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# Age of the Canada Basin, Arctic Ocean: Indications From High-Resolution Magnetic Data

Tao Zhang<sup>1,2</sup>, Jérôme Dyment<sup>3</sup>, and Jinyao Gao<sup>1,2</sup>

<sup>1</sup>Key Laboratory of Submarine Geosciences, State Oceanic Administration,  
Hangzhou 310012, China.

<sup>2</sup>Second Institute of Oceanography, Ministry of Natural Resources, Hangzhou  
310012, China.

<sup>3</sup>Institut de physique du globe de Paris, Université de Paris, CNRS, F-75005  
Paris, France.

## Key Points:

- High-resolution deep-tow and sea surface magnetic data in the ice-covered Canada Basin of the Arctic are presented.
- The crustal age of the Canada Basin is 139.5-128.6 Ma (142.4-132.8 Ma), with a spreading rate of ~32 (38) mm/yr.
- The opening of the Canada Basin was roughly contemporaneous with the closure of the ancient South Anyui Ocean.

**Abstract:** The origin and history of the Amerasia basin are long-running debates, which hinder our knowledge of the Mesozoic tectonic configurations and geodynamic processes in the Arctic. This lack of knowledge is due in part to the paucity of accurate magnetic data in the ice-covered basin. Here, we identify the crustal age of the Canada Basin, a major part of the Amerasia Basin, through high-resolution

23 deep-tow and sea surface magnetic data. The best fit of the four pairs of magnetic  
24 lineations revealed by the high-resolution magnetic data is 139.5-128.6 Ma (or  
25 142.4-132.8 Ma, depending on the geomagnetic polarity timescale). The crustal age  
26 provides crucial constraints on the evolution of the circum-Arctic tectonic features  
27 and generally supports the hypothesis that the opening of the Amerasia Basin is  
28 related to the subductions during the closure of the South Anyui Ocean.

29

### 30 **Plain Language Summary**

31 The Amerasia Basin of the Arctic Ocean is one of the last major puzzles of the plate  
32 reconstructions, due to the lack of age knowledge. The identification of magnetic  
33 anomalies is the routine method of acquiring the age of oceanic crust, yet floating ice  
34 in the basin makes it difficult to obtain magnetic data. We collected deep-tow  
35 magnetic data in the basin by lowering a magnetic sensor to a depth of ~2000 m,  
36 which provides high-resolution data and avoids floating ice. The identified magnetic  
37 lineations indicate the Amerasia Basin opened at 139.5-128.6 Ma (or 142.4-132.8 Ma,  
38 depending on the geomagnetic polarity timescale). The contemporaneous closure of  
39 the ancient South Anyui Ocean (~1000 km in the south) may have provided space for  
40 the opening of the Amerasia Basin. This interpretation then generally supports the  
41 existing hypothesis that the opening of the Amerasian Basin is associated with the  
42 subduction process in the South Anyui Ocean. Nevertheless, a more sophisticated  
43 geodynamical model is still needed.

## 44 **1 Introduction**

45 The opening of the Amerasia Basin shaped the Mesozoic configuration of major  
46 circum-Arctic geological features such as Arctic Alaska, Chukotka, and Arctic Canada.  
47 A robust tectonic model of the Amerasia Basin would yield insights into Arctic  
48 paleogeography, paleoclimate, the driving forces of the opening of the basin, and  
49 resource exploration within the dozens of circum-Arctic sedimentary basins  
50 [Shephard *et al.*, 2013]. However, the nature and age of the Amerasia Basin have been  
51 debated for decades [e.g., Embry, 1990, 2000; Grantz *et al.*, 1998, 2011; Lane, 1997;  
52 Miller *et al.*, 2006, 2017]. Recently, the oceanic nature of the crust in the Canada  
53 Basin, which forms the major part of the Amerasia Basin, was revealed by refraction  
54 seismic data [Chian *et al.*, 2016]. The roughly symmetric oceanic lithosphere about a  
55 N-S trending gravity low indicates that the Canada Basin was an E-W spreading  
56 oceanic basin. Nevertheless, the age of the Canada Basin remains elusive, with an  
57 inferred range of 160-72 Ma from investigations on the stratigraphy and volcanism  
58 along its margins and sparse geophysical data in the basin [e.g., Alvey *et al.*, 2008;  
59 Chian *et al.*, 2016; Døssing *et al.*, 2013; Embry, 1990; Gaina *et al.*, 2014; Grantz *et al.*,  
60 2011; Lane, 1997; Langseth *et al.*, 1990; Miller *et al.*, 2006, 2017; Taylor *et al.*, 1981].  
61 In particular, the low-resolution airborne magnetic data in the basin with an  
62 exceptionally thick (4-11 km) sedimentary cover impede the definitive identification  
63 of crustal age with magnetic anomalies. In this paper, we present recently sampled,  
64 high-resolution, deep-tow and sea surface magnetic data in the ice-covered Canada

65 Basin (Figure 1). We identify the crustal age of the Canada Basin through the  
66 amplitude, shape, and pairs of the magnetic lineations. We discuss the relationships  
67 between seafloor spreading of the Canada Basin and regional unconformities along  
68 the margins of the Canada Basin. We further suggest that the opening of the Canada  
69 Basin may be kinematically and geodynamically related to the demise of the South  
70 Anyui Ocean, which partly supports the existing models [e.g., *Koulakov et al.*, 2013;  
71 *Kuzmichev*, 2009].

## 72 **2 Geological Settings**

73 Located at the center of the Arctic region, the Amerasia Basin is bounded by Arctic  
74 Alaska, Chukotka, East Siberian Shelf, Lomonosov Ridge, and Arctic Canada (Figure  
75 1). Numerous models have been proposed to explain the formation of the Amerasia  
76 Basin [e.g., *Lawver and Scotese*, 1990; and references therein]. The floating ice, thick  
77 sediment, and presence of the High Arctic Large Igneous Province make it difficult to  
78 validate those models. Among them, the generally accepted anticlockwise rotation  
79 model proposes that the Chukotka-Alaska region rotated from Arctic Canada with a  
80 pole of rotation located near the Mackenzie Delta in Early Cretaceous [*Carey*, 1955;  
81 *Embry*, 1990, 2000; *Embry and Dixon*, 1990; *Grantz et al.*, 1998, 2011; *Halgedahl*  
82 *and Jarrard*, 1987; *Mickey et al.*, 2002]. Recent seismic [*Chian et al.*, 2016] and  
83 potential field data [*Andersen et al.*, 2010; *Gaina et al.*, 2011] from the Canada Basin  
84 revealed that the oceanic crust is roughly symmetrical about the N-S trending relict  
85 axis shown by a linear gravity low. The extent of oceanic crust and location of the

