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

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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

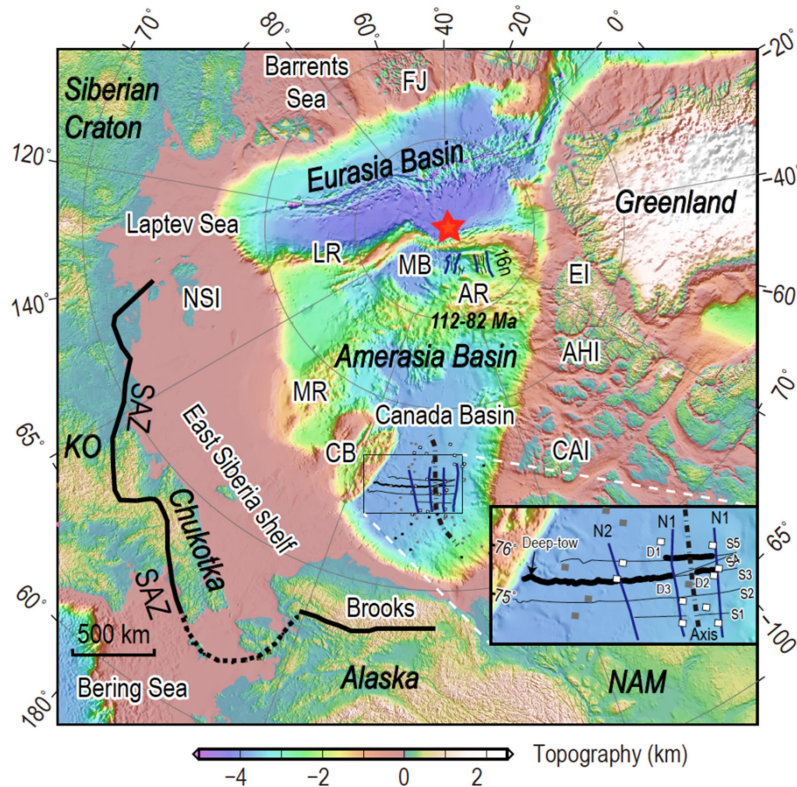
86 relict axis are broadly consistent with the depictions in the rotation model. However,
87 this model is challenged by alternative models [e.g., *Chian et al.*, 2016; *Døssing et al.*,
88 2018; *Hutchinson et al.*, 2017; *Koulakov et al.*, 2013; *Lane*, 199], in part due to the
89 uncertain age of 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 ± 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 ± 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 ~ 2 km above the sea floor and ~ 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 ~ 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 ~ 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**

