Seismic evidence for uniform crustal accretion along slow-spreading ridges in the equatorial Atlantic Ocean

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1 **Seismic evidence for uniform crustal accretion along slow-spreading**
2 **ridges in the equatorial Atlantic Ocean**

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6
7 The crustal accretion along mid-ocean ridges is known to be spreading-rate
8 dependent. Along fast-spreading ridges, two-dimensional sheet-like mantle
9 upwelling creates relatively uniform crust. In contrast, the crust formed along slow-
10 spreading ridges shows large along-axis thickness variations with thicker crust at
11 segment centres, which is hypothesised to be due a three-dimensional plume-like
12 mantle upwelling with melt focusing to segment centres. Using wide-angle seismic
13 data acquired from the equatorial Atlantic Ocean, we show that the crustal
14 thickness is nearly uniform (~5.5 km) across five segments for crust formed at the
15 slow-spreading Mid-Atlantic Ridge with age varying from 8 to 70 Ma. The crustal
16 velocities indicate that this crust is predominantly of magmatic origin. We suggest
17 that this uniform magmatic crustal accretion is due to a two-dimensional sheet-like
18 mantle upwelling, contradicting previous hypothesis. Our results also indicate that
19 the lateral extend of magmatic accretion could be used to define the sheet-like
20 mantle upwelling regions beneath the global ridge system.

The oceanic crust covers ~60% of the Earth's surface and is continuously generated along the ~65,000 km-long divergent plate boundaries at the Mid-Oceanic Ridges (MORs)¹. The MORs are partitioned into tens to hundreds kilometre-long first-order segments by oceanic transform faults (TFs)¹. Between two oceanic TFs, these ridge segments are further divided into smaller ridge segments (second-order segments) by non-transform offsets (NTOs) and overlapping spreading centres (OSCs)¹.

The crustal accretion along MORs and the resulting along-axis crustal thickness variation are known to be spreading-rate dependent. At the fast-spreading ridges, such as the East-Pacific Rise (EPR), the oceanic crust exhibits relatively uniform thickness with a modest crustal thinning (≤ 1.6 km; orange triangles in Fig. 1) at oceanic TFs, NTOs and OSCs²⁻⁶, interpreted to result from a two-dimensional (2-D) sheet-like mantle upwelling beneath these ridges³ and/or rapid ductile flow of hot lower crust along these ridges⁷. In contrast, the studies using gravity data^{3, 8-11} collected along the slow-spreading Mid-Atlantic Ridge (MAR) reveal systematically significant reduction (up to 50%) in crustal thickness from segment centres to the associated distal ends at TFs and NTOs. Crustal thickness estimation using active-source seismic data also shows large and systematic along-axis crustal thickness variations at MAR, where the crust at distal ends could be up to ~3.2-5.7 km thinner than that at the segment centres^{9, 12-18} (orange dots in Fig. 1). All these observations demonstrate focused magmatic crustal accretions along the slow-spreading ridges, which are interpreted to result from either a three-dimensional (3-D) plume-like mantle upwelling^{3, 8} or by a 3-D melt migration to the segment centres at the base of the lithosphere beneath slow-spreading ridges^{19, 20}. However, these previous studies on MAR are generally concentrated within or close to the spreading centres, sampling only young (≤ 2 Ma) oceanic crust. In contrast, old oceanic crusts formed along MAR show different patterns of segment-scale crustal thickness variation. For example, although large segment-scale crustal thickness variations are observed for ~5 and ~65 Ma old crust in the North Atlantic Ocean^{17, 21, 22}, the crustal thinning is only ~2.8-3.4 km (blue dots in Fig. 1), smaller than those observed along or near ridge axis. For much older crust (~130-150 Ma) in the North Atlantic Ocean, the crustal thinning is very modest^{23, 24}. A recent study

from the South Atlantic Ocean over 6.6-61.2 Ma crust at 31°S shows very little variations in crustal thickness over 40-60 km lateral distance²⁵.

The plate separation at the slow-spreading ridges is accommodated by two contrasting modes of accretion. For the regions where the melt supply is robust and sufficient, the magmatic accretion creates thick igneous crust. In contrast, for the regions where melt supply is limited, the plate separation is mainly accommodated through tectonic slips along low-angle detachment faults, emplacing lower crustal and upper mantle rocks to the seafloor forming oceanic core complexes (OCCs) and thin crust²⁶.

Crustal P-wave velocity (V_p) obtained using wide-angle seismic data can be used to discriminate between these two different modes of crustal accretion. For example, the magmatically accreted crust can be divided into two layers based on their velocities^{27, 28}. The upper crust (Layer 2) is characterised by a relatively low V_p (4.1-6.5 km/s) but a high vertical velocity gradient ($\sim 1\text{-}2\text{ s}^{-1}$) whereas the lower crust (Layer 3) has a high V_p (6.5-7.1 km/s) but a significantly reduced velocity gradient ($\sim 0.1\text{-}0.2\text{ s}^{-1}$). This seismic structure has been designated as the 'Penrose model' which equates these layers to an upper crust composed of extrusive basalts and sheeted dyke overlying a gabbroic lower crust¹. In contrast, the crustal V_p below the OCCs increases rapidly to $>6.5\text{-}7.0$ km/s within $\sim 1.0\text{-}2.5$ km depth below the seafloor^{22, 29-33}, suggesting the presence of lower crustal gabbroic rocks or partially serpentinised peridotite exhumed to shallow depths. However, any interpretation about the crustal structure based only on the V_p model is poorly constrained because of the ambiguity in velocity-lithology relationship. As the V_p increases from $\sim 4.5\text{-}5.0$ km/s for a completely serpentinised peridotite to ~ 8.0 km/s for an unaltered peridotite^{34, 35}, there is an overlap with the V_p (4.1-7.1 km/s) for magmatic oceanic crust²⁷. On the other hand, laboratory measured velocities of rocks from oceanic drillings show that the magmatic crustal rocks (basalt and gabbro) generally have P and S-wave (V_s) velocity (V_p/V_s) ratios <1.9 at the crustal pressure condition^{35, 36} while the V_p/V_s ratio of serpentinised mantle rocks varies from ~ 1.78 to ~ 2.21 with increasing

81 degree of alteration³⁴. Therefore, the Vp/Vs ratio is a useful physical parameter for
82 constraining the crustal composition³⁷, hence the thickness of the crust.