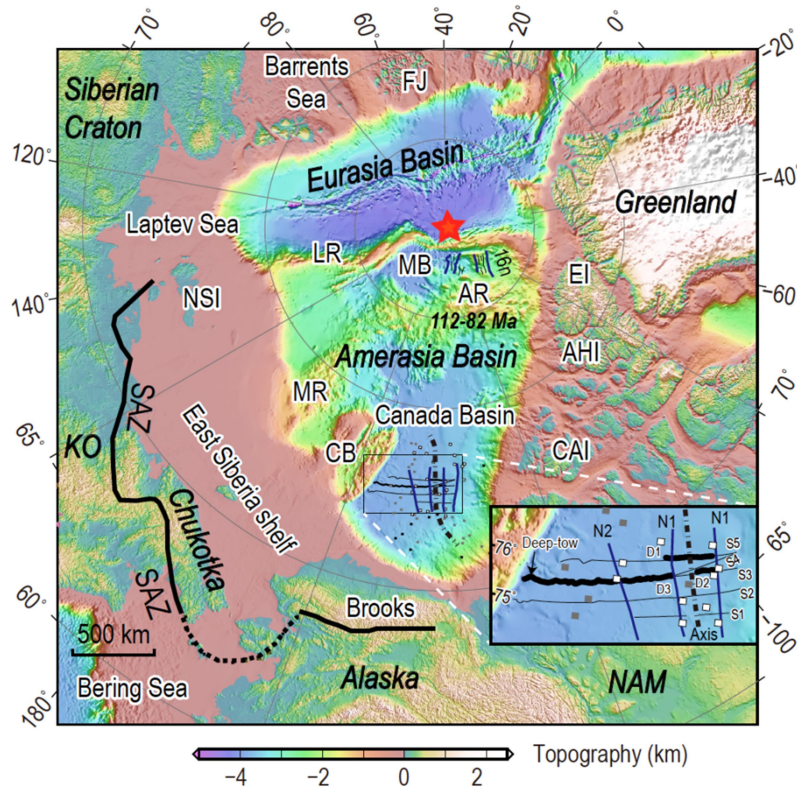
86 relict axis are broadly consistent with the depictions in the rotation model. However,  
87 this model is challenged by alternative models [e.g., *Chian et al.*, 2016; *Døssing et al.*,  
88 2018; *Hutchinson et al.*, 2017; *Koulakov et al.*, 2013; *Lane*, 199], in part due to the  
89 uncertain age of the Amerasia Basin.

90 In the anticlockwise rotation model, the Amerasia Basin opened in two stages  
91 [e.g., *Grantz et al.*, 2011]. The age of initial rifting ranges from the Early Jurassic  
92 (Hettangian) [*Hubbard et al.*, 1987] to early Middle Jurassic (Aalenian) [*Embry and*  
93 *Dixon*, 1990; *Mickey et al.*, 2002]. The second stage or main stage of opening is no  
94 older than Oxfordian-Tithonian (158-145.5), supported by the synrift sequence  
95 onlapping late Oxfordian-Tithonian strata at the Northwind Ridge [*Grantz et al.*,  
96 1998]. Nevertheless, the age of seafloor spreading is still highly controversial. *Grantz*  
97 *et al.*[2011] proposed that seafloor spreading was initiated in Hauterivian (~131 Ma)  
98 after correlating the beds of the late synrift sequences to the widely distributed Lower  
99 Cretaceous unconformity. *Halgedahl and Jarrard* [1987] suggested that the Alaskan  
100 North Slope was still adjacent to the Arctic Islands in Valanginian based on the  
101 paleomagnetic data from the North Slope Kuparuk Formation. *Embry and Dixon*  
102 [1990] interpreted the Albian-Cenomanian unconformity in the Sverdrup Basin as the  
103 breakup unconformity. Based on a petrologic study, *Miller et al.* [2017] suggested that  
104 spreading in the Amerasia Basin may have ended at ~90 Ma.

105 Several interpretations of the magnetic anomaly have been proposed in the  
106 Canada Basin based on available low-resolution airborne magnetic data. However, the

107 low amplitude and limited two pairs of conjugate positive magnetic anomalies made  
108 any oceanic crustal age identification uncertain and unreliable [*Chian et al.*, 2016;  
109 *Gaina et al.*, 2014; *Grantz et al.*, 2011; *Taylor et al.*, 1981]. *Taylor et al.* [1981]  
110 tentatively suggested that the crustal age of the Canada Basin ranges from the earliest  
111 Late Jurassic to Valanginian (CM25-CM12, 160-136 Ma). *Grantz et al.* [2011] and  
112 *Chian et al.* [2016] proposed similar identifications of chrons CM4n to CM2n  
113 (131-127.5 Ma), with a spreading rate up to 75 mm/yr. *Gaina et al.* [2014] identified  
114 CM16-CM4 (137.8-126.5 Ma) according to Channell (1995) and found a spreading  
115 rate of ~30 mm/yr for the younger stage of seafloor spreading in the northern part of  
116 the Canada Basin.

117 The morphology of the rift valley offers an independent constraint on the  
118 spreading rate of the Canada Basin. Reflection seismic data indicate that the valley of  
119 the relict ridge axis has depths of 1.0-1.5 km and widths of 30-40 km [*Chian et al.*,  
120 2016; *Grantz et al.*, 2011], which is typical for a slow spreading (< 75 mm/yr) ridge  
121 axis. The rough basement relief and relative thin crust (4-7 km) [*Chian et al.*, 2016]  
122 are also consistent with the characteristics of the slow to ultraslow spreading ridges  
123 [*Dick et al.*, 2003; *Malinverno*, 1991].



124

125 **Figure 1.** Topography of the Circum-Arctic Region. The location of SAZ is from *Shephard et al.*  
 126 [2013]. The location of the survey lines, relict ridge axis, and broad magnetic highs (N1 and N2)  
 127 are shown in the inset. AHI = Axel Heiberg Island; AR = Alpha Ridge; CAI = Canadian Arctic  
 128 Islands; CB = Chukchi Borderland; EI = Ellesmere Island; FJ = Franz Josef Land; KO =  
 129 Kolyma-Omolon; LR = Lomonosov Ridge; MB = Makarov Basin; MR = Mendeleev Ridge; NAM  
 130 = North American craton; NSI = New Siberian Islands; SAZ = South Anyui suture Zone.

### 131 **3 Data acquisition and processing**

132 We use one deep-tow magnetic profile (consisting of three sections D1-D3) and five  
 133 sea surface magnetic profiles (S1-S5) at 75°-76°N collected by Icebreaker “Xue Long”  
 134 in 2014 and 2016 to 2017, respectively (Figure 1). Most profiles are perpendicular to  
 135 the N-S trending gravity anomaly low near 142°W. In the summer of 2014, ~500 km



136 deep-tow magnetic data were sampled by a MarineMagnetics™ Overhauser  
137 magnetometer with a sensitivity of 0.015 nT mounted on a titanium-alloy frame and  
138 towed ~1.3 km above the seafloor at a speed of 2-3 knots (supporting information  
139 Figure S1). To measure the depth of the frame, a pressure sensor Sea-Bird™ SBE was  
140 mounted on the wire at 5 m above the frame. Controlled by the payoff of the winch,  
141 the depth of the magnetic sensor is ~2.5 km in average and varies within a relatively  
142 limited range of  $\sim\pm 0.5$  km (Figure S1).