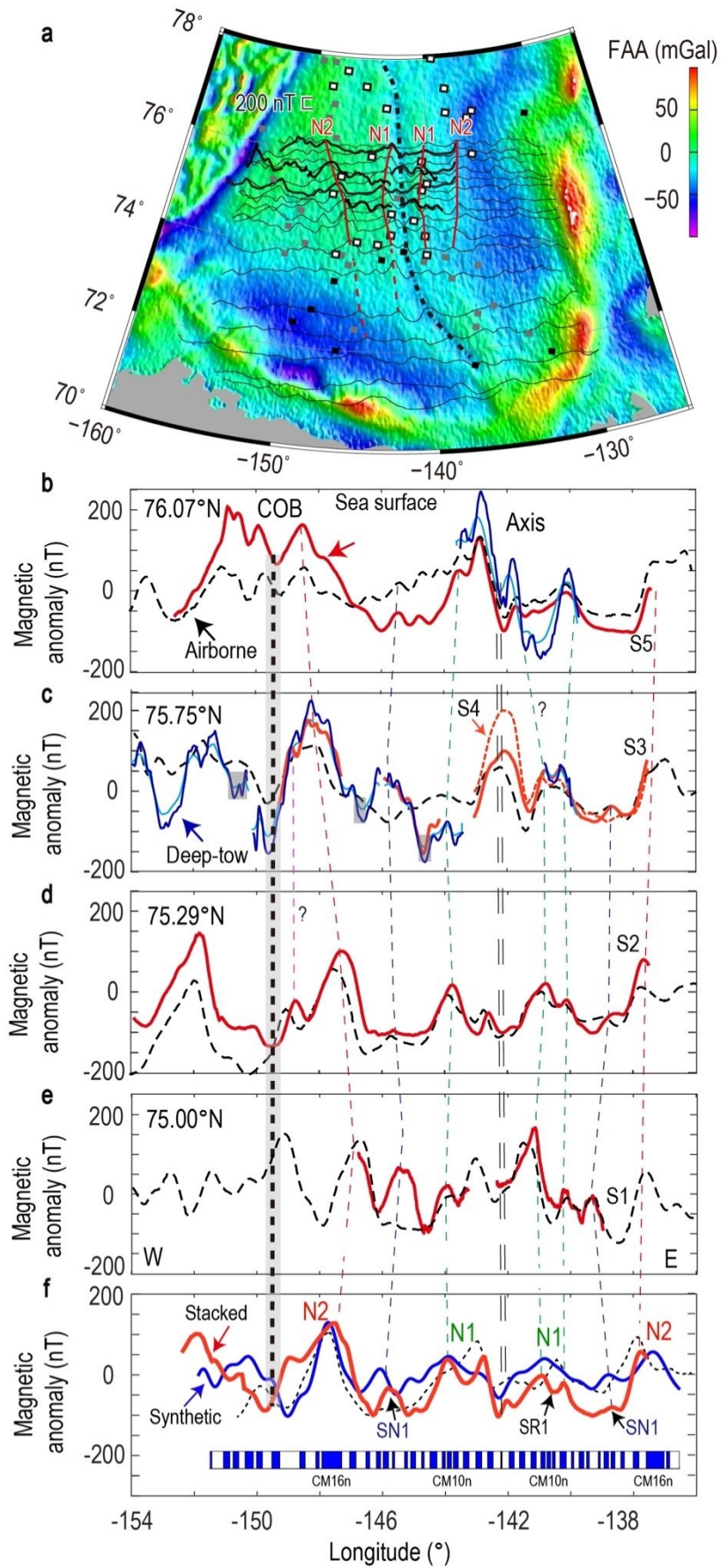
179 Two paired coherent broad magnetic highs are observed by the deep-tow, sea surface,
180 and airborne magnetic data (Figure 2). Here, we name them normal 1 (N1) and
181 normal 2 (N2) anomalies for the broad magnetic high close to and away from the
182 relict ridge axis, respectively. On the western 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 (~ 160 Ma) and not younger than the Late
252 Cretaceous (~ 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 (~ 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 ~ 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 ~ 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 (~ 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**

