

Article

Influence of Melt Supply on the Spreading State of a Slow–Ultraslow-Spreading Ridge: The Reykjanes Ridge, North Atlantic

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Abstract: Although recent research suggests that the morphology and crustal structure of slow–ultraslow-spreading ridges are mainly controlled by melt supply, there is a lack of quantitative understanding of the effect of systematic changes in melt supply on the seafloor spreading state of mid-ocean ridges. In this study, we used bathymetry, free-air gravity anomaly, and sediment thickness data to calculate the residual bathymetry, mantle Bouguer gravity and crustal thickness of the Reykjanes Ridge. According to the gradient of changes in crustal thickness and residual bathymetry along the axis, the influence of melt supply on the spreading state of the Reykjanes Ridge can be divided into three zones: ultra-strong effect zone (0–160 km), strong effect zone (160–610 km), and weak effect zone (610–930 km). In the ultra-strong effect zone, excess melt supply and a higher melting degree result in a strong upwelling and large melt eruption. The change in relative position between the Reykjanes Ridge and the Iceland hotspot results in the spreading state of the Reykjanes Ridge transforming from asymmetric spreading to symmetric spreading. In the strong effect zone, the decrease in melt supply and melting degree weakens the mantle upwelling and enhances the viscosity of the dehydrated mantle layer. Sufficient viscosity of the dehydrated mantle layer forces asymmetric asthenosphere rise along the sloping boundary of the lithosphere, resulting in symmetric spreading. In the weak effect zone, the pattern of magma upwelling becomes a focused magma supply pattern similar to that of the slow–ultraslow-spreading of the mid-ocean ridge, and tectonics dominate the spreading process. The asymmetry of this weak effect zone may be due to the concentration of tectonic and magmatic activity on one flank of the ridge.

Keywords: hotspots; Reykjanes Ridge; melt supply; mantle upwelling; asymmetric spreading



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1. Introduction

Melt supply is a key factor controlling the morphology, crustal structure, thermal regime, and crustal accretion pattern at mid-ocean ridges [1–5]. For the mid-ocean ridges with abundant melt supply, there exist continuous steady-state magma lenses along the axis [5,6], resulting in axial highs, relatively uniform crustal thickness and a lithospheric thermal state [7]. In contrast, for the mid-ocean ridges with limited melt supply, the continuous steady-state magma lens cannot be formed along the entire axis [5], resulting in the alternations of the magmatic zone and the amagmatic zone [4,8,9]. The along-axis variations in thermal structure led to dramatic variations in axial yield strength and hence in the along-axis variations in axial valley topography [5].

Numerical simulation results show that when the magma supply exceeds a threshold, the morphology and crustal structure of slow-spreading ridges can resemble those of

fast-spreading mid-ocean ridges [5,10]. Cannat et al. [3] proposed that variations in melt volume flux and depth of melt emplacement can cause sharp along-axis transitions in the lithospheric thermal state, resulting in a significant shift in lithospheric thickness and strength. Chen et al. [11] found that cyclic variations in melt supply have a significant impact on the morphology, crustal structure and thermal regime at mid-ocean ridges. During waxing phases, shallow axial melt lenses form in the segment center, resulting in smooth domal volcanos with traditional upper crustal structures. In contrast, during waning phases, the axial melt lens deepens or disappears, resulting in a cooler lithosphere, more faults and a thinner crust. Chen et al. [1] proposed that, for a given spreading rate and melt supply, if melt is emplaced in the shallow hydrothermally active zone, the thermal regime can be colder. Although scholars have conducted targeted research on the influence of melt supply on the mid-ocean ridges, the focus has mainly been on qualitative studies using numerical simulations, and there is a lack of quantitative observations and research on the effect of systematic changes in melt supply on the seafloor spreading state of mid-ocean ridges.

The northern part of the Reykjanes Ridge in the North Atlantic is connected to Iceland, which is formed by a mantle plume. It exhibits distinct morphology and crustal structure resulting from the interaction between the hotspot and the mid-ocean ridge. The hot mantle material of the Iceland hotspot directly feeds the Reykjanes Ridge, and the melt supply gradually decreases along the axis from north to south. Therefore, the Reykjanes Ridge is an ideal location for studying the influence of melt supply on the spreading state of mid-ocean ridges.

The Reykjanes Ridge is located between the Bight Fracture Zone and Iceland (Figure 1) and belongs to the slow-spreading mid-ocean ridge [12]. At about anomaly 24 (~55 Ma), the Reykjanes Ridge began seafloor spreading and connected with the already existing Mid-Atlantic Ridge and the Labrador Sea spreading center in a ridge–ridge–ridge triple junction [12]. The Reykjanes Ridge kept orthogonally spreading without transform faults as a linear geometry until anomaly 17 (~37 Ma) when the spreading direction abruptly changed ~30° as the Labrador Sea spreading creased [13,14]. At anomaly 17, the Reykjanes Ridge became the boundary between the North American and Eurasian plates, and the elimination of the triple junction initiated the Bight Fracture Zone [15]. This tectonic reconfiguration also led to the breakup of the linear spreading axis and the formation of a series of ridge segments perpendicular to the new spreading direction [13]. After the establishment of the segmented plate boundary of the Reykjanes Ridge, the stair-step ridge-transform configuration was eliminated from north to south and became a linear obliquely spreading ridge [12]. During the reconfiguration process, the Reykjanes Ridge formed the diachronous V-shaped crustal ridges resulting from the melt supply flowing outward from the mantle plume beneath Iceland [16]. At about anomaly 6 (~20 Ma), the Iceland hotspot crossed Greenland into the Atlantic Ocean near the ridge axis at anomaly 5 (~10 Ma). Subsequently, it moved eastward relative to the Reykjanes Ridge [17].

The melt supply significantly affects the morphology of a mid-ocean ridge [18]. In addition, the mantle Bouguer gravity anomaly (MBA) and crustal thickness can quantify the supply of excess melt and heat to a ridge [18]. Therefore, in this study, we used the latest topography data, free-air gravity anomaly data, and sediment thickness data to calculate the MBA, crustal thickness and residual bathymetry of the Reykjanes Ridge. By quantitatively analyzing the variations in residual bathymetry, MBA, and crustal thickness along the axis, as well as the across-axial asymmetry within the 20–0 Ma, we investigate the degree and mechanism of the influence of melt supply on the spreading state of the Reykjanes Ridge.