143       The deep-tow magnetic data are processed with the following five steps. (1)  
144 Magnetic data are merged with GPS position data and sensor depth data (Figure S1).  
145 (2) The International Geomagnetic Reference Field (IGRF) [Thébault *et al.*, 2015] is  
146 removed. (3) The diurnal variations are removed. (4) The data are resampled to  
147 equally spaced (50 m) points. (5) A Fourier transform method is used to  
148 upward-continue the data from an uneven level to a constant depth of 2 km below  
149 sea surface [Guspi, 1987]. Among these steps, steps 1, 2, and 4 have little effect on  
150 the characteristics of magnetic data. For step 3, we use the magnetic variations  
151 recorded at the Barrow and Resolution Bay magnetic observatories. Since the  
152 survey area is approximately one-fourth between the two observatories, we use a  
153 weighted average of Barrow (3/4) and Resolution Bay (1/4) magnetic data for the  
154 diurnal correction. During collection of the deep-tow magnetic data, the daily  
155 magnetic variation had amplitudes up to  $\pm 100$  nT (Figures S2 and 2), with a  
156 standard deviation of 40.8 nT. The daily magnetic variation is smaller but

157 comparable to the collected magnetic data with amplitudes up to 400 nT and an  
158 STD of 85.8 nT (Figure S2). We remove all data with diurnal variation exceeding  
159  $\pm 50$  nT, although the deep-tow magnetic data are probably much less affected by  
160 the ionospheric noise since this noise is attenuated by the conductive sea water  
161 above [Miller, 1977]. After the diurnal correction, the upward-continued magnetic  
162 anomaly to sea level fits well with the sea surface magnetic anomaly along the  
163 same track collected in 2016 (Figure 2), indicating that the diurnal correction  
164 efficiently reduce the associated external magnetic variations. In step 5, we remove  
165 the signals with wavelengths longer than 100 km or shorter than 2 km and  
166 upward-continued the magnetic data to 2 km below the sea surface, to obtain the  
167 deep-tow magnetic anomaly  $\sim 2$  km above the sea floor and  $\sim 7.5$  km above the  
168 igneous crust [Mosher and Hutchinson, 2019].

169       Along the deep-tow magnetic survey in 2014, most tracks ( $> 60\%$ ) were covered  
170 by floating ice. Nevertheless, in the areas with light ice-conditions, we collected  $\sim 110$   
171 km of sea surface magnetic data with a Cesium magnetometer towed 450 m behind  
172 the *R/V "Xue Long"*. In 2016 and 2017, the ice conditions were rather light ( $\sim 20\%$  ice  
173 coverage), which allowed us to collect  $\sim 1400$  km of sea surface magnetic data. The  
174 associated International Geomagnetic Reference Field model [Thébault *et al.*, 2015]  
175 and the diurnal variations are also removed from the sea surface magnetic data  
176 (Figure S2). For comparison, we also include the airborne magnetic anomaly data of  
177 Taylor *et al.* [1981] in Figure 2.

178 **4 Data presentation**

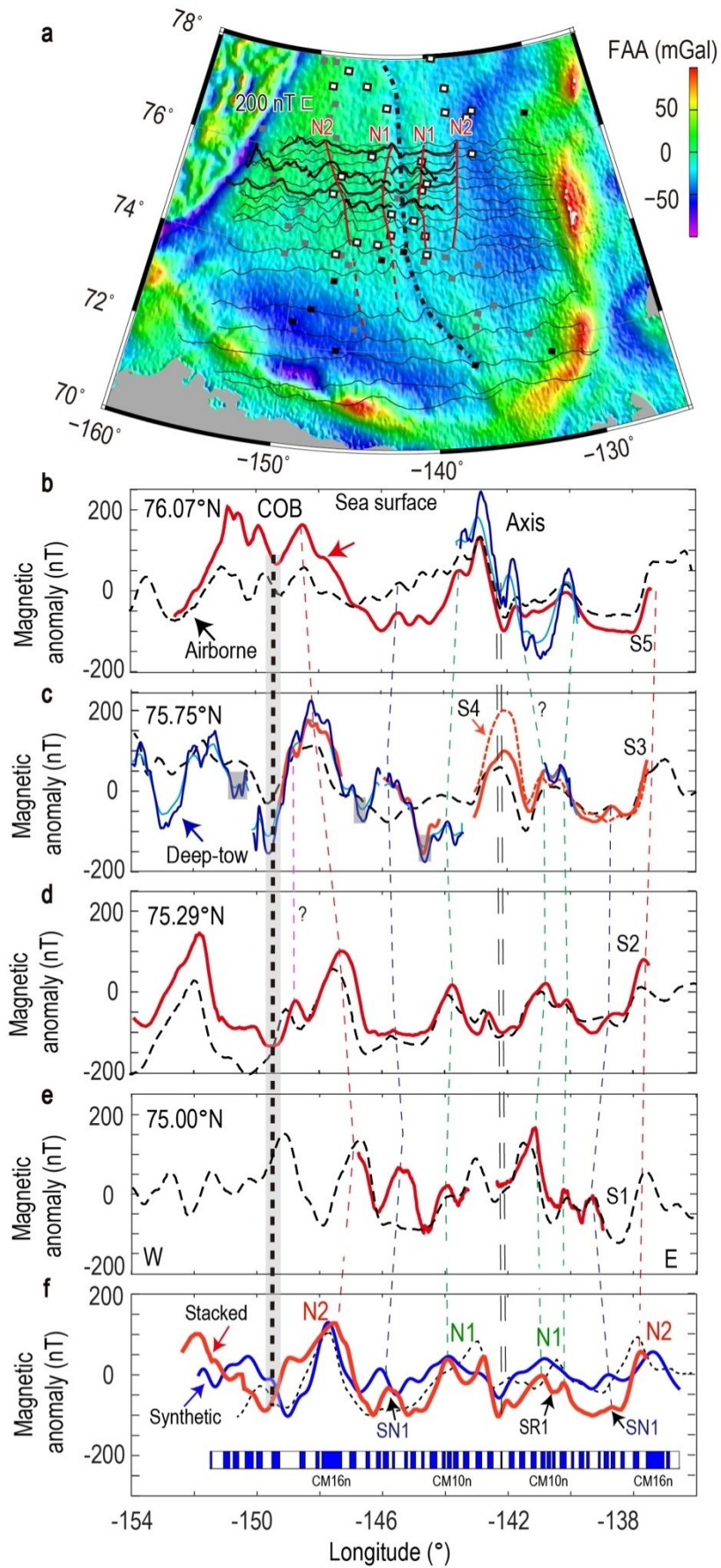
179 Two paired coherent broad magnetic highs are observed by the deep-tow, sea surface,  
180 and airborne magnetic data (Figure 2). Here, we name them normal 1 (N1) and  
181 normal 2 (N2) anomalies for the broad magnetic high close to and away from the  
182 relict ridge axis, respectively. On the western flank, N1 with peak-to-trough  
183 amplitudes of 150-200 nT straddles 50-80 km in the sea surface magnetic data. N2 on  
184 the western flank has larger amplitudes (up to 300 nT) and broader widths (90-120  
185 km) than N1. The amplitudes of N1 and N2 are comparable to the magnetic signals  
186 observed at other slow to ultraslow spreading ridges [e.g., *Gee and Kent, 2007*]. The  
187 two paired broad magnetic highs are roughly symmetrical with respect to the fossil  
188 axis, suggesting that these magnetic anomalies may reflect geomagnetic reversals and  
189 seafloor spreading in the Canada Basin. In addition, the power spectrum analysis of  
190 the deep-tow magnetic data and surface magnetic data suggests that the magnetic  
191 source layer is situated ~10 km below the sea surface and then resides within the  
192 igneous crust (Figure S3). Furthermore, the magnetic lineations are independent of the  
193 gravity anomalies (Figure S4), which further implies that the magnetic lineations are  
194 not associated with variations in the lithospheric structure. Therefore, the paired  
195 magnetic lineations (N1 and N2) reflect spatial variations in crustal magnetization  
196 associated with the record of magnetic field reversals within the oceanic crust.