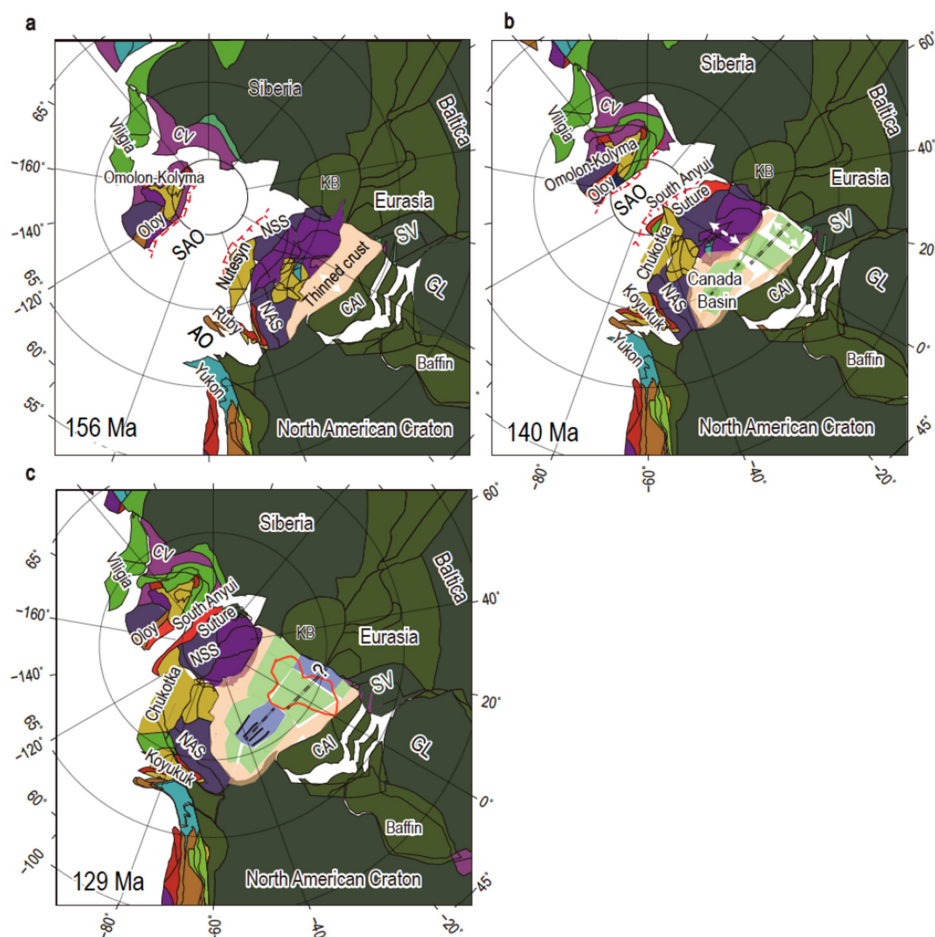
320 Three main regional unconformities (late Callovian-early Oxfordian, late

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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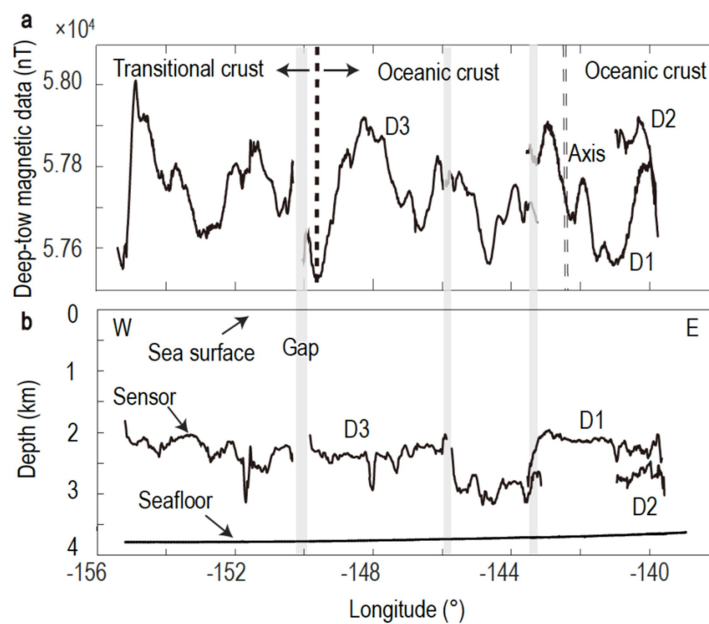
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594