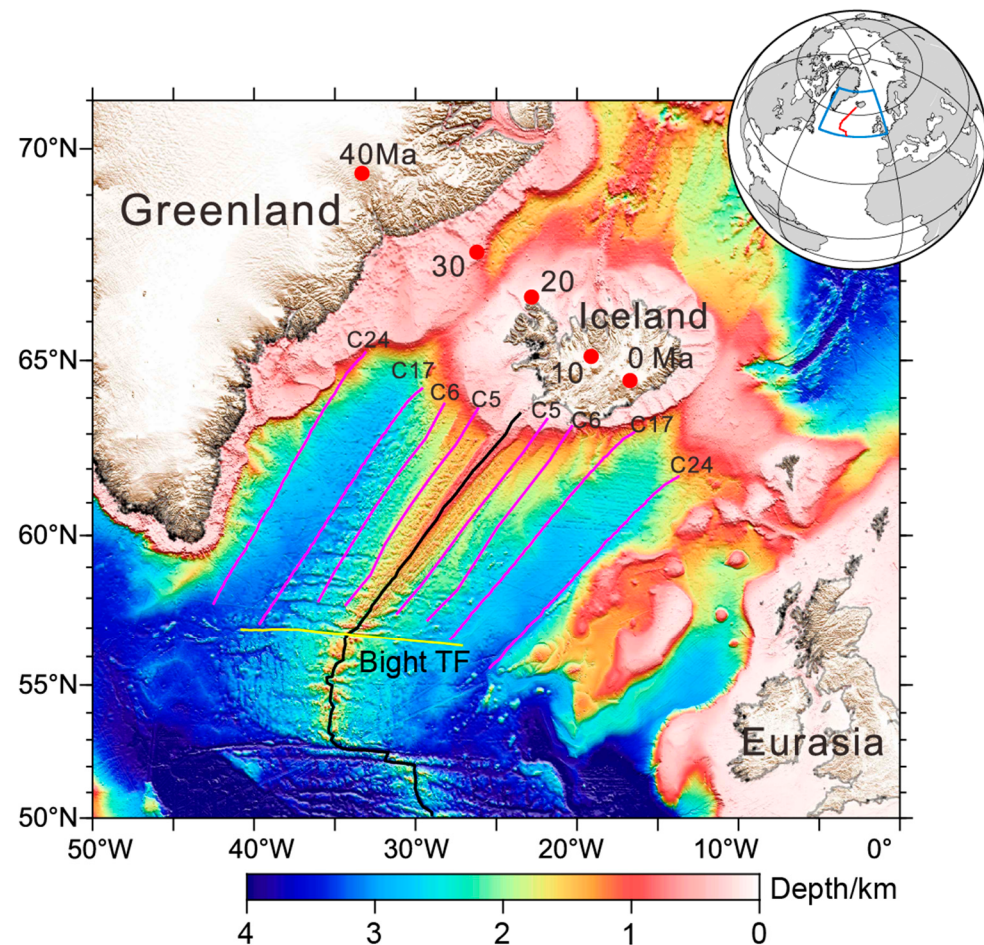


Figure 1. Map of the Reykjanes Ridge and flanking North Atlantic basin. Red circles indicate the track of the Iceland hotspot. Purple lines indicate magnetic anomaly strips. The black line indicates the mid-ocean ridge axis. The yellow line indicates the transform fault. Bight TF: Bight transform fault.

2. Methods and Data

2.1. Data

In this study, the bathymetry data (Figure 2a) were derived from the latest global bathymetric grid model GEBCO_2023 with a grid spacing of $15'' \times 15''$. The GEBCO_2023 is based on SRTM15+ [19] and comprises 3,732,480,000 data points encompassing elevation information of land and ice. Free-air gravity anomaly data (Figure 2b) were derived from the global gravity model DTU17 with a grid spacing of $1' \times 1'$ [20]. The DTU17 enhances the gravity field for shorter wavelengths (10–15 km) and incorporates an improved tidal model and a longer time series near coastal zones. Sediment data (Figure 2d) were derived from the latest global ocean sediment thickness model with a grid spacing of $5' \times 5'$ [21]. Compared to the previous model, the sediment thickness data based on multi-channel seismic data increased by 29.7%. Crustal age data (Figure 2c) were derived from the latest global oceanic crustal age model with a grid spacing of $1' \times 1'$ [22]. The model is primarily based on the latest magnetic strip identification results and plate tectonic models developed by Müller et al. [23].

We used a Mercator projection to project all data onto Cartesian coordinates to eliminate the errors introduced by the planar treatment of curvilinear coordinates. In order to quantitatively analyze the asymmetry in the RB, MBA and crustal thickness of the Reykjanes Ridge, six profiles were generated along the spreading direction at intervals of 140 km.