197 Two pairs of magnetic anomalies are not sufficient for a unique, unambiguous  
198 correlation with the geomagnetic polarity time scale (GPTS). In addition to N1 and

199 N2, some previously undetected, spatially coherent, low amplitudes and  
200 short-wavelength magnetic anomalies are also observed in the high-resolution  
201 deep-tow magnetic data and to a lesser extent in the sea surface magnetic data (Figure  
202 2). Such small magnetic anomalies could be related either to short geomagnetic  
203 polarity intervals, excursions (i.e. aborted reversals) or paleointensity variations,  
204 which would all be recorded in a similar way on both side of the (now fossil) ridge  
205 axis, or by crustal tectonic processes, short-period external magnetic field fluctuations,  
206 and/or artifacts during the data acquisition, which may have a different distribution.  
207 Among them, two short-wavelength low-amplitude magnetic anomalies could be  
208 ascribed to the geomagnetic variations, more likely to field reversals, considering  
209 their repeatability and consistency between profiles and their presence on both flanks  
210 of the relict ridge. Near the center of the broad magnetic low intervening N1 and N2  
211 on the western flank, a narrow magnetic high with an amplitude of ~200 nT is  
212 observed on the deep-tow magnetic profile (Figure 2). The amplitude of this anomaly  
213 decreases to ~50-100 nT on the sea surface magnetic profiles. On the conjugate  
214 eastern flank, a similar magnetic high is also present along three of the five sea  
215 surface profiles. Thus, this magnetic high may be ascribed to a short normal polarity  
216 interval in a relatively long period dominated by reversed polarities and is termed  
217 small normal anomaly 1 (SN1). Besides, a magnetic low with an amplitude of ~30 nT  
218 is observed near the center of N1 on the deep-tow magnetic anomaly on the eastern  
219 flank (Figure 2c). This magnetic low is also observed on almost all sea surface

220 magnetic profiles on the eastern flank of the ridge axis, except on profileS5. The  
221 consistency between profiles on the eastern flank suggests that this magnetic low is  
222 associated with one or a series of short reversed polarity intervals (termed SR1) in a  
223 relatively long period dominated by normal polarity.

224 Encouraged by the consistency of the magnetic anomalies between profiles and  
225 on conjugate flanks, we stack all the sea surface data and the upward-continued  
226 deep-tow data to the sea surface in an attempt to enhance the signal/noise ratio and  
227 better characterize the magnetic anomalies (Figure 2f). We also stack the airborne  
228 magnetic data by correlating N1 and N2 between these profiles (Figure 2f). Both  
229 stacked data show that N2 has higher amplitude than N1 on both flanks of the fossil  
230 ridge. We then identify the crustal age by fitting the observed anomaly and stacked  
231 anomaly with synthetic magnetic anomalies computed from GPTSs.



232

233 **Figure 2.** Deep-tow and sea surface magnetic anomalies in the Canada Basin. (a) Deep-tow, sea

234 surface, and airborne magnetic anomalies along their tracks. The background is based on the  
235 satellite-derived free air anomaly data [Sandwell *et al.* 2014]. The continental, transitional, and  
236 oceanic crust identified from sonobuoy data [Chian *et al.*, 2016] are shown in black, gray, and  
237 white squares, respectively. The relict ridge axis is marked with a dashed line. (b-e) Deep-tow, sea  
238 surface, and airborne magnetic anomalies at different latitudes. The upward continued deep-tow  
239 data to sea surface and to 2000 m below sea surface are shown in light blue and blue, respectively.  
240 The sea surface and airborne magnetic data are shown in red and black, respectively. The data  
241 associated with diurnal variation  $> 50$  nT are marked with gray boxes. The profiles S2-S5 and  
242 profile S1 are collected in 2016 and 2017, respectively. No reduction to the pole is necessary, as  
243 the data are collected at high latitude. (f) Stacked sea surface (red), stacked airborne (black), and  
244 best-fitting synthetic (blue) magnetic anomalies. The consistent magnetic anomalies between  
245 profiles are linked with dashed lines. The magnetic bodies in MHTC12 [Malinverno *et al.*, 2012]  
246 that produce the synthetic magnetic anomaly at the depth of basement are also shown. For more  
247 information on the correlation between the stacked magnetic anomaly and synthetic magnetic  
248 anomalies, see the text and Figures S5-S6.

## 249 **5 Identification of magnetic anomalies**

250 Based on the onshore and along-margin geological evidences, the age of the Canada  
251 Basin is not older than the Late Jurassic ( $\sim 160$  Ma) and not younger than the Late  
252 Cretaceous ( $\sim 72$  Ma) [e.g., Embry, 1990; Gaina *et al.*, 2014; Grantz *et al.*, 2011;  
253 Miller *et al.*, 2006, 2017; Taylor *et al.*, 1981]. Since the deep-tow magnetic anomalies  
254 are rather strong (up to 400 nT) and well-marked, we do not expect that they formed  
255 during the so-called “Jurassic quiet zone” ( $> 157$  Ma) characterized by numerous  
256 polarity reversals and a weak geomagnetic intensity, or during the Cretaceous quiet  
257 zone ( $\sim 120.6$  (124)-83 Ma) characterized by a constant (or very dominant) normal

258 polarity [Granot *et al.*, 2012]. Neither the relatively long polarity intervals of chrons  
259 C32n-C33n nor the intervening excursions or intensity variations (the “tiny wiggles”  
260 depicted by *Bouligand et al.* [2006]) produce the observed magnetic features. We  
261 therefore restrict our investigations and compare the observed and stacked anomalies  
262 with synthetic anomalies generated using the M-Series GPTS between the Jurassic  
263 quiet zone and the Cretaceous quiet zone.

264       The positions of the continent-ocean boundaries on both flanks of the relict ridge  
265 axis are derived from the sonobuoy data (Figure 1) [*Chian et al.*, 2016]. The location  
266 of the relict ridge axis is indicated by the ~15 mGal N-S trending gravity low  
267 at ~142°W. We select the upper boundary of layer 2 (igneous basement) from the  
268 sonobuoy data as the upper limit of the magnetic source [*Chian et al.*, 2018; *Mosher*  
269 *and Hutchinson*, 2019]. To produce sea surface anomalies with amplitudes up to 300  
270 nT at ~9.5 km above the magnetic sources, the thickness and magnetization of the  
271 magnetic sources are assumed to be 1 km and 5 A/m, respectively. The mean  
272 paleolatitude (72°N) of the magnetized bodies is based on the paleolatitude  
273 (~68°-76°N) of the Alaskan North Slope between 120 Ma and 150 Ma [*Seton et al.*,  
274 2012]. Since there are still controversies about the age of CM0r, we adopt two GPTSs,  
275 MHTC12 [*Malinverno et al.*, 2012] for which the age of CM0r is ~120.6 Ma, and  
276 GTS2012 [*Ogg*, 2012] for which it is ~125 Ma, to compute synthetic magnetic  
277 anomalies (Figures S5 and S6).