83
84 Here, we estimate the Vp and Vp/Vs ratio of five segments of the slow-spreading crust
85 covering ~8-70 Ma lithosphere in the equatorial Atlantic Ocean (Figs. 2a) using an
86 active-source seismic refraction dataset to characterize the nature of the oceanic crust and
87 to shed light on the crustal accretion and mantle upwelling processes along slow-
88 spreading ridges.

89
90 The equatorial Atlantic Ocean started opening ~100-140 Ma^{38, 39}. The average half-
91 spreading rate varied several times in the past between 15 and 25 mm/yr, and the current
92 half-spreading rate is ~16 mm/yr⁴⁰ (Fig. 2b), hence the equatorial MAR can be classified
93 as a slow-spreading ridge⁴¹. In this region, three east-west striking mega-transforms (St.
94 Paul, Romanche and Chain TFs) offset the MAR by a total of ~1,800 km (Fig. 2a). The
95 St. Paul TF system encompasses four TFs interrupted by three intra-transform ridge
96 segments, generating a total offset of ~600 km⁴². Away from its eastern ridge-transform
97 intersections (RTI), the traces of the four TFs persist up to ~20 Ma of seafloor and then
98 only two traces of the TFs can be identified on the older seafloor⁴³. Our seismic profile
99 transects the St. Paul fracture zone (FZ) where the crustal age is ~70 Ma in the north and
100 ~40 Ma in the south. The Romanche TF offsets the MAR by ~880 km⁴⁴. The portion of
101 the Romanche TF crossed by our seismic profile is characterized by a ~6 km-deep and
102 ~40 km-wide valley with bounding walls shallowing to <3 km depth below sea level. The
103 age contrast along our profile is ~32 Ma across the Romanche TF, with crustal age of ~40
104 Ma in the north and ~8 Ma in the south. Our seismic profile crosses the Chain FZ at ~160
105 km west of the western Chain RTI, where the crustal age is ~10 Ma in the north and ~24
106 Ma in the south. Along our profile, the Chain FZ is characterized by a ~10 km-wide
107 sedimentary basin, which is bounded by a transverse ridge to the south and shows gradual
108 seafloor shallowing to the north⁴⁵. An oblique pseudofault is observed between the Chain

and Charcot FZs, intersecting the Chain FZ at ~46 Ma to the west and NTO within the ~17 Ma⁴⁵, showing a typical half V-shaped feature on the seafloor.

The seismic data used in this study was acquired in 2018 during the ILAB-SPARC experiment^{42, 44, 45}. The ~855 km-long seismic profile is nearly parallel to the ridge axis in an approximately N-S direction, starting just north of the Charcot FZ in the south and extending to ~230 km distance north of the St. Paul FZ (Fig. 2a). Fifty OBSs were deployed along the ~700 km part of the profile at average 14.2 km interval. An airgun-array of 16 guns with a total volume of 4990 cubic inches was towed at 10 m depth below sea level and was fired every 300 m. Both P and S-wave arrivals were hand-picked on the pressure data recorded by hydrophones (Supplementary Figs. 1 and 2) and were inverted using a 2-D ray-based travel time tomographic method^{2, 46} to obtain Vp and Vs models (Figs. 3a,b; See Methods). Then the crustal and upper mantle Vp/Vs ratios (Fig. 3c) were calculated using the inverted Vp and Vs models. Using these results, we first determine the predominant mechanism of crustal formation for each crustal segment and then analyse the segment-scale crustal thickness variations to understand the crustal accretion process along the slow-spreading MAR in the equatorial region.

Results

Magmatically accreted crust versus tectonically controlled crust

To determine the dominant mode of crustal accretion, we use a Vp/Vs ratio of 1.9 as an indicator³⁷ to discriminate between the magmatic crust (Vp/Vs<1.9) and the serpentinised peridotite (Vp/Vs>1.9) at crustal depth. The tomographic results show two different types of crustal Vp structures for the five ridge segments characterized by distinct crustal Vp/Vs ratios (Fig 3), which can be interpreted resulting from two different modes of crustal accretion: magmatic accretion and tectonic accretion along detachment fault.

The oceanic crust within the Segments 1, 2, 3-S, 4 and 5 exhibits a two-layered Vp structure (Figs. 3a and 4a) characterized by distinct vertical Vp gradients. Using the vertical Vp gradient of 0.5 s^{-1} to define the boundary between the upper and lower crust⁴⁷, the upper crust is ~1.9-2.3 km thick showing high vertical Vp gradients of $\sim 0.66\text{-}0.80 \text{ s}^{-1}$ (Supplementary Table 1). Beneath the upper crust, the Vp in the lower crust, which is ~3.1-3.5 km thick, increases at a much-reduced velocity gradient of $\sim 0.13\text{-}0.17 \text{ s}^{-1}$ (Supplementary Table 1). The maximum crustal Vp within these segments is generally $<7.2 \text{ km/s}$ (Figs. 3a and 4a). These crustal Vp structures fall in the velocity range typical for older slow-spreading magmatic oceanic crust^{27, 28} (Fig. 4a). These crustal segments are generally characterized by crustal Vp/Vs ratios less than 1.9 (Fig. 3c), indicative of basaltic and gabbroic rocks in the crust³⁷. The Moho, the crust/mantle boundary, is well constrained using Moho wide-angle reflections. A rapid increase from crustal velocities to mantle velocities across the Moho is observed throughout these ridge segments, where the Vp increases abruptly by $\sim 0.7\text{-}0.9 \text{ km/s}$ across the Moho (Figs. 3a and 4a). Both the Vp and Vp/Vs ratio models support that the observed two-layered crustal velocity structure can be explained by the ‘Penrose’ model¹ indicating a predominantly magmatic accretion of crust throughout these ridge segments.