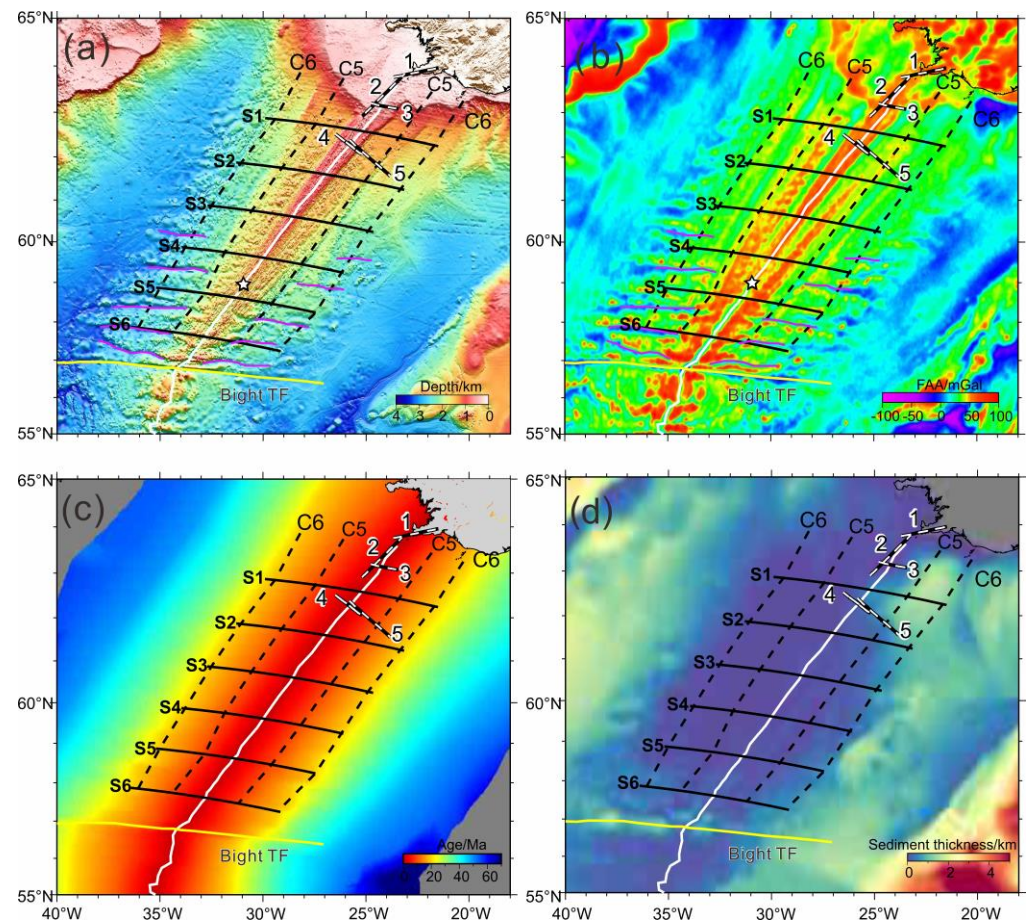


Figure 2. Overview maps of the Reykjanes Ridge. (a) Bathymetry. (b) Free-air gravity anomaly. (c) Crustal age. (d) Sediment thickness. The white line indicates the Reykjanes Ridge axis. Black lines indicate cross-axis profiles. Black dashed lines indicate magnetic strips. White–black lines indicate seismic profiles. Purple lines indicate transform faults. The white star indicates the transition from axial rise to rift valley. Bight TF: Bight transform fault.

2.2. Mantle Bouguer Anomaly

The MBA reflects the gravitational effect of crustal thickness, crustal density and the thermal state of the lithosphere. Because the ridge axis has a relatively uniform crustal thickness and crustal density, the MBA is mainly affected by the variation in the thermal structure. We used Parker’s [24] spectrum method to calculate the MBA by subtracting the gravitational effects of the water–sediment, sediment–crust, and crust–mantle interfaces from the free-air gravity anomaly [24,25]. The water density, crust density, and mantle density were assumed to be 1030, 2800, and 3300 kg/m³, respectively. The sediment density increases with depth due to compaction. Cowie and Karner [26] obtained the exponential density–depth curve. The exponential function fits well-log data in several sediment basins. We divided the sediment into 6 layers and assigned an average density to each layer according to Wang et al. [25]. The gravitational effects of the sediment are obtained by integrating the gravitational effects of each layer.

2.3. Thermal Correction

We calculated the gravitational effect of lithospheric cooling from a depth of 0 to 100 km using a 3D thermal model [27]. The thermal structure of the lithosphere was calculated using the thermal structure equation:

$$T = T_m \times \operatorname{erf}\left(y/2\sqrt{ut}\right) \quad (1)$$

where T_m , y , u , and t are the temperature at the bottom of the plate, lithospheric thickness, thermal diffusivity rate, and crustal age, respectively. We set $T_m = 1350$ °C and $u = 0.8$ mm²/s. The thermal structure was then transformed into a 3D density grid using the thermal expansion equation:

$$\Delta\rho = (T_0 - T)\alpha\rho_0 \quad (2)$$

where T_0 , α , and ρ_0 are the reference temperature, thermal expansion coefficient, and reference density, respectively. We set $T_0 = 1350$ °C, $\alpha = 3 \times 10^{-5}$ K, and $\rho_0 = 3300$ kg/m³. We divided the lithosphere into 10 equal layers and calculated the density structure of each layer. Then, we calculated the corresponding gravitational effect. Finally, the gravitational effect of lithospheric cooling was obtained by integrating the gravitational effects of each layer.

2.4. Residual Mantle Bouguer Anomaly and Crustal Thickness

We calculated the residual mantle Bouguer gravity anomaly (RMBA) by removing the gravitational effects of lithospheric cooling from the MBA [28]. The RMBA reflects deviation from the assumed crust–mantle structure [9,29,30]. We obtained the maximum crustal thickness model by assuming that the RMBA was only due to crustal thickness variations. Here, the crustal thickness was calculated by downward continuing the RMBA to a constant depth using Parker’s [24] frequency-domain inversion method. The downward continuation depth was set to be 10.5 km, i.e., the sum of the average water depth (3.5 km) and the average crustal thickness (7 km).

The crust–mantle density contrast is a key factor controlling the accuracy of the crustal thickness model. For the same RMBA, a larger crust–mantle density contrast can result in smaller lateral variations in crustal thickness [31]. To obtain the optimal crustal thickness model, we fixed the mantle density at 3300 kg/m³ and increased the crustal density from 2700 to 3000 kg/m³ in steps of 50 kg/m³ to obtain a series of crustal thickness models. These models were then compared with seismically determined crustal thickness, and the standard deviation of the differences was calculated (Table 1). The crustal thickness model with a crustal density of 2800 kg/m³, corresponding to a crust–mantle density contrast of 500 kg/m³, has the minimum standard deviation. To evaluate the roughness of the crustal thickness model, we compared the gravity-derived crustal thickness with the seismically determined crustal thickness values (Figure 3). The gravity-derived crustal thickness agreed relatively well with the seismic results: the average value and standard deviation were 0.28 and 0.92 km, respectively.

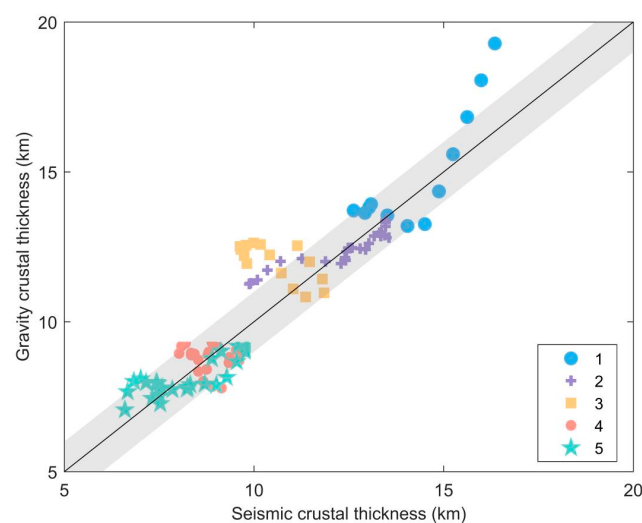


Figure 3. Comparison of gravity-derived crustal thickness values with seismically determined crustal thickness values from 5 seismic profiles shown in Figure 2.

Table 1. Sensitivity of the model results to assumed crustal density.