278       We adopt both the cross-correlation [*DeMets et al.*, 2010] and visual inspection



279 methods to determine the best-fitting polarity reversal sequences (Figures S5 and S6).  
280 Based on the least-square fitting criteria, the cross-correlation method quantitatively  
281 compares the amplitude and shape of the part of the stacked magnetic anomalies  
282 between the two broad magnetic highs (N1 and N2) with synthetic data (Figure S5).  
283 In visual inspection method, we fit N1, N2, SN1, and SR1 of the observed data with  
284 the synthetic magnetic anomalies.

285 Both methods give similar results: The synthetic data produced by the  
286 CM7r-CM16n (CM17n?) sequences for both GPTSs are the best fit of the stacked  
287 data (Figures 2f and S6). A series of normal polarity intervals of CM9n-11n and the  
288 long CM16n produce N1 and N2, respectively. The CM13n and negative polarity  
289 intervals between CM9 and CM11 are associated with the low-amplitude magnetic  
290 high (SN1) intervening N1 and N2 and the low-amplitude magnetic low (SR1) near  
291 the center of N1, respectively.

292 Therefore, the crustal age of the Canada Basin could be 139.5-128.6 Ma  
293 (142.4-132.8 Ma) according to the MHTC12 (GTS2012), and seafloor spreading  
294 occurred between Berriasian and Early Hauterivian (Figure S6). Near 75°N, the  
295 associated full spreading rate was ~32 (38) mm/year at the beginning of spreading and  
296 slowed down to ~30 (30) mm/year in the last ~3 Ma before the cessation. Seafloor  
297 spreading is slightly asymmetrical, with rates 5% faster on the western flank. Since  
298 the distance between the lineations N2 on both flank is slightly larger in the north than  
299 in the south (Figure 2a), the average spreading rates at the northern (~76.5°N) and

300 southern ( $\sim 74^\circ\text{N}$ ) limits of N2 are inferred to be 34.5 (39.0) and 28.3 (31.9) mm/year,  
301 respectively. The slow spreading rate is consistent with the presence of the  $\sim 1.5$   
302 km-deep rift valley and the 4-6 km thin crust in the Canada Basin. Since there are no  
303 robust constraints such as fracture zones on the spreading direction, we calculate the  
304 spreading rate assuming an orthogonal spreading. Recently, opening models of the  
305 Amerasia Basin involving a strike-slip component (oblique spreading or  
306 transtensional deformation) have been proposed based on northeast-trending structural  
307 fabrics [Døssing *et al.*, 2018; Hutchinson *et al.*, 2017]. The highly oblique spreading  
308 (up to  $\sim 50^\circ$ ) requires a spreading rate  $\sim 1.5$  times faster than that estimated from the  
309 magnetic lineations. In this case, the spreading rate we estimated corresponds to the  
310 effective spreading rate termed by Dick *et al.* [2003] and still matches the deep rift  
311 valley and thin crust in the Amerasia Basin. Among numerous proposed crustal ages  
312 from magnetic data, our result agrees better with the crustal age of 137.8-126.5 Ma  
313 proposed by Gaina *et al.* [2014]. As the seismic reflection data show that the synrift  
314 sequences overlap late Oxfordian-Tithonian ( $\sim 158$ -145.5 Ma) marine shelf or shelf  
315 basin deposits in three piston cores on the Northwind Ridge [Grantz *et al.*, 1998,  
316 2011], we further infer that the main stage of opening of the Canada Basin may had  
317 been fulfilled by rifting from late Oxfordian-Tithonian to Berriasian and the  
318 consequent seafloor spreading until early Hauterivian.

## 319 **6 Discussion**

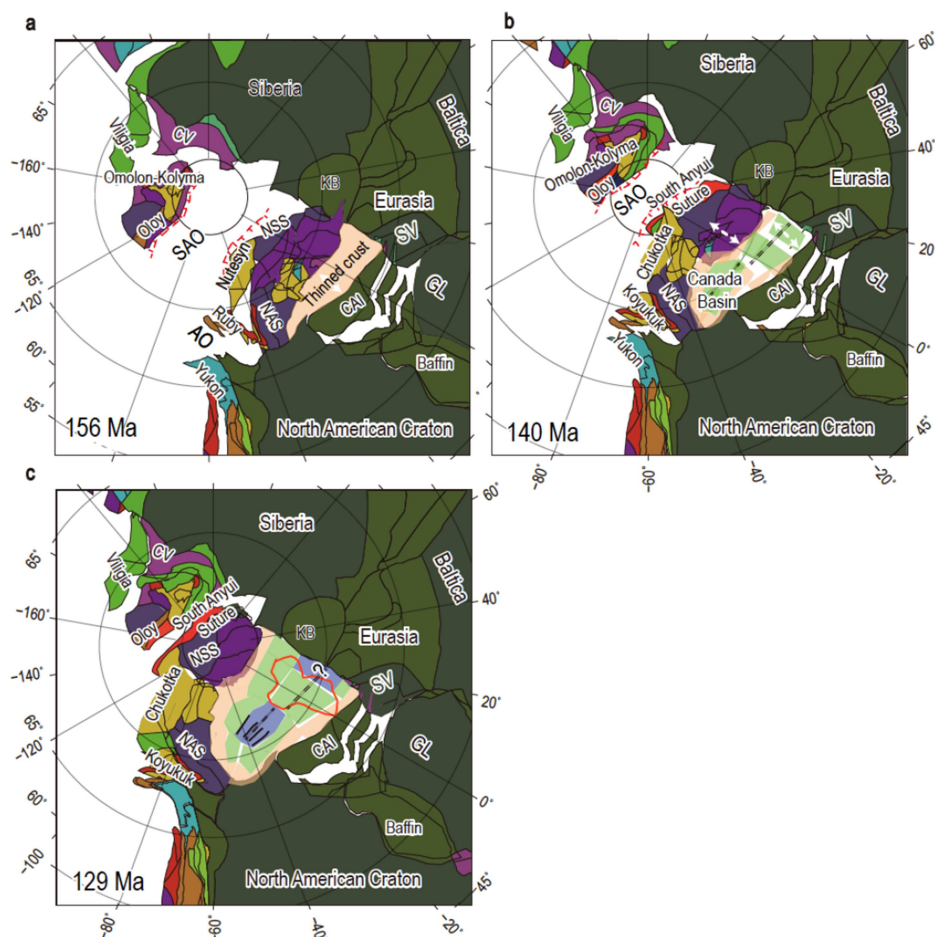
320 Three main regional unconformities (late Callovian-early Oxfordian, late

321 Hauterivian, and mid-Aptian) were interpreted as the breakup unconformity and were  
322 used to date the initial seafloor spreading of the Canada Basin by various authors  
323 [*Embry and Dixon, 1990; Grantz et al., 2011; Grantz and May, 1982; Hubbard et al.,*  
324 *1987*]. However, the seafloor spreading during Berriasian-early Hauterivian of the  
325 Canada Basin suggests that the relationships between those regional unconformities  
326 along the margins and the initiation of seafloor spreading in the Amerasia Basin  
327 cannot be steadily associated. The thinned continental crust and transitional crust in  
328 the Canada Basin could be as wide as 300 km [*Chian et al., 2016*]. During the  
329 formation of such wide margins, sequential active faulting may have migrated toward  
330 the future oceanic crust [*Brune et al., 2014*]. Therefore, further evidences, such as the  
331 reflection seismic data from the Canada Basin to the areas of the regional  
332 unconformities, are needed to address the relationship between the regional  
333 unconformities and the breakup event. Even so, part of the rifted margins of the  
334 Canada Basin were already subaquatic at the time of the breakup [*Grantz et al., 2011*],  
335 which may further obscure the identification of the breakup unconformity [*Franke,*  
336 *2013*].