In contrast, the Segment 3-N between 340 and 400 km profile distance immediately south of the Romanche TF shows a rapid increase in the crustal Vp of $\sim 7.7 \text{ km/s}$ at $\sim 2.2 \text{ km}$ sub-basement depth (Figs. 3a and 4b). This anomalously high Vp exceeds the typical velocity for the normal oceanic crust ($<7.2 \text{ km/s}$ ^{27, 28}). Conversely, the upper crustal Vs within Segment 3-N shows a similar velocity-depth variation as that within Segment 3-S. The crustal Vp/Vs ratios are larger than 1.9 in the upper crust (Fig. 3c), suggesting the presence of serpentinised peridotite, indicative of predominantly tectonic extensions³⁷. The rapid increase in the crustal Vp with depth can be interpreted as a gradual decrease in mantle serpentinisation. Assuming a Vp of 8.0 km/s and 5.0 km/s for 0% and 100% serpentinised peridotite^{34, 35}, respectively, the observed $\sim 7.7 \text{ km/s}$ velocities represent $\sim 10\%$ of mantle serpentinisation at $\sim 2.2 \text{ km}$ sub-basement depth. A recent petrological study of dredged rocks from the eastern Romanche RTI region identified fragments of old OCCs

on the south flank of the Romanche TF⁴⁸. This is consistent with our interpretation that immediately south of the Romanche TF, the plate separation is mainly accommodated by tectonic slip along detachment faults.

Intra-segment crustal thickness variation within magmatic segments

The thickness of the igneous crust is defined as the thickness between the top basement and the seismically constrained Moho. Based on the crustal Vp and Vp/Vs structures, Segments 1, 2, 3-S, 4 and 5 (Fig. 3) are interpreted as consisting of magmatic crust. The average crustal thicknesses within these magmatic segments are ~5.4-5.6 km (Fig. 5), consistent with some previous estimates in this region^{42, 44, 45}. The average crustal thicknesses of the five segments are ~500-700 m thinner than the global average of the old (>7.5 Ma) slow-spreading crust but still fall into the standard deviation (6.1 ± 1.0 km²⁷). The oceanic crust shows a slight thinning at FZ, TF and pseudofault regions, which is confined in a narrow zone in the vicinity of the trough of these tectonic discontinuities (Fig. 5 and blue squares in Fig. 1). At the southern end of Segment 1, the crust thins by ~600 m at the centre of the northern St. Paul FZ valley over <5 km distance and the average crustal thickness within the St. Paul FZ is only ~200-400 m thinner than the average thickness of the adjacent segments⁴² (Figs. 5a,b). The crust within the ~20 km-wide Romanche transform valley is ~5.1 km thick on average, which is only ~500 m thinner than the average of Segment 2 to the north (Fig. 5b). Though the crustal Vp/Vs ratio is not constrained, Gregory et al.⁴⁴ argued the thick crust at the Romanche transform zone is likely composed of fractured mafic rocks, and hence can support the supershear rupture during the 2016 M_w 7.1 earthquake⁴⁹. The crust beneath the ~10 km-wide Chain FZ trough has an average thickness of ~4.7 km, which is ~700-800 m thinner than the averages of the adjacent Segments 3-S and 4 (Figs. 5c,d). The thinnest crust within the Chain FZ trough is ~4.1 km, which is ~500 m thinner than that measured in ref. (45). The average crustal thickness within the pseudo-fault trough is ~5.2 km with the thinnest crust of ~4.8 km thick, which is ~200-400 m thinner than the adjacent segments (Fig. 5e). The crustal Vp and Vs (Figs. 3a,b) are lower at the FZ, TF and pseudofault regions relative to the surrounding crust, which could be due to the presence of faults and fluids.

196

197 We take the standard deviation of mean crustal thickness as a measure of the intra-
198 segment crustal thickness variation^{13, 14, 25, 50, 51}. The standard deviations of the mean
199 crustal thickness within these five segments are 0.1-0.4 km (Fig. 5), which is much
200 smaller than those (≥ 1.0 km^{13, 14}) for other MAR segments where large along-axis
201 crustal thinning is observed. However, they are very close to those (≤ 0.3 km^{50, 51})
202 observed along the fast-spreading EPR, suggesting relatively uniform crustal thickness
203 within these segments. Though the locations of NTOs are difficult to identify over the old
204 crust in this region, there is no evidence supporting thicker crust at the segment centre
205 and >2.8 km gradual crustal thinning towards the tectonic discontinuities within any of
206 the five segments. The relatively low crustal Vp at the FZ, TF and pseudofault regions
207 (Fig. 3a) indicates the thick crust at these tectonic discontinuities is not due to the
208 inherent positive velocity-Moho depth trade-off in the travel time tomography⁴⁶. A
209 Monte-Carlo based uncertainty analysis yields the variance in the Moho depth of <330 m
210 (see Methods; Supplementary Fig. 5b), demonstrating that the observed uniform crust
211 does not result from a particular starting Vp model used in tomography. The
212 checkerboard tests (see Methods; Supplementary Figs. 6 and 7) demonstrate that the used
213 tomography method can recover ~ 1.5 - 2.0 km crustal thickness variations when the real
214 variations in crustal thickness are ~ 2.5 - 3.0 km over ~ 70 km lateral distance. These tests
215 demonstrate that the observed uniform crust within these five ridge segments is real.