Assumed Crustal Density (kg/m ³)	Root Mean Square (RMS)
2700	1.246
2750	1.113
2800	1.045
2850	1.119
2900	1.406
2950	1.941
3000	2.767

2.5. Residual Bathymetry

The residual bathymetry (RB) reflects the initial topography during mid-ocean ridge formation [30,32]. The RB is calculated by removing of the predicted effects of plate cooling [27] and sediment loading from the observed bathymetry [25]. Assuming the Airy isostatic compensation, the effect of the sediment loading is calculated using the following equation:

$$\Delta z = h_s(\rho_m - \rho_s)/(\rho_m - \rho_w) \quad (3)$$

where Δz is the correction to the bathymetry, h_s is the sediment thickness, and ρ_m , ρ_s , and ρ_w are the mantle density, sediment density, and water density, respectively, which are the same as those used in the gravitational calculations. The effect of plate cooling was calculated using the plate-cooling equation:

$$\begin{cases} D = 2600 + 365t^{1/2}, t < 20 \text{ Ma} \\ D = 5651 - 2473e^{-0.0278t}, t \geq 20 \text{ Ma} \end{cases} \quad (4)$$

where D is the thermal subsidence (m), and t is the crustal age (Ma).

3. Results

3.1. Mantle Bouguer Gravity Anomaly

The MBA of the Reykjanes Ridge and its adjacent area ranges from −200 to 160 mGal and gradually increases from −125 mGal in the north to 0 mGal in the south along the ridge axis (Figure 4a). The MBA curve is divided into two segments: at the 0–160 km range, the MBA increases rapidly by 65 mGal with a gradient of 0.406 mGal/km; at the 160–930 km range, the MBA increases by 65 mGal with a gradient of 0.08 mGal/km (Figure 4b). The width of the MBA low-value zone (<0 mGal) narrows from 282 km in profile 1 to 14 km in profile 6 (Figure 4c). In the north of profile S2, the western side of the ridge has a wider MBA low-value zone than that of the eastern side of the ridge. In the south of profile S2, the Reykjanes Ridge has a symmetric MBA low-value zone. In profile S4, the width of the MBA low-value zone narrows rapidly. In the south of profile S6, the width of the MBA low-value zone approaches 0 km.

3.2. Residual Bathymetry

The RB of the Reykjanes Ridge and its adjacent area ranges from 0 to 4000 m and gradually decreases from ~2700 m in the north to ~300 m in the south along the ridge axis (Figure 5a). The RB curve is divided into three segments: at the 0–160 km range, the RB decreases rapidly by ~700 m with a gradient of 4.375 m/km; at the 160–610 km range, the RB decreases by ~700 m with a gradient of 1.556 m/km; at the 610–930 km range, the RB decreases by ~1200 m with a gradient of 3.750 m/km (Figure 5b). The transition zone between the second and third segments corresponds to the area where the axial morphology changes from rise to rift valley, and the RB decreases from 1600 to 1100 m.

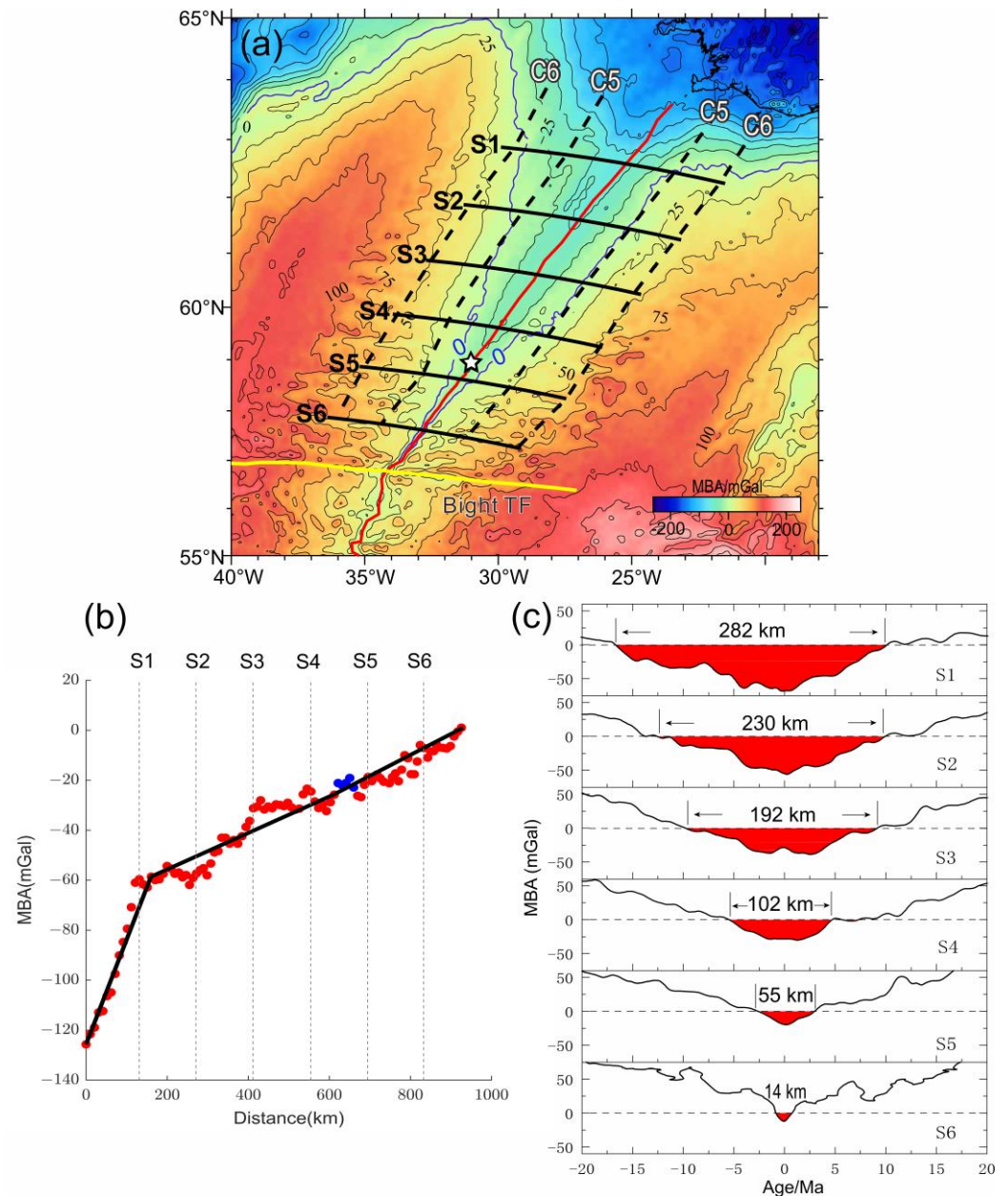


Figure 4. (a) Mantle Bouguer gravity anomaly. The red line indicates the Reykjanes Ridge axis. Black lines indicate cross-axis profiles. Black dashed lines indicate magnetic strips. The white star indicates the transition from axial rise to rift valley. (b) The variation in MBA along the axis. Blue scatters indicate the transition from axial rise to rift valley. (c) The variation in MBA along the six profiles. The red shadow zone indicates the MBA low-value zone (<0 mGal).