SUPPLEMENTARY INFORMATION



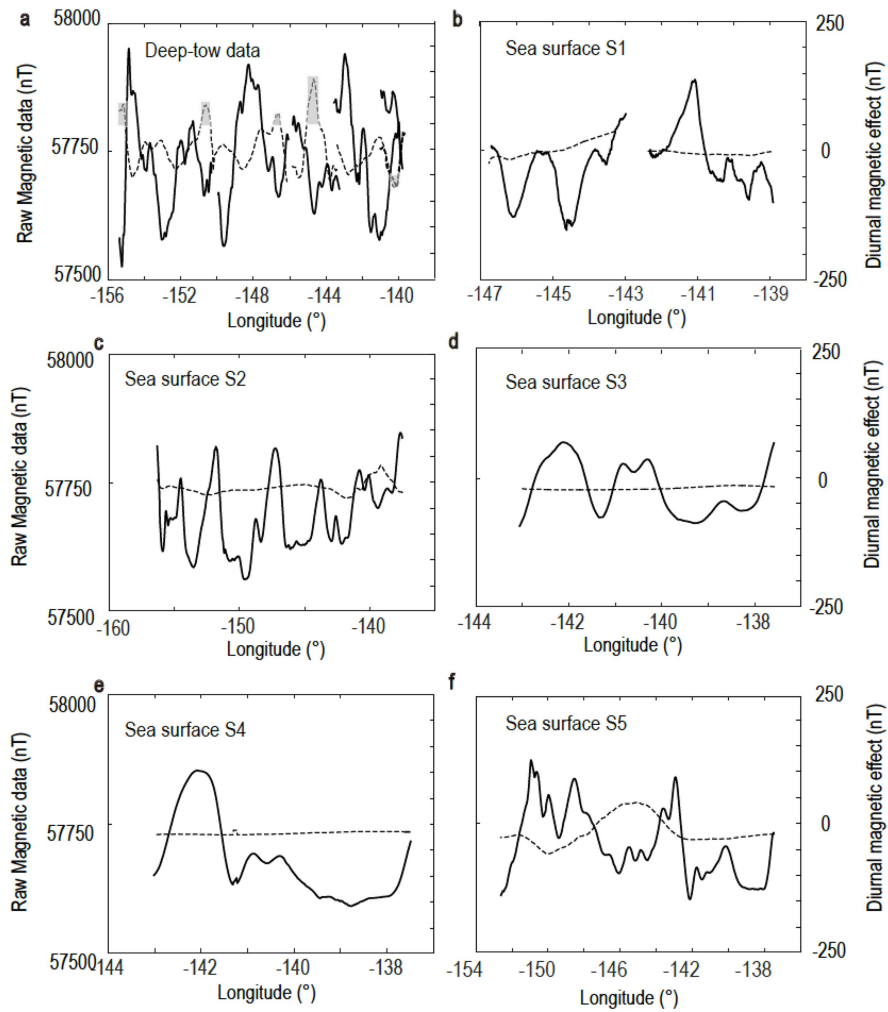
596

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

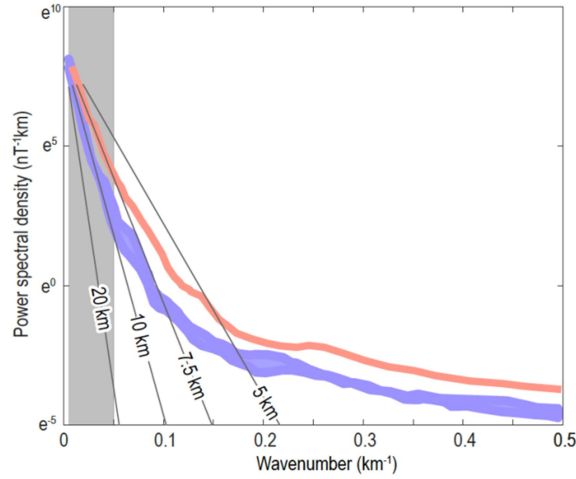
602 **Figure S2.** Comparisons between the raw magnetic data (solid lines) and the diurnal effects