216

217 **Discussions**

218 Most of previous studies using gravity and seismic data collected on the slow-spreading
219 MAR reveal systematic and significant along-axis reduction in crustal thickness from the
220 centre of segments towards the oceanic TFs, FZs and NTOs^{3, 8-18, 21, 22}, suggesting a 3-D
221 plume-like mantle upwelling^{3, 8} and/or a focused melt concentration to segment centres^{19,}
222 ²⁰ beneath the slow-spreading ridges. However, the five segments with predominantly
223 magmatically accreted crust in the equatorial Atlantic Ocean show slightly thin crust with
224 little intra-segment crustal thickness variation throughout each segment (Figs. 1 and 5),

which is remarkably different from the most previous observations in the Atlantic Ocean but similar to some recent observations on 6.6-61.2 Ma oceanic crust at 31°S in the South Atlantic Ocean²⁵. The crust within these segments is either (1) originally formed uniformly at the ridge axis, or (2) has been modified after crustal formation by tectonic extension/stretching during the amagmatic period, or (3) by a second-stage crustal accretion at RTIs. Here, we analyse all these hypotheses and propose the best mechanism that explains the formation of uniformly thick crust in the equatorial Atlantic Ocean.

The tectonic extension and stretching account for ~10% of plate separation at the slow-spreading ridges^{52, 53}, which could thin the oceanic crust. The fault spacing and heave are generally larger at segment ends than at segment centres of slow-spreading ridges, indicating more tectonic extension should occur at segment ends due to the decreasing magma supply⁵⁴. If the focused mantle upwelling and melt supply did occur at the time of crustal formation, the tectonic extension and stretching would increase the magnitude of the intra-segment crustal thinning^{10, 54} within these five ridge segments, rather than diminish this trend. Furthermore, numerical modelling shows that the gravity anomaly caused by tectonic faulting increases with increasing fault throw and spacing⁵⁵, further supporting the idea that the tectonic extension and stretching will enhance the along-axis variation in crustal thickness. All the evidence demonstrates that the tectonic extension and stretching cannot explain the uniform crust observed along our profile.

Some recent studies indicate that the crust at oceanic TF can get augmented by a second stage of magmatic accretion at RTI^{45, 56}. This second-stage crustal accretion at RTIs could laterally emplace dykes at the segment ends and into the transform valley⁴⁵, forming J-shaped structures on the seafloor and thickening the crust. However, the gravity data reveal a ~5 km-thick crust within the Chain transform zone⁵⁷, which is comparable with the average ~5.1 km-thick crust within the Chain FZ trough along our profile (Fig. 5c). Although we do not have any information about the crustal thickness for the Romanche FZ, the crust within the Romanche TF has similar thickness to those within St. Paul and

Chain FZs along our profile (Fig 5). These observations suggest the second-stage magmatic accretion may have limited augmentation to the crust at RTIs. Furthermore, the J-shaped structures have very short extensions over the old ocean floor⁵⁶, suggesting the second-stage crustal augmentation occurs within a limited region in the vicinity of the RTI. Thus, the second-stage magmatic accretion at RTIs is not likely to change the segment-scale variations in the crustal thickness.

Another possibility is that the uniform crust might reflect the patterns of crustal and mantle flows and the melt migration along the ridge axis at the time of crustal formation. The slow-spreading (~10 mm/yr half-spreading rate) Reykjanes Ridge in the North Atlantic Ocean produces 8-10 km-thick crust showing an absence of significant along-axis variation in thickness⁵⁸. This is interpreted to be due to a rapid ductile flow within the hot lower crust because of the influence of the Iceland hotspot⁷. Though the ridge between 5°N and the St. Paul TF is suggested to be influenced by the Sierra Leone plume^{59, 60}, the crustal thicknesses north and south of the St. Paul FZ are similar, ruling out any significant influence of the thermal anomaly to crustal accretion in our study area. Furthermore, the average crustal thickness of 5.4-5.6 km is not thick enough, and hence not hot enough to enable rapid ductile flow in the lower crust. On the other hand, a 3-D diapiric mantle upwelling would produce a large along-axis variation in crustal thickness^{3, 8}, which is inconsistent with our observations of the five crustal segments studied here.

Finally, the relatively uniform crustal thickness observed in the equatorial Atlantic Ocean suggests a relatively uniform (nearly 2-D) mantle upwelling beneath the ridge axis at the time of the crustal accretion (Fig. 6), contradicting the previously proposed 3-D plume-like mantle upwelling model for the slow-spreading ridges⁸ based on the observations from the North and South Atlantic Ocean, instead it is more similar to the uniform mantle flow pattern occurring at the fast-spreading ridges³. The petrological studies of basaltic glasses dredged along the equatorial MAR between the St. Paul and Charcot TFs show a long-wavelength trend (~600 km) in the mantle potential temperature, the mean degree of

partial melting and the maximum depth of the decompression melting. This long-wavelength trend is independent of the offset and the location of TFs^{59, 60}. While north of the St. Paul TF to ~5°N, the mantle potential temperature, the initial depth and the mean degree of mantle melting have little variations over ~400 km lateral distance due to the influence of the Sierra Leone plume^{59, 60}. These observations suggest that the extent of mantle melting and mantle isotherms are not significantly suppressed approaching TFs and NTOs as required in the 3-D diapiric mantle upwelling^{3, 8}, supporting a relatively uniform mantle upwelling in the study region.