On the whole, the RB asymmetry curve exhibits an asymmetric–symmetric–asymmetric pattern from north to south during 20–5 Ma and a symmetric pattern during 5–0 Ma (Figure 5c). Profile S1 is asymmetry during 20–4 Ma and symmetry during 4–0 Ma. During 20–4 Ma, the RB of the western side of the ridge is 286 m higher than that of the eastern side. At 12 Ma, the asymmetry amplitude reaches a maximum value of 666 m. During 4–0 Ma, the RB asymmetry amplitude is -39 ± 70 m. Profiles S2 and S3 exhibit a low RB asymmetry amplitude (51 ± 101 m and -9 ± 121 m, respectively) during 20–0 Ma. During 20–12 Ma, the RB of the western side of profile S4 is 166 m lower than that of the eastern side. During 12–6 Ma and 4–0 Ma, the RB is relatively symmetry, with an asymmetry amplitude of 51 ± 101 m. During 6–4 Ma, the RB of the western side of the ridge is 284 m lower than that of the eastern side. Similar to profile S1, profile S5 is asymmetric during 20–4.5 Ma

and symmetric during 4.5–0 Ma. During 20–13 Ma, the RB of the western side of the ridge is 292 m lower than that of the eastern side. During 13–10.5 Ma, the asymmetric polarity of the RB reversed twice. During 10.5–4.5 Ma, the RB of the western side of the ridge is 215 m higher than that of the eastern side. During 4.5–0 Ma, the RB is relatively symmetric, with an asymmetry amplitude of 33 ± 104 m. The RB of profile S6 is asymmetric during 17–5 Ma and symmetric during 5–0 Ma. The RB asymmetry amplitude decreases rapidly from 718 m to -16 m during 20–17 Ma. During 17–10 Ma, the RB of the western side of the ridge is 253 m lower than that of the eastern side. During 10–5 Ma, there are two RB asymmetry polarity reversals. During 5–0 Ma, the RB exhibits relative symmetry, with an asymmetry amplitude of 63 ± 124 m.

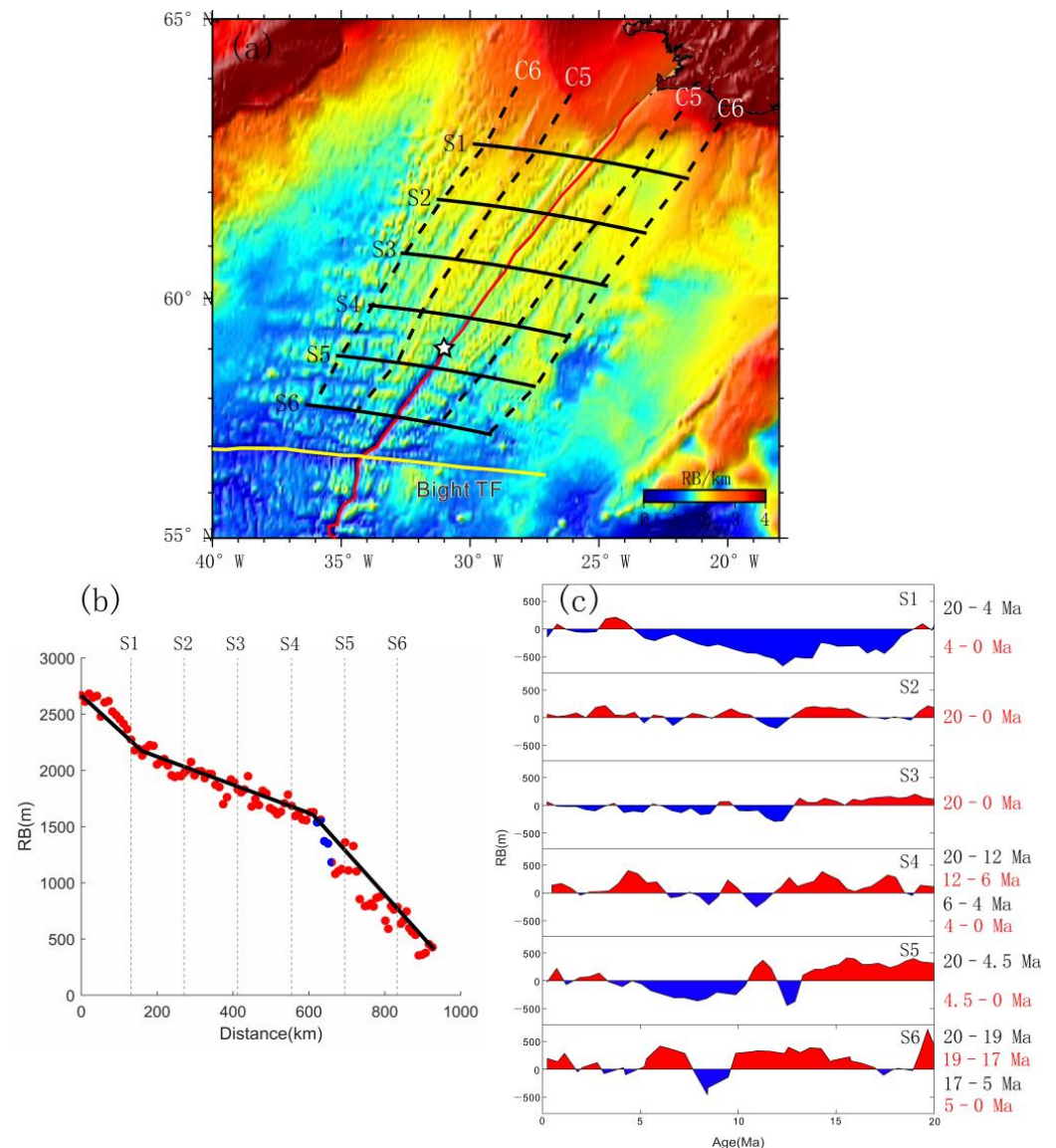


Figure 5. (a) Residual bathymetry. The red line indicates the Reykjanes Ridge axis. Black lines indicate cross-axis profiles. Black dashed lines indicate magnetic strips. The white star indicates the transition from axial rise to rift valley. (b) The variation in residual bathymetry along the axis. Blue scatters indicate the transition from axial rise to rift valley. (c) The asymmetry of residual bathymetry (the residual bathymetry of the western side of the ridge minus that of the eastern side of the ridge). The black numbers and red numbers in the sidebar represent asymmetric time periods and symmetric time periods, respectively.