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

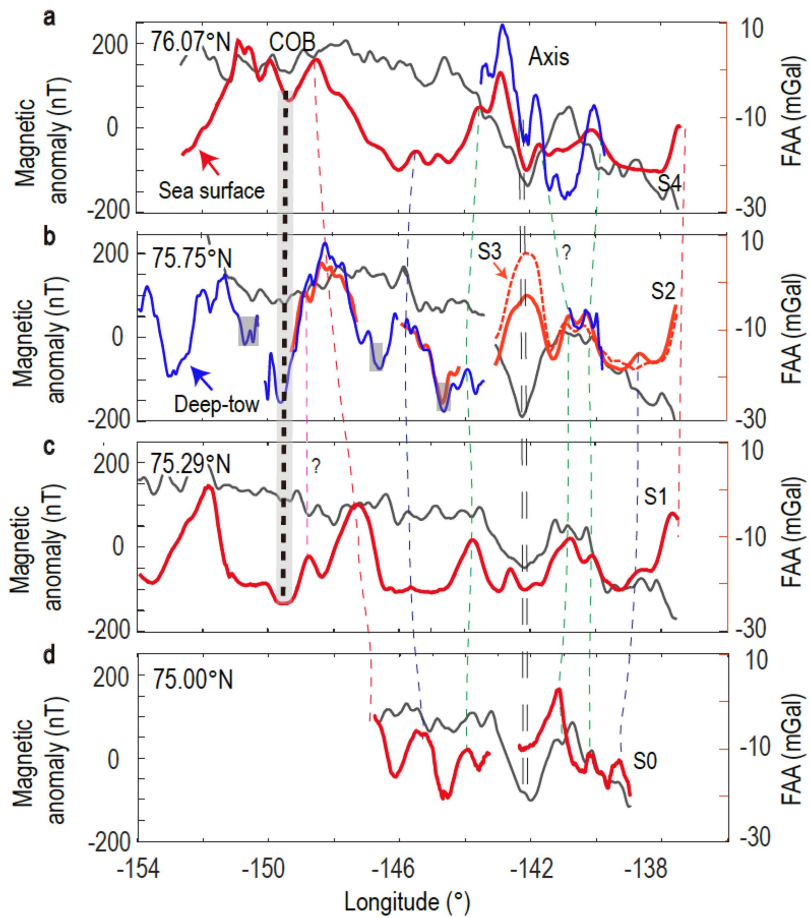
605 with gray boxes.

