minshull et al. (2006) used a layered modeling and inversion  
833 procedure constrained by model parametrization of two crustal layers. The wide-angle data were  
834 collected on 8 OBSs (~10–30 km spacing) with no coincident MCS data to guide the layered  
835 inversion. Therefore, we speculate that a fair amount of variably serpentinized peridotites may be  
836 present in the subsurface at the SWIR at 66°E based on the overlap of the NS profile velocity  
837 envelope with Minshull et al. (2006) velocities, and that a denser seismic survey followed by  
838 first arrival traveltimes tomography for a single model layer would provide a better and more  
839 detailed comparison.

840 The SWIR at 57°E (Fig. 11c), across the Atlantic Bank, shows velocities ranging from  
841 3.5 km/s to ~5.8 km/s at the seafloor (Muller et al., 2000). This velocity range gradually narrows  
842 to 6.4–6.9 km/s for depths ~2.2–4.1 km dbf as the vertical velocity gradient decreases. For  
843 depths greater than 4.1 km dbf the velocities range from ~6.8–7.8 km/s. The authors interpreted  
844 magmatic oceanic crust composed of layers 2 and 3 on top of the uppermost mantle rocks. The  
845 NS profile velocities are overall in disagreement showing slower velocities for depths up to ~2.6  
846 km dbf and higher velocities for greater depths. We interpret the topmost ~2.6 km dbf on the  
847 NS profile, characterized by lower velocities than those from Muller et al. (2000), to represent  
848 fully serpentinized and highly fractured peridotites at the top that gradually transition to ~40%  
849 serpentinized peridotites with minor fractures at the bottom. The NS profile velocities continue  
850 to increase with increasing depth, though at a reduced gradient, due to continued fracture closure  
851 and further diminishing serpentinization. The Moho at the SWIR at 57°E is constrained by *PmP*  
852 reflections and gravity modeling, in contrast to our study area where the amagmatic seismic crust  
853 lacks any shreds of evidence of a Moho. However, the results from Muller et al. (2000) likely  
854 have greater uncertainties than our results since they use only 8 OBSs (~5–15 km spacing) and  
855 do not have coincident reflection data, which are used to guide layered modeling and inversion.

856 The velocity field at the segment center at 50°E (Fig. 11d, Niu et al., 2015) can be  
857 divided in 4 sections: (1) top one (0–0.6 km dbf), with a mostly moderate gradient and  
858 velocities ranging from ~2 km/s to ~4 km/s; (2) a high gradient section underneath (0.6–2.8 km  
859 dbf) with velocities of ~4.4–6.5 km/s; (3) a low gradient section (2.8–5.5 km dbf) further down  
860 with a narrow range of velocities (6.5–7 km/s); and (4) a low gradient section (depth > 5.5 km  
861 dbf) at the bottom with velocities ranging from 6.6 km/s to 8.1 km/s. The NS profile velocities  
862 are mostly in agreement within the top section, lower than velocities of Niu et al. (2015) in the

863 top part and in agreement with the bottom part of the second section, are significantly higher in  
 864 the third section, and somewhat higher in the fourth section. Niu et al. (2015) interpreted layers  
 865 2A, 2B, and 3 overlying the uppermost mantle. The investigated segment center at 50°E has an  
 866 anomalously thick crust, ranging from ~5.5 km to 10.2 km, and a robust and well-imaged axial  
 867 magma chamber (Niu et al., 2015; Jian et al., 2016, 2017). This interpretation is consistent with a  
 868 magma-rich spreading center and, as such, it differs from magma-starved spreading interpreted at  
 869 64°30'E.