3.3. Crustal Thickness

The crustal thickness of the Reykjanes Ridge and its adjacent area ranges from 0 to 24 km and gradually thins from ~12 km in the north to ~3 km in the south along the ridge axis (Figure 6a). Similar to the RB, the crustal thickness curve is divided into three segments: at the 0–160 km range, crustal thickness decreases rapidly by ~4 km with a gradient of 0.025 km/km; at the 160–610 km range, crustal thickness decreases by ~3 km with a gradient of 0.007 km/km; at the 610–930 km range, crustal thickness decreases by ~3.5 km with a gradient of 0.011 km/km (Figure 6b). The crustal thickness in the area where the ridge morphology changes from rise to rift valley decreases from 6.2 km to 5 km.

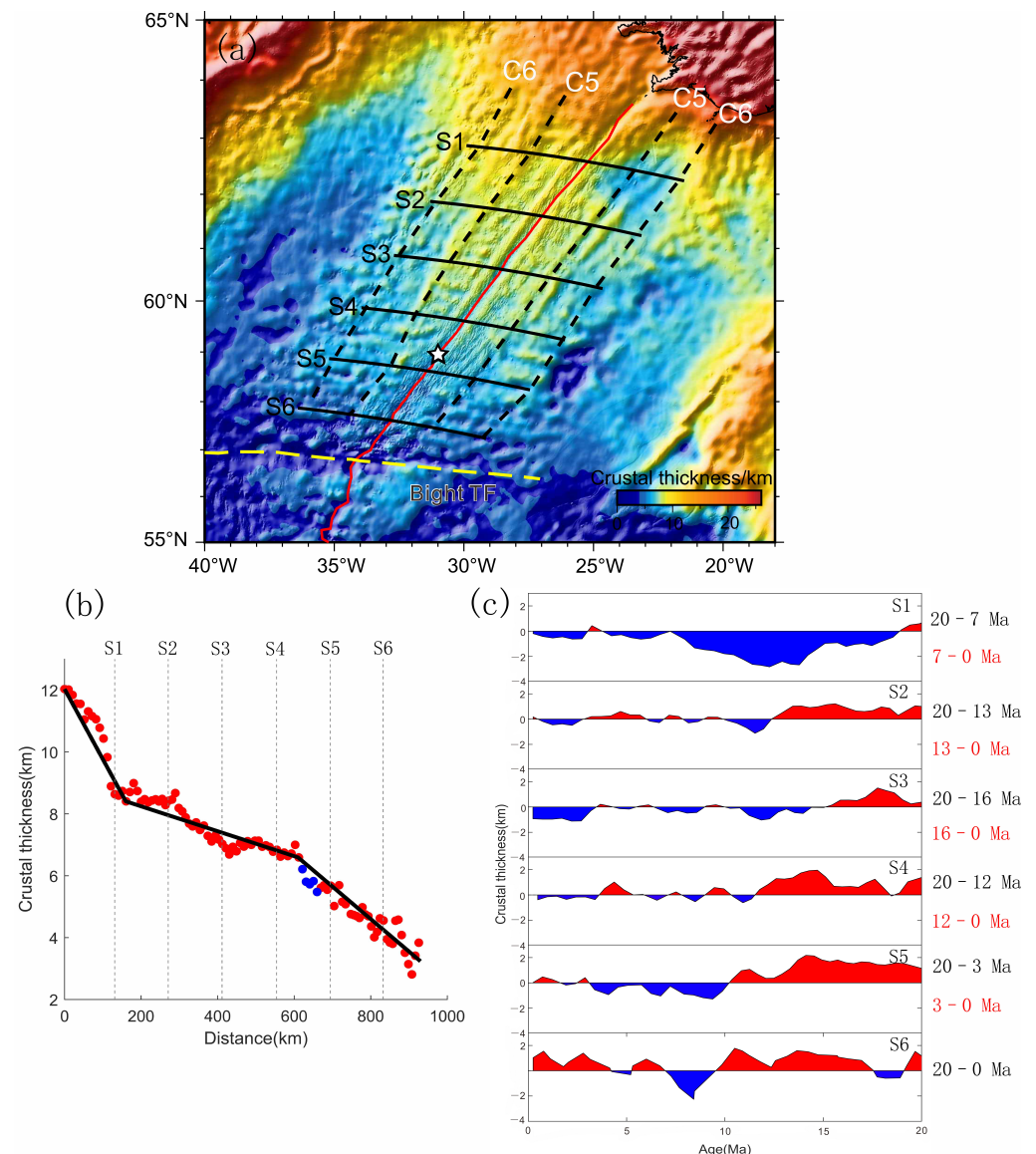


Figure 6. (a) Crustal thickness. The red line indicates the Reykjanes Ridge axis. Black lines indicate cross-axis profiles. Black dashed lines indicate magnetic strips. The white star indicates the transition from axial rise to rift valley. (b) The variation in crustal thickness along the axis. Blue scatters indicate the transition from axial rise to rift valley. (c) Asymmetry of crustal thickness (the crustal thickness of the western side of the ridge minus that of the eastern side of the ridge). The black numbers and red numbers in the sidebar represent asymmetric time periods and symmetric time periods, respectively.

The crustal thickness asymmetry curve exhibits an asymmetric–symmetric–asymmetric pattern and from north to south during 20–10 Ma and symmetric pattern during 10–0 Ma

(Figure 6c). The crustal thickness of profile S1 is asymmetric during 20–7 Ma and symmetric during 7–0 Ma. During 20–7 Ma, the crust of the western side of the ridge is 1.33 km thicker than that of the eastern side. The crustal thickness asymmetry reaches a maximum amplitude of ~ 2.82 km at 12 Ma. During 7–0 Ma, profile S1 has symmetric crustal thickness with an asymmetry amplitude of -0.33 ± 0.29 km. The asymmetry of the crustal thickness of profiles S2–S3 is significantly lower (-0.11 ± 0.40 km and -0.11 ± 0.63 km, respectively) than that of profile S1. During 20–12 Ma, the crustal thickness asymmetry of profile S4 became strong. The crust of the western side of the ridge is 1.01 km thinner than that of the eastern side. During 12–0 Ma, profile S4 has a symmetric crustal thickness (-0.03 ± 0.40 km). The crustal thickness of profile S5 is asymmetric during 20–3 Ma and symmetric during 3–0 Ma. The crustal thickness of the western side of the ridge is 1.32 km thinner during 20–10 Ma but 0.72 km thicker during 10–3 Ma than that of the eastern side. During 3–0 Ma, crustal thickness became symmetric, with an asymmetry amplitude of 0.15 ± 0.25 km. The crustal thickness of profile S6 shows a clear asymmetry. During 20–17.5 Ma, there are two asymmetry polarity reversals, with an asymmetry amplitude of 0.17 ± 0.76 km. Compared to the eastern side, the crust of the western side of the ridge was 1.10 km thinner during 17.5–9.5 Ma, 1.42 km thicker during 9.5–7 Ma, and 0.64 km thinner during 7–0 Ma.