337         The crustal age of the Canada Basin also provides further kinematic and  
338 geodynamic implications for the Mesozoic circum-Arctic region. Between Late  
339 Jurassic and Early Cretaceous, the Arctic Alaska and Chukotka blocks experienced  
340 intense tectonic activity, including collision between the Alaska- Chukotka and the  
341 Kolyma-Omolon blocks to the south, the associated closure of the South Anyui Ocean

342 (SAO), and the closure of the Angayucham Ocean (Figure 3). The subductions that  
343 consumed the SAO have been postulated as the source of driving force for the  
344 opening of the Amerasia Basin based on tomography images [e.g., *Gaina et al.*, 2014;  
345 *Koulakov et al.*, 2013] and the range of South Anyui suture zone [*Kuzmichev*, 2009].  
346 Our results further show that the opening of the Canada Basin was roughly  
347 contemporaneous with the closure of the SAO and the associated subductions. At the  
348 beginning of the main rifting stage of the Canada Basin (~158 Ma), the closure of  
349 SAO was initiated by southward and northward subductions (Figure 3a), as indicated  
350 by the ages of the Oloy Arc (~160–140 Ma) in the south and the Nutesyn Arc  
351 (~160–150 Ma) in the north, respectively [*Amato et al.*, 2015; *Layer et al.*, 2001;  
352 *Shephard et al.*, 2013]. The final stage of the closure of the SAO was fulfilled by the  
353 collision between the Chukotka and Kolyma-Omolon blocks. Although the collision  
354 mainly occurred between 119 and 106 Ma [*Amato et al.*, 2015; *Miller et al.*, 2009;  
355 *Sokolov et al.*, 2002, 2009], it may started as early as 130–124 Ma [*Layer et al.*, 2001;  
356 *Toro et al.*, 2003], which also coincide with the cessation of the seafloor spreading in  
357 the Canada Basin at 128.6 Ma. This temporal and spatial consistency lead us to  
358 suggest that the closure of the SAO provided space for the opening of the Canada  
359 Basin, and the collision between the Chukotka and Kolyma-Omolon blocks at the  
360 final stage of the SAO closure terminated the seafloor spreading of the Canada Basin.  
361 This inference then generally supports the idea that the opening of the Canada Basin  
362 is associated with the subduction process in the SAO [*Koulakov et al.*, 2013;

363 *Kuzmichev, 2009*]. Nevertheless, the clockwise rotation of the “Arctida” plate requires  
 364 an overall higher opening rate in the southern part of the Canada Basin [*Koulakov et*  
 365 *al.*, 2013]. As “a common back-arc basin” proposed by *Kuzmichev* [2009], the Canada  
 366 Basin was also too far away (more than 1,000 km) from the subduction zones in the  
 367 SAO, since a back-arc spreading center is usually limited to a distance of 200–300 km  
 368 from a trench [e.g., *Toksöz & Hsui, 1978*]. Further information on the geometry,  
 369 extent, and polarity reversals of the subductions in the SAO may help to refine a more  
 370 comprehensive geodynamic model to address these issues.



371  
 372 **Figure 3.** Plate reconstructions of the circum-Arctic Region between 156 Ma and 129 Ma. The  
 373 shapes of the geological features are based on *Müller et al.* [2016]. (a) Rifting created the

374 Amerasia Basin. The South Anyui Ocean was then subducting to the south and north. East of the  
375 SAO, the Angayucham Ocean (AO) and associated Koyukuk Arc (initiated at ~160-145 Ma) are  
376 believed to be the eastern extensions or the counterparts of the SAO and Nutesyn Arc [*Amato,*  
377 2004, 2015; *Churkin et al., 1981; Nokleberg et al., 2000*], respectively. (b) Initial seafloor  
378 spreading in the Amerasia Basin. (c) Seafloor spreading cessation in the Amerasia Basin. The  
379 position of the future Alpha Ridge is marked by a red line. Since the position and geometry of the  
380 Chukchi Borderland in Mesozoic remain controversial [*Grantz et al., 1998; Hutchinson et al.,*  
381 2017; *Miller et al., 2006*], we only place the CBL in (c) according to its present configuration. The  
382 rifting direction in strike-slip models is shown in dashed arrow along the eastern boundary of the  
383 Northwind Ridge [*Døssing et al., 2018; Hutchinson et al., 2017*]. Note the northeast-trending  
384 strike slip is subparallel to the subduction zones in the SAO, which may require a new explanation  
385 about the dynamic relationship between them. The inferred thinned continental crust, transitional  
386 crust, and oceanic crust in the Canada Basin are marked with brown, green, and blue, respectively.  
387 CBL = Chukchi Borderland; CV = Central Verkhoyansk; GL = Greenland; KB = Kara Block;  
388 NAS = North Alaska Slope; NSS = Northern Siberia Shelf; SAO =South Anyui Ocean; SV =  
389 Svalbard.

390

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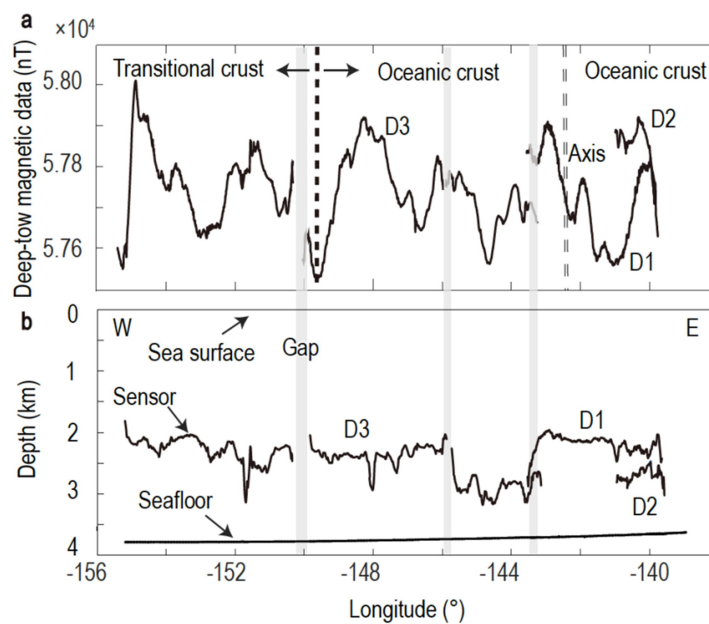
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## SUPPLEMENTARY INFORMATION