606



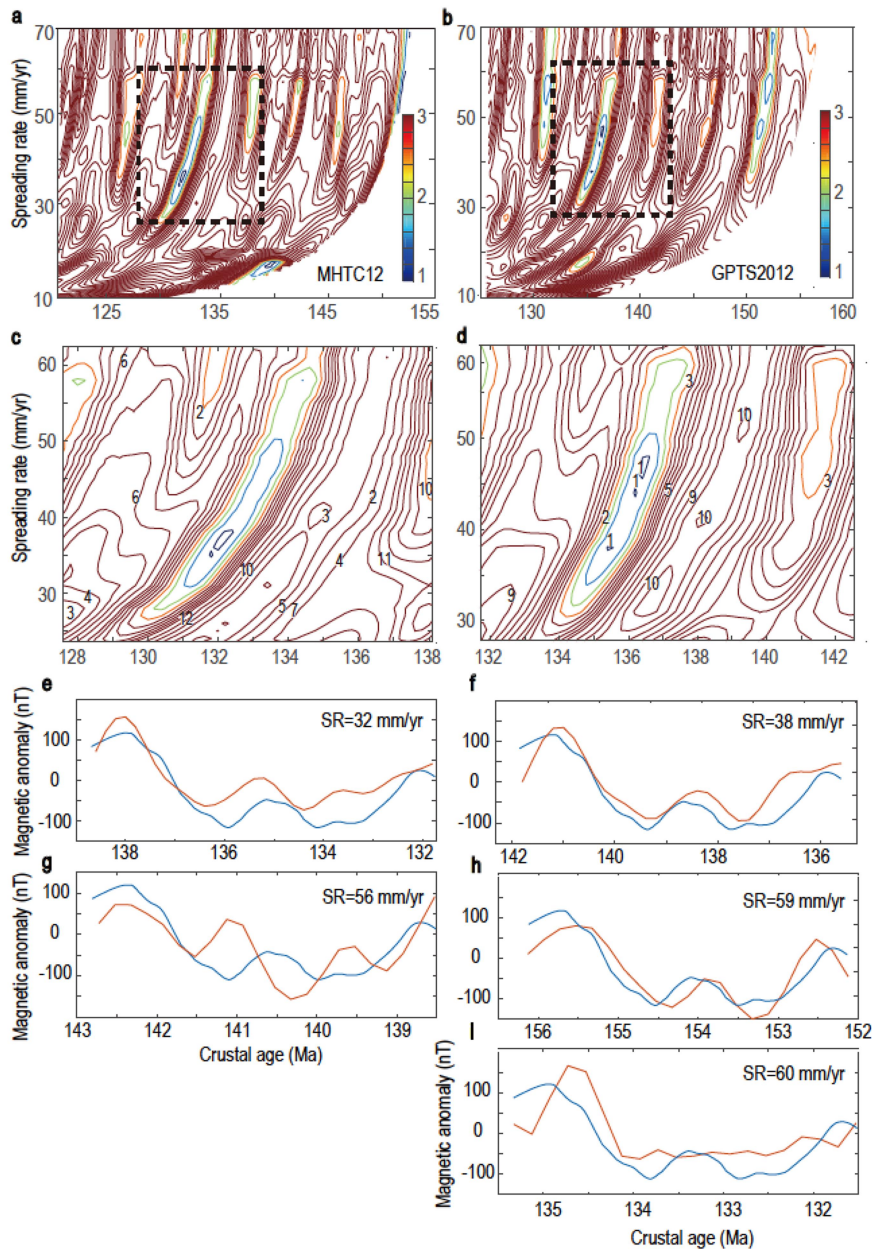
607

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

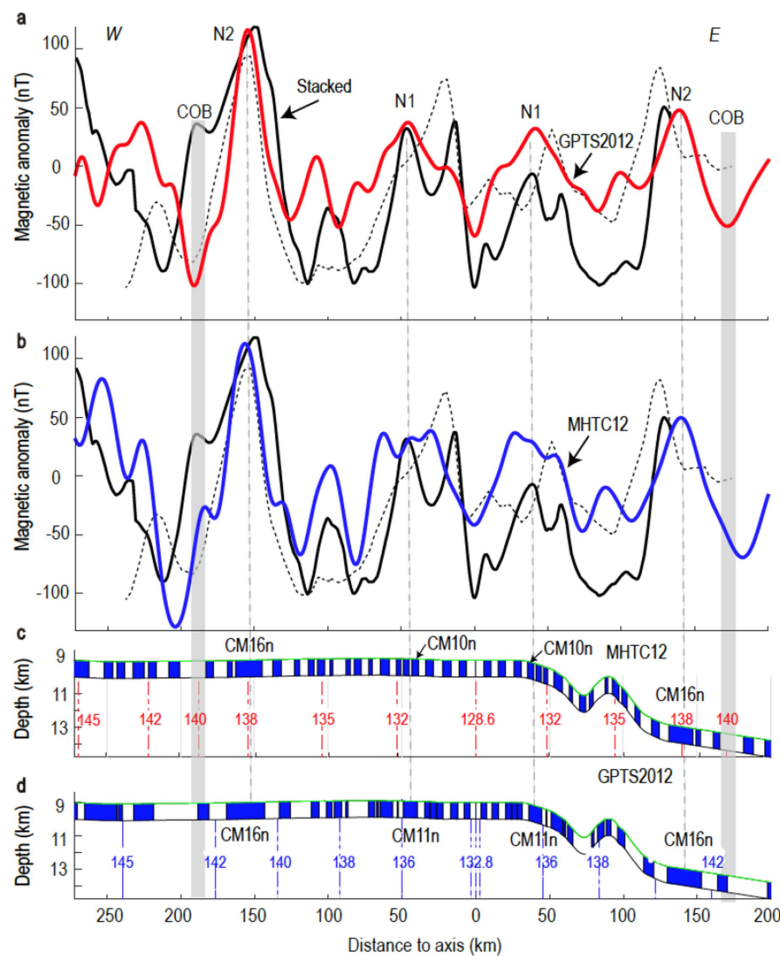
613 **Figure S4.** Free air anomaly (FAA) along the tracks of magnetic data. The FAA data (black lines)
 614 are from *Sandwell et al.* [2014]. Note that the two broad magnetic highs (N1 and N2) are not
 615 correlated with the FAA.
 616



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.

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 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).
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