4. Discussion

4.1. Hotspot Effect Zones: Variation in the Mantle Upwelling Pattern along the Ridge Axis

The RB and crustal thickness of the Reykjanes Ridge exhibit three distinct segments along the ridge axis (Figures 5b and 6b), which may reflect the change in the mantle upwelling pattern caused by the change in melt supply. We classified the influence of the Iceland hotspot on the spreading state of the Reykjanes Ridge into three zones (Figure 7): ultra-strong effect zone (0–160 km), strong effect zone (160–610 km), and weak effect zone (610–930 km).

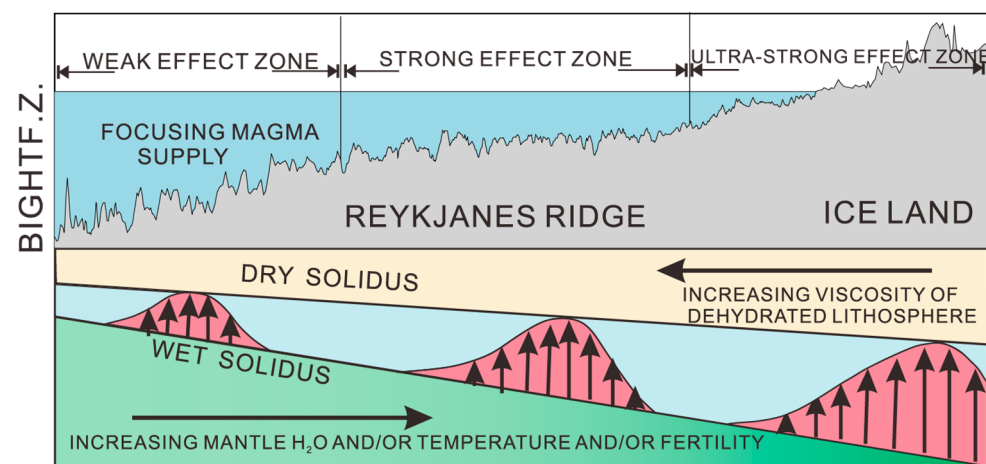


Figure 7. Conceptual model of propagating buoyant upwelling instabilities beneath Iceland and along the Reykjanes Ridge (modified from Martinez and Hey [15]). The pink area with upward-pointing arrows indicates the upwelling instability.

The topography and crustal structure of the Reykjanes Ridge are attributed to mantle plumes radiating outward from the Iceland hotspot [15,16,33]. Martinez and Hey [15] suggested that a low-viscosity melt zone persists beneath the Reykjanes Ridge, which results in the propagation of an upwelling instability cell in the shallow buoyant mantle along the ridge axis forming the V-shaped topography (Figure 7). Our results show that the ultra-strong effect zone has a thick crust (>8 km), a wide low MBA zone and a high RB (>2000 m), indicating that it has adequate mantle melt supply, a wide shallow melting zone and an isostatic compensation lithosphere. In addition, the RB and crustal thickness

have large variation gradients along the axis, which may reflect the rapid decrease in the amount of mantle material. We suggest that the Iceland plume provides an excess mantle melt material for the ultra-strong effect zone, which exceeds the consumption of mantle upwelling, thus forming a wider mantle melting zone. Nichols et al. [34] measured the water contents of basaltic glasses along the Reykjanes Ridge and Iceland and found that the concentration of water increased from 165 ppm at the southern end of the Reykjanes Ridge to between 620 and 920 ppm beneath Iceland. Such a rise in the mantle water content increases the degree of melting by up to 10%. Sufficient melt and a high melting degree reduce the mantle viscosity and enhance the buoyancy of the mantle upwelling instability cell, resulting in a faster vertical advection rate. In addition, thermal erosion can thin the thickness of the lithosphere lid so that the resistance of mantle upwelling instability cell is weakened. These obviously enhance the mantle upwelling, resulting in a large amount of mantle melt material eruption. The total amount of mantle melts and mantle volatiles decrease rapidly along the Reykjanes Ridge axis.

With the continuous decrease in melt amount and mantle volatiles, the mantle melting degree decreased and the mantle viscosity increased (Figure 7), resulting in a decrease in the axial propagation rate and vertical advection rate of the buoyant mantle upwelling instability cell. The density structure resulting from the cooling disrupts the local continuity of buoyant upwelling. Continuous upwelling requires the accumulation of residual buoyant mantle material, resulting in temporal intermittency [35,36]. These significantly weakened the mantle upwelling and reduced the amount of mantle melt material eruption and the mantle melt flux along the axis, resulting in a decrease in the variation gradient of RB and crust thickness along the axis in the strong effect zone. In the transitional zone between the strong effect zone and weak effect zone, the axial morphology of the Reykjanes Ridge changed from the high to rift valley (white star in Figures 4–6). Our results show that the crustal thickness decreases rapidly from 6.2 km to 5 km and the width of the low MBA zone narrows rapidly, indicating that the shallow mantle melting zone became narrow and the magma supply became more concentrated. In the weak effect zone, the variation gradient of RB and crust thickness becomes large again, and the spatial variation gradually increases from north to south, which is consistent with the traditional slow–ultraslow-spreading mid-ocean ridge (full spreading rate less than 50 mm/yr). Lee and Searle [37] found that the Reykjanes Ridge axis south of 59°10'N has the highest magnetization anomaly with the greatest variability and exhibits the characteristics of a slow-spreading ridge unaffected by a plume. We suggest that the mantle upwelling becomes the focusing magma supply pattern similar to that of a slow–ultraslow-spreading mid-ocean ridge. The small melt supply cannot maintain a continuous magma lens along an entire segment but forms many magma concentration centers, and the magma gradually decreases to both sides of the center [8,9]. This results in great variation in topography and crustal thickness along the axis, such as the Gakkel Ridge and Southwest Indian Ridge. In a magma-deficient environment, a large number of normal faults, including detachment faults, will develop to accommodate plate separation [38–40]. The thermal structure of the lithosphere has an important effect on mantle melting and the axial morphology of the ridge [5,41,42]. Faulting can bring seawater into the crust, thereby enhancing hydrothermal circulation and taking away more heat. This reduces mantle buoyancy and magmatism, resulting in the rapid thinning of the crust and the deepening of the axial valley.