## 870 **6 Conclusions**

871 We constrain tomographic velocity structure across- and along-axis the ultraslow-  
 872 spreading SWIR at 64°30'E by inverting first arrivals from two ~150-km-long orthogonal wide-  
 873 angle OBS profiles. Our major findings and conclusions are the following:

- 874 1. About 75% of the investigated uppermost lithosphere appears to be composed of highly  
 875 fractured and fully serpentinized peridotites at the top, with a gradual decrease in pore  
 876 space volume and serpentinization degree to unaltered peridotites at depth. Key evidence  
 877 for this are seismic velocities that increase rapidly with depth, changing from 3–4 km/s at  
 878 the seafloor to 7 km/s at depths ranging from 1.5–6 km dbfs, much lower vertical velocity  
 879 gradient for velocities >7 km/s that gradually reduces with increasing depth, and no  
 880 distinct and seismically well-characterized Moho observable in the data.
- 881 2. A system of detachment faults is imaged in the subsurface for the first time in our study  
 882 area by the profile that crosses the ridge axis. We interpret a sharp lateral change  
 883 (horizontal gradient  $\sim 1 \text{ s}^{-1}$ ) in velocities, switch in polarity of the largest velocity  
 884 anomaly, from 1.5 km/s to -1.5 km/s, and high vertical gradient ( $\sim 2 \text{ s}^{-1}$ ) in the velocities  
 885 coincident with the shallow section (top 2 km) of the highest topographic feature as the  
 886 seismic expression of an active axial detachment fault. Five abandoned detachment faults  
 887 are also identified based on anomalously high velocities and vertical gradients that  
 888 characterize the footwalls on all identified detachment faults, with the opposite  
 889 characterizing the hanging walls. Serpentinization across-axis is controlled by the  
 890 longevity of detachments and fault block movement, with longer-lived faults showing  
 891 deeper serpentinization extent and footwalls showing less pervasive serpentinization due  
 892 to continuous exhumation.
- 893 3. Comparison of the 1D velocity-depth functions at their crossing point between the two  
 894 orthogonal profiles suggests an up to 5% ridge-parallel fast-axis anisotropy from ~0.5 to  
 895 ~2.2 km dbfs attributed to the preferential distribution of cracks parallel to the ridge, and  
 896 a similar magnitude but reversed polarity ridge-normal fast-axis anisotropy at >~2.2 km  
 897 dbfs that we attribute to the lattice-preference orientation of olivine minerals in the less  
 898 serpentinized peridotites.
- 899 4. Upper lithospheric composition along the western half of the axis-parallel profile seems  
 900 to transition over a distance of ~10 km from the variably serpentinized peridotites  
 901 domains in the East to predominately mafic magmatic domains in the West. The  
 902 exhumed mantle domains show high and laterally variable seismic velocities at shallow  
 903 depths, reaching 4–4.5 km/s within the first 0.5 km dbfs and 7 km/s at depths 2–3.5 km  
 904 dbfs, a high vertical velocity gradient of  $1\text{--}2 \text{ s}^{-1}$ , and an overall positive velocity anomaly  
 905 (up to 0.5-1 km/s). The mafic domains show sharply lower seismic velocities that reach

906 3.5 km/s at 0.5 km dbf and 7 km/s only at 4–5.3 km dbf, a moderate to low and  
 907 smoother vertical velocity gradient of  $0.5\text{--}1\text{ s}^{-1}$ , and a large negative velocity anomaly of  
 908  $-1.5\text{--}1\text{ km/s}$ . We suggest that the change in the seafloor accretion mode is propelled by a  
 909 westward increase in melt supply.

910 5. Comparison of velocity structure from our results at the ultraslow-spreading SWIR at  
 911  $64^{\circ}30'E$  with the velocity structure elsewhere at the SWIR has shown an overall  
 912 agreement of our results with studies at amagmatic segments, a disagreement with  
 913 identified anomalously magma-rich segments, and mixed results for magma-starved  
 914 segments, such as the SWIR at  $66^{\circ}$  and  $57^{\circ}$  E. This work has also demonstrated the  
 915 challenges in carrying out meaningful velocity structure comparisons when the data  
 916 resolution and techniques used to determine the velocities are not the same (e.g. layered  
 917 modeling vs. tomography).

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 931 Tools (GMT) software (Wessel and Smith, 1998).

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