The lithosphere-asthenosphere boundary beneath the slow-spreading ridges is expected to progressively deepen from the segment centre towards the oceanic TFs⁶¹ due to the juxtaposition of ridge axis with cold lithosphere. The lithospheric thinning away from the TFs towards the centre of segments would shallow the top surface of the melting region, and subsequently enhance the melting at the centre of the ridge segment and produce a thick crust. Magde et al.^{19, 20} argued that the segment-scale crustal thickness variations observed at the slow-spreading MAR between 33°N and 35°N are primarily produced by the focused melt migration to the centre of segments along the slope of the top surface of the melting region, though the mantle upwelling is roughly uniform. In our study region, however, the basaltic glasses dredged from the vicinities of the western RTI of the St. Paul TF, the eastern RTIs of the Romanche and Chain TFs show large local geochemical and isotopic dispersions, suggesting the melt did not pool and mix in the magma chamber but segregated and erupted rapidly from the mantle⁵⁹. This implies that the melt would migrate vertically and rapidly beneath the ridge axis in the equatorial Atlantic Ocean (Fig. 6), rather than focusing to the segment centres before eruption. The nearly 2-D sheet-like mantle upwelling together with nearly vertical melt migration and rapid eruption would facilitate the formation of relatively uniform crust along the ridge axis at the time these five crustal segments were formed. Although a possible existence of a relatively uniform mantle upwelling can be made on the petrological observations along the present-day ridge axis in the equatorial Atlantic Ocean, our results indicate that this process has persisted over the last 70 Myr.

313

314 The observed uniformity in crustal thickness within a second-order ridge segment
315 between the Rio Grande and Cox FZs in the South Atlantic Ocean is suggested to be
316 related to a long first-order ridge segment (~320-350 km) facilitating large-scale mantle
317 upwelling²⁵. To investigate the influence of the first-order ridge segment length on the
318 crustal accretion, we compiled the maximum crustal thickness variation between the
319 centre and ends of second-order ridge segments and the length of the corresponding first-
320 order ridge segments in the Atlantic Ocean (Fig. 7). Our compilation demonstrates that
321 there is no positive or negative correlation between the first-order segment lengths and
322 the crustal thickness variations. For example, the ~220 km-long first-order segment
323 between the Hayes and Oceanographer TFs shows a maximum of 2.8-4.3 km along-axis
324 crustal thickness variations over the 0-5 Ma crust^{13, 14, 17} while the ~800 km-long segment
325 between the Kane and Atlantis TFs shows a maximum crustal thickness variation of 3.5
326 km along the ridge axis⁸. The lengths of the first-order ridge segments in the equatorial
327 Atlantic Ocean are similar to that of the segment between the Hayes and Oceanographer
328 TFs, but relatively uniform crust is observed. Lin et al.⁸ and Detrick et al.¹⁰ argued that at
329 the slow-spreading ridges, the magnitude of the along-axis crustal thickness variation
330 within a second-order ridge segment is positively correlated with the length of the
331 second-order segment. We cannot identify the locations of the fossil NTOs (except the
332 pseudofault between the Chain and Charcot FZs) along the seismic profile⁴³, but the
333 present-day spreading centres in the equatorial Atlantic Ocean have similar second-order
334 segment length (for example ref. 45) as those observed in other regions in the Atlantic
335 Ocean. We therefore argue that the lengths of first- and second-order ridge segments have
336 no influence on the crustal accretion process.

337

338 We envisage that the formation of the uniform crust might be associated with the brittle
339 mantle weakening and hydrous mantle melting at the transform zone and the enhanced
340 hydrothermal circulations on the spreading centres. By considering a more realistic brittle
341 mantle weakening, Behn et al.⁶² argued the thermal structure of the transform zone is
342 much warmer than predicted from the half-space cooling model, which can better fit the

depth of seismicity on the oceanic TFs. Their modelling also demonstrates an enhanced mantle upwelling along the transform zone and a much thinner lithosphere at the transform zone, especially at the centre of the transform, than estimated in previous studies using simplified rheologic laws. The base of the lithosphere at the transform zone can be further uplifted by the low-temperature hydrous mantle melting due to the presence of water (Wang et al., 2022). Meanwhile, the lithosphere beneath the ridge axis in the equatorial Atlantic Ocean is imaged to be thickened due to the enhanced hydrothermal cooling^{63, 64}. The combined effect of all these processes could significantly decrease the cold edge effect of large oceanic TF, leading to a relatively flat isotherm (lithospheric base) beneath the ridge axis with only a slight deepening of the lithospheric base in the vicinity of the transform zone. Using the crustal thickness inferred from gravity data, Detrick et al.¹⁰ argued that the magnitude of the along-axis crustal thinning is approximately proportional to the offset of the adjacent TF. However, only three oceanic TFs of length <120 km are investigated in their studies, which are much shorter than the three mega-transforms (~300-880 km) in the equatorial Atlantic Ocean. In contrast, the compilation of crustal thickness variations (Fig. 8) inferred from active-source seismic data demonstrate that there is no positive correlation between the magnitude of the along-axis crustal thickness variation and the length of the adjacent TF or NTO. The magnitude of the crustal thickness variation is more scattered and larger when the adjacent TF or NTO is short but is smaller for longer TFs. This suggests the mega-transform could facilitate a stable 2-D sheet-like mantle upwelling and a relatively uniform crustal accretion along the slow-spreading ridges.