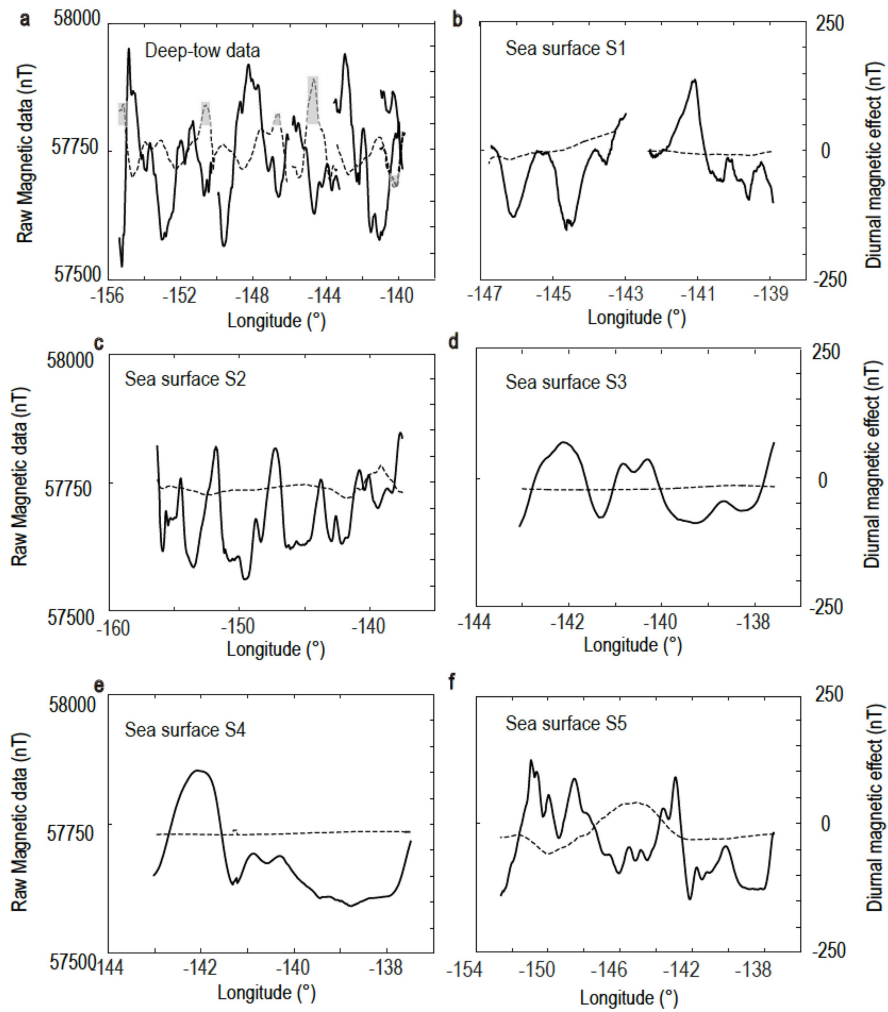
596

597 **Figure S1.** Raw deep-tow magnetic data (a) and depth of the sensor (b). The near E-W trending

598 deep-tow magnetic profile consists of three sections (D1-D3). Section D1 is along ~76.07°N and

599 is located ~40 km north of the other sections.

600



601

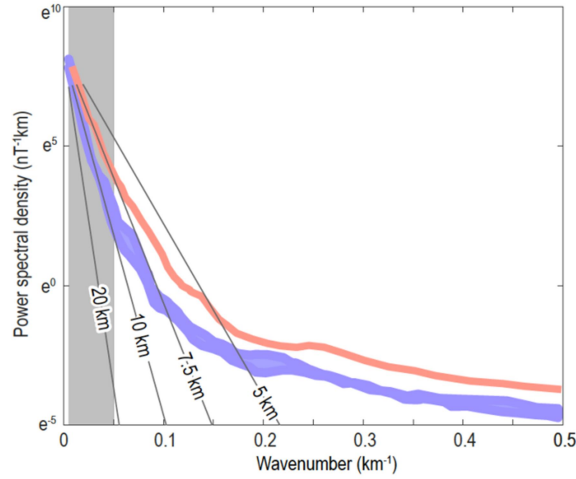
602 **Figure S2.** Comparisons between the raw magnetic data (solid lines) and the diurnal effects

603 (dashed lines). The STD of the diurnal variation for the sea surface surveys in 2016 and 2017 is

604 11.3 nT and 14.3 nT, respectively. The diurnal variations with amplitudes  $> 50$  nT are covered

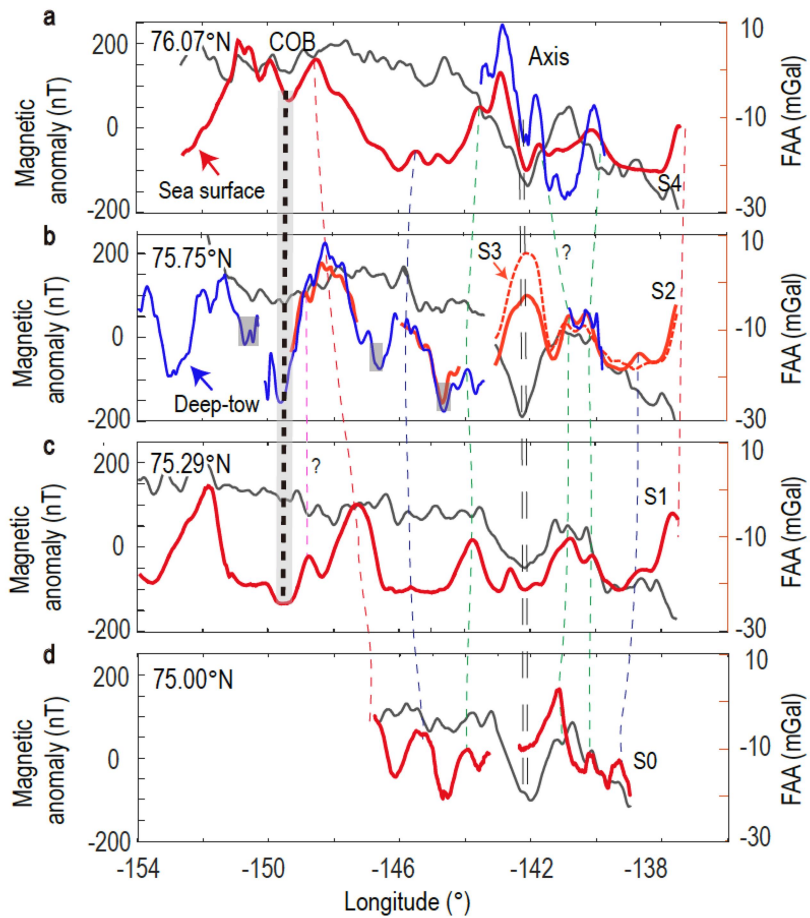
605 with gray boxes.