4.2. The Effect of Magma Supply on Asymmetric Spreading in the Mid-Ocean Ridge

The asymmetric spreading of mid-ocean ridges is generally thought to be related to hot spots [18,43]. When the ridge is situated near a hotspot (usually several hundred kilometers or less), a fraction of the hot mantle material [44,45] may migrate toward the ridge along the base of the asthenosphere. During this migration, the hot mantle material heats the lithosphere on the side near the hotspot [46–48]. Melts tend to move toward the hotter and weaker side of the ridge, forming the asymmetric topography and crustal structure [25,30].

Seafloor spreading along profile S1, in the ultra-strong effect zone, was distinctly asymmetric during 20–7 Ma, but symmetric during 7–0 Ma. The evolution process of asymmetric spreading of profile S1 is in good agreement with the changes in the relative position between the Iceland hotspot and Reykjanes Ridge (Figure 1): about 30 Ma ago, the Iceland hotspot crossed Greenland; ~10 Ma, it was located near the ridge axis; subsequently, it moved eastward relative to the Reykjanes Ridge [17]. The Iceland hotspot began to interact with the Reykjanes Ridge at ~20 Ma ago and heated the lithosphere on the western side of the mid-ocean ridge. This allowed magma to distribute more west in the shallow, resulting in an asymmetric crustal structure and lithospheric thermal structure. As the Iceland hotspot came closer to the Reykjanes Ridge, it provided more mantle plume material to the ridge, resulting in a gradual increase in asymmetric amplitude. At ~10 Ma ago, the Iceland hotspot may have begun to move eastward relative to the Reykjanes Ridge. However, because the lithosphere on the western side of the ridge had previously been hotter, and received more melt distribution, the lithosphere could remain relatively hotter even if the hot spot stopped heating the western lithosphere. This causes the magma to continue to have a more westward distribution, but with the enhancement of the thermal state of the eastern lithosphere, the distribution of magma becomes more equal, resulting in the gradual symmetry of the topography and crustal structure.

Seafloor spreading along profiles S2 and S3, in the strong effect zone, was clearly symmetrical, indicating the Iceland hot spot has little influence on the asymmetric spreading of the Reykjanes Ridge. The dehydrated mantle layer above the solidus has a high viscosity, which decreases the rate of asthenosphere upwelling and results in a symmetric mantle corner flow [49]. We suggest that the decrease in the melt supply and water content of the mantle causes the dehydrated mantle layer to have sufficient viscosity so that the asymmetric asthenosphere only rises along the sloping boundary of the lithosphere, which forms a symmetric topography and crustal structure.

Seafloor spreading along profiles S4–S6 was asymmetric during 20–10 Ma, which is consistent with the existence of the transform fault (Figure 2a,b). Thus, we suggest that the asymmetric spreading may be related to the process of plate boundary reconstruction. At ~37 Ma, an abrupt change in the opening direction led to the mechanical breakup of the brittle lithosphere and the formation of a stair-step topography orthogonal to the new opening direction [15]. In the Iceland hotspot reference frame, both the North American and Eurasian plates near the Reykjanes Ridge have a westward motion component, which results in an eastward buoyant upwelling lag [15,50,51]. Within this plate boundary reconstruction, new spreading segments may develop more easily toward the eastern side of the ridge or directly over the east-lagging melt zone [52], resulting in a thicker crust and higher RB. With the elimination of ridge shifts and the crustal segmentation, the shallow mantle upwelling regathers to the linear spreading center. Profile S4 is in the strong effect zone and should also have a more viscous dehydrated mantle, forming a symmetrical topography and crustal structure like profiles S2–S3.

However, profiles S5–S6 located in the weak effect zone still show obvious asymmetric spreading after the elimination of the transform faults (~10 Ma), which may be related to the focused magma supply pattern. Unlike the ultra-effect zone and the strong effect zone, the weak effect zone cannot form a continuous and stable magma lens due to less magma supply, and the magma concentration center may be located on one side of the mid-ocean ridge axis. Electromagnetic data showed that mantle upwelling of the slow-ultraslow-spreading mid-ocean ridge concentrates in a narrow and highly asymmetric region [53]. This indicates that although magma will eventually converge to the spreading center in the shallow areas, deep mantle convection and crystallization heat release processes may be concentrated on one side [31], thus affecting magma distribution [30]. We suggest that during 10–5 Ma, the fault developed on the west side of the ridge in profile S5 to elevate the terrain. At the same time, the magma center may be located on the west side of the ridge and provide more melts, so that topographic uplift is compensated to a certain extent. Therefore, the RB on the west side of the ridge has larger asymmetric amplitude than that

of the crust thickness. In different magmatic activity cycles, magma distribution and fault development may be concentrated on different sides of the mid-ocean ridge axis, which leads to the continuous reversal of asymmetric polarity of RB and crustal thickness.

5. Conclusions

(1) The influence of melt supply on the spreading state of the Reykjanes Ridge is not simply linear and can be divided into three zones: an ultra-strong effect zone (0–160 km), a strong effect zone (160–610 km), and a weak effect zone (610–930 km).

(2) In the ultra-strong effect zone, the excess melt supply and higher melting degree result in a wide shallow mantle melting zone and large melt eruption. During 20–12 Ma, the Iceland hotspot heated the lithosphere on the western side of the ridge and induced the preferential distribution of melt to the western side, resulting in asymmetric spreading. During 12–0 Ma, the thermal state of the lithosphere on the eastern side of the ridge was enhanced and the amount of magma distributed on either side of the ridge gradually equalized, resulting in symmetric spreading.

(3) In the strong effect zone, the decrease in melt supply and melting degree reduces the melt eruption and enhances the viscosity of the dehydrated mantle layer. This forces an asymmetric asthenosphere increase along the sloping boundary of the lithosphere forming a symmetric topography and crustal structure. Eastward buoyant upwelling lag means that new spreading segments may develop more easily toward the eastern side of the ridge or directly over the east-lagging melt zone, resulting in a thicker crust and higher RB on the eastern side of the ridge.

(4) In the weak effect zone, the pattern of magma upwelling becomes the focusing magma supply pattern similar to that of a slow–ultraslow-spreading mid-ocean ridge. This resulted in a strong spatial variation in RB and crustal thickness along the ridge axis. The concentration of tectonism and/or magma on one flank of the ridge resulted in asymmetric RB and crustal thickness.

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