The tectonic fabric of the ocean basin also reflects the different magmatic accretion patterns (focused or uniform) along the MAR. Though NTOs are observed on the ridge axis in the equatorial Atlantic Ocean^{43, 45}, the seafloor fabric⁴³ indicates that these NTOs are more transient features with nearly no imprint away from the ridge axis, suggesting that the base of the lithosphere beneath these short-lived NTOs does not deepen much and remains nearly flat, hindering the melt focusing towards segment centres and leading to a relatively uniform crustal accretion.

373

374 The equatorial Atlantic Ocean has been suggested to be a cold spot^{60, 65}, where the
375 mantle temperature is at least 150°C lower than that beneath other parts of the MAR⁶⁵.
376 Our results suggest that the dominantly uniform magmatic accretion can occur not only at
377 fast-spreading ridges³ and at slow-spreading ridges influenced by hotspots⁷, but can also
378 occur at slow-spreading ridges influenced by cold spots over a long geological period.
379 The average crustal thicknesses of the five magmatic segments in the equatorial Atlantic
380 Ocean are thinner than the global average²⁷, which can be attributed to a low degree of
381 mantle melting due to the cold mantle^{59, 60}. Studies using bathymetry and seismic
382 refraction datasets found that the detachment fault accommodating plate separation
383 occurs along ~50% of the ridge axis in the North Atlantic Ocean^{22, 66}. However, along our
384 seismic profile, the magmatically accreted crustal segments comprise >90% of the total
385 profile length, which is much higher than the previous estimates for the slow-spreading
386 ridges, suggesting that the mode of accretion plays a more important role in defining the
387 2-D versus 3-D mantle upwelling, and the lateral extent of the magmatic accretion could
388 be used to map the 2-D sheet-like mantle upwelling regions along the global spreading
389 ridge system.

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Methods

Seismic data analysis and travel time picking. The travel times of crustal refractions (Pg), wide-angle reflections from the Moho (PmP) and the mantle refractions (Pn) have been hand-picked on the seismic data after band-pass filtering (4-15 Hz for Pn arrivals and 4-20 Hz for Pg, PmP and S-wave arrivals). The Pg phases are identified as first arrivals outside of the water wave on 50 OBSs from ~4 to ~35 km offsets (Supplementary Fig. 1, orange dots). The picking uncertainty of the Pg arrivals varies between 30 and 50 ms. The Pn phases are identified from ~15-35 to ~250 km offsets for most OBSs (Wang et al., 2022). In this work, only the Pn arrivals within 40 km offset are shown (Supplementary Fig. 1, blue dots). The picking uncertainty of the Pn arrivals varies between 45 and 100 ms. The PmP phases reflected off the Moho are identified at ~10-40 km offsets as second arrivals after the Pg and Pn phases (Supplementary Fig. 1, cyan dots). The picking uncertainty of the PmP arrivals is 50 or 70 ms.

The crustal and mantle S-wave arrivals are observed on seismic data recorded by 43 OBSs as secondary arrivals after the P-wave first arrivals and codas (Supplementary Fig. 2). In this study, we use Sg, SmS and Sn to represent the S-wave crustal refractions, Moho reflections and mantle refractions, respectively. We first pick the S-wave arrivals that have high signal-to-noise ratios using the trace-to-trace coherency of phases and then we invert these picks to obtain a smooth velocity model for the crust and uppermost mantle. Synthetic travel times of S-wave phases are modelled using this smooth velocity model, which is then used as a guide to identify and pick more S-wave arrivals that have low signal-to-noise ratios^{25, 67}. The Sg arrivals are identified and picked between 5 and 35 km offsets (Supplementary Fig. 2, orange dots) with picking uncertainty ranging from 30 to 60 ms. The SmS phases are picked at offset ranges similar to the PmP arrivals (Supplementary Fig. 2, cyan dots), with picking uncertainty of 60-80 ms. The Sn arrivals are picked on 38 OBSs (Supplementary Fig. 2, blue dots) with picking uncertainty of 50-100 ms.

Starting velocity models for travel time tomography. The seafloor depth and sediment thickness along the seismic profile are determined using coincident high-resolution bathymetry data and seismic reflection data^{44, 45}. The velocities of water and sediment are set to 1.5 km/s and 1.86 km/s in tomography^{44, 45}, respectively. The initial crustal Vp model is constructed using a simplified one-dimensional (1-D) velocity profile^{44, 45} hanging from the basement. We introduce a smooth Moho interface into the starting Vp model at ~5.5 km depth below the basement. The initial structure of the Moho is obtained by smoothing the basement reflector within a 12 km-wide sliding window. The starting mantle Vp model is constructed utilizing a 1-D velocity profile hanged from the smooth Moho, where the mantle Vp increases from 7.8 km/s just below the Moho with a vertical velocity gradient of 0.014 s⁻¹ to 20 km sub-Moho depth and then increases gently (gradient of 0.0035 s⁻¹) to the base of the model. The model is 867 km wide and 90 km deep, and is discretized by 300 m grid spacing horizontally and 60 m grid spacing vertically. In this study, we first model and invert the picked travel times of P-wave arrivals to obtain the crustal and mantle Vp. The starting Vs model for inverting the picked travel times of S-wave arrivals is converted from the final Vp model, assuming a Vp/Vs ratio of 1.74 for both crust and mantle. The depth of the Moho constrained by the PmP travel times is fixed during the inversion of the travel times of S-wave arrivals.