606



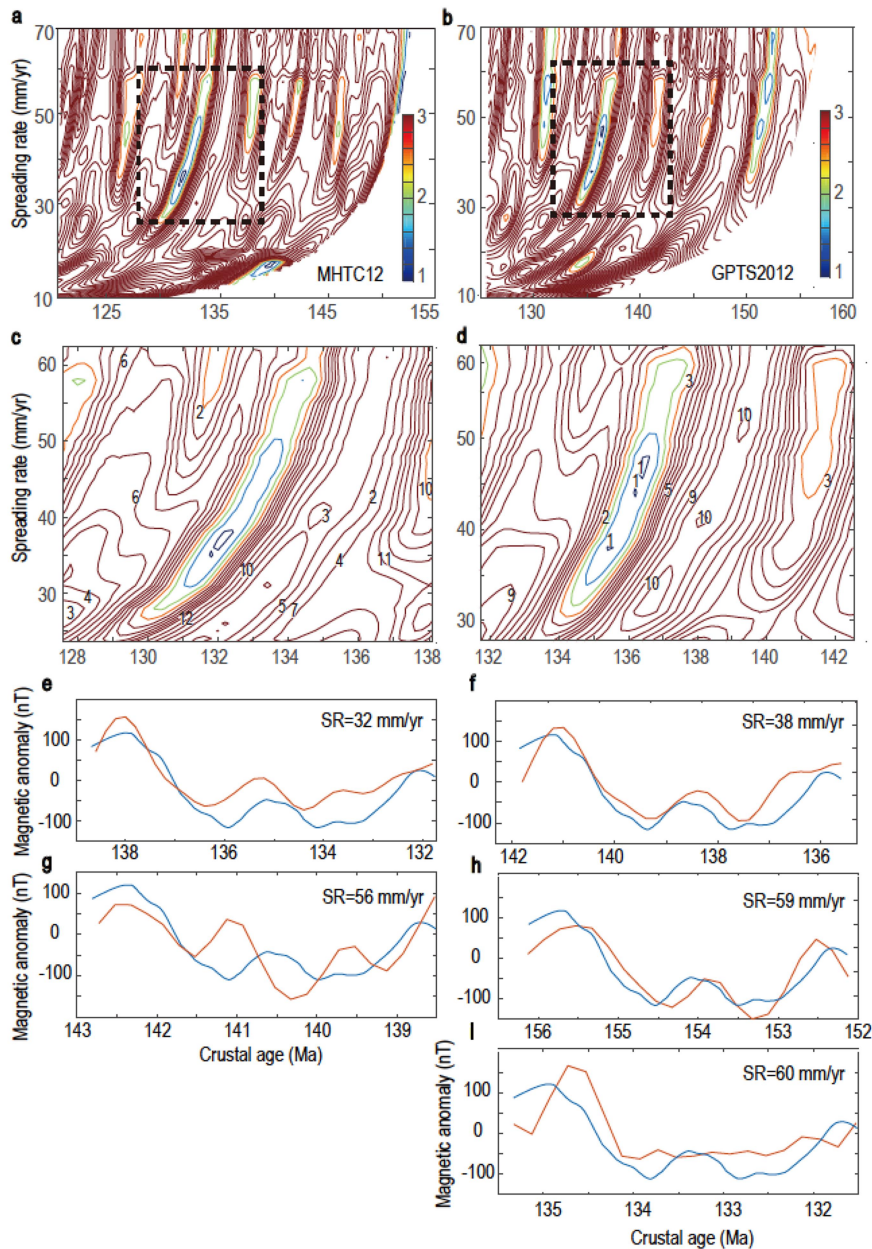
607

608 **Figure S3.** Power spectral densities computed from the sea surface (blue) and deep-tow upward  
 609 continued to sea surface data (red). The blue shaded area represents variance. The gray shaded  
 610 area indicates the signals with wavelengths between 20 km and 300 km, which is most likely due  
 611 to crustal sources.



612

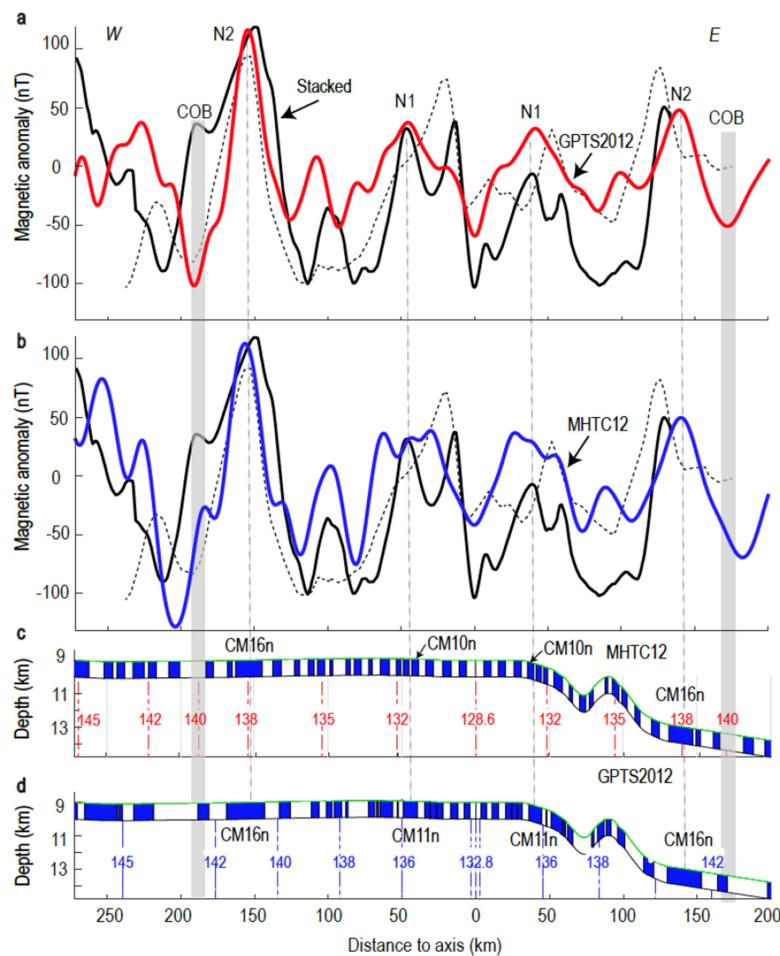
613 **Figure S4.** Free air anomaly (FAA) along the tracks of magnetic data. The FAA data (black lines)  
 614 are from *Sandwell et al.* [2014]. Note that the two broad magnetic highs (N1 and N2) are not  
 615 correlated with the FAA.  
 616



617  
 618 **Figure S5.** Estimation procedure of the best fit sequences. (a-b) Contours of the least-squares  
 619 misfit of the anomaly between the characteristic stacked data and synthetic data based on

620 MHTC12 (a) and GTS2012 (b). The cross-correlation method compares the stacked and synthetic  
 621 data based on least-squares fitting criteria [DeMets *et al.*, 2010]. In each comparison, the  
 622 amplitude scale of the synthetic data is adjusted to match the peak-to-trough amplitude of N1 and  
 623 N2. The contours are normalized by the misfit of the best-fitting least-squares model. (c-d) The  
 624 contours around the best-fitting model in the dashed frames of (a-b). (e-f) Comparison of the  
 625 stacked data and best-fitting data based on MHTC12 (e) and GTS2012 (f). (g-i) Comparison of the  
 626 stacked data and other candidates of the best-fitting models based on MHTC12 (g) and GTS2012  
 627 (h and i). The spreading rates (SR) are also shown.

628



629  
 630 **Figure S6.** Identification of the crustal age in the Canada Basin. (a-b) Comparison between the

631 stacked magnetic anomalies (black lines) and the best-fitting synthetic magnetic anomalies. The  
632 synthetic magnetic anomalies produced by the magnetic bodies in MHTC12 and GTS2012 are  
633 shown in red and blue, respectively. Stacked airborne magnetic anomaly is also shown (dashed  
634 line). (c-d) The best-fitting magnetic sequences in MHTC12 (c) and GTS2012 (d).  
635