Ray-based travel time modelling and tomography. We model and invert all the picked travel times using a two-dimensional ray-based travel time tomography method^{2, 46}. The ray paths and travel times of the crustal and mantle arrivals are calculated using the shortest path method⁶⁸. We take a top-down inversion strategy in the tomography of P-wave arrivals. The Pg arrivals are inverted first to constrain the velocity of the upper crust, followed by a joint inversion of Pg and PmP arrivals simultaneously to constrain the crustal velocity and the Moho depth. After the joint inversion of Pg and PmP arrivals, we fix the crustal velocity and the Moho depth, and invert the travel times of Pn arrivals only to update the mantle velocity. First- and second-order velocity regularizations are imposed to obtain a smooth velocity model⁴⁶. The regularization parameters are tested and selected in each iteration step to avoid the introduction of artefacts. We use the

standard χ^2 value⁴⁶ to measure the mismatch between the modelled and manually picked travel times. We stop the inversion when the χ^2 value approaches 1.0 or when non-physical artefacts appear when further decreasing the χ^2 value. The final χ^2 value is 1.6 for Pg and PmP arrivals (Supplementary Fig. 3a) and is 2.2 for Pn arrivals (Supplementary Fig. 3b). The final RMS misfits⁴⁶ are 42, 65 and 74 ms for Pg, PmP and Pn arrivals, respectively.

The same inversion method and strategies are used in the S-wave tomography, except the Moho depth constrained by the PmP arrivals is fixed in the joint inversion of Sg and SmS arrivals. Doing so, we have assumed that P- and S-waves share the same boundary between crust and upper mantle. The final χ^2 value of combined Sg and SmS arrivals is 1.2 (Supplementary Fig. 3c) and that of Sn arrivals is 1.5 (Supplementary Fig. 3d). The final RMS misfits are 51, 63 and 98 ms for Sg, SmS and Sn arrivals, respectively.

Ray coverage. We use the derivative weight sum (DWS)⁶⁹ to represent the density of ray coverage through the final tomographic models. The DWS of the P-wave arrivals throughout the crust and upper mantle is shown in Supplementary Fig. 4a. The Vp of the upper crust is constrained by dense rays of Pg arrivals between 60 and 760 horizontal distances along the profile, while the lower crust has relatively sparse PmP ray coverage. In the mantle, the Pn arrivals sample the upper mantle down to ~60 km depth below sea level beneath the Romanche transform valley and the maximum sampled depth gradually decreases to south and north (Wang et al., 2022), but here we only show the model down to 12.5 km depth (Supplementary Fig. 4a).

Supplementary Fig. 4b shows the DWS of S-wave arrivals through the crust and mantle. There is good ray coverage throughout the crust between the horizontal distance ~50 to 385 km and ~460 to 760 km. The crustal Vs within the Romanche transform zone is not constrained by any S-wave rays.

Monte-Carlo analysis. To assess the accuracy of the crustal and mantle velocities and the Moho depth, we perform Monte-Carlo analyses⁷⁰ starting from different initial models to produce different inverted models and estimate the standard deviation of these models from the mean model to measure the model variance. Here we only describe the details of the Monte-Carlo analysis for crustal Vp and Moho depth. We create 50 starting crustal models by randomly perturbing the minimum and maximum velocity of the one-dimensional crustal velocity profile by $\pm 5\%$ and the initial crustal thickness by ± 750 m. These starting crustal models are then inverted using the same tomography algorithm and inversion parameters as described before. These 50 models show similar crustal Vp structure, suggesting the inversion of the picked Pg and PmP travel times is robust. The variance in the final crustal Vp model is less than 0.06 km/s in the upper crust and is less than 0.18 km/s in the lower crust (Supplementary Fig. 5a). The maximum standard deviation of the Moho depth is ~ 330 m (Supplementary Fig. 5b). The preferred Moho (Fig. 3a) falls in the standard deviation of the average Moho depth from the Monte-Carlo analysis (see red curve in Supplementary Fig. 5b), and the maximum depth difference between the preferred Moho and the averaged Moho from Monte Carlo analysis is ~ 158 m (Supplementary Fig. 5b).

Similar Monte-Carlo analyses are performed to assess the variance in the crustal Vs. Supplementary Fig. 5c shows the variance of the crustal Vs calculated using 50 final inverted models. For most portion of the crust, the crustal Vs has a variance < 0.06 km/s, and large variances up to ~ 0.18 km/s are observed around the TF and FZs and at the southern and northern extremity of the model. The variance in the crustal Vp/Vs ratio is estimated using the preferred crustal Vp model (Fig. 3a) and the 50 crustal Vs models from the Monte-Carlo analysis. These models are used because they share the same Moho. We calculate the average of the 50 Vp/Vs models and take the standard deviation as the variance (Supplementary Fig. 5d). The variance in crustal Vp/Vs ratio is < 0.025 for most portion of the crust, and large variances up to ~ 0.1 only occur at local regions where the crustal Vs shows large variance.

Checkerboard tests. We also use the checkerboard test described in ref. (4) to assess the resolution of the final crustal Vp and Vs models. We first perform the checkerboard test to assess the crustal Vp. The checkerboard input models are designed by adding 2-D sinusoidal velocity perturbation of $\pm 8\%$ into the starting velocity model after slight smoothing. The anomaly size in the checkerboard models is 20×3 (Supplementary Fig. 6a) and 40×2 km (Supplementary Fig. 7a), respectively. A sinusoidal perturbation with a half-wavelength of 70 km and the vertical variation of 1.25 km was added to the initial Moho to examine the resolvability in the Moho depth^{4,45}. The checkerboard models show ~ 2.5 - 3.0 km crustal thickness variations over ~ 70 km distance (red curves in Supplementary Figs. 6c and 7c). Synthetic seismic travel times are calculated using the checkerboard models and the same receiver-source geometry as the Pg and PmP picks. Some random noise is added to these synthetic picks to represent the picking uncertainty. We invert these synthetic travel times using the same workflow as that for the OBS dataset starting from the same initial model. The results show that velocity anomalies of sizes 20×3 and 40×2 km are almost completely resolved (Supplementary Figs. 6b and 7b) with better recovery for anomaly with 20×3 km size. The crustal thickness within the recovered model shows ~ 1.5 - 2.0 km variations over ~ 70 km distance (magenta curve in Supplementary Figs. 6c and 7c), suggesting the tomography method can partially recover the large along-axis variations in crustal thickness. A similar checkerboard test is applied to assess the resolution of crustal Vs model. The checkerboard test shows a higher resolution of the crustal Vs model than the crustal Vp model, where the 8% velocity anomaly with the 15×3 km size can be well recovered (Supplementary Fig. 8). Higher resolution is obtained in crustal Vs model because no velocity-Moho depth trade-off occurs in the tomography of S-wave arrivals.

Figures

Fig. 1 Crustal thickness (H) difference between segment centres and segment ends as a function of spreading rate. Only the crustal thickness constrained by active-source seismic data is considered. The crustal thickness data from the Atlantic Ocean are selected following two criteria: (1) systematically along-axis crustal thinning is observed within the second-order ridge segment and (2) the crustal thicknesses at segment centre and at least one segment end are measured. The blue squares show the crustal thickness reductions at TF, FZ and pseudofault region relative to the average crustal thickness of segments obtained in this study. When the crustal thicknesses of both ends of a segment are available, the one with thinner crust is plotted. The crust at the ends of slow-spreading ridges is generally ≥ 2.8 km (the thick black dashed line) thinner than that at the associated segment centres. Data for EPR, South and North MAR are given in Supplementary Table 2. MAR: Mid-Atlantic Ridge, EPR: East Pacific Rise. The thin dashed vertical lines mark the boundaries of fast- and slow-spreading ridges⁴¹. The thick dashed horizontal line indicates the boundary between uniform and segment centre focused crust.

Fig. 2 Study area. a, Bathymetry map showing the Mid-Atlantic Ridge (MAR; red lines), transform faults (TFs; white solid lines) and fracture zones (FZs; white dashed lines) in the equatorial Atlantic Ocean. The seismic profile is shown as a brown line, with the location of every tenth OBS marked by cyan dots. The age of the oceanic crust⁴⁰ is contoured and labelled every 10 Myr. The black rectangle in the globe inset shows the location of the study area. **b**, Variation in average half-spreading rate⁴⁰ of the equatorial MAR in the past 80 Myr. The dashed blue lines from left to right indicate the ages of 8, 24, 40 and 70 Myr, respectively. As the half-spreading rate varied between 15 and 25 mm/yr in the past 80 Myr, the equatorial MAR can be classified as the slow-spreading ridge (half-spreading rate between 10 and 27.5 mm/yr⁴¹).

Fig. 3 Crustal and upper mantle velocity models. Inverted P-wave velocity (V_p) model (a), S-wave velocity (V_s) model (b) and V_p/V_s ratio (c). The distance ranges of Segment 1 to 5 are marked on the top of a.

Fig. 4 One-dimensional P-wave velocity-depth profiles. a, Profiles extracted from the Segments 1, 2, 3-S, 4, 5 with predominantly magmatic accreted crust. **b**, Profiles extracted from the Segment 3-N with predominantly tectonically formed crust. The black dashed lines represent the velocity envelope for the oceanic crust aged 59-150 Myr in the Atlantic Ocean²⁸. The grey shadings represent the velocity envelope for the slow-spreading oceanic crust aged >7.5 Myr²⁷. The black solid lines represent the 7.2 km/s velocity.

Fig. 5 Crustal thickness variation (black curves) and average crustal thickness (red lines) of each ridge segment. The dashed blue lines in each panel indicate the boundaries of ridge segments. The numbers in red show the average crustal thickness and the standard deviation.

Fig. 6 Schematic diagram showing the relatively uniform crustal accretion and the two-dimensional sheet-like mantle upwelling. The black thick arrows represent the mantle flow. The red solid arrows represent the melt migration in the melting region and the red dashed arrows represent the melt migration in the lithospheric mantle. In this model, the melt is not focused to segment centres at the base of the lithosphere^{19, 20}, but migrates upward vertically.

Fig. 7 Maximum crustal thickness variation between centres and ends of second-order ridge segments versus the length of the corresponding first-order ridge segment in the Atlantic Ocean. The crustal thickness constrained by both active-source seismic data and gravity data is considered. Data are given in Supplementary Table 3.

Fig. 8 Crustal thinning towards the oceanic TF or NTO in the Atlantic Ocean as a function of the length of TF or NTO. Only the crustal thickness constrained by active-source seismic data is considered. The blue squares show the crustal thickness reductions at TF, FZ and NTO relative to the average crustal thickness of segments obtained in this study. For a transform fault composed of several intra-transform faults, the total length of the transform fault is used. Data are given in Supplementary Table 4. TF: transform fault; NTO: non-transform offset.

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Author contributions

Z.W. processed and analysed the OBS data and wrote the paper. S.C.S. developed the project, led the data acquisition and supervised the data processing, interpretation and writing.

Competing interests

The authors declare no competing interests.

Data availability

The high-resolution bathymetry data, the multichannel seismic reflection data and the OBS data from the OBS26 to OBS50 are available online (<https://doi.pangaea.de/10.1594/PANGAEA.922331>) under the condition of acknowledging Marjanović et al., 2020. The OBS data from the OBS01 to OBS15 are available online (<https://doi.pangaea.de/10.1594/PANGAEA.937195>) under the condition of acknowledging Growe et al., 2021. The OBS data from the OBS16 to OBS25 will be available on request from the authors after the publication of this paper.

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